The Effects on Ocean Models of Relaxation toward Observations at the Surface

PETER D. KILLWORTH, DAVID A. SMEED, AND A. J. GEORGE NURSER

Southampton Oceanography Centre, Southampton, United Kingdom

(Manuscript received 24 August 1998, in final form 29 March 1999)

ABSTRACT

This paper discusses the errors in surface tracer and flux fields in ocean models induced by using approximate surface boundary conditions involving relaxation toward observed values rather than more physically realistic conditions that involve (often inaccurate) surface fluxes. The authors show theoretically and with a global model example that where there is a net annual surface flux of tracer (balanced by advection), (i) the annual mean surface tracer field is biased compared with the observations and (ii) the annual mean tracer flux is also biased if the surface tracer field has a feedback on the surface tracer advection or diffusion. As previously shown, the amplitude of the annual cycle of tracers is also decreased. The global model indicates that temperature offsets of 1°–2°C (or even greater) and heat flux errors of 30 W m^{-2} occur in regions of strong advection, such as the equatorial upwelling zone, western boundary currents, and the Antarctic Circumpolar Current. These are all areas crucial for the thermohaline circulation, so that the use of such boundary conditions is likely to yield incorrect estimates for climate simulation models. Zonally integrated meridional heat fluxes may be in error by up to 25%.

1. Introduction

Haney (1971) is a much misquoted man. Since his innovative paper formulating a surface flux condition relating effective air temperatures to the surface water temperature, his general concept has been used—and misused—in a variety of ways by ocean modelers. Always, however, his name has been attached to simplified surface forcings in regions where there are no reliable observational flux estimates. Haney originally suggested that the surface heat flux could be well approximated by a term proportional to the difference between an effective air temperature (which varied with latitude) and the surface temperature, with a coefficient that also varied with latitude. This is equivalent to modeling the heat flux as an estimated flux (e.g., from some climatology) plus a relaxation toward the observed surface temperature (Barnier et al. 1995; cf. also the DYNAMO Group 1997; Oberhuber 1988).

Haney's work has been misapplied in several ways. Sometimes the heat flux is taken to be proportional to a difference between a true air temperature and the ocean temperature. Commonly, however, "Haney conditions" have been taken as relaxation of surface temperatures—and salinities, which Haney (1971) did not address—toward observed values (rather than to an effective air temperature), usually with a constant relaxation timescale. Such a relaxation has been used as a convenient proxy for reliable surface fluxes.

This has been necessary for numerical ocean models since, despite recent advances (Josey et al. 1998), oceanographers do not possess completely consistent surface flux fields with which to force their models. For example, globally integrated annual mean surface fluxes are nonzero so that a perfect model, forced entirely by fluxes, would not give realistic solutions over long integrations. Conversely, the many shortcomings in the models provide other reasons why at present models would be likely to fail if forced entirely with surface fluxes. For example, coarse resolution ocean-only climate models have western boundary currents that are far too broad and slow. As a result, applying "correct" surface fluxes to such a model induces incorrect surface temperatures; the converse also holds, in that enforcing "correct" surface temperatures would induce incorrect surface fluxes, of course. At the very least, surface fluxes currently need adjustment to "fit" the model being used (cf. Large et al. 1997). Relaxation toward observations can be seen as a way that this could be done, and we may ask how well or badly such an approach works.

Lacking accurate flux fields (and maybe sufficiently accurate models to permit the use of unadjusted fluxes), basin and global ocean modelers have reluctantly used relaxation to observations. This immediately leads to a well-known difficulty. Using a modified Haney condi-

Corresponding author address: Dr. Peter Killworth, James Rennell Division for Ocean Circulation, Southampton Oceanography Centre, Empress Dock, Southampton SO14 3ZH, United Kingdom.
E-mail: p.killworth@soc.soton.ac.uk

tion—a common terminology for relaxation toward observations—implies that the model heat flux into the ocean H_M is given by

$$H_M = \lambda(T - T_M),$$

where λ is a feedback coefficient, often assumed constant; T is the ''observed'' surface temperature to which we are relaxing; and T_M is the model surface temperature. Trivially, then,

$$T_M = T - \lambda^{-1}H_M,$$

so the surface temperature must be in error by an amount $\lambda^{-1}H$ in order to generate the model flux H.

This leads to the well-known contradiction (e.g., Oberhuber 1988) that, *if the model achieves a ''correct'' surface temperature, there can be no heat flux through the surface.* Since almost everywhere in the ocean receives either net heating or cooling, the model would have to predict incorrect surface temperatures if surface fluxes were to be correct. (Conversely, models with correct sea surface temperature could have no net northward heat flux.) Nonetheless, modelers frequently seek to diagnose poleward heat fluxes from models run with relaxation (e.g., Saunders et al. 1999).[1]

Typical values of the Haney coefficient λ, which gives the heat flux per degree of deviation of the surface temperature from climatology, might be 30–50 W m^{-2} °C^{-1}. Thus sea surface temperature errors of up to 5°C are implied in order to generate heat fluxes of 100–200 W m^{-2} such as are found either (i) in the seasonal cycle in midlatitudes or (ii) in the annual mean in regions of strong heat loss or gain such as the equatorial Pacific and the Gulf Stream (Josey et al. 1998). More intense relaxation, with $\lambda > 100$ W m^{-2} °C^{-1}, is unattractive since (i) features such as eddies and planetary waves are damped unrealistically rapidly and (ii) western boundary currents, whose physics is not always well represented by coarse models, will be severely degraded.

Now it could be argued that, because the model physics is in error, whether the surface fluxes implied by relaxation are correct in some sense is of small importance. Forcing a model by relaxation, especially strong relaxation, is implicitly adding in the right amount of heat to force the (incorrect) model to reproduce some form of reality. While true, most modelers seek to validate their models by computing poleward heat fluxes, which are essentially northward integrations of surface fluxes if storage terms are neglected. Incorrect poleward heat fluxes would thus lead to rejection of the model; local heat flux errors would also be produced. Thus when relaxation is employed, we need to understand both the errors induced in the surface tracers and also the errors induced in the surface fluxes.

The errors in the seasonal cycle have been discussed by Pierce (1996) at individual locations, neglecting advection. He demonstrated how such relaxation underestimated the seasonal cycle and damped out high-frequency variability. He suggested relaxation toward a modified surface temperature, computed from an amplified and lagged version of the original surface temperature. In the case of no advection, this would yield the correct temperature field at all times—indeed, it was constructed to do so—although the surface heat fluxes, which Pierce does not discuss, could still be in error. Pierce shows results from running an OGCM with and without his modified condition, and a clear improvement is visible.

Pierce's results are extended by Chu et al. (1998), who examine the validity of relaxation towards monthly mean temperatures (neither Chu et al. nor Pierce consider salinity) by comparing predictions from relaxation with NCEP heat flux analyses. They are unable to find a correlation between the two anywhere in the world ocean. However, approaches using the original Haney (1971) method perform well. Greatbatch et al. (1995) used a variety of surface-forcing formulations compared with a control model whose fluxes were known (a method we shall employ below) for an examination of the survival of interdecadal climate variability.

Thus relaxation toward observed values of temperature (and, presumably, salinity) gives a poor reproduction of reality. But how poor? Even if instantaneous heat fluxes are poor, it is possible that the annually averaged poleward heat fluxes deduced from these would be acceptable. Relaxation toward a steadily warming ocean surface temperature will tend to produce a net warming since the modeled temperature will lag the observed value. The converse will occur in times of cooling. The annual average of this is unclear. The above papers, in addition, did not discuss the role of advection in the maintenance of surface tracer values.

This paper concentrates on two features not discussed in the earlier papers: the effect of advection and the size of, and errors in, surface fluxes (Chu et al. concentrate on the correlation patterns, rather than the sizes, in the diagnosed heat fluxes). Heat flux climatologies differ between sources at precisely the level we seek to examine here. Rather than rely on any given climatology, we work first with a simple purely time-dependent model. A simple example involving heat fluxes is discussed in section 2, and in section 3 we briefly discuss the case of a steady situation. In section 4, we extend the argument to include a time-varying surface forcing and show that the response is a biased, lagged, smaller, and less variable temperature than the observed. Some aspects of this have been derived by Pierce (1996). Section 5 examines the effects of relaxation at single points in the ocean, ignoring advective effects. The simple model

[1] A referee notes that simply changing the sea surface temperature dataset that is relaxed toward can modify poleward heat flux estimates by 0.3 PW, showing the sensitivity of models to this type of surface forcing.

used in these sections employs an equally simple representation of advection, which is validated later.

In section 6 we use a coarse-resolution global model to study the modeled surface temperature including the effects of advection and bias. By running this model to a seasonal equilibrium over a millennium, we have a model whose physics can be regarded, for perturbation purposes, as ''truth'' (e.g., its treatment of western boundary currents is poor, but the surface fluxes in the model are entirely consistent with the boundary currents). We then force a second version of the model toward the surface tracers observed in the first experiment. The differences between the models can then be examined, without the distractions of needing to modify fluxes to satisfy model infelicities. We show that surface temperatures and surface fluxes are both seriously in error using relaxation conditions. We conclude with a brief discussion of alternative approaches.

2. Heat fluxes

In the steady state, the observed correct surface sea surface temperature (SST), or more generally some tracer, is maintained by a balance between surface heating H and advection and diffusion of heat A:

$$\rho c_P h \frac{dT}{dt} = H + A = 0, \tag{1}$$

where c_P is the specific heat, ρ is the density of seawater, and h is the mixed layer depth. Here ''advection'' is used as a shorthand for flux divergence, including any eddy parameterizations used. A perfect model, driven by the correct surface flux H, will reproduce the correct SST, T, and advection A.

A similar balance will hold in the steady state of a model being driven by relaxation toward the observed temperature (using subscript M to denote model values):

$$H_M + A_M = \lambda(T - T_M) + A_M = 0. \tag{2}$$

As discussed earlier, in general the relaxed SST, T_M, $\neq T$ in order to generate a nonzero flux H_M. However, there is a further problem: the advective fluxes in the model may be sensitive to the surface temperature field. Hence, even if the model is perfect, because the SST $T_M \neq T$, the advective flux A_M may differ from the correct advective flux A. This difference will occur both because fluxes depend directly on T (e.g., in upwelling regions) but also because advective velocities will change through density changes that depend on T. Thus, in the steady state, the surface flux H_M will differ from the correct flux H.

For simplicity let us assume a crude linear dependence of the advection upon the local SST:

$$A_M = A + \lambda_A(T - T_M). \tag{3}$$

(This is particularly applicable for upwelling regions, but more generally might be relevant for small departures from observations, by assuming a Taylor series.) Since our model has zero dimension, we cannot permit the advective flux to depend on lateral gradients, for example. In section 6 we examine the behavior of the advection in a three-dimensional model and find that it is well described by (3) above.

It then follows that

$$H_M = -A_M \Rightarrow$$

$$H - \frac{\lambda_A}{\lambda} H_M = -A - \lambda_A(T - T_M), \tag{4}$$

so

$$H_M = \frac{H}{(1 + \lambda_A/\lambda)}. \tag{5}$$

The inaccuracy of the relaxation fluxes H_M then depends on the ratio between the feedback coefficient λ (which regulates the errors in SST) and the advective feedback sensitivity λ_A (which regulates the sensitivity of the fluxes to the SST). It is unclear how to estimate this advective feedback sensitivity λ_A, which will be model and process dependent; it will be discussed in the next section.

We now extend these arguments to examine temperature itself, followed by the addition of time varying forcing.

3. Steady forcing

Let us measure the surface temperature (or, more generally, some tracer) relative to some background value of $\overline{T}$. We now assume that the observed surface temperature takes a steady value, at some location. Without loss of generality, this value can be taken as zero (its value being absorbed into $\overline{T}$). We shall retain the symbol T so that the shape of the formulas below is recognizable, though its value is zero. This value is maintained by two equal and opposite effects: surface forcing $F_S(T)$ and advective effects $F_A(T)$, so that

$$\frac{\partial T}{\partial t} = 0 = F_S + F_A, \tag{6}$$

where t represents time. Both effects are of course constant in this simple case. However, we assume that, if a modeled surface temperature T_M is used to simulate T, then the circulation, and hence the local advective effects, will be modified. We again choose the simplest possible dependence, relevant for small departures from observations, by assuming a Taylor series, so that we write

$$F_A(T) = -F - \frac{T}{\tau_A}. \tag{7}$$

(We again refer the reader to section 6 for some justification of this choice.) Here F is a constant forcing due to the mean current, and the modulus of τ_A is an advective feedback timescale corresponding to the expansion in T (fn 2). If τ_A is large, the advection depends only weakly on the surface temperature; if τ_A is small, the advection is strongly altered by changing the surface temperature, and the linear expansion (7) would be inaccurate. Since τ_A is just an expansion coefficient for the advection, about which we know little at this stage, it could theoretically take either sign. It will appear from our later results that τ_A is positive.

The model of surface temperature replaces F_S with a relaxation term, but retains the advection, so giving

$$\frac{\partial T_M}{\partial t} = 0 = \frac{T - T_M}{\tau} + F_A(T_M), \tag{8}$$

where τ is the relaxation coefficient pulling T_M toward the observed T. Thus the solution T_M satisfies

$$0 = \frac{T - T_M}{\tau} - F - \frac{T_M}{\tau_A}$$

or

$$T_M = \frac{\tau^*}{\tau}T - \tau^*F. \tag{9}$$

Here $\tau^{*-1} = \tau^{-1} + \tau_A^{-1}$ is a combined timescale.

We see immediately that the response to steady relaxation forcing has two terms, though of course the first term is zero by our choice of reference (this term will become relevant in the following section). The other term, which is nonzero, is a bias proportional to F, which reduces as the relaxation timescale reduces. However, it is not possible to ensure that the modeled and observed surface temperatures are identical when advective effects are included.

This bias, $-\tau^*F$, induces an erroneous surface flux. The correct surface flux is simply F, which cancels the true advective component. The modeled surface flux is $(-T_M)/\tau = (\tau^*/\tau)F$ so that the error in surface flux is $-\tau F/(\tau + \tau_A)$. This error decreases with τ_A, reaching zero for very long advective feedback times, but is of the order of the mean advection when $\tau \approx \tau_A$.

It is unclear how to estimate quantities such as feedback times. If we took the simplest approach, the advective component would be of order vT_y, where v is a typical wind-induced surface drift and T_y a northward gradient. Taking $v = 0.05$ m s^{-1} and $T_y = 2°/1000$ km, $F \approx 10^{-7}$ in SI units. Over a mixed layer of depth 50 m, this is equivalent to a flux of 20 W m^{-2}. If we took $\tau_A = L/v$, where L is the gyre length scale (here 1000 km), we would have $\tau_A = 7$ months, and hence $\tau^* = 0.875$ months if τ has a value of one month.[3] Such values would give a surface flux error of 2.5 W m^{-2}, certainly much smaller than any surface fluxes that could be observed with confidence. Nonetheless, apparently small errors such as this can still generate noticeable errors in poleward heat flux: spread over the Southern Ocean, this error is equivalent to 0.1 PW northward heat flux error. In active regions such as boundary layers, errors would be far larger.

4. Time-varying forcing

We now extend the argument to time-varying flows and assume for simplicity that the observed surface temperature (or, more generally, some tracer) is

$$T = T_0 \sin\omega t, \tag{10}$$

where T is again measured relative to some value $\overline{T}$; T_0 is the amplitude of the observed value. Phase is measured relative to that for the surface temperature. The problem posed here is linear, so solutions can be superposed, for example, other Fourier components. The frequency ω may be annual or may correspond to some shorter timescale such as spring warming.

We then suppose that the surface temperature is produced by a combination of surface forcing $F_S(t, T)$ and advective forcing $F_A(t, T)$, where now dependency on time is explicitly included, so that

$$\frac{\partial T}{\partial t} = F_S(t, T) + F_A(t, T). \tag{11}$$

From climatology and measurements, we may know some aspects of the surface forcing, which will be deduced below. However, details of the advective forcing remain unknown; we continue to permit dependence on T in F_A to include a measure of generality. We suppose that

$$F_A(t, T) = -F + A\sin(\omega t + \phi_A) - \frac{T}{\tau_A}, \tag{12}$$

where F remains the steady forcing due to the mean current system, A and ϕ_A are the amplitude and phase of the seasonal advective cycle, and τ_A is the feedback timescale of the surface temperature on the advective forcing, assumed independent of time for simplicity. Again, if there is no temperature dependency in the advective forcing, τ_A is infinite. None of these quantities are known in general.

We then have

[2] The τ_A later similar terms replace the λ factors earlier; when heat fluxes are considered, feedback coefficients (e.g., in W m^{-2} °C^{-1}) are relevant, whereas when temperatures is considered, a simple relaxation timescale is relevant.

[3] Note that the simple estimate in section 6 gives $\tau_A \approx 100$ days, which is of the same order as the 7 months assumed above; this gives $\tau^* = 0.75$ months, with answers similar to those following.

$$\frac{\partial T}{\partial t} = \omega T_0 \cos\omega t = F_S + F_A$$
$$= F_S - F + A\sin(\omega t + \phi_A) - \frac{T}{\tau_A} \quad (13)$$

from which F_S can be deduced. We permit a surface feedback timescale τ_S in a similar way, by writing

$$F_S = S(t) - \frac{T}{\tau_S}, \quad (14)$$

and $S(t)$ is found by substituting for T, giving

$$S(t) = F - A\sin(\omega t + \phi_A) + T_0\left\{\frac{\sin\omega t}{\tau_F} + \omega\cos\omega t\right\} \quad (15)$$

as the forcing at the surface necessary to reproduce the observed seasonal cycle. Here $\tau_F = 1/(\tau_A^{-1} + \tau_S^{-1})$ is a combined feedback timescale, shorter than either advective or surface timescales. There are two terms in braces in (15). The second term is dominant if $\omega\tau_F \gg 1$; lacking detailed knowledge of the advection it is unclear if this would be the case. (With similar estimates to those in the last section, we find that for the annual frequency, $\omega\tau_A \sim 4$, which is quite large. If ω is the annual frequency, then $\tau_S > 2$ months is enough to satisfy this.) The time average of $F_S + F_A$ is zero, as required. The time-averaged surface forcing (including the feedback term) is F.

The modeled surface temperature, T_M, is assumed to satisfy

$$\frac{\partial T_M}{\partial t} = \frac{(T - T_M)}{\tau} + F_A(t, T_M), \quad (16)$$

where τ is again a relaxation time scale pulling the surface temperature toward observations. For most purposes, this timescale will be small compared with the seasonal timescale; that is, $\omega\tau \ll 1$. The same advective field (apart from changes in temperature) applies to the modeled temperature. Thus

$$\frac{\partial T_M}{\partial t} + \frac{T_M}{\tau} = \frac{T}{\tau} - F + A\sin(\omega t + \phi_A) - \frac{T_M}{\tau_A} \quad (17)$$

or

$$\frac{\partial T_M}{\partial t} + \frac{T_M}{\tau^*} = -F + \frac{T_0 \sin\omega t}{\tau} + A\sin(\omega t + \phi_A), \quad (18)$$

where $\tau^{*-1} = \tau^{-1} + \tau_A^{-1}$. Note that Eqs. (13) and (18) imply

$$\left(\frac{\partial}{\partial t} + \frac{1}{\tau^*}\right)(T_M - T) = -F_S(t, T). \quad (18a)$$

Equation (12) has solution

$$T_M = -F\tau^* + \frac{\tau^*}{[1 + (\omega\tau^*)^2]^{1/2}} \times \left\{\frac{T_0}{\tau}\sin(\omega t - \psi^*) + A\sin(\omega t + \phi_A - \psi^*)\right\} \quad (19)$$

where $\tan\psi^* = \omega\tau^*$. The modeled surface forcing is then

$$\frac{T - T_M}{\tau} = \frac{F\tau^*}{\tau} + \frac{T_0}{\tau}\left\{\sin\omega t - \frac{\tau^*}{\tau[1 + (\omega\tau^*)^2]^{1/2}}\sin(\omega t - \psi^*)\right\} - \frac{A\tau^*}{\tau[1 + (\omega\tau^*)^2]^{1/2}}\sin(\omega t + \phi_A - \psi^*). \quad (20)$$

After a little algebra, this can be rewritten as

$$\frac{T - T_M}{\tau^*} = F - \frac{A}{[1 + (\omega\tau^*)^2]^{1/2}}\sin(\omega t + \phi_A - \psi^*) + \frac{T_0}{\tau_A}\frac{\sin(\omega t - \psi^*)}{[1 + (\omega\tau^*)^2]^{1/2}} + \frac{T_0\omega}{[1 + (\omega\tau^*)^2]^{1/2}}\cos(\omega t - \psi^*) \quad (21)$$

$$= \frac{1}{[1 + (\omega\tau^*)^2]^{1/2}}\left\{F_S\left[t - \frac{\psi^*}{\omega}, T\left(t - \frac{\psi^*}{\omega}\right)\right] - F\right\} + F. \quad (22)$$

Discussion

Several points emerge.

1) We see immediately that the annual mean T_M remains $-F\tau^*$, which represents a *bias* in the modeled response. This results from the need to deviate from T in order to generate a flux. The amount of bias decreases as τ decreases; that is, stronger forcing toward observations reduces the bias.
2) The oscillatory component of T has two terms. The first is similar to the observed temperature T, but with a reduced amplitude and a lag ψ^*. If indeed $\omega\tau$ is small, both damping and lag are small also. The second term appears because the relaxation toward observations cannot directly account for advective effects and tends to zero as the relaxation time τ tends to zero. Pierce (1996) found a modified form of this part of the solution.

3) The time-averaged diagnosed surface forcing is $F\tau^*/\tau$. If τ is much the shortest of the timescales, this is approximately F, so the mean surface forcing is reproduced. If this is not the case, there is a bias in the mean surface forcing, of $F\tau^*/\tau - F = -F\tau/(\tau + \tau_A)$. This is as seen in the steady state for the heat flux (5), where the timescale ratio $\tau/\tau_A = \lambda_A/\lambda$ in (5).
4) There are additional terms in the diagnosed surface forcing.
   - The time varying part of the diagnosed surface flux is lagged and damped in relation to the true surface flux; see (22). The result is very similar to that found by Pierce (1996); however, the inclusion of advection modifies both the phase lag and the damping.
   - The phase lag $\psi^* = \tan^{-1}\omega\tau^*$. When the timescale for advection τ_A is long compared with the relaxation time τ, the effect is small and $\psi^* \sim \tan^{-1}\omega\tau$.
   - The amplitude is damped by the factor $(\tau^*/\tau)[1 + (\omega\tau^*)^2]^{-1/2}$, the same factor as for the surface temperature signal. If $\tau_A \ll \tau$, which is not usually the case, the damping would be approximately τ_A/τ, which can be very significant.
5) For very strong relaxation ($\tau \rightarrow 0$), it is easy to see that $T_M \rightarrow T$, and $(T - T_M)/\tau \rightarrow S(t)$, provided merely that in the latter, any surface feedback time $\tau_S \gg \tau_A$. In other words, strong relaxation recovers the original solution.

We can examine the meaning of these estimates by inserting approximate numbers. Let us take values used in the global model below: $\tau = 37$ days. In section 6 we show that a typical value for τ_A is about 111 days, or $\tau_A \approx 3\tau$. This gives $\tau^* \approx 3\tau/4 \approx 27$ days. We consider the points in order.

The annual mean temperature (point 1) has a bias of $-F\tau^*$. Using our estimates, and taking $F = 10^{-7}$ in SI units as before, gives a mean bias of 0.2°C. Thus for the fairly rapid relaxation that we have assumed, the mean temperature offset is fairly small. The phase lag is 25° (point 2), and the amplitude of the surface temperature signal is reduced to 68% of the original. The time-averaged surface flux (point 3) is 75% of the original. The amplitude of the annual surface flux is also reduced to 68% of the original (point 4).

These terms can contribute toward a strong surface forcing. Assuming the temperature tendency occurs over a mixed layer depth h, then a rate of temperature change γ becomes a heat flux $\rho c_p h\gamma$. Taking $h = 50$ m and an annual amplitude of 2.5 C, the term ωT_0 above yields a heat flux of about 100 W m^{-2}. Thus the errors produced by relaxation can be *large.*

5. A simple time-dependent model

These effects can be seen more realistically by considering the time-dependent problem without advection or other effects. (The temperature aspect of the solutions here thus parallels, but extends, Pierce's 1996 results.) We take the 12-month surface temperature data from Levitus and Boyer (1994) at five typical locations in the World Ocean [North Atlantic, North Pacific, equatorial Pacific, South Atlantic, and South Ocean (Indian sector)]. Each of these records is treated by the method of Killworth (1996) to produce a continuous piecewise linear pseudo-surface temperature, whose monthly means are those of Levitus and Boyer.

We then solve (16) at each location, with the advective contribution F_A set to zero. This is clearly not the case at any of these locations, which possess either annual average heating or cooling. However, this simple model enables us to further quantify the above analysis.

After at most two years, the system settles to a recurring annual cycle (because the memory time is about τ, which is much less than a year). The solution is computed daily; the solution between any two days can be written explicitly as a combination of an exponential decay on timescale τ and a linear function of time (caused by the forcing, which is always linear within a day). Figure 1 shows the annual signal at the five locations, using a relaxation time τ of one month. The observed temperature signal is usually dominated by the annual cycle so that the results of the previous section apply well. Even when there is variability on the one-month timescale [e.g., in the equatorial area shown; the matrix premultiplication (Killworth 1996) of necessity enlarges the variation], the modeled temperature continues to show a lag of about a month. However, the amplitude of the response is noticeably smaller for rapidly varying observed temperatures, as predicted above.

Figure 2 shows the implied heat fluxes from the solutions in Fig. 1, assuming a uniform mixed layer depth of 50 m. The heat fluxes all have no annual mean, by construction, but easily reach values of 100 W m^{-2} as suggested by the analysis. It is interesting to note that the behavior of the North Atlantic and North Pacific modeled heat fluxes is similar; there is no reason that this should be the case for the real ocean.

Figure 3 shows the sensitivity of the surface temperature to the relaxation time τ for the North Atlantic location. The amplitude and phase of the signal for small τ are that of the observed surface temperature, from (16). Conversely, for large τ, the amplitude tends to zero (because the observed value is essentially changing too rapidly for the modeled temperature then to respond). Note, however, that variations of τ over one decade—using values which have appeared in the literature—induce variations in the amplitude of the modeled temperature of a factor of 2. Changing τ also strongly modifies the heat flux, as Fig. 4 shows; changes over one decade in τ induce changes of heat flux by a factor of 2. For very small values of τ, the heat flux is uniform within a given month of linearly interpolated forcing, with steps at the month boundaries.

A good argument can be made from these results that

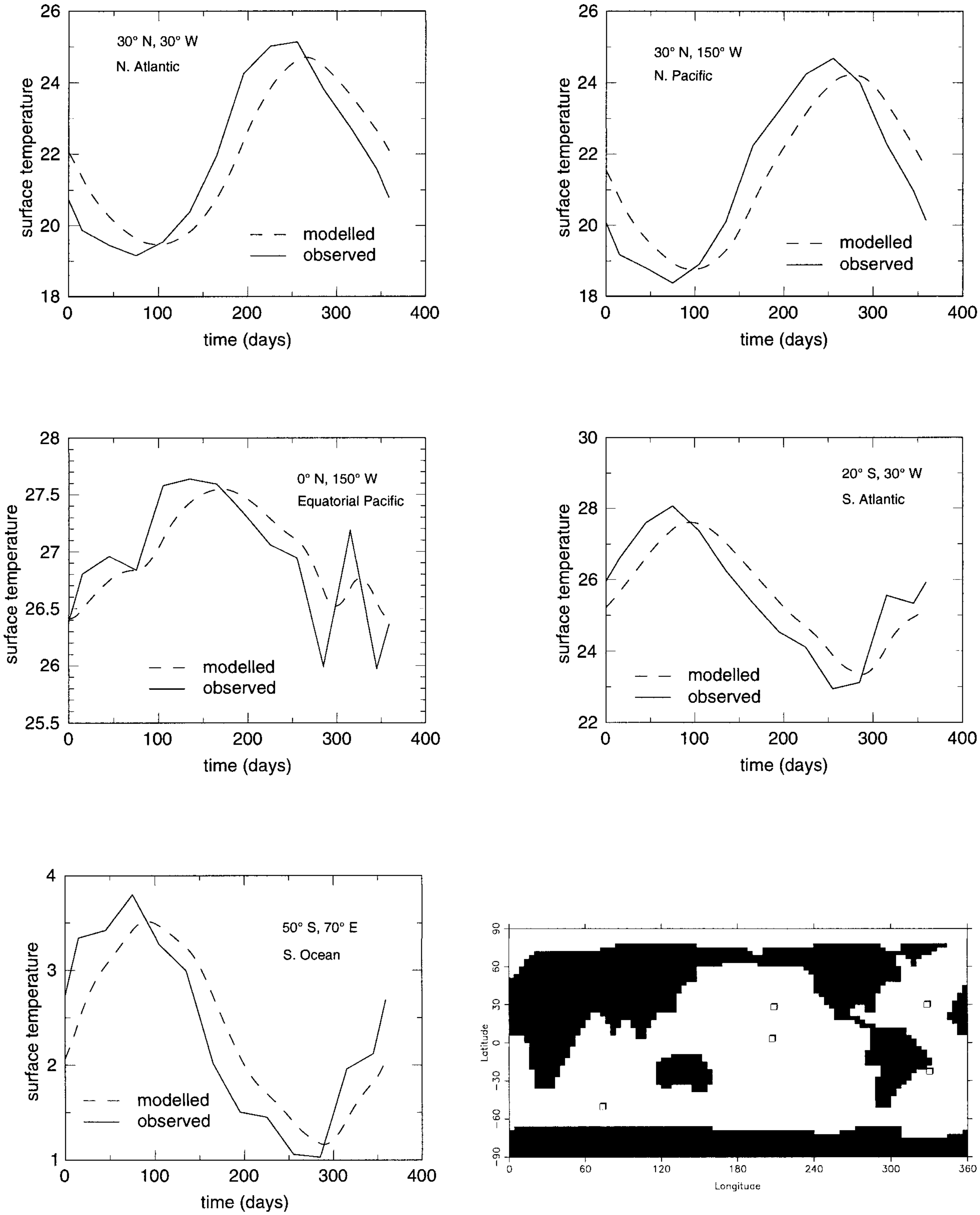

FIG. 1. The seasonal cycle of temperature at five locations in the world ocean induced by simple relaxation toward an effective surface temperature field (shown by the full line) whose monthly average is that given by Levitus and Boyer (1994). The response, for a relaxation time of one month, is shown dashed. The last diagram shows the locations superimposed on the coarse global model used later in the paper.

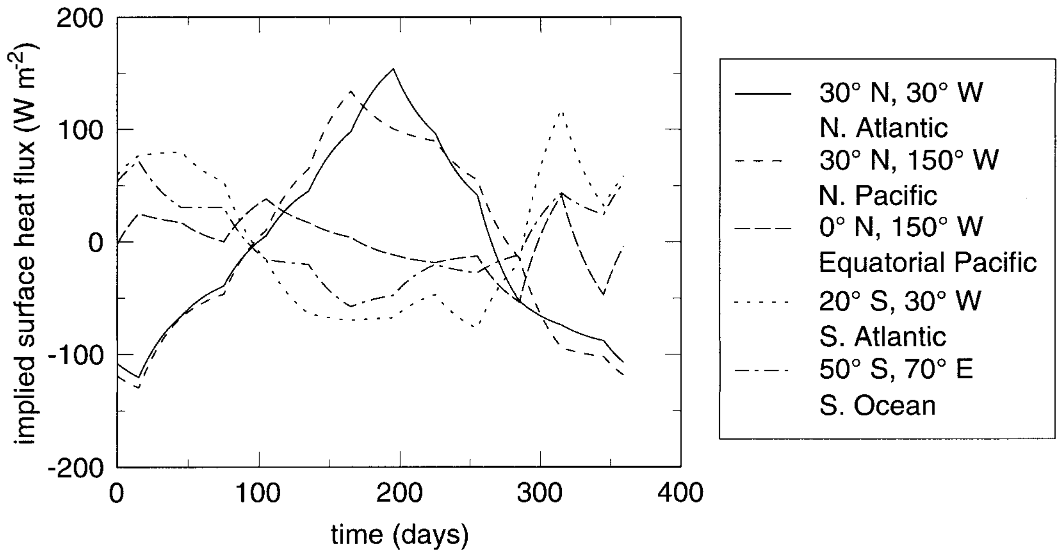

FIG. 2. The surface heat flux at five locations induced by relaxation towards observed temperatures, assuming a uniform mixed layer depth of 50 m.

strong forcing, with τ of order a few days at most, has the benefit that it would reproduce the observed surface tracer fields and, to a good degree of accuracy, the surface fluxes. Thus this forcing has automatically made the results consistent with whatever surface forcing produced those surface fields. In some sense, the surface fluxes become ''correct'' to within the model physics. Several difficulties should be noted. The model physics are themselves incorrect, which may cause internal difficulties; strong damping rapidly destroys mesoscale features in fine-resolution models (and, indeed, also destroys anomaly propagation in coarse-resolution models; Zhang et al. 1993); and flow in regions of strong advection, for example, strong western boundary currents, will be heavily distorted by strong relaxation.

6. A test with a global ocean model

In order to find out how important the advective terms are, a model similar to that of Pierce (1996) was run. We used the MOM2 model (Pacanowski 1995) in its test case 2 configuration. This is a pseudo-global model with a grid size of 3° in latitude and 4° in longitude. The surface forcing uses the Hellerman and Rosenstein (1983) wind field and relaxes toward monthly values of the Levitus (1982) temperature and salinity using the simplest interpolation scheme (whose effects will be evident below). We chose a relaxation timescale of 37 days, which, for a surface layer thick-

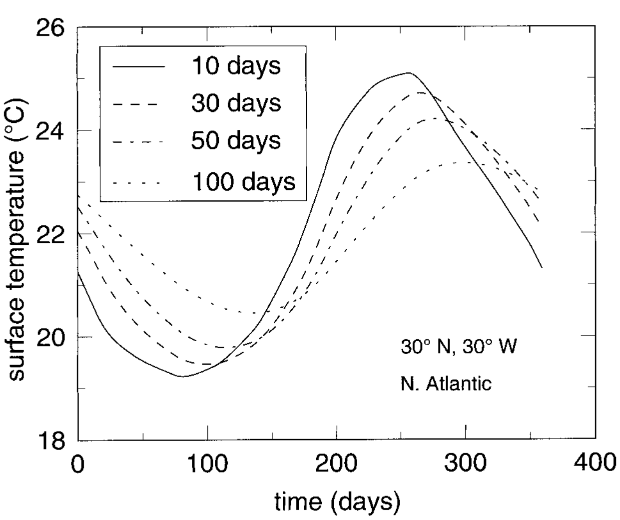

FIG. 3. Annually varying modelled surface temperature at 30°N, 30°W in the North Atlantic, as the relaxation time τ varies.

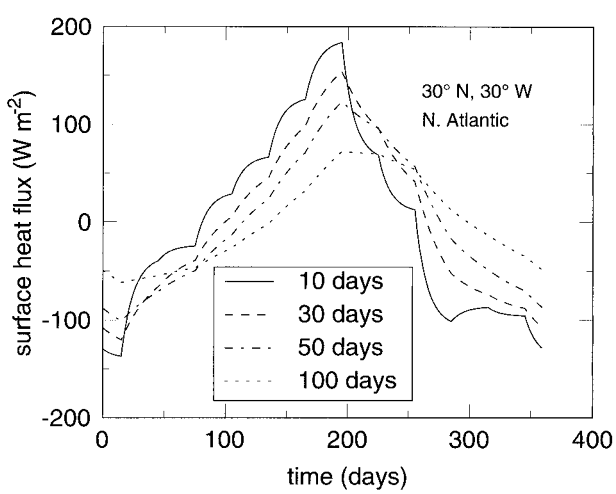

FIG. 4. The surface heat flux modeled at 30°N, 30°W in the North Atlantic, assuming a uniform mixed layer depth of 50 m, as the relaxation time τ varies.

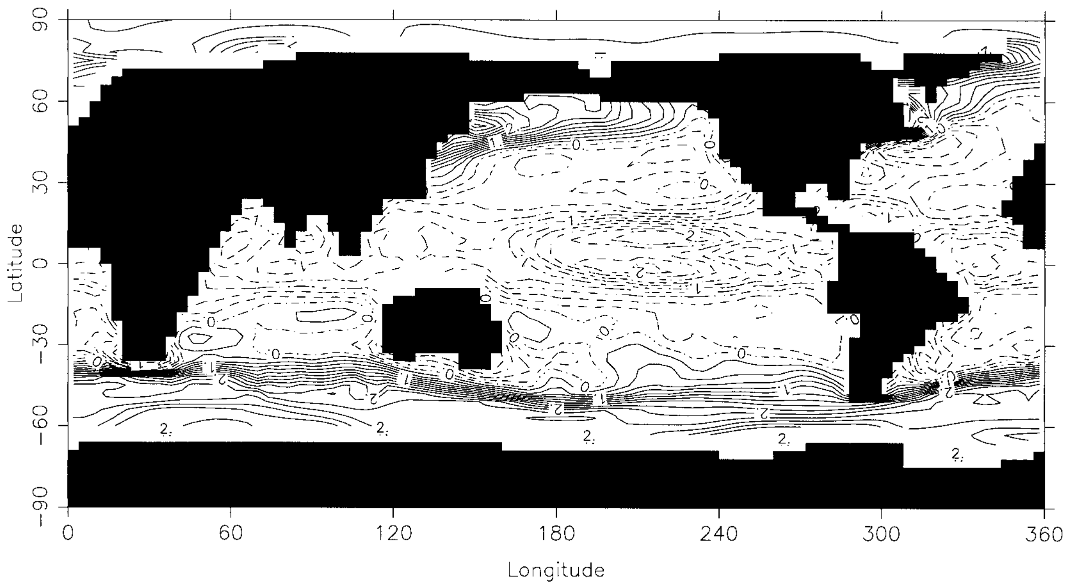

FIG. 5. Error in annually averaged surface temperature in a global model (relaxed case − original case) when forced by relaxation toward observed temperature and salinity values. Contour interval 0.2°C; firm lines are positive, dashed lines negative, and the zero contour is dash-dotted.

ness[4] of 25 m, corresponds to a feedback coefficient of 31 W m^{-2} °C^{-1}. Semtner and Chervin (1992) used 30 days with a 25-m surface layer, while OCCAM (Webb et al. 1997) used 30 days with a surface layer thickness of 20.55 m. These correspond to feedback coefficients of 39 and 31 W m^{-2} °C^{-1}, respectively (OCCAM has no mixed layer parameterization); we chose arbitrarily to follow the latter value of feedback coefficient.

The model fields are initialized with the Levitus (1982) fields and the model then ran for 960 years, which is long enough for the surface fields to have come into approximate equilibrium (the annually averaged rate of heat storage is 0.1 PW). This run was continued for another single year, and the surface heat and salinity fluxes, and the surface temperature and salinity, were stored globally each day.

These values are taken as the ''truth,'' or ''observed,'' fields for the run, although of course they were produced by relaxation itself. Nonetheless, the flux fields stored do produce the surface temperatures stored, which is all we need for later comparison. Furthermore, changing the surface forcing to relaxation, which will now be done, occurs within the same model physics as the original calculation. Thus for comparison purposes, the model is ''perfect'': questions such as overly wide western boundary layers and their effects on flux estimates become irrelevant since both model and perturbation handle such things identically. In this sense, the approach is similar to that used by Greatbatch et al. (1995).

A simple test was to rerun the model from the start, forcing it by the annual cycle of ''observed'' fluxes. Within 50 years, the surface cycle of temperature had appeared everywhere and was maintained during the rest of the integration. Thus the model, forced correctly, was able to reproduce itself.

The model was now run from the same initial (Levitus) condition for another 960 years, but using surface fluxes that were the relaxations toward the daily, annually repeating observed fields as discussed above. The same relaxation time τ of 37 days was used, giving a feedback coefficient of 31 W m^{-2} °C^{-1}, as noted above, which is well within normally used values. Again an extra year was then run, with surface fluxes and fields again stored daily, to permit a comparison.

The question that we may then pose is, how well did the relaxation toward observations reproduce either surface values or surface fluxes from the first run? We first examine annual averages. Our analysis suggests that surface fields will be biased by an amount proportional to the mean advection present so that this bias will be largest in active current systems. Figure 5 shows the annually averaged error in surface temperature (relaxed − original).[5] In midlatitudes, away from western bound-

[4] This calculation is merely one designed to provide fields and forcings for later comparison, so that the value of the timescale is irrelevant. However, we naturally prefer to use forcing fields and parameters similar to that used by other authors, but we note that it is unclear whether it is better to follow the choices of previous general circulation model calculations in terms of feedback coefficient or relaxation timescale.

[5] Since the original model is ''perfect,'' there is no need to compare with actual observations—this is purely a model–model comparison.

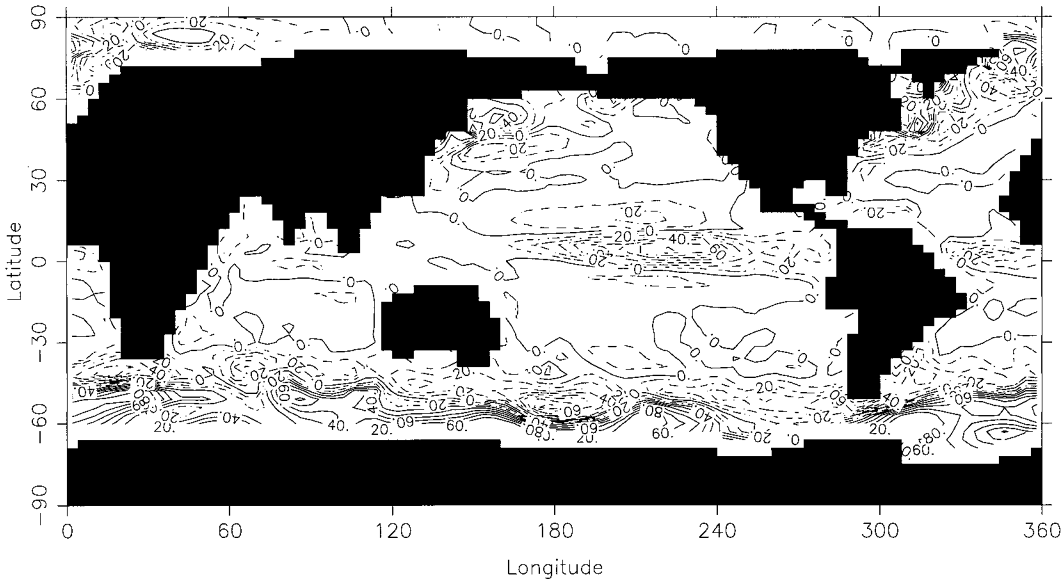

FIG. 6. Error in annually averaged heat flux for the global model in Fig. 5. Contour interval 10 W m^{-2}; contour style as in Fig. 5.

ary currents, the error is usually small (up to 0.4°C). However, in equatorial areas, the surface is too cold by at least 1°C. In western boundary layers (or the coarse version thereof in this model) temperatures are again 1°C or more in error, and much of the Southern Ocean is up to 2°C too warm. Thus our analysis is qualitatively confirmed. Salinity errors are of order 0.1 psu in most regions, with again larger values in the dynamically active regions.

Our prediction that surface fluxes show an annual bias that depends on the timescale for the advective feedback can also be tested. Figure 6 shows the error in annually averaged heat flux. The net heat flux errors are indeed small in most areas, but are typically at least 20–30 W m^{-2} in the active regions, suggesting that feedback times there are not large compared with the relaxation time of just over one month. Counterintuitively, the equatorial heat flux error is negative: the ocean with relaxation is cooler than the "observed" ocean, but since the relaxation is toward the observed temperatures, it is perfectly possible for the heat flux error to remain negative. In the Southern Ocean, large areas receive much more heat than in the "observed" case.

The source for these errors can be seen in Fig. 7, which shows annual cycles at points approximately in the same location as those in Fig. 1. Flaws in the forcing of the observed system are clearly visible, with steps clearly evident in the observed fields at month boundaries; the steps at the Southern Ocean point are presumably related to winter convection. The observed fields also show a much weaker annual cycle than the Levitus and Boyer (1994) data in Fig. 1 (and sometimes with an offset). Nonetheless, defining the observed fields again to be truth, Fig. 7 shows the similar reduction in amplitude and phase shift of the relaxed signal, with (in some cases) a well-defined bias over the year of up to 1°C. The heat fluxes at these locations can differ from the original by 20–40 W m^{-2}, with a net bias near the equator and in the Southern Ocean as noted above. Not all the responses follow the simple advection-free lagged and weaker amplitude pattern, especially in the Southern Ocean. Figure 8 shows these fluxes (and demonstrates the unevenness of the default interpolation routine in MOM2, where end-of-month changes are clearly visible). Again, the seasonal signal in the relaxed case is damped compared with the observed, and is 20–30 W m^{-2} less than the observed in the equatorial and Southern Ocean areas.

These results permit a belated check on the reliability of our simple parameterization of the advective feedback term. For the original calculation, averaging (11) over the annual cycle gives

$$\overline{F_S} + \overline{F_A(T)} = 0, \tag{23}$$

while averaging (16) gives

$$\overline{F_R} + \overline{F'_A(T_M)} = 0, \tag{24}$$

where $F_R = (T - T_M)/\tau$ is the surface flux estimate using relaxation and F'_A is the modified advective flux. Thus the error in surface flux is associated with the error in advective flux by

$$\begin{aligned}\text{error in surface flux} &= \overline{F_R} - \overline{F_S} = \overline{F_A(T)} - \overline{F'_A(T_M)} \\ &= (\overline{T_M} - \overline{T})/\tau_A, \end{aligned} \tag{25}$$

if our parameterization is correct. A plot of surface flux error against SST error is given in Fig. 9, using zonally averaged values. The best straight line fit is shown: the

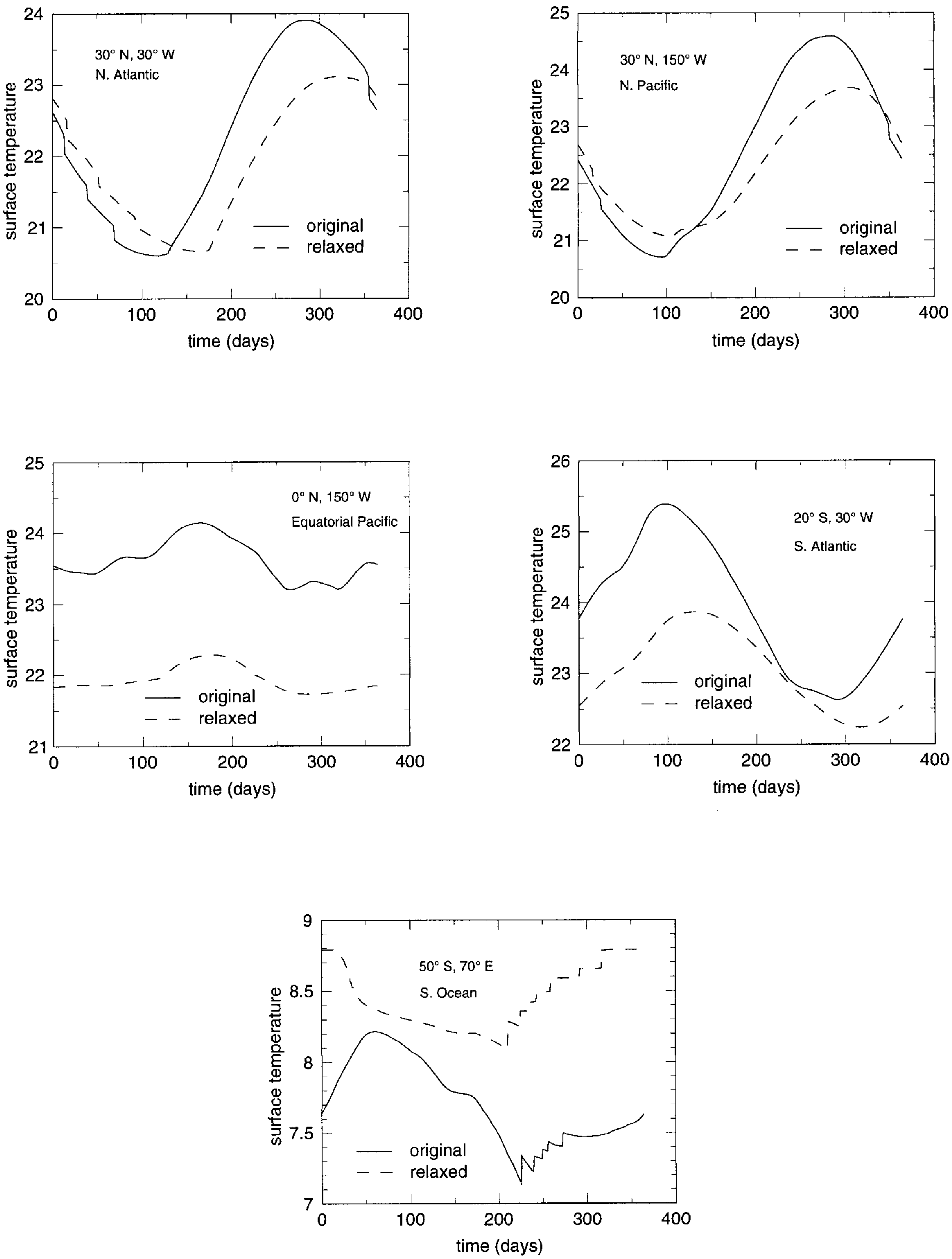

FIG. 7. The seasonal cycle of the observed (firm line) and relaxed (dashed line) surface temperatures at the five locations in Fig. 1.

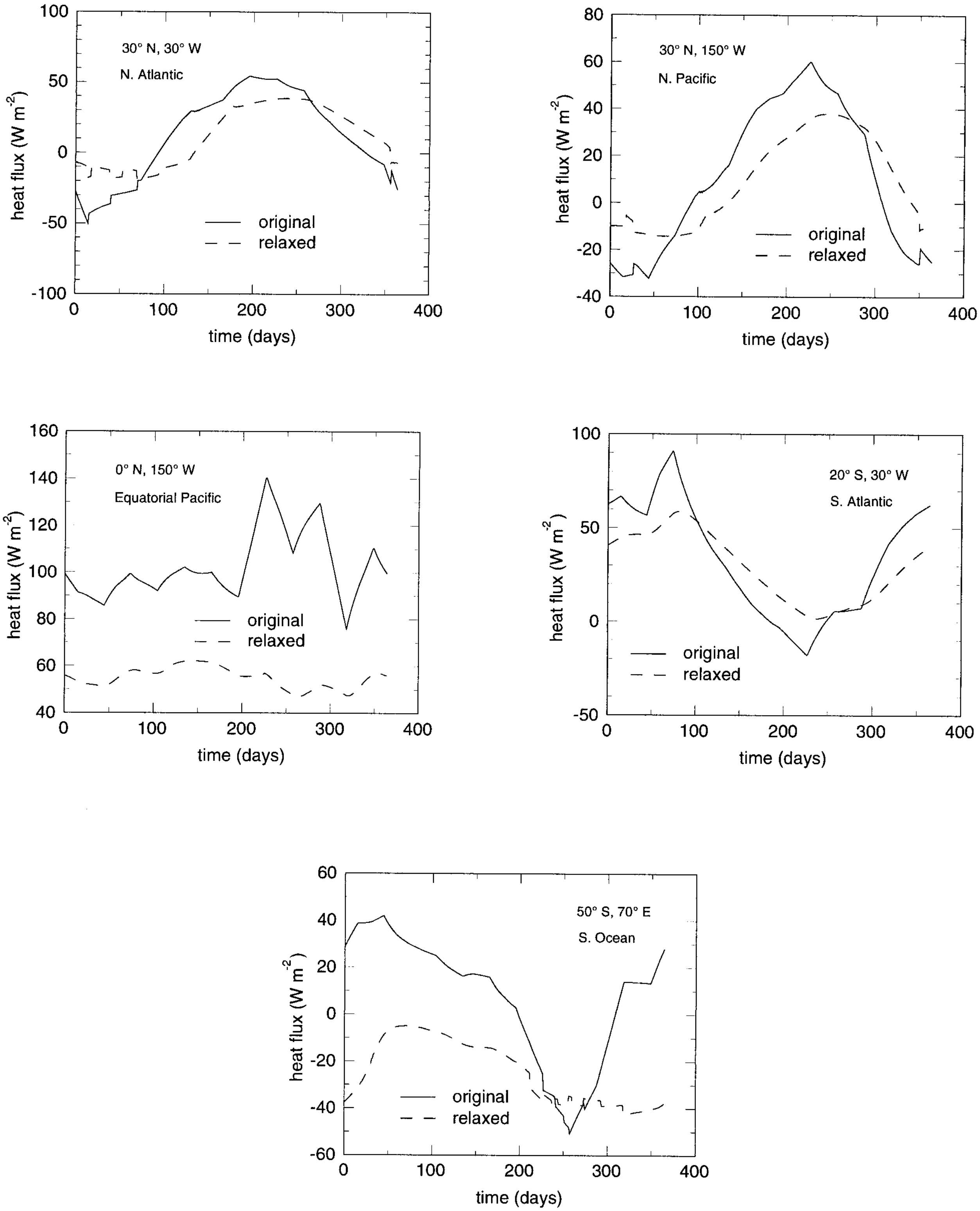

FIG. 8. The seasonal cycle of the observed (firm line) and relaxed (dashed line) surface heat fluxes at the five locations in Fig. 1.

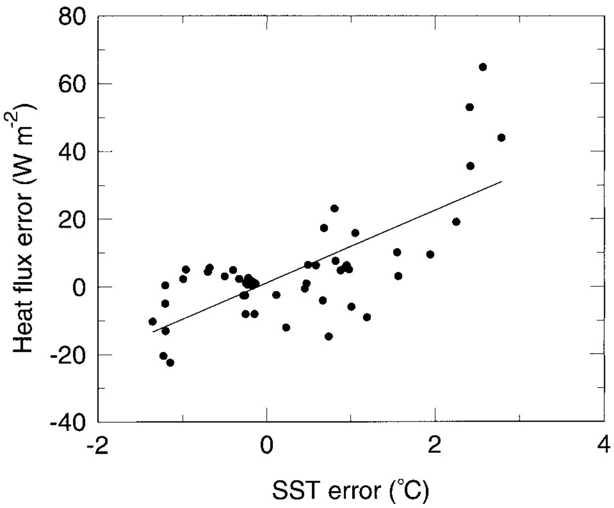

FIG. 9. A scatterplot of zonally averaged surface heat flux error against SST error. The correlation is 0.72. The best fit line is shown, corresponding to an advective feedback coefficient of 10.7 W m^{-2} °C^{-1}, or a feedback time of 111 days.

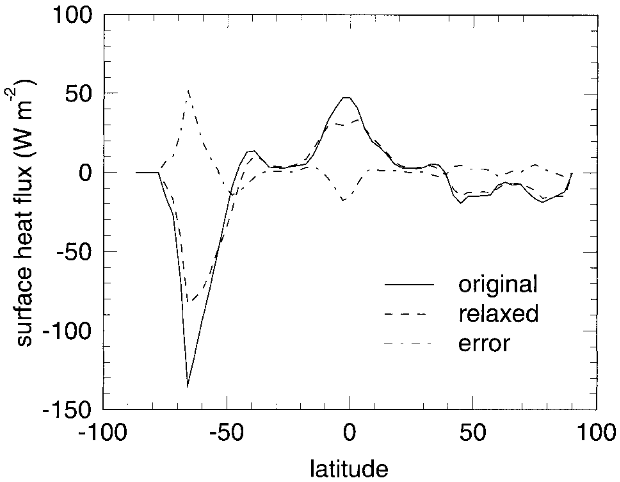

FIG. 10. Longitudinal and annual average of the surface fluxes in observed (firm line), relaxed (dashed line), and relaxed–observed (dash–dotted line) as a function of latitude.

correlation is 0.72, and the gradient of the line corresponds to an advective feedback coefficient of 10.7 W m^{-2} °C^{-1}, or a feedback timescale of 111 days. (The plot is similar if each grid point value is plotted, with predictably more scatter, and a reduced correlation of 0.49.)

Thus our simple parameterization, enforced for the zero-dimensional problems earlier, is confirmed, at least for this coarse resolution model. It is unclear whether this result would hold for finer resolution models. Other possibilities could hold: a suggestion would be a dependence on a northward mean gradient. A similar plot (not shown) reveals no correlation of heat flux error with the northward gradient, however, so for this coarse model, this suggestion does not hold.

The result of the heat flux errors can be quite large even when averaged longitudinally. Figure 10 shows the average surface heat flux (the average includes land points where fluxes are zero); averaged errors reach 20 W m^{-2} again, though they tend to cancel each other longitudinally in the Northern Hemisphere. When integrated northward from the southern boundary, to produce effective poleward heat fluxes (Fig. 11; note that the combination of coarse resolution and large horizontal diffusion produces far too large a southward heat flux in the Southern Hemisphere), the error reaches 0.5 PW in the Southern Hemisphere and is largely cancelled at the equator; thus the error is a noticeable fraction of the signal.[6]

[6] Neither observed nor relaxed northward heat fluxes integrate exactly to zero at the northern boundary even after a 960-year integration; for comparison, errors of up to 0.5 PW were observed after 100 years of integration. However, although the system is not in seasonal equilibrium, the errors in heat flux are far larger than the net error observed by north–south integration.

The salinity signal, and accompanying implied $E - P$ fluxes, have similar behavior and are not shown here.

7. Conclusions

This paper has continued an examination of the utility of boundary conditions involving a relaxation toward observed seasonal values of temperature and salinity, begun in Pierce (1996) and Chu et al. (1998). We have produced a formal method to study the errors induced with such boundary conditions, and shown that in the absence of advection the problem reduces to Pierce's (1996) discussion. In particular, it is then possible to adjust the surface values that are relaxed towards so that the correct answer is found.

However, our main thrust lies in the inclusion of ad-

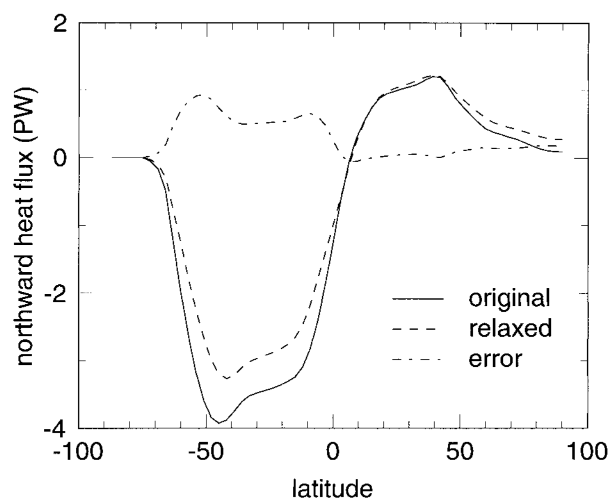

FIG. 11. Implied northward heat fluxes (PW) obtained by integrating the values in Fig. 9 northward.

vective effects. We have shown two things when advection is present: (i) the annual mean tracer field is biased compared with the observations and (ii) the annual mean tracer flux is biased if the surface tracer field has a feedback on the surface advection. Tests with a coarse-resolution global model show that errors are strong (over 1°C, or 30–40 W m^{-2} for heat, with similar findings for salinity) in dynamically active regions, such as the equator, western boundary currents, and the Southern Ocean, and lead to implied northward heat flux errors of up to 0.5 PW. These errors were obtained with feedback coefficients of 31 W m^{-2} °C^{-1} (relaxation times of 37 days for a surface layer 25 m thick). Weaker feedback would give larger errors. Presumably, also, the errors will be even larger in models with finer resolution, and hence stronger advection. Unfortunately the regions where the errors are concentrated are also those most important for water mass transformation and the thermohaline circulation so that use of boundary conditions that relax toward observations will give biased results and may misrepresent the thermohaline circulation.

These results are negative, in the sense that they imply that current practice is unlikely to yield accurate results. But, as we noted, there are no fully reliable and consistent flux datasets available yet, so modelers must use some less than optimal surface condition. Four main possibilities currently exist.

First, the use of relaxation toward an effective surface value, that is, a return to Haney's (1971) original suggestion. This has proven successful in many cases, and was adopted by Barnier et al. (1995), Oberhuber (1988), and others. Second, an alternative approach is to emulate a simple atmospheric boundary layer (which was to some extent behind Haney's approach) and so permit the feedback of surface temperature on the atmosphere. Such thinking was behind Rahmstorf and Willebrand's (1995) formulation of surface flux, for example. Large et al. (1997) find that a boundary condition employing bulk formulae gives more accurate results than a relaxation condition; however Wadley et al. (1996) found the opposite, unless they included freshwater, rather than a saline, flux in their calculations. A slightly different approach was advocated by Schopf (1983), who suggested a relaxation to the atmospheric radiative temperature with a very long relaxation timescale (he did not include salinity effects) using an atmosphere with zero heat capacity.

Third, the system can be coupled to an atmospheric model, thus permitting a complete feedback from ocean to atmosphere. Here, though, the success will depend crucially on the reliability of not just the ocean model, but the atmospheric model as well. If either model is poor, the results will be poor also. Fourth, as discussed earlier, strong surface relaxation could be employed, thus ''forcing'' the model to reproduce observed surface tracers and, to a good degree of approximation, the surface fluxes. However, the physical disadvantages of this scheme, for example, the damping of disturbances and the degradation of western boundary current physics, argue against its use in many cases.

The discussion as to how best to formulate surface boundary conditions for ocean models will continue. At its heart must be continual attempts to improve our knowledge of the surface fluxes themselves.

Acknowledgments. Our thanks to Jeff Blundell, who provided the output from the MOM2 model, and to the referees for many helpful suggestions.

REFERENCES

Barnier, B., L. Siefridt, and P. Marchesiello, 1995: Thermal forcing for a global ocean circulation model using a three-year climatology of ECMWF analyses. *J. Mar. Sys.,* **6,** 363–380.

Chu, P. C., Y. C. Chen, and S. H. Lu, 1998: On Haney-type surface thermal boundary conditions for ocean circulation models. *J. Phys. Oceanogr.,* **28,** 890–901.

DYNAMO Group (Barnard, S., Barnier, B., Beckmann, A., Böning, C. W., Coulibaly, M., De Cuevas, D'A., Dengg, J., Dieterich, C., Ernst, U., Herrmann, P., Jia, Y., Killworth, P. D., Krüger, J., Lee, M.-M., Le Provost, C., Molines, J.-M., New, A. L., Oschlies, A., Reynaud, T., West, L. J., Willebrand, J.), 1997: Final Scientific Report. Berichte aus dem Institut für Meereskunde an der Universität Kiel, Nr. 294, 334 pp. [Available from Institut für Meereskunde, Düstenbrooker Weg 20, 24105 Kiel, Germany.]

Greatbatch, R. J., G. Li, and S. Zhang, 1995: Hindcasting ocean climate variability using time-dependent surface data to drive a model: An idealized study. *J. Phys. Oceanogr.,* **25,** 2715–2725.

Haney, R. L., 1971: Surface thermal boundary conditions for ocean circulation models. *J. Phys. Oceanogr.,* **1,** 241–248.

Hellerman, S., and M. Rosenstein, 1983: Normal monthly wind stress over the world ocean with error estimates. *J. Phys. Oceanogr.,* **13,** 1093–1104.

Josey, S. A., E. C. Kent, and P. K. Taylor, 1998: New insights into the Ocean Heat Budget Closure Problem from analysis of the SOC Air–Sea Flux Climatology. *J. Climate,* **12,** 2856–2880.

Killworth, P. D., 1996: Time interpolation of forcing fields in ocean models. *J. Phys. Oceanogr.,* **26,** 136–143.

Large, W. G., G. Danabasoglu, S. C. Doney, and J. C. McWilliams, 1997: Sensitivity to surface forcing and boundary layer mixing in a global ocean model: Annual-mean climatology. *J. Phys. Oceanogr.,* **27,** 2418–2447.

Levitus, S., 1982: *Climatological Atlas of the World Ocean.* NOAA Prof. Paper No. 13, U.S. Govt. Printing Office, 173 pp.

——, and T. Boyer, 1994: *World Ocean Atlas 1994.* Vol. 4: *Temperature. NOAA NESDIS 4.* ?? U.S. Government Printing Office, 117 pp.

Oberhuber, J. M., 1988: *An Atlas Based on the COADS Data Set: The Budgets of Heat, Buoyancy and Turbulent Kinetic Energy at the Surface of the Global Ocean.* Report 15 Max-Planck Institut für Meteorologie, 20 pp. [Available from Max-Planck-Institut für Meteorologie, Bundestrasse 55, D-20146, Hamburg, Germany.]

Pacanowski, R. C., 1995: MOM 2 documentation, user's guide and reference manual. GFDL Ocean Tech. Rep. 3, 232 pp.

Pierce, D. W., 1996: Reducing phase and amplitude errors in restoring boundary conditions. *J. Phys. Oceanogr.,* **26,** 1552–1560.

Rahmstorf, S., and J. Willebrand, 1995: The role of temperature feedback in stabilizing the thermohaline circulation. *J. Phys. Oceanogr.,* **25,** 787–805.

Saunders, P. M., A. C. Coward, and B. A. de Cuevas, 1999: The circulation of the Pacific Ocean as seen in a global ocean model (OCCAM). *J. Geophys. Res.,* **104,** 18 281–18 299.

Schopf, P. S., 1983: On equatorial waves and El Niño. II: Effects of air–sea thermal coupling. *J. Phys. Oceanogr.,* **13,** 1878–1893.

Semtner, A. J., and R. M. Chervin, 1992: Ocean general circulation from a global eddy-resolving model. *J. Geophys. Res.,* **97,** 5493–5550.

Wadley, M. R., G. R. Bigg, D. P. Stevens, and J. A. Johnson, 1996: Sensitivity of the North Atlantic to surface forcing in an ocean general circulation model. *J. Phys. Oceanogr.,* **26,** 1129–1141.

Webb, D. J., A. C. Coward, B. A. De Cuevas, and C. S. Gwilliam, 1997: A multiprocessor Ocean General Circulation Model using message passing. *J. Atmos. Oceanic Technol.,* **14,** 175–183.

Zhang, S., R. J. Greatbatch, and C. A. Lin, 1993: A re-examination of the polar halocline catastrophe and implications for coupled ocean–atmosphere models. *J. Phys. Oceanogr.,* **23,** 287–299.