An Advective Atmospheric Mixed Layer Model for Ocean Modeling Purposes: Global Simulation of Surface Heat Fluxes*

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ABSTRACT

A simple model of the lowest layer of the atmosphere is developed for coupling to ocean models used to simulate sea surface temperature (SST). The model calculates the turbulent fluxes of sensible and latent heat in terms of variables that an ocean model either calculates (SST) or is forced by (winds). It is designed to avoid the need to specify observed atmospheric data (other than surface winds), or the SST, in the surface flux calculations of ocean models and, hence, to allow a realistic representation of the feedbacks between SST and the fluxes. The modeled layer is considered to be either a dry convective layer or the subcloud layer that underlies marine clouds. The turbulent fluxes are determined through a balance of horizontal advection and diffusion, the surface flux and the flux at the mixed layer top, and, for temperature, radiative cooling. Reasonable simulations of the global distribution of latent and sensible heat flux are obtained. This includes the large fluxes that occur east of the Northern Hemisphere continents in winter that were found to be related to both diffusion (taken to be a parameterization of baroclinic eddies) and advection of cold, dry air from the continent. However, east of North America during winter the sensible heat flux is underestimated and, generally, the region of enhanced fluxes does not extend far enough east compared to observations. Reasons for these discrepancies are discussed and remedies suggested.

1. Introduction

Simulation of sea surface temperature (SST) is both the most important and one of the most difficult tasks demanded of ocean models. It is important because, except for surface roughness, it is the only oceanic variable that affects the atmosphere directly. Prediction of future climate variations depends on our ability to predict SST.

SST is determined by the net effect of surface heat fluxes and ocean dynamics and mixing processes. Determining the surface fluxes has proved to be a particularly thorny problem for ocean modelers. In a coupled atmosphere–ocean model they are determined internally and are fully interactive with the simulated SST and near-surface atmospheric fields. In stand-alone atmosphere models the fluxes over the ocean are calculated with the simulated atmospheric quantities and the imposed SST. In either of these cases the principal problem is the choice, or computation, of the exchange coefficients that appear in the formulas for the fluxes of sensible heat, moisture, and momentum. Atmosphere simulations have been shown to be sensitive to how this is done (e.g., M. J. Miller et al. 1992) and there are a large number of schemes in use (e.g., Sommeria 1988). It is fair to say that it is a source of considerable uncertainty regarding the reliability of model simulations and forecasts.

The ocean modeling case retains all of these problems and adds some. The ocean model needs to know what the fluxes of heat and moisture are at the surface but does not know the atmospheric state that contributes to their determination. The whole problem has frequently been avoided by simply relaxing the simulated SST to its observed value (e.g., Han 1984; Haney 1971). While this may be useful if one is primarily interested in examining the model's circulation and deep-ocean thermodynamic structure, it is totally uninteresting if one's principal goal is to simulate SST. Even in the former case it can present problems. Any systematic error in the ocean dynamics, or near-surface mixing, that would create errors in the simulated SST is masked by generation of an equally incorrect surface flux. Errors in the surface flux will inevitably lead to further errors in the simulation of near-surface mixing processes that depend on the surface buoyancy flux.

An alternative approach is to force the model with either observed heat fluxes (i.e., calculated with the bulk formulas and observed atmospheric variables and SST) or model-generated ones. This inevitably causes

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problems. For most of the ocean advection is small and the SST is determined by a balance of the heat flux and heat storage in the mixed layer. If the flux is imposed then the only way the model can adjust to errors in the flux is through advection and large errors in the SST will result. To avoid this the observed fluxes are usually combined with a term that relaxes the SST to its observed value. This, however, reintroduces the issue of specifying the SST. For example, Gordon and Corry (1991) used this method to illustrate some problems with the dynamics and mixing processes in their ocean model. To that extent it is an acceptable procedure. If we are interested in modeling and predicting the SST, then it is clearly not the right approach.

A third approach (e.g., A. J. Miller et al. 1992) is to compute the fluxes with the full bulk formulas and to specify from data the air temperature and air humidity that appear in these formulas. (Unless otherwise noted, the term humidity refers to specific humidity.) This avoids specifying the SST and would appear to be suitable for cases where the model's simulation of SST is of primary interest. This approach has been criticized by Seager et al. (1988) and Seager and Blumenthal (1994). They argue that, since the observed air temperature and air humidity are observed to closely track the SST, to specify these quantities is tantamount to putting in the answer; that is, it comes close to specifying the SST. This is so because the much greater heat capacity and moisture content of the ocean-mixed layer relative to the atmospheric boundary layer means that surface fluxes adjust the air temperature and humidity to the SST and not the other way around. To specify the thermodynamic state of the near-surface air is to constrain the SST much like a "relaxation to observed SST" term does.

A more useful approach is to compute the surface fluxes in terms of the SST and other variables, such as cloud cover and winds, over which the ocean has no *direct* control. In this case, the air temperature and air humidity will have to be modeled in terms of the SST relying on some assumption about turbulent processes in the atmospheric planetary boundary layer (PBL). Seager et al. (1988) introduced this approach and made the assumption that the air humidity was a fixed proportion of the saturation humidity evaluated at the SST. This assumes the relative humidity is approximately constant. Implicitly it assumes the humidity is given by a balance of surface fluxes and entrainment of dry air from above the PBL. Since Seager et al. (1988) were interested in the tropical Pacific they were able to assume the sensible heat flux was small, and combined this and the longwave radiative cooling into a cooling that was, somewhat arbitrarily, proportional to the SST in degrees centigrade. The solar flux was calculated from a bulk formula and ship-observed cloud cover.

This surface heat flux model, in tandem with a linear, reduced-gravity ocean model that also contained a constant depth mixed layer, produced a reasonable simulation of the seasonal cycle of tropical Pacific SST. Seager (1989) had similar success when the model was applied to simulation of interannual variability in the Pacific. Blumenthal and Cane (1989) introduced a statistical optimization technique to find the optimal values of the uncertain coefficients in the heat flux model. They concluded that in the Pacific the errors in the model SST were of the size that they could be accounted for by errors in the forcing fields (most notably cloud cover). In the Atlantic the simulation was less successful and it was concluded that either the heat flux model or the ocean model were in error.

Seager and Blumenthal (1994) have recently presented improved simulations of the tropical Pacific SST climatology. They used two satellite-derived solar radiation datasets, which considerably reduced the uncertainties in this field compared to previous estimates. They also introduced a new heat flux model that separately accounted for the latent, sensible, and longwave radiative heat losses. The latter was modeled with a bulk formula. The air temperature and air humidity that appear in the calculation of the sensible and latent heat fluxes were modeled in terms of the SST and following the scheme introduced by Seager et al. (1988), rests on the assumption of a one-dimensional balance in the atmospheric PBL. Uncertain parameters were optimized. The SST simulation was noticeably better than those obtained by Seager et al. (1988) or Blumenthal and Cane (1989). This improvement was attributed to the more complete heat flux model (longwave radiative cooling was found to not be a small term) and the satellite solar radiation data. The reduced uncertainties in the heat flux allowed unambiguous identification of errors in the ocean model's treatment of entrainment and ocean mixed layer processes in the eastern equatorial Pacific.

This progress has however been restricted to the tropical oceans. Simulating global SST introduces new problems. For example, the parameterization of the air humidity in terms of the SST used by Seager and Blumenthal (1994) is valid for the Tropics only. Further, the most striking feature of heat fluxes in the midlatitudes is the enhancement of the latent and sensible fluxes over the Kuroshio Current and Gulf Stream during northern winter. This occurs, at least in part, as a result of advection of cold, dry air off the continents and its passage over relatively warm water. In these cases the one-dimensional equilibrium assumption of our earlier work is simply wrong. Even in the Tropics fluxes can be enhanced by flow of dry air off Africa and over the Atlantic (Blumenthal 1990).

To extend our method of heat flux calculation globally we need to replace our PBL equilibrium assumption with a model that includes advective processes.

Such a model was introduced by Blumenthal (1990). This model calculated the air humidity only from a balance of surface fluxes, a flux at the top of the well-mixed layer, and advection by the observed winds. Luksch and von Storch (1992) introduced an advective PBL model of the air temperature over the North Pacific and calculated the specific humidity using observed relative humidity and the modeled air temperature. Kleeman and Power (1995) have also presented an advective model of air temperature and then calculate the specific humidity assuming a fixed relative humidity.

Here we present yet another model in the same spirit. The advantage of ours is that it calculates both the air humidity and temperature and hence determines its own relative humidity. It also does not need to specify the temperature above the PBL as Kleeman and Power (1995) did. The model is conceived as a means for calculating in a realistic way the heat fluxes needed by ocean models. It is designed for use in experiments that examine the ocean's role in climate change, and because of this it adopts a number of quasi-equilibrium assumptions that are valid on monthly and longer timescales. We should make clear that it is not intended as a replacement for atmosphere models in coupled integrations. Instead, it is designed for coupling to an ocean model that is being forced by observed winds, solar radiation, and cloud cover. Hence, together with an ocean model, it can be used to examine the mechanisms of SST variability that arise in climate variability. It cannot be used to unravel the causes of coupled variability or to look at feedbacks between the SST and atmospheric dynamics or cloud cover. For that a fully coupled atmosphere–ocean model is needed.

We show that this model produces a reasonable simulation of the surface fluxes of sensible and latent heat over the global oceans. We argue that it provides a simple way to calculate the fluxes in terms of model SST that is applicable to a wide range of ocean modeling studies. The model is described in the next section, the nonadvective solutions in section 3, the numerical procedure in section 4, and the flux simulations in section 5. Probable causes of errors are discussed in section 6 and a summary provided in section 7.

2. The advective atmospheric mixed layer model

The first issue to deal with is which part of the lower atmosphere to model. Over the tropical and subtropical oceans the PBL is most clearly defined by the trade wind inversion that occurs at the top of a layer of shallow cumulus convection (e.g., Augstein 1978). In regions of the ocean where shallow convection is absent, the PBL is instead defined by a dry convective layer. If we were to define the PBL as including the cloud layer, we would be forced to deal with the subsidence through the trade inversion that balances the radiative cooling, and the evaporative cooling and moistening, at this level (Betts and Ridgway 1988, 1989). That would require a quite complex PBL model coupled to some assumption for the troposphere above.

To avoid this we note that the regions of shallow convection divide into a cloud layer overlying a subcloud mixed layer. In the subcloud layer the potential temperature and humidity are approximately constant, justifying use of the term "mixed layer" to describe this layer (e.g., Nicholls and LeMone 1980). The mixed layer is approximately 600 m deep, which compares well to the depth of dry convective layers. Further, work on dry and subcloud mixed layers shows that certain similarities between the two layers occur, particularly in the relationship between fluxes at the mixed layer top and at the surface (Betts 1976). This suggests adopting a model that represents a dry convective or subcloud layer and uses the same closure assumption in both cases. Indeed it never distinguishes between the two. We use the term "mixed layer" to describe this layer and to distinguish it from the entire PBL.

a. Mixed layer equations

The following derivation closely follows that of Betts (1976). The potential temperature and moisture equations, in pressure coordinates, can be written as

$$\frac{\partial \theta}{\partial t} + \mathbf{u}\cdot\nabla\theta + \omega\frac{\partial \theta}{\partial p} = -\frac{\partial(\overline{\omega'\theta'})}{\partial p} + R, \qquad (1)$$

$$\frac{\partial q}{\partial t} + \mathbf{u}\cdot\nabla q + \omega\frac{\partial q}{\partial p} = -\frac{\partial(\overline{\omega'q'})}{\partial p}. \qquad (2)$$

Here, θ is the potential temperature, q is the specific humidity, $\mathbf{u}$ is the horizontal velocity, ω is the vertical velocity (all time means), and primed quantities denote deviations from the time mean. The overbar indicates that the turbulent quantities are time averaged. The first terms on the right therefore represent the time-integrated effects of vertical transports by turbulence. The term R is the radiative cooling. Assuming that θ, q, and $\mathbf{u}$ are constant in the mixed layer, we can integrate these equations from the surface pressure p_0 to the pressure at the mixed layer top p_B. The mixed layer equations become

$$P\left(\frac{\partial \theta}{\partial t} + \mathbf{u}\cdot\nabla\theta\right) = -\left(\frac{\partial P}{\partial t} + \mathbf{u}\cdot\nabla P - \omega_B\right)\Delta\theta - (\overline{\omega'\theta'})_0 + (\overline{\omega'\theta'})_B + PR, \qquad (3)$$

$$P\left(\frac{\partial q}{\partial t} + \mathbf{u}\cdot\nabla q\right) = -\left(\frac{\partial P}{\partial t} + \mathbf{u}\cdot\nabla P - \omega_B\right)\Delta q - (\overline{\omega'q'})_0 + (\overline{\omega'q'})_B, \qquad (4)$$

where $P = p_0 - p_B$, $\Delta q = q_B - q$, $\Delta\theta = \theta_B - \theta$, and the subscripts 0 and B denote values at the surface and immediately above the mixed layer top, respectively. The turbulent fluxes are represented as (Betts 1975, 1976)

$$F_{0\theta} = -(\overline{\omega'\theta'})_0 = C_0\omega_0(\theta_0 - \theta), \tag{5}$$

$$F_{0q} = -(\overline{\omega'q'})_0 = C_0\omega_0(q_0 - q), \tag{6}$$

$$F_{B\theta} = -(\overline{\omega'\theta'})_B = \omega_B^*\Delta\theta, \tag{7}$$

$$F_{Bq} = -(\overline{\omega'q'})_B = \omega_B^*\Delta q, \tag{8}$$

where ω_B^* is a convective mass flux. We have adopted the traditional bulk formula for the surface fluxes using a velocity scale $\omega_0 = |\mathbf{u}|P/h$, where $|\mathbf{u}|$ is the mean surface wind velocity and h is the depth of the layer. Then substituting (5)–(8) into (3) and (4) we have

$$P\left(\frac{\partial\theta}{\partial t} + \mathbf{u}\cdot\nabla\theta\right) = F_{0\theta} + \left(\frac{\partial P}{\partial t} + \mathbf{u}\cdot\nabla P + \omega_B + \omega_B^*\right)\Delta\theta + PR, \tag{9}$$

$$P\left(\frac{\partial q}{\partial t} + \mathbf{u}\cdot\nabla q\right) = \mathrm{F}_{0q} + \left(\frac{\partial P}{\partial t} + \mathbf{u}\cdot\nabla P + \omega_B + \omega_B^*\right)\Delta q. \tag{10}$$

These equations differ from Kraus and Turner (1967) and Tennekes (1973). In those studies subsidence and advection were neglected and the vertical integral of the time derivative term was assumed to be given by $P\partial\theta/\partial t$, and $(\overline{\omega'\theta'})_B$ by $\Delta\theta\partial P/\partial t$. Hence, the convective mass flux does not appear in those formulations. The formulation here, deriving as it does from Betts (1976), is complete.

The second terms on the right-hand side of these equations represent the fluxes at the mixed layer top. A closure scheme is needed to determine these fluxes.

b. The fluxes at the mixed layer top

First we combine the potential temperature and moisture equations into an equation for virtual potential temperature $\theta_V = \theta(1 + 0.61q)$:

$$P\left(\frac{\partial\theta_V}{\partial t} + \mathbf{u}\cdot\nabla\theta_V\right) = F_{0\theta_V} + \left(\frac{\partial P}{\partial t} + \mathbf{u}\cdot\nabla P + \omega_B + \omega_B^*\right)\Delta\theta_V + PR', \tag{11}$$

where $R' = (1 + 0.61q)R$. The relationship between the surface fluxes and those at the PBL top has been examined by a number of authors. For a dry convective boundary layer, in the absence of advection and subsidence, Tennekes (1973) found that

$$\frac{dP}{dt}\Delta\theta = 0.2(\overline{\theta'w'})_0. \tag{12}$$

That is, the flux at the mixed layer top was a fixed proportion of the surface flux. He justified this on the basis of a turbulent kinetic energy budget and similarity considerations. Note that the flux of heat at the mixed layer top is *downward,* so that both the surface flux and the turbulent flux at the mixed layer top act to warm the layer. They are balanced by radiative cooling. The flux at the mixed layer top is downward because it is assumed, reasonably, that the atmosphere is stably stratified above the mixed layer with the potential temperature increasing with height. Turbulence then brings higher potential temperature air into the mixed layer.

Tennekes' case was for a dry convective boundary layer. What happens in the subcloud layer that underlies areas of shallow convection? It turns out that (12) does not hold very consistently. However, Lilly (1968), also based on consideration of the turbulent energy equation, suggested a closure of the same form but defined using the *virtual* potential temperature, which is the relevant quantity when considering a moist layer. Lilly's paper was the first to suggest such a closure and he considered two limiting cases, one of maximum entrainment in which the flux at the mixed layer top equaled the surface flux, and a minimum entrainment case in which the flux at mixed layer top was zero. Betts (1973) suggested a value between these two extremes. Subsequently, Betts (1976) and Nicholls and LeMone (1980) confirmed such a relation for the virtual potential temperature fluxes, in the presence of subsidence but in the absence of advection, implying

$$\left(\frac{dP}{dt} + \omega_B + \omega_B^*\right)\Delta\theta_V = \beta_V F_{0\theta_V}. \tag{13}$$

Betts suggested $\beta_V = 0.21 \pm 0.03$, while Nicholls and LeMone suggested $\beta_V = 0.17$. These values fall between Lilly's (1968) limits and are close to those found for the dry convective layer by Tennekes (1973). While in the subcloud layer, the sensible heat flux does not decrease linearly with height as in a dry layer, it appears that the virtual potential temperature flux behaves much like the potential temperature flux in the dry layer. [However, Betts and Ridgway (1989) have used the Tennekes relation to close the sensible heat flux in the subcloud layer of their model of shallow convection, apparently with success.]

Using the relation (13) and letting dP/dt denote a Lagrangian derivative (i.e., $d/dt = \partial/\partial t + \mathbf{u}\cdot\nabla$), the virtual potential temperature equation becomes

$$P\left(\frac{\partial\theta_V}{\partial t} + \mathbf{u}\cdot\nabla\theta_V\right) = (1 + \beta_V)F_{0\theta_V} + PR'. \tag{14}$$

This can be solved for θ_V if P, the winds, and the radiative cooling are known. To get the latent and sensible heat fluxes we solve the q equation and use the relationship between θ_V and q to derive θ. To solve the

q equation requires a closure for the moisture flux at the PBL top.

There is little guidance about how to do this. Indeed Nicholls and LeMone (1980) found no simple relation between the fluxes of moisture at the surface and cloud base. Betts (1976), Albrecht et al. (1979), Albrecht (1979) and Bretherton (1993) all avoid the need to specify this relation by solving for the humidity in the cloud layer. However, this introduces the need to know the fluxes at the inversion base that are derived, in part, by specifying the thermodynamic structure above the inversion. We wish to avoid the need to specify the atmospheric structure because, for the nonidealized situations we are considering, it is not clear how to do so without resorting to a complete atmosphere model.

Our approach is based on the observation that the decrease in humidity between the subcloud layer and the cloud layer, or across the top of a dry convective layer, is a characteristic feature of the marine mixed layer. Thus, by setting $\Delta q = q_B - q = (\gamma - 1)q$ with γ a constant, we simply relate q above the mixed layer to its mixed layer value. This is at least qualitatively realistic and, as we shall see, can be made quantitatively so as well.

The models mentioned above also solve for the cloud mass flux and mixed layer depth, which again requires knowledge of the atmosphere above the mixed layer. Wishing to avoid this, we adopt a more empirical approach and assume that

$$\left(\frac{\partial P}{\partial t} + \mathbf{u}\cdot\nabla P + \omega_B + \omega_B^*\right) = \delta C_0\omega_0. \qquad (15)$$

This relates the net entrainment at the mixed layer top to the generation of turbulence at the surface. With these assumptions, and setting $\mu = -(\gamma - 1)\delta$, the humidity equation becomes

$$P\left(\frac{\partial q}{\partial t} + \mathbf{u}\cdot\nabla q\right) = C_0\omega_0 q_0 - C_0\omega_0(1 + \mu)q. \qquad (16)$$

It should be noted that we only need to specify a value for the parameter μ and do not need to specify values for γ and δ. In the next section we describe how we chose a value for μ. We will consider steady solutions for a mixed layer of fixed pressure depth (i.e., $\partial P/\partial t = \mathbf{u}\cdot\nabla P = 0$). In this case the equations for q, θ_V, and θ reduce to

$$P\mathbf{u}\cdot\nabla\theta_v = (1 + \beta_V)C_0\omega_0(\theta_{V0} - \theta_V) + PR', \qquad (17)$$

$$P\mathbf{u}\cdot\nabla q = C_0\omega_0 q_0 - C_0\omega_0(1 + \mu)q, \qquad (18)$$

$$\theta = \frac{\theta_V}{(1 + 0.61q)}. \qquad (19)$$

These equations can be solved if the surface temperature, winds, and radiative cooling are known. Since the radiative cooling will be a fixed heat loss, or perhaps an imposed cooling that depends on cloud cover only, this achieves our goal of a means of calculating the fluxes without needing to specify anything about the atmospheric thermodynamic structure but including the effects of advection. It contains an uncertain parameter μ, and to provide a value for this we will first consider the nonadvective solutions.

3. Consideration of nonadvective solutions

Over most of the oceans, away from coasts and regions of strong SST gradients, the advection is small and the surface fluxes are balanced one-dimensionally by fluxes at the mixed layer top and, for temperature, by radiative cooling. In this case Eq. (17) for the flux of virtual temperature reduces to

$$C_0\omega_0(\theta_{V0} - \theta_V) = \frac{PR'}{(1 + \beta_V)}, \qquad (20)$$

and (18) gives the equilibrium humidity as

$$q = \frac{q_0}{(1 + \mu)}. \qquad (21)$$

Since β_V is a constant, the first of these two equations tells us that the spatial pattern of the virtual temperature flux (buoyancy flux) is given by the spatial pattern of the radiative cooling. In equilibrium situations the radiative cooling does not vary much in space so the virtual temperature flux is also relatively constant in space. If we were to use the same closure for the potential temperature equation (e.g., Betts and Ridgway 1989), we would find that the sensible heat flux, and by implication the latent heat flux, would also have the spatial pattern of the radiative cooling. We know from observations that this is not the case since both of these have maximums in regions of strong winds (e.g., Esbensen and Kushnir 1981; Weare et al. 1980). The closure for the moisture flux at the mixed layer top that we adopted was chosen to capture this observed correlation as we now illustrate.

Over most of the oceans the relative humidity is observed to be about 75%–80% (e.g., Kleeman and Power 1995). The equilibrium relative humidity, expressed as a fraction, in this model is approximated by $(1 + \mu)^{-1}$, suggesting $\mu = 0.25$ is a reasonable choice (cf. Blumenthal 1990). The evaporative flux is then given by

$$C_0\omega_0(q_0 - q) = \mu C_0\omega_0 q, \qquad (22)$$

which has the desired proportionality to wind speed. This will also mean the sensible heat flux will be related to the wind speed.

By adopting the usual closure for θ_V, but a different closure for q, both the sensible and latent heat fluxes depend on the wind speed as observed. The pattern of the θ_V flux is still given by the radiative cooling. This indicates the close coupling between the energy gains by turbulent entrainment and surface fluxes, and energy loss by radiation (Sarachik 1978). The closure for humidity can hence be made both qualitatively and quantitatively reasonable in the sense that the equilibrium relative humidity will be close to the mean observed value of 80%.

4. Solution of the advective mixed layer equations

Equations (17)–(18) are nonlocal and must be solved by a numerical procedure. Since, when coupled to an ocean model, they will need to be solved quite frequently, we must find an efficient means for their solution. Here we show how, at the cost of introducing some small erroneous terms, the two equations can each be transformed into two tridiagonal systems that can be solved with little computational expense.

First we include Laplacian diffusion of virtual potential temperature and humidity with an eddy diffusivity ν. This is included to parameterize the effects of baroclinic eddies in midlatitudes. We then define the following scaled quantities:

$$(u', v', \nu') = \frac{(u, v, \nu)P}{[(1 + \beta_V)C_0\omega_0]}, \quad (23)$$

$$(u'', v'', \nu'') = \frac{(u, v, \nu)P}{[(1 + \mu)C_0\omega_0]}. \quad (24)$$

Using these, and with a little rearrangement, the θ_V and q equations can be written as

$$(1 + \mathbf{u}'\cdot\nabla + v'\nabla^2)\theta_V = \theta_{V0} + \frac{PR'}{(1 + \beta_v)C_0\omega_0}, \quad (25)$$

$$(1 + \mathbf{u}''\cdot\nabla + v''\nabla^2)q = \frac{q_0}{1 + \mu}. \quad (26)$$

We illustrate our solution procedure with the q equation. We propose to solve this equation by factoring as

$$\left(1 + u''\frac{\partial}{\partial x} + \nu''\frac{\partial^2}{\partial x^2}\right)\left(1 + v''\frac{\partial}{\partial y} + \nu''\frac{\partial^2}{\partial y^2}\right)q = \frac{q_0}{1 + \mu}. \quad (27)$$

This introduces a number of erroneous cross terms. It is easy to show by scaling arguments that, for typical spatial scales of 1000 km, the erroneous terms are all at least four orders of magnitude smaller than the terms in the original equation. For much smaller scales, as may be encountered in regions of advection off the coasts of North America and Asia, the erroneous terms are still three orders of magnitude smaller. The advantage of this factorization is that the advection–diffusion equation can be solved quickly and efficiently as two tridiagonal systems. After discretization with an upwind difference scheme, both of the brackets in (27) form a tridiagonal matrix that can be inverted cheaply.

Observed values of θ_V and q are specified around the coasts and are taken from the climatology of European Centre for Medium-Range Weather Forecasts (ECMWF) analyses for 1985–92. The value of the exchange coefficient C_0 is taken to be 0.0014, and the value for ν is 4×10^6 m^2 s^{-1}. This value is well within the range of observed values, as interpreted from the results of Lau and Wallace (1979) by Gill (1982, 591). The radiative cooling is a fixed 2°C day^{-1}. We choose a mixed layer 60 mb deep in accord with that observed. The SST, advecting winds, and the wind speed are also monthly climatologies of the ECMWF analyses. Apart from the SST all values are for the 1000-mb level of the ECMWF analyses. If the first solution produces regions with $\theta_V \geqslant \theta_{V0}$, then we assign a smaller exchange coefficient of 0.00075 to account for the atmospheric stability and complete a single iteration. The calculation is performed on a uniform 2.5° × 2.5° grid.

5. Simulation of latent and sensible heat fluxes

For reference, the surface fluxes of sensible and latent heat for January and July, as calculated from ship-observed variables by Esbensen and Kushnir (1981),

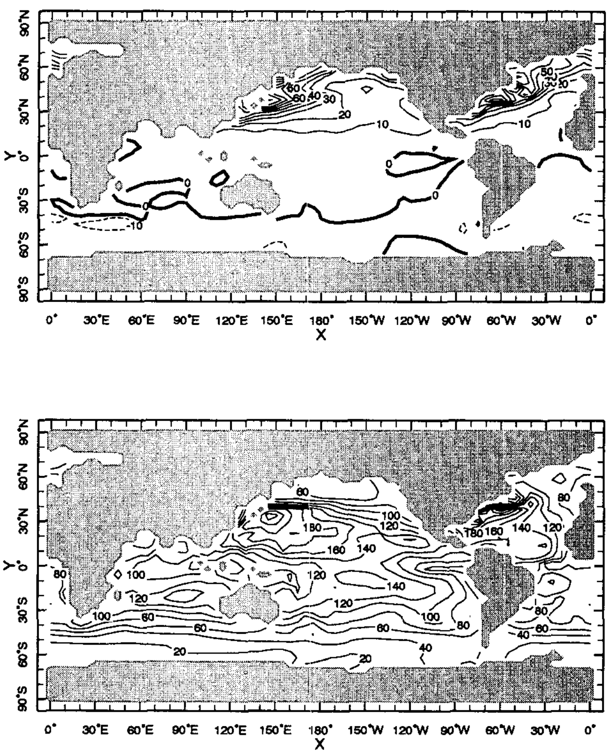
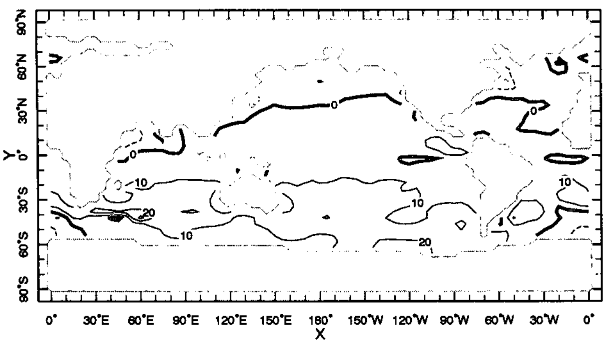


FIG. 1. (a) Sensible heat flux and (b) latent heat flux for January according to Esbensen and Kushnir (1981).



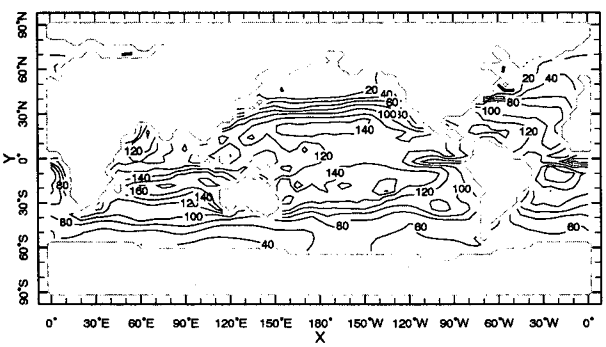


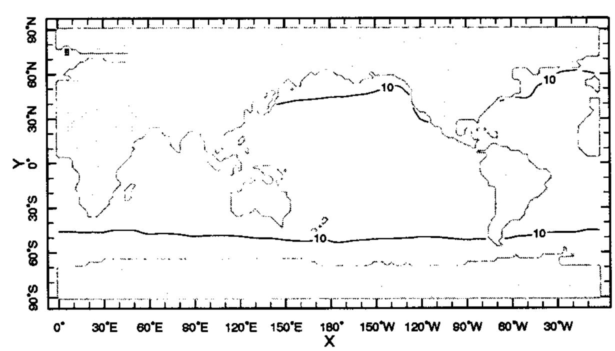
FIG. 2. (a) Sensible heat flux and (b) latent heat flux for July according to Esbensen and Kushnir (1981).

are shown in Figs. 1 and 2. There are many problems involved in calculating fluxes (e.g., Blanc 1987), and we do not present these as the "truth" but merely as a rough indication of what the fluxes are.

The first case we will consider is one in which there is no diffusion and no advection. In this case the fluxes are in local one-dimensional equilibrium. The fluxes of sensible and latent heat are shown in Figs. 3 and 4 for January and July. Comparing first the sensible fluxes, we see that they are approximately correct in July and in the Southern Hemisphere in January, but completely miss the observed maxima of the heat flux off the eastern coasts of Asia and North America in northern winter. This is also where the largest errors occur in the latent heat flux, although the underestimation is less severe. Clearly, the large fluxes off the wintertime Northern Hemisphere continents are a nonlocal phenomena.

These localized and expected errors aside, the local balance produces reasonable fluxes almost everywhere else. For example, in the tropical Pacific the model produces maximums in the latent heat flux north and south of the equator (although the southern one is too small) and minimums over the cold tongue and in the west Pacific. The same kind of patterns appear in the tropical Atlantic. Fluxes over the Indian Ocean are also reasonably simulated, although the modeled Gulf of Arabia maximum is larger than the observed one. Wintertime fluxes over the Southern Ocean also agree with those observed, although the paucity of data here means that the data are highly unreliable. In summer, the fluxes over the Southern Ocean are too large, which may be related to a higher observed relative humidity than in our equilibrium assumption (see, for example, Kleeman and Power 1995). In all these places, model–data differences are within the expected uncertainty of the data. Apparently the areas of active advection are limited, and the equilibrium assumptions that have previously been invoked in our ocean modeling work are justified for most areas of the ocean. Clearly, the climatically interesting and active western boundary current regions are not among them.

Next we include horizontal diffusion in the model. Boundary conditions are set so that there is no diffusion across the coasts. The fluxes are shown for the same months in Figs. 5 and 6. Maximums in the sensible heat flux now occur off the coasts of Asia and North America in January, although they are a lot weaker than observed. Since there is no diffusion across the coast, this does not occur by spreading of dry continental air over the nearby ocean. Instead it arises from the mixing of air across the very sharp SST gradients. In the previous local equilibrium case, the air temperature and humidity track the SST and exhibit the same



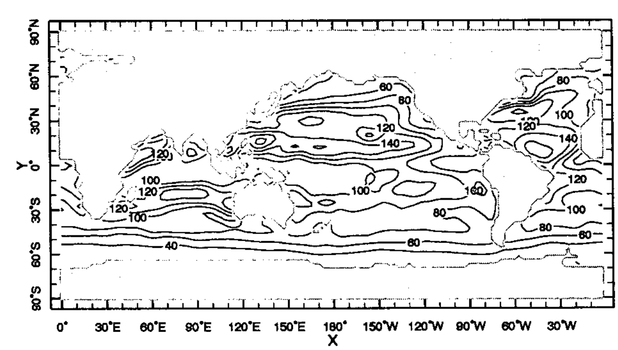


FIG. 3. (a) Sensible heat flux and (b) latent heat flux for January as computed by the model with no diffusion or advection.

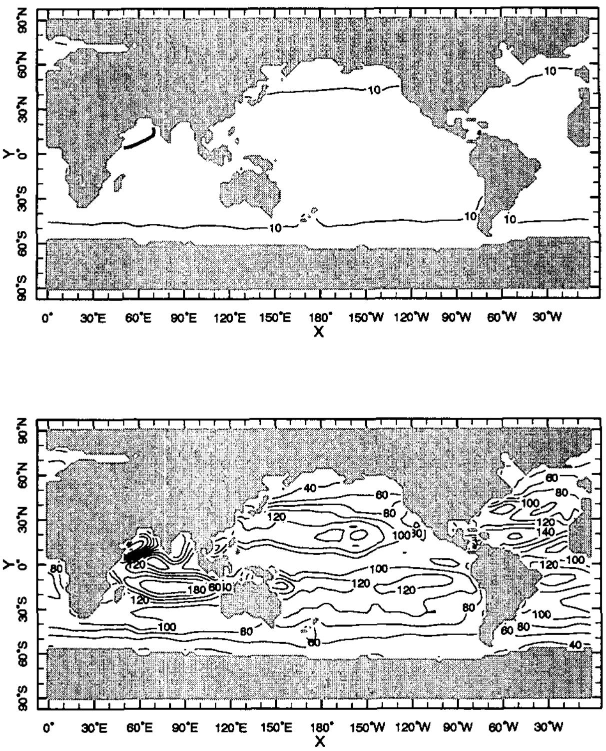


FIG. 4. Same as Fig. 3 but for July.

gradients. The addition of horizontal mixing moves some colder, drier air over the warmer waters to the south and substantially enhances the fluxes. There is little doubt that during winter the very active baroclinic eddies in this region will act in this way. Problems still remain and include an insufficient intensification of the sensible heat fluxes in each northern ocean and of the latent heat flux over the Kuroshio in January.

Next we include advection of temperature and humidity. The fluxes for this case are shown in Figs. 7 and 8 for the same months. It is clear that this results in a substantial enhancement of the fluxes off the Asian coast. In the North Atlantic, the effect is more marked on the sensible heat flux than on the latent heat flux. Over the Kuroshio, the latent heat flux has about the same magnitude as the estimates of Esbensen and Kushnir (1981), although the sensible flux reaches only 60 W m^2 as opposed to 90 W m^2. But even that level of discrepancy is within the range of uncertainty of the flux estimates (see Blanc 1987; Weare and Strub 1981; Weare et al. 1981).

There is a systematic overestimation of the latent heat flux over the tropical Pacific cold tongue. The equilibrium solution is most in error in this region. This is because the observed relative humidity over the cold tongue is higher than our assumed equilibrium value and reduces the latent heat flux. Addition of either diffusion or advection does reduce the fluxes, although not to the magnitude observed. This occurs because of transfer to the cold tongue region of warmer and moister air from surrounding areas. This is a process similar to that suggested by Wallace et al. (1989). They also suggested that stabilization of the lower atmosphere by warm, moist advection leads to decoupling of the lower part of the boundary layer from that above. Their results indicate a highly complex atmospheric PBL in this region and it is not surprising that our simple model is only able to capture some of the processes operating.

We now turn to a closer examination of the fluxes over the North Atlantic in winter. Figure 9 shows the Esbensen and Kushnir (1981) estimates for the North Atlantic, and Fig. 10 repeats the results shown in Fig. 7 but also for the North Atlantic only. Clearly, the general pattern of the latent heat flux is captured with a maximum following the course of the Gulf Stream and including the minimum off the coast of Nova Scotia. The most notable differences are that the observed maximum latent flux is farther off the coast than modeled. The tongue of low fluxes at around 30°N is more marked than in the data, but this comparison may be related to the low spatial resolution (5° × 5°) of the observed data. Another noticeable difference is that the modeled latent heat flux is a maximum off the west coast of Africa, but this does not appear in the data.

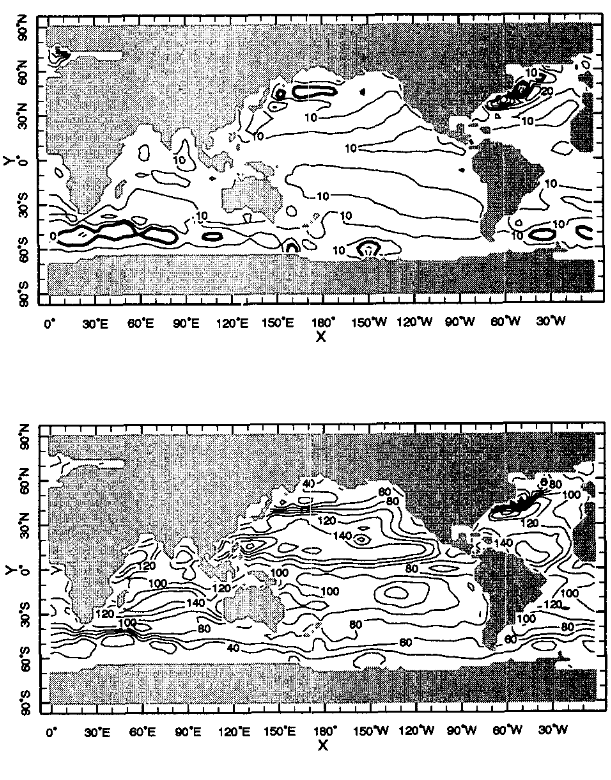


FIG. 5. (a) Sensible heat flux and (b) latent heat flux for January as computed by the model with diffusion but no advection.

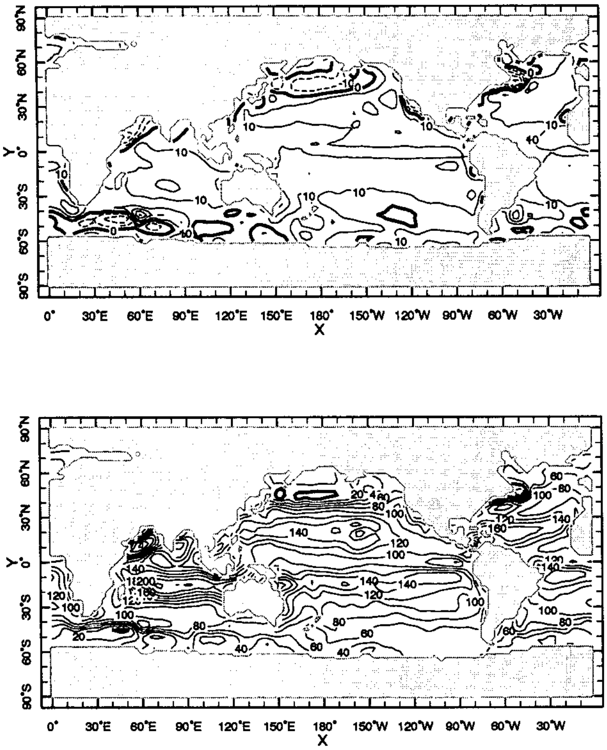


FIG. 6. Same as Fig. 5 but for July.

The modeled maximum is unavoidable. The African landmass is very dry at these latitudes and the ECMWF-analyzed winds blow off the coast and over the ocean with large latent heat fluxes resulting. It is likely that the data analysis has inadvertently smoothed out this maximum. Corroborating evidence that the large fluxes off Africa may be real are provided by the ocean model simulations of Blumenthal and Cane (1989). They used the equilibrium assumption of Seager et al. (1988) to model the latent heat fluxes and found that the model SST was too warm off West Africa (amongst other problems). Lower humidity resulting from advection of dry air would increase the fluxes and cool the SST.

Off North America the simulated sensible heat fluxes are in very good agreement with observed. The only error is that, together with the latent heat fluxes, the near-coastal maximum does not extend far enough east over the ocean.

We have calculated the area-weighted root-mean-square (rms) difference between the modeled and Esbensen and Kushnir (1981) observed latent heat fluxes. It was found to be 28 W m^{-2} in January and 34 W m^{-2} in July. The values for a comparison of the latent heat flux estimates of Esbensen and Kushnir (1981) with those of Oberhuber (1988) for the same months were 17 W m^{-2} and 15 W m^{-2}. Over western boundary currents the difference between the two datasets can reach 90 W m^{-2}. It should be remembered that the two datasets are not independent because they both use the same ship-reported data.

Weare et al. (1981) have estimated from an error analysis that the standard deviation of the error in bulk formula estimated latent heat fluxes is 20 W m^{-2}. However, this puts more faith in the accuracy of individual estimates than Blanc (1987) suggests is appropriate. Blanc offers an analysis of errors introduced by the accuracy of the original flux measurements used as the basis for the bulk method, the choice of bulk scheme, the accuracy of the sensors aboard the ship, and ship distortions. He concludes the rms error of individual latent heat flux estimates, for typical magnitudes of about 120 W m^{-2}, is around 50 W m^{-2}. Such a large value suggests that the error estimate of Weare et al. (1981) is a lower bound. We conclude that the differences between the fluxes modeled here and the observed data are within the uncertainties of the data.

The sensitivity of the modeled fluxes to uncertain parameters are easily understood. For example, changing the value of μ alters the equilibrium relative humidity. The latent heat flux increases as the equilibrium relative humidity decreases. Increasing the pressure depth P of the mixed layer increases the effects of advection and enhances the fluxes over the western

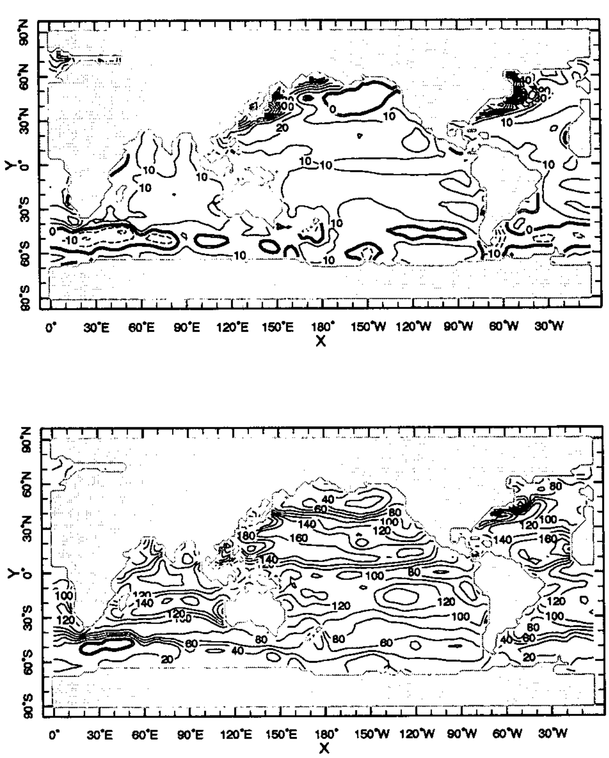


FIG. 7. (a) Sensible heat flux and (b) latent heat flux for January as computed by the model with diffusion and advection.

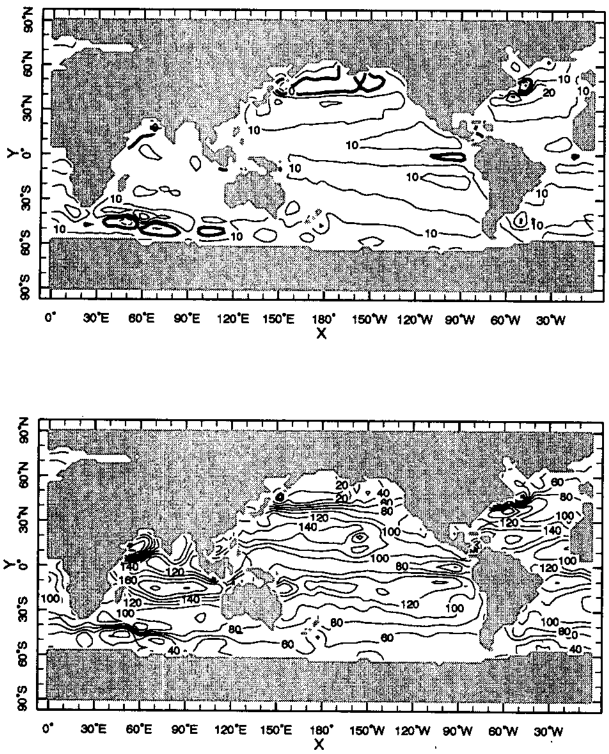


FIG. 8. Same as Fig. 7 but for July.

dient of moisture, improves the latent heat fluxes but degraded the sensible heat fluxes.

This last observation suggests some other problem with the model construction. A number of possibilities come to mind. First there is the possibility that the radiative cooling is erroneous. We assume a fixed 2°C day^{-1} cooling. In reality the cooling depends on the absolute values and vertical structure of temperature and moisture. For example, it is possible that in regions of dry advection the more rapid moistening of the mixed layer increases the vertical gradient of moisture. As a result, an optically thickening layer lies below a relatively transparent layer, which may enhance the radiative cooling of the layer. There is no observational data available to test this. We instead examined the results of a long run of the Geophysical Fluid Dynamics Laboratory GCM (Lau and Nath 1994) and computed the radiative cooling of the lowest 60 mb. Off Japan there was a definite maximum in radiative cooling in winter (up to 4°C day^{-1}) but there was no such maximum off North America. In general, the horizontal structure of the radiative cooling was indistinct, showing no simple correlation with cloud cover or anything else. Of course this may be a model artifact and it would be interesting to see the results of other models.

Even if there was some structure to the radiative cooling it is unlikely to be of much significance over boundary currents. While this brings the fluxes more into agreement with the observations, it is obtained at a cost. Referring back to the discussion of the fluxes, in regions where the advective effects are small, we note from (20) that the flux of virtual potential temperature will increase with P. However, from (22) we see that the latent heat flux does not increase. Since the virtual potential temperature flux is comprised of both the latent and the sensible heat flux, the latter must increase with P. This occurs because the mass being radiatively cooled increases, which requires that the fluxes of sensible heat at the surface and mixed layer top increase to balance. In summary, mixed layers thicker than 60 mb improve the fluxes near the western boundaries but lead to sensible heat fluxes over the open ocean in excess of those observed. In future work we intend to use the optimization technique of Blumenthal and Cane (1989) to find optimal values for these parameters.

6. Discussion of remaining errors

The heat flux errors in the model are concentrated in regions of cold and dry advection over warm western boundary currents. Elsewhere in the ocean the errors are not particularly significant. We have found that inclusion of an equation for the humidity above the mixed layer, which allows for a variable vertical gra-

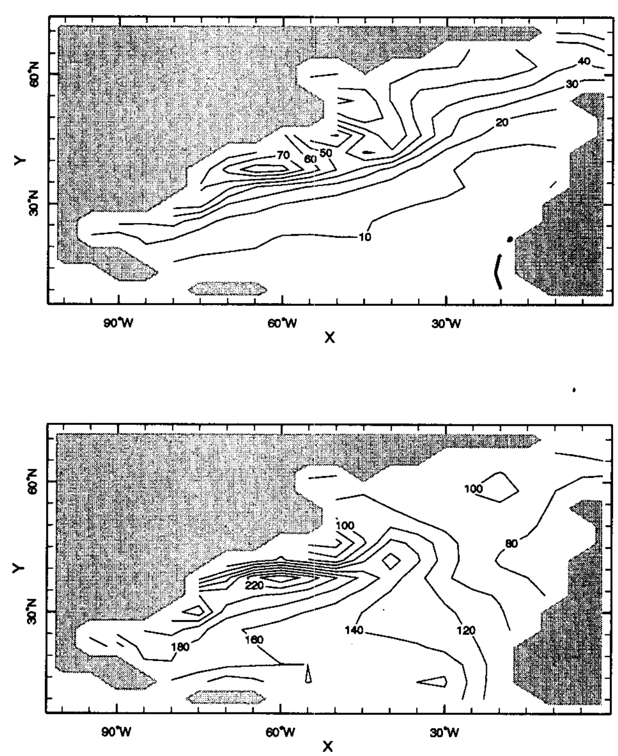


FIG. 9. (a) Sensible and (b) latent heat fluxes over the North Atlantic for January according to Esbensen and Kushnir (1981).

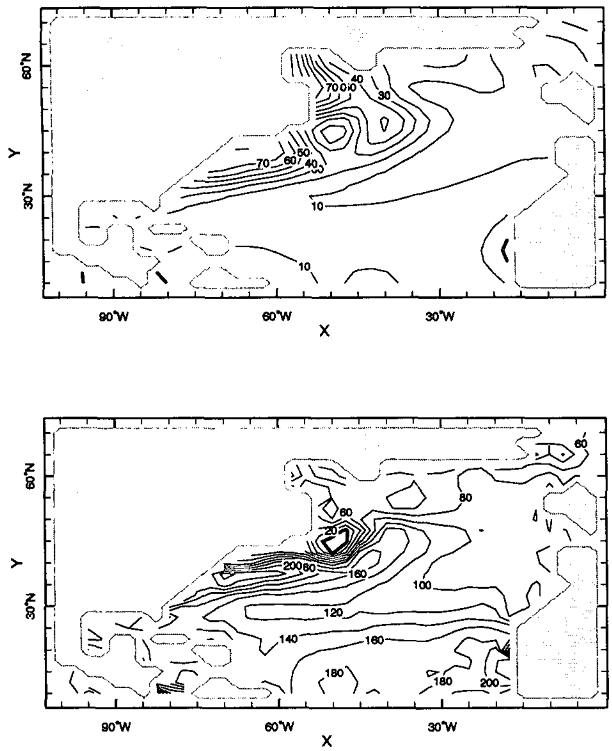


FIG. 10. (a) Sensible and (b) latent heat fluxes over the North Atlantic as computed by the model with advection and diffusion.

the wintertime western boundary currents. The observed surface sensible heat fluxes can reach 100 W m^{-2}, which would warm a 60-mb thick layer by about 15°C day^{-1}, which is much larger than can be balanced by radiative cooling. Instead the primary balance is between cold advection and warming by turbulent fluxes. At this point, within the kind of model we use here, we cannot justify a more complex treatment of the low-level radiative cooling.

We have shown how the diffusion is responsible for some of the observed structure in the modeled heat fluxes. This raises the question of whether the assumed diffusivity form is adequate for parameterizing the effects of horizontal fluxes by transient eddies. The data analysis of Lau and Wallace (1979) did show a correlation between the eddy flux of heat and the local temperature gradient. However, they found no simple functional form that could express the relation between the two and hypothesized that the flux also depended on some measure of the vertical stratification. Clearly, the diffusivity parameterization we use is only a crude approximation to the effects of transients and will introduce errors that are difficult to quantify.

Another possible cause of error is the constant-depth mixed layer. Consideration of the equilibrium solution shows that, as the mixed layer depth increases, the mass being radiatively cooled increases and the surface flux must increase to balance it. The equilibrium humidity does not depend on the layer depth within our approximation. We can derive realistic sensible heat fluxes over the Gulf Stream by increasing the mixed layer depth, but a globally constant depth increase would raise the fluxes everywhere, which is not desired. A reasonable sensible heat flux would result if the mixed layer were deepest at the edges of the continents and then shallowed eastward. In fact this is observed to be the case. Grossman and Betts (1990) have examined aircraft flight data during a cold air outbreak off the Carolina coast. The mixed layer does shallow from 1000 m to 500 m as a shallow cloud layer develops above it off the coast. With our approximations, such a variation of mixed layer depth would lead to increased fluxes off the wintertime coasts. This suggests that a variable-depth mixed layer should be the next improvement attempted.

A third possibility is that the closure on the fluxes at mixed layer top is simply wrong in regions of strong advection; that is, there is no simple relationship between the surface fluxes and those at the mixed layer top. Most of the data on which our closure is based was taken in regions of weak advection so this certainly cannot be excluded. However, both Chou and Zimmerman (1989) and Grossman and Betts (1990) found the same closure on θ_V did apply in the highly advective environments of extreme cold air outbreaks. In summary, the neglect of mixed layer depth variations seems the most likely cause of error.

7. Conclusions

We have presented a simple model of the atmospheric mixed layer that can be used to compute surface fluxes of sensible and latent heat. The model is intended for coupling to ocean models used for examining the ocean's role in climate. The fluxes are calculated in terms of the wind direction and speed and the SST only. The mixed layer is considered to be either a dry convective layer or a shallow subcloud layer. The closure relates the fluxes of virtual potential temperature and humidity at the mixed layer top to the surface fluxes. The closure on temperature, while empirical, is a quite-well-established relationship based on a number of independent observations (e.g., Betts 1976; Nicholls and LeMone 1980). The closure for moisture flux is also empirical but has little observational evidence to support it. It is largely justified on the basis of its success.

The model balances the fluxes at the mixed layer top and surface with radiative cooling (assumed constant), horizontal advection, and diffusion. The latter is assumed to parameterize horizontal eddy fluxes. The addition of advection is a distinct advance on the air temperature and humidity models that have been used in the ocean modeling work of Seager et al. (1988) and Seager and Blumenthal (1994). Together with a means of obtaining the precipitation and runoff, it provides a much more reliable way of dealing with the ocean's boundary conditions than is currently adopted in many models used for simulating low-frequency ocean variability (e.g., Weaver and Sarachik 1991; Marotzke and Willebrand 1991).

The model presented here differs somewhat from the advective models for flux calculation developed by Luksch and von Storch (1992) and Kleeman and Power (1995). Both those models solved for the air temperature and then used some assumption on the relative humidity to derive the specific humidity. By solving for both the virtual potential temperature and the air specific humidity, our model determines the relative humidity internally, although its equilibrium value is much the same as that assumed by Kleeman and Power (1995). The Luksch and von Storch (1992) model was for air temperature anomalies only, whereas ours is for the total temperature (including the climatology) and humidity. Further, Kleeman and Power specified the air temperature above the mixed layer from data, but our model uses no external information other than winds and the SST.

We simulate the global fluxes for January and July. Over most of the ocean the fluxes are in reasonable agreement with the fluxes estimated on the basis of ship observations (e.g., Esbensen and Kushnir 1981). Also the model simulates the wintertime enhancement of the fluxes over the Kuroshio and Gulf Stream and another, smaller maximum off the West African coast. This enhancement does not occur where the mixed layer quantities are determined by a one-dimensional vertical balance. The enhancement is found to result from both diffusion and advection. Diffusion mixes lower temperature and drier air from north of the western boundary currents southward over warmer waters and increases the fluxes. It smooths the air temperature and humidity fields but has the opposite effect on the fluxes. Advection of cold, dry air off the continent further enhances the fluxes, especially by the coast. If diffusion is reduced then the enhancement due to advection is greater.

The next step in advancing the model's realism must address the problem of the underestimation of the sensible heat fluxes over the winter western boundary currents. For now we will exclude the possibility that radiative cooling is the problem because the only data we have (from a GFDL model run) are inconclusive on this point. We believe the fixed-depth mixed layer is more likely to be the problem and note that the mixed layer (as opposed to the entire PBL) is observed to shallow offshore (Grossman and Betts 1990). Including a variable-depth mixed layer will require a method for predicting its depth. A possibility is the simple Richardson number formulation introduced by Troehn and Mahrt (1986). This scheme has been shown to be of use in simulations of airmass transformation over the North Sea (Holtslag et al. 1990) and in a GCM (Holtslag and Boville 1993). However, it requires the temperature and moisture profile above the mixed layer. These can probably be determined by assuming a one-dimensional equilibrium profile and then solving by balancing a relaxation to this value with horizontal advection. Inclusion of a variable-depth mixed layer will require either an iterative or time-marching procedure. Even so, the simplicity of this model means that coupling to an ocean model would introduce insignificant extra computation.

Even as is the current model is capable of calculating reasonable heat fluxes solely on the basis of quantities that an ocean model either calculates (SST) or is forced by (winds), it's success relies on an accurate treatment of the feedbacks that relate the fluxes to the SST. When coupled to an ocean, it will allow the ocean model to compute its own fluxes without any "relaxation to climatology" terms or specification of near-surface atmospheric quantities, which are undesirable if the primary goal is to simulate SST.

So far we have coupled the model to an ocean GCM of the entire Pacific Ocean that has been developed from the Gent and Cane (1989) model but with an isopycnal vertical coordinate (Murtugudde et al. 1995). Early results are encouraging and the tropical simulation compares favorably with that of Seager and Blumenthal (1994). We also intend to use an Atlantic version of this model to examine the origins of the SST anomalies identified by, among others, Kushnir (1994). We will also couple it to the simple ocean model of Blumenthal and Cane (1989) to look at tropical Atlantic SST and whether advection off the coast of West Africa is important to the ocean heat budget there. We also intend to modify the model, as outlined above, by including a variable-depth mixed layer.

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APPENDIX

Subroutine HTFLUX

The mixed layer model described in this paper is available from the authors as a FORTRAN subroutine that can be called as

```
CALL HTFLUX(SST,U,V,WSPD,LSM,Q,T,SH,-
              RLH,SLAT,DXD,DYD,NX,NY)
```

Here NX is the number of grid points in the east–west direction and NY is the number in the north–south direction, SST is an array containing the SST (observed or model calculated), U and V are arrays containing the zonal and meridional winds, WSPD is an array containing the wind speed, and LSM is an array containing a land–sea mask with values of 1 denoting land and 0 denoting ocean. The terms Q and T are arrays containing the observed values of the air humidity and temperature. These are used to specify the boundary conditions around the continental margins. (Essentially, over all land points, the equilibrium humidity and virtual potential temperature are set to the observed values and the winds and diffusion are set to zero ensuring that the model values are equal to those observed.) On output, the subroutine returns the arrays SH and RLH, which are the sensible and latent heat fluxes in W m^{-2}. All the arrays are dimensioned (NX, NY). The term SLAT is the southern latitude of the domain. The term DXD is an array containing the east–west grid spacing, and DYD is a vector containing the north–south grid spacing. The model does allow certain irregular grids but is not completely general in this respect. The model parameters are set within the subroutine. The subroutine calls several other subroutines that set up the tridiagonal matrices and do the inversion. These are also available. The subroutine can be obtained by sending a request via e-mail to rich@seppie.ldeo.columbia.edu or by regular mail.

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