A One-Dimensional Mixed Layer Ocean Model for Use in Three-Dimensional Climate Simulations: Control Simulation Compared to Observations

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ABSTRACT

A study has been made of the dynamic interactions between the surface layer of the ocean and the atmosphere using a climate model that contains a new approach to predicting the sea surface temperature (SST). The atmospheric conditions are simulated numerically with the NCAR Community Climate Model (CCM3). The SST is determined by a modified Kraus–Turner-type one-dimensional mixed layer ocean model (MLOM) for the upper ocean that has been coupled to CCM3. The MLOM simulates vertical ocean dynamics and demonstrates the effects of the seasonal variation of mixed layer depth and convective instability on the SST. A purely thermodynamic slab ocean model (SOM) is currently available for use with CCM3 to predict the SST. A large-scale ocean general circulation model (OGCM) may also be coupled to CCM3; however, the OGCM is computationally intensive and is therefore not a good tool for conducting multiple sensitivity studies. The MLOM provides an alternative to the SOM that contains seasonally and spatially specified mixed layer depths. The SOM also contains a heat flux correction called Q-flux that crudely accounts for ocean heat transport by artificially specifying a heat flux that forces the SOM to replicate the observed SST. The results of the coupled MLOM–CCM3 reveal that the MLOM may be used on a global scale and can therefore replace the standard coupled SOM–CCM3 that contains no explicit ocean dynamics. Additionally, stand-alone experiments of the MLOM that are forced with realistic winds, heat, and moisture fluxes show that the MLOM closely approximates the observed seasonal cycle of SST.

1. Introduction

Many climate sensitivity studies of past and future climates are conducted with atmospheric general circulation models (AGCMs) that contain physically simple representations of the ocean. These ocean parameterizations are used in lieu of the more complex ocean general circulation models (OGCMs) that include the full three-dimensional ocean circulation, temperature, and salinity. The complexity of the OGCMs makes them less practical when multiple sensitivity analyses are required.

The two simplistic ocean representations that are most often used in these climate sensitivity studies are observed sea surface temperatures (SSTs) that are seasonally and spatially specified and SSTs determined from one-dimensional thermodynamics, where the SST calculation is based on the surface heat balance. This latter model, known as a slab ocean model (SOM), represents the surface ocean as a slab of specified depth. AGCMs run with fixed SSTs are overspecified. They reproduce the modern climatology, but may not be sufficiently flexible in simulating conditions that vastly differ from modern conditions. The SOM contains vertical thermodynamics but does not contain an explicit calculation of horizontal or vertical dynamics. The dynamics within the SOM are effectively specified by a heat flux correction. As an alternative, we propose...
the use of a one-dimensional mixed layer ocean model (MLOM), which allows the specification of deep ocean temperature and salinity and calculates the mixed layer depth and the entrainment of waters from the pycnocline into the mixed layer, thus affecting the mixed layer temperature.

The MLOM follows the basic physics of the mixed layer ocean model of Kraus and Turner (1967) and Niiler and Kraus (1977) that calculates the ocean mixed layer temperature $T_m$ (a proxy for the SST), mixed layer salinity $S_m$, and mixed layer depth $h_m$ using both vertical thermodynamics and dynamics. Battisti et al. (1995) describes the use of a mixed layer ocean model in climate change experiments, particularly the importance of mixed layer dynamics to high-latitude SST. We take this idea a step further by coupling the MLOM to a state-of-the-art AGCM, the Community Climate Model, version 3.2 (CCM3) from the National Center of Atmospheric Research (NCAR). The goal of this coupling is to provide a means of exploring the influence of the deep ocean in a climate sensitivity study and to investigate the importance of changes in the ocean’s thermal inertia and entrainment on the SST of a climate simulation.

The motivation for using the MLOM in climate simulations made with CCM3 is twofold. First, the MLOM contains vertical dynamics of the mixed layer in the form of mixed layer deepening and entrainment that are not included in the SOM. It also allows for a determination of the impact of the deep ocean on climate that is not possible with the SOM. Second, the MLOM, although less physically realistic than the OGCM, contains less computational complexity. This allows the climate simulation to reach a steady state more quickly, thus making it a useful tool for climate sensitivity analysis.

The MLOM prognostically determines the mixed layer depth and the rate of entrainment of subsurface waters into the mixed layer. Seasonal changes in mixed layer depth affect the thermal inertia of the mixed layer. This change in thermal inertia impacts the SST with surface warming/cooling having a greater impact on SST during spring/summer when the mixed layer is shallow, while having a lesser impact during fall/winter when the mixed layer is deep. It is also believed that the deep mixed layers in fall may delay the onset of sea ice growth (Martinson 1990). Entrainment of subsurface waters into the mixed layer during mixed layer deepening may also affect the mixed layer temperature through the vertical transport of heat. Subsurface waters have different thermal properties from surface waters. In the Tropics and subtropics, subsurface waters are generally colder than the water within the mixed layer. Hence, mixed layer deepening leads to the entrainment of cold subsurface water into the mixed layer, which causes the transport of cold water into the mixed layer. In the high latitudes where subsurface waters may be warmer than that in the mixed layer, entrainment causes the transport of warm water into the mixed layer. The subsurface entrainment at high latitudes is thought to be a likely cause of polynyas in the Southern Ocean sea ice (Martinson 1990).

In this paper, we describe the physics of the MLOM and its coupling to the NCAR CCM3. The goal is not to have the coupled MLOM–CCM3 reproduce an exact duplicate of any observational dataset. However, to assess the effectiveness of the MLOM within the climate simulation, we will compare the results of the coupled MLOM–CCM3 simulation to the coupled SOM–CCM3 simulation and to observations. Section 2 gives an overview of the model equations for the MLOM and the SOM and the steps required to couple the MLOM to CCM3. Section 3 describes the setup of the coupled experiments. Section 4 compares the results of the coupled MLOM–CCM3 simulation to the coupled SOM–CCM3 simulation, and to observations. Last, section 5 provides a discussion and conclusions to our results and the motivation for a study of the impact of deep ocean temperatures on climate (see Oglesby et al. 2005).

2. Models

The SOM and MLOM are approximations to the governing equations for the ocean that include conservation of momentum, heat, mass, salt, and an equation of state that describes the density of the ocean. The general equations that govern the ocean can be found in Gill (1982).

a. MLOM description

The MLOM belongs to a class of models called bulk turbulence models (Kraus and Turner 1967; Niiler and Kraus 1977; Niiler 1975; Kim 1976; Garwood 1977; Davis et al. 1981). These models are depth averaged and only consider momentum transfer at the surface and bottom of the mixed layer. They contain rather simplistic physics, and the MLOM, in particular, is not able to reproduce diurnal dynamics. Several more complex models exist that are able to predict the diurnal cycle and the small-scale features of the oceanic mixed layer (Mellor and Yamada 1982; Price et al. 1986; Gaspar 1988; Kantha and Clayson 1994). Kantha and Clayson (1994) and Martin (1985) provide a comparison of several mixed layer models, including a comparison of
bulk turbulence models to the more complex models listed above. We choose a bulk turbulence model because of its simplicity and ease in coupling to the NCAR CCM3.

The MLOM follows the basic physics of Kraus and Turner (1967) and Niiler and Kraus (1977) with some slight modifications to allow the model to account for buoyant instability. The model contains a mixed layer with constant temperature and salinity underlain by a pycnocline with linearly varying temperature and salinity and a deep ocean with constant temperature and salinity. Figure 1 shows the idealized vertical structure of the ocean water column used in the MLOM. The pycnocline temperature gradient $\Delta T/\Delta z$ and salinity gradient $\Delta S/\Delta z$ are given by the approximate annual-averaged linear variation of temperature and salinity with depth from the Levitus dataset (Levitus and Boyer 1994). These values are determined from zonal averages along different latitude bands. The deep ocean temperature $T_d$, salinity $S_d$, and depth $h_d$ are also specified from annual-averaged Levitus values. A finite temperature and salinity discontinuity may occur between the temperature in the mixed layer and the temperature at the top of the pycnocline. The pycnocline temperature and salinity are determined from the following linear relationships given the appropriate gradients and deep ocean values:

$$T(z) = \gamma_T(z - h_d) + T_d,$$

$$S(z) = \gamma_S(z - h_d) + S_d,$$

where $z$ is measured from the ocean surface to the ocean bottom (as is conventional in oceanography).

The MLOM contains simplifying assumptions to the governing equations that make it applicable to a horizontally homogeneous vertical water column that includes turbulent mixing processes. These assumptions take into account the interaction of the ocean mixed layer with the atmosphere above and the pycnocline below. Since the layer is turbulent, perturbations to the flow are considered to determine their effect on the mean flow. The underlying assumption of the model is that the mixed layer is homogenous through its depth. This allows the model equations to be integrated through the depth of the mixed layer. The governing equations are therefore changed from a set of partial differential equations to a more simple set of prognostic, ordinary differential equations for temperature ($T$), salinity ($S$), and depth ($h$).

The governing equations are simplified under the following assumptions: 1) The mixed layer is horizontally homogenous. Therefore, variations in the east/west and north/south directions are small compared to vertical variations. Also, the horizontal momentum is small compared to vertical momentum. Hence,

$$\nabla \cdot \mathbf{u} = 0,$$

where $\mathbf{u} = (u, v, w)$ is the velocity vector. 3) The mixed layer is hydrostatic and Boussinesq so that the vertical momentum equation becomes

$$\frac{1}{\rho_o} \frac{\partial P}{\partial z} = -g \frac{\rho}{\rho_o}.$$

4) The mixed layer is turbulent. The turbulent fluxes of momentum, temperature, and salinity dominate the molecular diffusion processes. A process of Reynolds averaging is therefore applied, and the momentum equations are combined into an equation for the turbulent kinetic energy (TKE). The resultant equations for mixed layer temperature $T_m$, salinity $S_m$, density $\rho$, and entrainment velocity $w_e$ from the TKE balance, are [note: A complete derivation of the depth-integrated and Reynolds averaged mixed layer equations may be found in Stephens (1998)] as follows:

$$\rho_o c_p h_m \frac{dT_m}{dt} = w_e \Delta T + F_A + F_{SW}(1 - e^{-\phi_m}),$$

FIG. 1. Idealized vertical structure of temperature. The temperature in the mixed layer ($T_m$) and the temperature of the deep ocean are constant, while the temperature within the pycnocline varies linearly with depth. A finite temperature jump may exist at the base of the mixed layer. The mixed layer temperature and depth evolve according to mixed layer physics. The mixed layer salinity and density have a similar vertical profile.
\begin{align}
\rho &= \rho_o [1 - \alpha \Delta T + \beta \Delta S], \\
2m e^{-\rho_o \gamma u^*} + \frac{h_m}{2} [\frac{1 - n}{\rho_p} B_o + (1 + n) B_o] \\
\frac{dr}{dt} &= \Delta h m + \Delta u^2,
\end{align}

where \(\frac{dr}{dt}\), not to be confused with the total derivative \(D\Delta t\), means the local rate of change (note: partial derivatives are not used in these equations since only temporal derivatives, and not spatial derivatives, are taken); \(\rho_o\) is the mean density of the ocean; \(c_p\) is the specific heat capacity; \(\alpha\) and \(\beta\) are the thermal expansion and haline contraction coefficients, respectively; \(\Delta T\) and \(\Delta S\) are the difference in temperature and salinity between the mixed layer and the top of the pycnocline (see Fig. 1), respectively; \(F_{SW}, F_A\), and \(F_S\) are the surface solar heat flux, the surface nonsolar heat flux, and the surface salt flux into the mixed layer, respectively; \(u^*\) is the surface friction velocity caused by winds; \(\Delta u\) is the shear between the mixed layer and pycnocline specified as the greater of 3 cm s\(^{-1}\) or the friction velocity; and \(\gamma, r,\) and \(n\) are the dissipation parameters for solar radiation, wind, and buoyancy, respectively. The net surface buoyancy flux \(B_o\) and \(\Delta b\) are defined below:

\begin{align}
B_o &= \frac{g}{\rho_o} \left[ \frac{\alpha}{c_p} F_A - \beta F_S \right] \\
&+ \frac{ga}{\rho_o c_p} \left[ 1 - e^{-\gamma h_m} - \frac{2}{\gamma h_m} (1 - e^{-\gamma h_m}) \right], \\
\Delta b &= \frac{\rho_o}{c_p} [\alpha \Delta T + \beta \Delta S].
\end{align}

The buoyancy term in Eq. (6) is written so that stabilizing buoyancy, \(B_o > 0\), is not dissipated.

The MLOM undergoes three distinct regimes of flow. The mixed layer is either buoyantly stable and deepening (entrainment regime), stable and shallowing (shallowing regime), or buoyantly unstable and deepening (convective regime).

In the entrainment regime, the mixed layer is buoyantly stable. However, wind-induced mixing and mixing caused by surface cooling or evaporation cause the mixed layer to entrain a portion of the pycnocline. The equations that govern the mixed layer temperature, salinity, and the extent of mixed layer deepening are Eqs. (3), (4), and (6) with \(w_e = \frac{dh_m}{dt}\) in Eq. (7).

In the shallowing regime, the stabilizing surface buoyancy flux from surface heating or precipitation is greater than the destabilizing effect of surface wind forcing. The mixed layer, in this case, does not “de-train” but rather achieves a depth, known as the Monin–Obukov length scale, where the stabilizing influence of buoyancy and the destabilizing influence of winds are in balance (Turner 1973). Equations (3), (4), and (6) that govern the mixed layer temperature, salinity, and depth are used with \(w_e = 0\) in Eq. (7). The entrainment equation [Eq. (6)] is then solved implicitly for the mixed layer depth \(h_m\).

If the water column is buoyantly unstable, the MLOM enters the convective regime and the mixed layer depth, temperature, and salinity are adjusted to conserve mass and render the layer stable in a manner similar to Kim (1976). In this case, the entrainment equation [Eq. (6)] does not apply because this equation assumes that the mixed layer is buoyantly stable. Convective adjustment is therefore used to stabilize the mixed layer through deepening. The equation that governs the convective adjustment is

\[ (\rho^* - \rho)h_m = \int_{h_m}^{h} [\rho - \rho(z)] dz, \]

where \(\rho\) and \(h_m\) are the mixed layer density and depth before the adjustment, \(\rho^*\) and \(h^*\) are the density and depth after adjustment, and \(\rho(z)\) is the vertical distribution of density in the pycnocline and is constructed from the linear pycnocline temperature and salinity using the density equation [Eq. (5)]. The discontinuity between the mixed layer and pycnocline density is eliminated in the adjustment and hence the tendency for mixed layer deepening is enhanced. Buoyant instability occurs when there is intense cooling or evaporation, which causes the mixed layer density to exceed the density of the pycnocline, a feature that is not accounted for by the “normal” mixed layer physics.

The MLOM, unlike the SOM, is able to capture important mixed layer dynamics. The model explicitly determines the mixed layer depth, based on physics, and is able to account for the impact of the underlying ocean on the mixed layer temperature. However, there are some phenomena that the MLOM is unable to capture. Processes such as deep convection and large-scale gyre circulation are not included. The implication of assumptions 1 and 2 is that the upwelling velocity is zero. Hence, the model does not account for upwelling.
that may create east/west temperature differences in the major basins. Additionally, the structure of the underlying ocean thermocline is crudely approximated by the specification of linear temperature and salinity gradients. The MLOM does not calculate horizontal currents or have a horizontal heat flux correction to account for such currents. The addition of a horizontal heat flux correction will be a goal of future improvements to the model. At present, the specification of such fluxes complicates the model’s ability to maintain convective stability in light of the specified temperature and salinity gradients that are unable to adjust. The fixed pycnocline temperature and salinity gradients cause the mixed layer to have no memory of temperature and salinity changes that may have resulted from the previous years during deepening. Hence, there is a heat and salt loss/gain from year to year because of the absence of a seasonal pycnocline.

b. SOM description

The SOM contains seasonally and spatially specified mixed layer depths. It neglects the explicit calculation of currents and therefore only contains a single equation for the mixed layer temperature given by

$$\rho_o c_p h_o \frac{dT}{dt} = F_T,$$

where $h_o$ is the mixed layer depth, and $F_T$ is the total incident heat forcing at the surface, solar and nonsolar. The SOM contains seasonally and spatially specified mixed layer depths. In addition, a heat transport term, $Q$-flux, may be added to the right-hand side of the equation to crudely account for vertical and horizontal heat transport (Kiehl et al. 1998). With the addition of a $Q$-flux correction, the SOM is able to generate many of the horizontal temperature differences that are observed in the ocean. However, these variations are only implicit in the $Q$-flux parameterization and are not explicitly determined by the model, which contains no dynamics. In fact, the SOM without $Q$-flux performs poorly compared to observations.

c. CCM3 description

The global climate model used for this study is the NCAR Climate System Model (CSM; Boville and Gent 1998). The CSM includes atmospheric, oceanic, land surface, and sea ice components. The atmospheric component (CCM3) has a horizontal resolution of T42, or an equivalent gridpoint resolution of 2.8° latitude by 2.8° longitude, and the vertical is resolved by 18 layers. The standard land surface option for CSM (and hence CCM3) used in our study is the Land Surface Model (LSM; Bonan 1998). Kiehl et al. (1998) and Hack et al. (1998) summarize important characteristics of the model-generated climate.

The CCM3 communicates with the mixed layer/sea ice model through exchanges between the two systems in the planetary boundary layer (PBL). Momentum from wind stress, thermal energy from solar radiation, terrestrial radiation, and sensible and latent heats, and salt fluxes from precipitation, evaporation, and surface runoff are taken from the AGCM and used to force the mixed layer/sea ice model. From this interaction, SST and snow and ice depths are returned to the AGCM for future calculations. The climate system is allowed to evolve in this manner until equilibrium is achieved. The values of $T_a$, $S_a$, and $h_a$ for the present day were adapted from Levitus and Boyer (1994) and are prescribed as zonal averages (Fig. 2). As shown in Fig. 2, the $T_a$ values are highest along and just north of the equator (though still no greater than 7°C) and lowest near the poles (just 2°–3°C above the freezing point of seawater). Interestingly, values in the midlatitudes are more than 2°C warmer in the Southern Hemisphere than in the Northern Hemisphere, which suggests that the Southern Hemisphere is less stably stratified than the Northern Hemisphere and is therefore more prone to convection. The pycnocline temperature and salinity gradients are specified along the same zonal bands as the deep ocean temperature and salinity.

3. Experiment plan

In this section, we describe the experiment setup for each of the models. A stand-alone simulation of the MLOM was run with winds, heat, and moisture fluxes taken from a modern control simulation of the SOM–CCM3 with the horizontal heat flux correction, $Q$-flux. An additional control simulation of the SOM–CCM3 was run with modern conditions, but without the $Q$-flux correction. Both SOM–CCM3 runs were conducted with seasonally and spatially specified mixed layer depths. The final simulation, with MLOM coupled to CCM3 (MLOM–CCM3), was run without a $Q$-flux correction. The concept of a heat flux correction in the MLOM–CCM3 would be different from that of the SOM–CCM3. In the MLOM–CCM3, the inclusion of vertical dynamics allows for an explicit calculation of vertical heat transport. Therefore, any heat flux correction that would be added to the MLOM–CCM3 should only include horizontal heat transports. However, we do not include a horizontal heat flux correction in our simulation of MLOM–CCM3 because the addition of such a correction could affect the stability of the mixed layer.
Fig. 2. The zonal deep ocean (a) temperature, (b) salinity, and (c) depth are specified across latitude bands and are based on the observations of Levitus and Boyer (1994).
**a. The MLOM stand-alone simulation**

The MLOM stand-alone simulation was forced with the control simulation of SOM–CCM3 with $Q$-flux (horizontal heat transport) turned on. The results of the SOM–CCM3 have SSTs that differ little from observations. A full exploration of these results will be shown in a later section. The forcings used are the surface wind stress, the net surface heat flux minus the surface upward longwave heat flux, and the net surface moisture flux (evaporation minus precipitation). A sensitivity study of the MLOM to surface moisture fluxes reveals that the model is only moderately sensitive to changes in evaporation and precipitation. The moisture flux changes the SST values by, on average, less than 1°C. Hence, we have run the stand-alone simulations without the addition of moisture. We conduct a separate sensitivity analysis that shows the lack of sensitivity of the model to moisture fluxes.

Three sensitivity studies were conducted with the MLOM stand-alone simulation. The sensitivity of the MLOM to dissipation processes was conducted by varying the dissipation parameters for wind stress ($r$) and buoyancy ($\beta$). The sensitivity of the MLOM moisture fluxes was conducted by specifying maximum and minimum values for $E - P$ from Peixoto and Oort.

An additional stand-alone simulation of the MLOM was also run to test the ability of the MLOM to reproduce the proper seasonal cycle. We chose several latitudes along the 150°W longitude line (in the Pacific Ocean) and compared the results of the seasonal cycle for these points to the MLOM–CCM3, the SOM–CCM3, and observations. These results are shown after the results of the coupled simulation described below.

**b. The coupled SOM–CCM3 and MLOM–CCM3 simulations**

The MLOM–CCM3 and SOM–CCM3 each were run at T42 resolution (2.8° latitude by 2.8° longitude) and a 20-min time step. The atmospheric component of the model was run with 18 vertical atmospheric layers.

1) **THE COUPLED SOM–CCM3 SIMULATION**

Two simulations of the SOM–CCM3 were run, one with a $Q$-flux correction and a second with no $Q$-flux correction. The simulation without $Q$-flux will be referred to as SOM/NO-$Q$FLUX and the simulation with $Q$-flux will be referred to as SOM/$Q$FLUX. The SOM/NO-$Q$FLUX was run for 15 model years. The SOM/$Q$FLUX was run for 55 model years. Both models contain seasonally and spatially specified mixed layer depths. The SOM/$Q$FLUX has $Q$-flux values that are also seasonally and spatially specified.

The SOM is directly coupled to the atmospheric component of CCM3 through surface fluxes of momentum, heat, and moisture. These fluxes are determined by the AGCM using bulk parameterizations in which the exchange rates are determined by differences between quantities (momentum, temperature, and moisture) in the mixed layer and in the first atmospheric layer of the AGCM. The surface fluxes of heat are used by the SOM to calculate the SST. In grid points that contain sea ice, the slab ocean temperature is kept at the freezing temperature (−1.87°C), and the sea ice model, which depends solely on thermodynamics, determines the thickness of the sea ice cover and the surface temperature of the ice (Semtner 1984).

2) **THE COUPLED MLOM–CCM3 SIMULATION**

The MLOM–CCM3 was run for 19 model years. No $Q$-flux correction was used in the model because the addition or extraction of heat from a grid point could compromise the buoyant stability of the grid point. This could lead to an additional vertical heat transport as the model adjusts to a convectively stable state. Although the model was run with a time step of 20 min, the MLOM was implemented only once per day. As in the SOM–CCM3, surface fluxes of heat, momentum, and moisture were calculated by the AGCM. In the MLOM–CCM3, the fluxes were accumulated at every time step. After 72 time steps (equivalent to one day), the accumulated heat and momentum fluxes were averaged and used to determine the mixed layer temperature and depth. The surface precipitation minus evaporation and surface runoff are also accumulated, and the total accumulated moisture flux is used to determine the mixed layer salinity and density. The MLOM–CCM3 is run with a 1-day time step because it does not adequately simulate the diurnal cycle. It is not unreasonable to couple the MLOM to the atmospheric model once per day since the diurnal variation of SST is of little consequence for the length of simulation we are performing.

In locations where there is sea ice, the MLOM is turned off. The mixed layer is treated as a slab with fixed depth while sea ice is present, and the mixed layer temperature is held at the freezing temperature of seawater. Once the mixed layer becomes ice free, the MLOM is again turned on. Moisture fluxes from snowmelt are accumulated and used to adjust the mixed layer density once the mixed layer becomes ice free. The model does not contain sea ice advection, therefore, for grid points with seasonal ice cover, the net ice growth is exactly equal to the net ice melt. Hence, ice growth/melt does not factor into the mixing processes within the mixed layer. We have not taken into account...
the impact of ice growth/melt on the mixing processes in regions of permanent sea ice coverage.

4. Results

a. Sensitivity of the MLOM to dissipation parameters and moisture fluxes (stand alone)

The choice of dissipation parameters is based on the values chosen by Niiler and Kraus (1977). Dissipation of turbulence generated by wind shear decays exponentially with depth with an e-folding scale of 20 m, \( r = 0.05 \). The linear dissipation parameter for destabilizing buoyancy is \( n = 0.2 \).

Sensitivity tests with the MLOM were run by varying the dissipation parameters over a range of values. This range represents values of \( r \) and \( n \) that are physically relevant. Each sensitivity experiment was run for 10 yr with model equilibration having been reached by the third year. We show the results from the fifth year of each simulation. The model initial and boundary conditions, which include the initial mixed layer temperature, salinity and depth, and the deep ocean temperature and salinity, are taken from 50°N latitude and 145°W longitude, commonly referred to as Ocean Station Papa. We chose this data location because the net horizontal heat transport is minimal, eliminating the need for a \( Q \)-flux correction. Data from Station Papa have been used in several studies of mixed layer modeling for this same reason (Martin 1985; Price et al. 1986; Large et al. 1994; Kantha and Clayson 1994). The MLOM was forced with winds and heat forcing from the SOM–CCM3 with \( Q \)-flux.

We conduct a separate study that shows the impact of freshwater forcing from precipitation and evaporation on the mixed layer depth, entrainment, and SST. These fluxes have a minimal effect at our data location, which we have forced with the maximum and minimum yearly averaged precipitation from Peixoto and Oort (1992). One might expect a greater sensitivity in regions such as the ITCZ. However, this region is characterized by greater vertical stability and is therefore less likely to convect with similar moisture fluxes.

1) Wind sensitivity analysis

Figures 3a,b show the model results for temperature and mixed layer depth for the case in which the dissipation parameter for winds was varied. The parameter \( r = 0.05 \) (default) was varied from \( r = 0.0 \) (peak wind case), representing an infinite e-folding depth, to \( r = 100 \) (minimal wind case), representing an e-folding depth of 1 cm. For smaller values of the dissipation parameter, the effect of the winds is distributed over a deeper depth. This causes the winds to have a greater impact on mixed layer deepening. For larger values of the dissipation parameter, the effect of the winds is distributed over a smaller depth and the winds have a lesser impact on deepening.

The temperature of the mixed layer decreases as the dissipation parameter decreases. For values of the e-folding depth that are smaller than 20 m \(( r = 0.05 )\), the difference in mixed layer temperature and depth from the default values is minimal. For larger values of \( r \) (smaller e-folding depths), the seasonality of the mixed layer temperature is slightly affected with colder mixed layer temperatures in winter and warmer mixed layer temperatures in summer. For values of the e-folding depth that are larger than 20 m, the difference in mixed layer depth and temperature from the default values is more significant.

The temperature difference between the “peak-wind” and the “minimal-wind” cases is most pronounced during summer with a maximum temperature difference of 7°C in July. This corresponds to a difference in mixed layer depth of approximately 100 m. The shallower mixed layer of the minimal wind case has less thermal inertia than the deeper mixed layer of the peak wind case. Since entrainment is essentially shut off during summer, the mixed layer temperature is primarily controlled by surface fluxes of heat with the shallow mixed layer achieving a substantially higher temperature than the deeper mixed layer. Therefore, it is the change of thermal inertia that is responsible for the marked change in temperature among the different values of the dissipation parameter.

The temperature differences between the minimal-wind and the peak-wind cases are much less pronounced in winter. Although the difference in mixed layer depth, approximately 250 m in January, is larger than the summer difference, the mixed layer temperatures differ by only 3°C compared to the 8°C temperature difference in summer. In winter, the mixed layer temperatures are influenced by entrainment. Therefore, entrainment of colder water from the pycnocline has some control on the mixed layer temperature. Since the temperature of the pycnocline is determined by the specified mixed layer temperature gradients and the deep ocean temperature [Eq. (1)], the extent that the mixed layer temperature varies from the prescribed deep ocean temperature is more limited in the winter season. Ultimately a predictive scheme for pycnocline temperature and salinity would be required to remove this constraint.

2) Buoyancy sensitivity analysis

The mixed layer temperature and depth for variations in the buoyancy dissipation parameter are shown
The result of increasing the buoyancy dissipation parameter is to further destabilize the mixed layer, which causes the mixed layer temperature to decrease and the mixed layer depth to increase because of the increase in entrainment.

The buoyancy dissipation parameter was varied from a value of 0 (no destabilizing buoyancy) to a value of 1 (maximal destabilizing buoyancy). When the dissipation parameter equals 1, there is no dissipation of destabilizing buoyancy. As a general rule, the temperature decreases as the dissipation parameter increases for the reasons mentioned above. However, there is a period of approximately 3 months, during winter when the temperatures for $n = 0.2$ (default) and
(n = 0.5) are warmer than the “no destabilizing buoyancy” case.

The largest temperature differences occur from summer to fall, between May and October, with differences as large as 3°C between the no destabilizing buoyancy case and the “maximal destabilizing buoyancy” case. This temperature difference is approximately half that of the wind sensitivity analysis. The difference in mixed layer depth is also smaller, 20 m in the buoyancy sensitivity analysis compared to 100 m in the wind sensitivity analysis. We therefore conclude that the model is significantly more sensitive to changes in the e-folding scale of wind mixing than to changes in the dissipation of buoyancy.

Similar to the wind sensitivity analysis, the winter mixed layer temperature is only slightly affected by variations in the dissipation parameter, while much larger variations occur in mixed layer depth. Lastly, while variations in r affect the mixed layer depth throughout the annual cycle, variations in n only significantly affect the mixed layer depth during fall and winter. During spring and summer, the buoyancy is primarily stabilizing and stabilizing buoyancy is not dissipated [Eq. (6)]. Therefore, the mixed layer depth during the shallowing phase of the mixed layer should not change significantly from the default values.

3) Moisture Sensitivity Analysis

In the moisture sensitivity analysis, we force our data point (Ocean Station Papa) with the zonally averaged annual evaporation minus precipitation, E – P, from Peixoto and Oort of the latitude bands with maximum and minimum E – P values. We use the maximum zonally averaged E – P (evaporation greater than precipitation) of –635 mm yr⁻¹ from 0° to 10°N and the minimum E – P (precipitation greater than evaporation) of 528 mm yr⁻¹ from 20° to 30°S. As expected, the temperature decreases and the mixed layer depth increases when evaporation is greater than precipitation because the increase in surface salinity increases the instability of the mixed layer. The opposite effect occurs when precipitation exceeds evaporation with decreased surface salinity and increased stability. Although we have used tropical E – P results to force our model at Ocean Station Papa, these moisture fluxes would have an even more limited impact on the Tropics since the stability of the tropical water column in our study is greater than that of the subtropics. When the net evaporation exceeds the net precipitation, the salinity of the mixed layer increases. If the salinity becomes so large that the mixed layer is convectively unstable, convective adjustment will cause the layer to adjust in such a way that density is conserved. The mixed layer temperature and salinity adjust to the specified values at the new convective depth.

The moisture flux has a minimal effect on the mixed layer depth and temperature. The mixed layer temperature changes, on average, less than 1°C. The least change in mixed layer depth and temperature occurs during mixed layer shallowing. During shallowing, the mixed layer depth is not governed by entrainment, but rather by Monin–Obukov scaling. The rate of shallowing, after the seventh month, is the same for all three cases since the mixed layer depth is specified by the balance between wind and buoyancy forcing. The buoyancy imparted by moisture is a negligible contribution to the total buoyancy forcing since it is an order of magnitude smaller than the buoyancy contributed by heating.

b. Comparison of the MLOM–CCM3 and the SOM–CCM3 to observations

1) Surface Temperature

In general, the MLOM–CCM3 and the SOM/QFLUX, simulate the basic structure of the observed SST well, while the SOM/NO-QFLUX predicts temperatures that are, overall, significantly warmer than observations. For the annual-averaged conditions we fully describe the results of the MLOM–CCM3 and the SOM/NO-QFLUX. The results of the SOM/QFLUX simulation are statistically similar to observations. We will explore these results more fully in the description of the seasonal surface temperature.

The annual-averaged global mean temperature (UTC), over the oceans, for the MLOM–CCM3 is within 1°C of the observed UTC of 22.1°C. The SOM/NO-QFLUX is 4°C warmer than the observed UTC. The warmer temperatures of the SOM/NO-QFLUX are due to the absence of Q-flux, in the model, that artificially corrects the surface temperature to approximate the observed temperature. Since the SOM/QFLUX predicts temperatures that are similar to observations, the SOM/QFLUX reveals the reliance of the SOM on the Q-flux correction in the coupled SOM–CCM3 simulation.

Figure 4 shows the annual-averaged temperature and temperature differences from observations for the MLOM–CCM3. The SST in the MLOM–CCM3 simulation shows a distinct lack of meridional structure (Fig. 4a). This is not surprising because in the absence of horizontal heat transports, the affect of heat transport to higher latitudes by intense western boundary currents is lost. The zonal nature of the SST is also caused
FIG. 4. (a) Annual-averaged surface temperature for MLOM–CCM3 (°C) and (b) MLOM–CCM3 minus observations' SST differences (°C). Dashed lines show negative values and solid lines show positive values.
by the specification of a deep ocean temperature \((T_d)\) that is zonally invariant. The Gulf Stream in the North Atlantic Ocean and the Kuroshio in the North Pacific Ocean are not present in the simulation. The MLOM–CCM3 SST is, on average, \(4^\circ\)C colder than observations in the North Atlantic and \(2^\circ\)C colder than observations in the North Pacific (Fig. 4b). Similarly, the absence of western boundary currents in the Southern Hemisphere such as the Brazil Current in the South Atlantic Ocean and the Agulhas Current in the South Pacific Ocean, causes the MLOM–CCM3 temperatures to be colder than observations. In addition to the lack of poleward heat transport by western boundary currents, the MLOM–CCM3 contains large regions of SST that are warmer than observations on the eastern boundaries of the continents caused by the absence of the broad eastern boundary currents that transport cold water equatorward. In the Antarctic Circumpolar Current (ACC), where the winds are particularly strong, the SST is also colder than observations. The most obvious cause of this reduced SST is entrainment, which brings the cold subsurface waters to the surface in the presence of strong vertical mixing by winds. The mixed layer depths are also deep in this region, which is evidence of the role of entrainment in determining the MLOM SST.

The annual-averaged SST for the SOM/NO-QFLUX, on the other hand, appears to show some evidence of the northward transport of heat by western boundary currents (Fig. 5a). For example, the isotherms in the North Atlantic branch toward the northeast off the coast of North America, which appears similar to the structure of the isotherms seen when the Gulf Stream circulation is present. Given that there is no \(Q\)-flux, the meridional structure must be connected to the specification of the mixed layer depths, perhaps in conjunction with factors such as cold air blowing from land over the nearby ocean, which cools the ocean adjacent to the land relative to the open ocean. The differences in temperature between the SOM/NO-QFLUX and observations show extensive regions where the SST is warmer than observations (Fig. 5b). The only negative differences (SOM/NO-QFLUX colder than observations) occur in the Southern Hemisphere, just north of Antarctica, where SOM/NO-QFLUX predicts too much sea ice coverage. Again, it is only with the use of \(Q\)-flux that the SOM can be made to predict the observed SST.

Figures 6 and 7 show the December–January–February (DJF) and the June–July–August (JJA) averaged surface temperatures and SST differences from observations for the MLOM–CCM3, the SOM/NO-QFLUX, and the SOM/QFLUX. The MLOM–CCM3 captures, reasonably well, the seasonal difference in SST between DJF and JJA (Figs. 6a,b and 7a,b). There is an obvious seasonal cycle of SST with an equatorward migration of the isotherms from summer (JJA for the Northern Hemisphere and DJF for the Southern Hemisphere) to winter (DJF for the Northern Hemisphere and JJA for the Southern Hemisphere). In summer, the isotherms contain more meridional structure than the isotherms in winter. During winter, the SST is determined by the atmospheric heat fluxes and by the entrainment of subsurface water into the surface ocean during mixed layer deepening. From the sensitivity analysis, it is clear that the mixed layer temperature, in winter, is inexorably linked to the prescribed deep ocean temperatures because of entrainment. The zonal prescription of these deep ocean temperatures therefore causes the winter SST to also be zonally situated. During summer, the mixed layer is shallow and entrainment is at a minimum. Therefore, the control of the SST occurs mainly through surface fluxes of heat, which as a result of cloud coverage and heat advection are not strictly zonal. This leads to a “less” zonal response in the summer SST.

The SOM/NO-QFLUX predicts temperatures that are warmer than observations for both DJF and JJA (Figs. 6c,d and 7c,d). Nonetheless, the seasonal variation of SST is similar to the observed seasonal variation.

The SSTs for the SOM/QFLUX for both DJF and JJA are very similar to observations (Figs. 6e,f and 7e,f). In fact, the temperatures for the SOM/QFLUX are the observed temperatures except in a few isolated regions. The most notable differences occur in the North Pacific and North Atlantic Oceans. In JJA, the North Pacific is on average \(2^\circ\)C warmer than the observed temperature. The North Atlantic is approximately \(1.5^\circ\)C warmer. In DJF, the temperature differences are smaller, with an SST difference of \(1^\circ\)C in the North Pacific and \(0.5^\circ\)C in the North Atlantic. In the Tropics of the Northern Hemisphere and in the Southern Hemisphere Tropics and subtropics, there are no discernable differences between the SSTs of the SOM/QFLUX and those of observations. There are slight negative differences (SOM/QFLUX colder than observations) for JJA and DJF in the Southern Hemisphere midlatitudes.

The marked similarity of the SOM/QFLUX to observations, and the apparent inability of the SOM–CCM3 without \(Q\)-flux to adequately simulate the observed SST, is evidence that the coupled SOM–CCM3 must rely on the prescribed \(Q\)-flux to generate a realistic climatology. Without such a \(Q\)-flux correction, the
Fig. 5. Same as in Fig. 4, but for SOM–CCM3.
The land surface temperatures of the MLOM–CCM3 appear to adequately reproduce the observed surface temperature except in regions where horizontal currents and therefore horizontal heat transports are important.

The MLOM–CCM3 surface temperature is significantly warmer than the observed surface temperature, while the SOM–CCM3 surface temperature appears to adequately reproduce the observed surface temperature.
are compared to the observations of Legates and Willmott (1990; observed temperatures not shown). Despite the zonal nature of the SSTs in the MLOM–CCM3, the land surface temperatures only differ from observations in a few regions. In JJA the MLOM–CCM3 is 3°–5°C colder than observations in western North America and at the southern tip of South America (Fig. 7c). More significant temperature differences occur in DJF with colder temperatures in the MLOM–CCM3 in North Africa, South America, and in the western and

Fig. 7. Same as in Fig. 6, but for JJA.
southern United States and warmer temperatures in South Africa (Fig. 6).

2) SEA LEVEL PRESSURE

The distribution of sea level pressure (SLP) is fundamentally linked to key features in the atmospheric circulation and therefore can be used to summarize the overall model performance. Any model that purports to improve the representation of the ocean should, at minimum, perform at least as well as existing models in simulating the atmosphere. We therefore compare the SLP of the MLOM–CCM3 (Figs. 8a,b), to the SOM/NO-QFLUX (Figs. 8c,d), the SOM/QFLUX (Figs. 8e,d), and observations (results not shown) for DJF and JJA. The SLP of the SOM/QFLUX is a close approximation to the observed SLP, with only slight differences from observations.

The annual-averaged distribution of SLP in the MLOM–CCM3, the SOM/NO-QFLUX, and the SOM/QFLUX is similar to observations (results not shown). There are high-pressure cells in the subtropics, in both the Northern and Southern Hemispheres. The Aleutian and Icelandic lows can be found in the Northern Hemisphere, poleward of the subtropics. There is fairly continuous low pressure in the Tropics indicating the position of the ITCZ. In addition, there is a zonal belt of SLP south of 40°S, in the Southern Hemisphere, that decreases toward Antarctica.

In DJF the primary differences between the model simulations and the observed SLP are the intensity of the Aleutian and Icelandic lows and the subtropical highs. In DJF, the Aleutian low in the MLOM–CCM3 and the SOM/NO-QFLUX is more intense than in the observations. The MLOM–CCM3 pressure is approximately 6 mb less than the observed pressure, and the SOM/NO-QFLUX has an even greater difference of approximately 16 mb. The Aleutian low in the SOM/QFLUX is slightly less intense with pressure that is approximately 2 mb greater than the observed pressure. While the position of the Aleutian low-pressure center is close to the observed location in the MLOM–CCM3 and the SOM/QFLUX, it is east of its observed location in the SOM/NO-QFLUX. The Icelandic low and subtropical highs in the MLOM–CCM3 and the SOM/QFLUX, during DJF, are both more intense than observations. The SOM/NO-QFLUX, surprisingly, predicts both the position and intensity of the Icelandic low correctly. However, the subtropical highs in the Northern Hemisphere of the SOM/NO-QFLUX are nearly absent. In their place is a broad band of high pressure that encircles the globe.

The seasonal migration of the subtropical highs from their DJF locations to their JJA locations is seen in all three simulations (Figs. 8b,d,f). In JJA the subtropical highs dominate the pressure in the Northern Hemisphere. The intensity of these subtropical highs is greater than observations in all three experiments. The subtropical highs in the Southern Hemisphere also have greater intensity than observations.

Overall, we can conclude that while still containing problems, the simulation of the SLP in MLOM–CCM3 is at least as good as in coupled SOM–CCM3 and should be adequate for most purposes.

3) SEA ICE

The extent and thickness of sea ice for DJF and JJA is shown in Fig. 9 for the MLOM–CCM3 (Figs. 9a,b), the SOM/NO-QFLUX (Figs. 9c,d), and the SOM/QFLUX (Figs. 9e,f). Since data concerning the thickness of the observed sea ice coverage are limited, we tend to concentrate on the extent of the sea ice coverage with only minimal discussion of the thickness.

In DJF, the extent of the Northern Hemisphere ice coverage in the MLOM–CCM3 is less than observations, while the SOM/NO-QFLUX and the SOM/QFLUX ice extent replicates the observed extent. The lack of sea ice at these “low” high latitudes, in the MLOM–CCM3, is connected to the zonally averaged values of T_s, which cause the subsurface waters to be too warm for sea ice to form. Hence, the MLOM–CCM3 does not contain ice to the south of the Bering Strait, along the eastern Eurasian coast, and within Lake Huron.

During JJA, the observed Northern Hemisphere sea ice coverage is restricted to the Arctic Ocean. The sea ice extent in both the MLOM–CCM3 and the SOM/NO-QFLUX approximates the observed extent, although the thickness of the sea ice coverage for the SOM/NO-QFLUX is one-fifth the thickness of the MLOM–CCM3 because of the warmer surface temperature of the SOM/NO-QFLUX. The SOM/QFLUX overestimates the sea ice extent. In JJA, the ice extends down the western coast of Greenland and into Lake Huron. It is relatively easy to correct the sea ice extent in the SOM/QFLUX by specifying a value of Q-flux under the sea ice that will melt the ice cover.

The ice is too extensive in the Southern Hemisphere, during DJF (Southern Hemisphere summer), in the MLOM–CCM3 and in the SOM/QFLUX. The ice extent in the SOM/NO-QFLUX is the most similar to the observed extent. Actually, the winter sea ice coverage (JJA), in the SOM/NO-QFLUX, is only slightly different from the summer coverage (DJF). This suggests that the model does not predict the correct summer ice extent but rather contains less sea ice because of its high surface temperature. In the MLOM–CCM3, the
Fig. 8. DJF- and JJA-averaged sea level pressure in mb: (a) MLOM–CCM3 for DJF; (b) MLOM–CCM3 for JJA; (c) SOM–CCM3 for DJF; (d) SOM–CCM3 for JJA; (e) SOM–CCM3 with Q-flux correction DJF; and (f) SOM–CCM3 with Q-flux for JJA.
winter sea ice extent (JJA) is less than the summer sea ice extent (DJF). Although this appears to be rather unphysical, it occurs because of entrainment of the relatively warm subsurface waters during mixed layer deepening in the MLOM. The Southern Hemisphere, although stably stratified, is near a neutrally stable state. Hence, a small increase in destabilizing buoyancy or in wind forcing, will cause the MLOM to convect. Since

Fig. 9. Same as in Fig. 8, but for sea ice thickness (in m).
the high latitudes are characterized by warm subsurface water, the entrainment of this warm water restricts the model’s ability to form ice.

4) Mixed Layer Depth

Figure 10 shows the MLOM–CCM3 mixed layer depths for JJA and DJF and the corresponding observations for the same period. The observed mixed layer depths are from the Levitus dataset (Levitus and Boyer 1994). A comparison of MLOM–CCM3 to observations shows that, like observations, the mixed layer depths are deepest in the ACC during winter (JJA). While qualitatively getting this feature correct, the MLOM does not accurately predict the full extent of deepening in the ACC in either summer or winter. This is likely caused by the lack of sensitivity of the model to wind forcing, which was shown in a sensitivity analysis in which the strength of the winds was varied (results not shown). The observations reflect the importance of wind mixing on the depth of the mixed layer in the ACC. The MLOM appears to adequately capture the seasonal cycle of mixed layer depths in the extratropical regions. It contains a seasonal transition of mixed layer depths from shallow depths in summer to deep depths in winter. The seasonal cycle will be further investigated in the following section. The tropical mixed layer depths, in the MLOM–CCM3, are consistent with observations, with depths of approximately 20 to 60 m.

c. Seasonal cycle of sea surface temperature and mixed layer depth: Comparison of MLOM–CCM3, SOM–CCM3, MLOM stand-alone, and observations

Figure 11 shows the seasonal distribution of SST at 50°S, 30°S, 10°S, 0°, 10°N, 30°N, and 50°N along the 150°W longitude line, in the Pacific Ocean for the MLOM–CCM3, the SOM–CCM3, the MLOM stand-alone, and observations. The SOM–CCM3 is warmer
Fig. 11. Seasonal mixed layer temperature at several latitudes along 150°W (°C). The vertical axis is temperature in °C.
than observations but has a seasonal cycle that closely resembles observations. The MLOM stand-alone is closest to observations, particularly at the equator, 50°N, 30°N, and 30°S. The forcing in this simulation is taken from SOM–CCM3 with the horizontal heat flux correction (Q-flux), which has a climatology that is very similar to observations. This implies that the winds, heat fluxes, and moisture fluxes also closely resemble observations. The MLOM–CCM3 shows a lack of seasonality compared to observations. The temperature at 30°N and 30°S show the “most lack of seasonality” with temperatures that are close to observations during summer and fall, but which are significantly warmer than observations during winter and spring. However, these are the latitudes at which the horizontal heat transport is a maximum (Peixoto and Oort 1992). The lack of seasonality may therefore be attributed to the lack of horizontal heat transport. In addition, it is probable that the wind, heat, and moisture fluxes in the model differ from observations, which results in a difference in the temperatures predicted by the MLOM–CCM3. The relative accuracy of the MLOM stand-alone simulation speaks to the capability of the MLOM to generate an accurate representation of the observed SST.

Figure 12 shows the mixed layer depths for the locations specified above. The MLOM contains a pronounced seasonal cycle with shallowing in spring and summer and deepening in fall and winter. The seasonal cycle of the MLOM stand-alone simulation appears to represent the approximate behavior of the observed seasonal cycle in most cases. For example, the amplitude of the seasonal cycle for 50°, 30°, and 10° are close to observations, although the overall depth of the mixed layer is larger than observations for 10° and 30°. The MLOM–CCM3 overestimates the extent of mixed layer deepening in winter, except at 50°S where the winds probably play an important role in deepening. However, the MLOM is not as reactive to deepening from winds as it is from deepening from buoyancy forcing.

5. Discussion and conclusions

The mixed layer ocean model (MLOM) is a viable alternative to the more widely used slab ocean model, particularly for studies that involve the analysis of the impact of the deep ocean on climate. The MLOM is more physically realistic than the slab ocean model. It includes vertical ocean dynamics that explicitly determine the mixed layer depth and mixed layer entrainment. Both the mixed layer depth and entrainment directly affect the SST calculation through the effects of thermal inertia and vertical heat transport. The SOM is unable to accurately predict the observed SST without the specification of a heat flux correction, Q-flux, that artificially forces the model to replicate the observed SST. This causes the SOM to be limited in making predictions of SST that vary significantly from the observed SST. The MLOM directly calculates the vertical heat transport although it presently lacks a mechanism to determine the transport of heat by strong horizontal currents. It can therefore best be used for climate sensitivity and change studies other than observed climate, although it is limited by the specification of the deep ocean temperature and depth.

The coupling of the MLOM to the NCAR Community Climate Model (CCM3) shows that with the MLOM, CCM3 is able to generate a reasonable prediction of the observed climate given modern conditions. The general structure of the SST and the mixed layer depths from the MLOM are sufficiently close to observations. The MLOM, however, does have a problem replicating the meridional structure of the SST. This is likely caused by the lack of horizontal heat transport in the model and the specification of the deep ocean temperature, which is zonally invariant. Also, the coupled MLOM–CCM3 simulation does not determine the proper seasonal cycle of SST at all latitudes. However, the MLOM stand-alone simulation, which was forced with realistic winds and surface fluxes of heat and moisture, predicts a more accurate seasonal cycle. We therefore conclude that the inaccuracy of the seasonal cycle in the MLOM–CCM3 is linked to the surface forcings in the simulation, which are not guaranteed to approximate the observed forcings.

The MLOM–CCM3, like the coupled SOM–CCM3, has difficulty predicting the correct extent and depth of the sea ice coverage. This problem can be addressed in the MLOM–CCM3 by using a more accurate salinity and density calculation. At present, the salinity in the MLOM–CCM3 is only used to ensure the stability of the mixed layer. More attention should be paid to the specification of salinity. Additionally, we use a crude linear approximation of density in the model. The convective adjustment in the MLOM–CCM3 could be made more physical if a more realistic prediction of density was included.

In conclusion, the coupled MLOM–CCM3 simulation shows that the MLOM, like the SOM, may be used on a global scale to predict the SST in a coupled climate simulation. Although problems still exist with the model, many of them may be easily corrected. For example, one could specify a nonzonal deep ocean temperature to more accurately reflect the actual T_d values. Also, the density calculation can be made more physical so that the high-latitude calculation of SST and sea
Fig. 12. Seasonal mixed layer depth at several latitudes along 150°W (m). The vertical axis is depth in meters.
ice extent can be made more accurate. In a companion paper (Oglesby et al. 2005), we perform a paleoclimate sensitivity study with CCM3–MLOM, in which we evaluate the climate response to imposed changes in the deep ocean. This study thus serves as one example of how MLOM–CCM3 can be a useful tool for studies of past climates.

REFERENCES