Effects of penetrative radiation on the upper tropical ocean circulation

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Abstract: The effects of penetrative radiation on the upper tropical ocean circulation have been investigated with an ocean general circulation model (OGCM) with attenuation depths derived from remotely sensed ocean color data. The OGCM is a reduced gravity, primitive equation, sigma coordinate model coupled to an advective atmospheric mixed layer model. Our simulations use a single exponential profile for radiation attenuation in the water column which is quite accurate for OGCMs with fairly coarse vertical resolution. The control runs use an attenuation depth of 17 m while the simulations use spatially variable attenuation depths. When a variable depth oceanic mixed layer is explicitly represented with interactive surface heat fluxes, the results can be counterintuitive. In the eastern equatorial Pacific, a tropical ocean region with one of the strongest biological activity, the realistic attenuation depths result in increased loss of radiation to the subsurface, but increased sea surface temperatures (SSTs) compared to the control run. Enhanced subsurface heating leads to weaker stratification, deeper mixed layers, reduced surface divergence and hence less upwelling and entrainment. Thus, some of the systematic deficiencies in the present day climate models, such as the colder than observed cold-tongue in the equatorial Pacific may simply be related to inaccurate representation of the penetrative radiation and can be improved by the formulation presented here. The differences in ecosystems in each of the tropical oceans are clearly manifested in the manner in which biological heat trapping affects the upper ocean. While the tropical Atlantic has many similarities to the Pacific, the Amazon, Congo and Niger river discharges dominate the attenuation of radiation. In the Indian Ocean, elevated biological activity and heat trapping are away from the equator in the Arabian Sea and the southern tropics. For climate models, in view of their sensitivity to the zonal distribution of SST, using a basin mean of the ocean color derived attenuation depth reduces the SST errors significantly in the Pacific although they occur in regions of high mean SST and may have potential feedbacks in coupled climate models. On the other hand, the spatial variations of attenuation depths in the Atlantic are crucial since using the basin mean produces significant errors. Thus the simplest and the most economic formulation is to simply employ the annual mean spatially variable attenuation depths derived from ocean color.
1 Introduction

Variations in the attenuation of visible radiation in the upper ocean alters the vertical distribution of local heating, and has potential implications for thermal and dynamical processes as well as for ocean-atmosphere interactions. These variations are largely controlled by the concentration of light-absorbing pigments associated with phytoplankton that vary over a wide range of time and space scales. These variations have been investigated in detail, not only for their influence on the upper ocean structure, but also for generating full spectral radiance profiles over a range of oceanic conditions (Denman 1973, Simpson and Dickey 1981a, Lewis et al. 1990, Siegel et al. 1995, Ohlmann et al. 1996, Ohlmann et al. 2000). A potential for positive feedback between stabilization of the mixed layer due to biological heat trapping which can lead to enhanced stratification, and in turn to favorable conditions for phytoplankton growth was suggested by Sathyendranath et al. (1991). While this hypothesis can only be tested in a coupled ecosystem model, the influence of light attenuation due to phytoplankton growth on ocean circulation can be studied in ocean general circulation models (OGCMs). Such modeling studies have either been sophisticated solar transmission formulations in a one-dimensional (1D) setting (e.g., Ohlmann and Siegel 2000) or simplistic representations of penetrative radiation in realistic three-dimensional (3D) models (for e.g., Chen et al. 1994a, Schneider and Zhu 1998). One recent study in an isopycnal OGCM addressed the modulation of SST by surface chlorophyll (Nakamoto et al. 2000), but with relaxation to climatological quantities, and only for the Arabian Sea.

Simpson and Dickey (1981b) used a 1D, second-moment turbulent closure scheme with modification to account for solar flux divergence. Their investigation of the role of downward irradiance in determining the upper ocean structure showed that SST, mixed layer depths (MLDs), eddy diffusivity of heat, and the mean horizontal velocities were dependent on the particular form of solar irradiance divergence and the influence of the vertical profile of radiation depended also on the wind-speeds. One dimensional studies such as these have been instrumental in driving home the point that proper representation of solar irradiance is important for accurate computation of SSTs. The simulated profiles of full spectral radiance by Ohlmann and Siegel (2000) show that solar transmission at around 20 m of the ocean can exceed 40 W m$^{-2}$ for a surface irradiance of 200 W m$^{-2}$ as predicted by Lewis et al. (1990) and confirmed by Siegel.
et al. (1995). While these numbers are significant, state-of-the-art coupled climate prediction models, or even stand-alone OGCMs used for process studies, do not resolve the upper ocean to better than order ~ 10 m. Layer models such as the one employed here only resolve a bulk mixed layer and level models such as the one used by Schneider and Zhu (1998) have levels of order 10 m near the surface. Thus, in a practical sense, especially for climate prediction models, we only need to worry about solar radiation that escapes the top layer or level of the model. As pointed out by Ohlmann et al. (1998), solar transmission in the visible band within the oceanic mixed layer below 10 m, can be represented with a single exponential profile to within 10% accuracy.

Our goal is mainly to address the potential for enhancing the simulation of tropical SSTs and in turn, ENSO prediction models via improved representation of vertical profiles of solar radiation without incurring excessive computational costs. We thus focus on the tropics and investigate a single exponential profile with constant attenuation depth and compare it to model simulations with spatially variable attenuation depths derived from satellite ocean color data. Schneider and Zhu (1998) used a single exponential profile with an e-folding depth of 15 m in a coupled model to demonstrate several improvements in their simulation compared to when the penetrative radiation was completely ignored. Schneider et al. (1996) analyzed a coupled GCM run to conclude that penetrative radiation was important in their model in maintaining the vertical structure of the western Pacific warm pool. The OGCM study of Nakamoto et al. (2000) shows that the seasonal variability in surface chlorophyll results in seasonally dependent absorption of solar irradiance and heating rate in the upper ocean. Our simulations are more comprehensive and address dynamical and thermodynamical effects of heat trapping by surface chlorophyll in all three tropical oceans with realistic interactions between mixed layer depth variations and surface heat fluxes, and also between mixed layer depths and subsurface variabilities. Based on our simulations, we conclude that the spatially variable annual mean of the attenuation depths derived from satellite ocean color data are sufficient for capturing the largest part of the heat trapping.

The Coastal Zone Color Scanner (CZCS) data are only accurate enough for a composite annual mean over the global tropics and the newly available SeaWiFS ocean color data are not sufficiently long for computing climatological monthly mean attenuation depths. Some details of the CZCS and SeaWiFS data can be found in McClain et al. (1993) and McClain et al. (1998).
Interannual simulations of the tropical Pacific with constant and spatially variable annual mean attenuation depths are analyzed to understand the effects of interannual variability in ecosystems and attenuation depths. Access to routine space-based observations of ocean color and optical property data related to attenuation depth (McClain et al. 1998) make the present study timely and feasible.

The paper is organized as follows. Section 2 contains a brief description of the model and the formulation of penetrative radiation. Section 3 contains detailed comparisons of climatological model simulations with variable and constant attenuation depths. Discussion of interannual runs for the Pacific Ocean and effects of time-varying attenuation depths are presented in section 4. A summary of results is reported in section 5.

2 Model ocean

The ocean model is a reduced-gravity, primitive-equation OGCM configured separately for each of the tropical oceans. It consists of 19 sigma-coordinate layers beneath a variable-thickness, surface mixed layer (Chen et al. 1994b). Surface heat fluxes are determined by coupling a simple model of the atmospheric boundary layer to the OGCM and freshwater forcing is included as a natural boundary condition. Model details are presented in Murtugudde et al. (1996, 1998) and Murtugudde and Busalacchi (1999). Model salinity, temperature and layer thicknesses are relaxed to Levitus (1994) data near the open southern and northern boundaries. For our control run, the model is forced by Hellerman and Rosenstein (1983) winds, Xie and Arkin (1996) precipitation, ISCCP cloudiness, and ERBE all-sky radiation. Due to a lack of sufficiently long time series of reliable cloudiness and solar-radiation data, climatological ERBE radiation and ISCCP cloudiness are used in the interannual simulations to determine surface heat fluxes (also see Hackert et al. 2000). The 10 m winds from NCEP reanalyses (Kalnay et al. 1996) winds with a drag coefficient of 0.0015 are used to construct wind-stresses for the interannual simulations and wind speeds for latent and sensible heat fluxes are also computed from the 10 m winds.

The Pacific Ocean extends over 124°E-76°W×30°S-30°N with a resolution of 1° in longitude and a stretched grid in latitude (1/3° within 10°S-10°N and 1° near the boundaries). The Atlantic covers 90°W-14°E×40°S-40°N with 1/2° longitudinal resolution and stretched latitudinally (1/3° within 10°S-10°N and 1° near the boundaries). The Indian Ocean has uniform
$1/2^\circ \times 1/3^\circ$ resolution over $32^\circ$E-$126^\circ$W $\times 30^\circ$S-$26^\circ$N. Further details are not presented here since similar configurations have been used in our previous studies reported in Murtugudde et al. (1996), Murtugudde and Busalacchi (1998), and Hackert et al. (2000). The model performance in terms of simulations of SST, currents and thermocline structure are reasonable for the control runs as reported in earlier papers. The purpose here is not to demonstrate further improvements in model performance although annual mean model SSTs are compared to Levitus SSTs to determine if basin mean values for attenuation depths can be employed in climate models to simplify the formulation of penetrative radiation. It is shown that while this appears to be reasonable in the Pacific, using the basin basin mean attenuation depth does degrade SST simulation in the Atlantic at the mouths of Congo and Niger rivers.

Solar radiation impinging just below the surface of the ocean ($I(0)$) consists of energy distributed across a wide spectral range. For wavelengths longer than 700 nm, the ocean is strongly absorbing, and these wavelengths contribute to heating the upper few centimeters. For shorter wavelengths, the ocean is more transparent, and solar energy crossing the air-sea interface propagates to depths where it contributes to a penetrating flux of solar energy and to local heating at depths removed from the surface. The fraction of total solar flux that resides in these penetrative wavebands is approximately 47% (Frouin et al. 1989). This value varies slightly depending on atmospheric aerosols, but these variations are relatively small especially for the problem addressed here. This fraction we designate as $\gamma$, and assume it to be constant for our model runs below. The remaining fraction $(1 - \gamma)$ is fully absorbed in the upper layer of the model ocean.

The penetration of solar irradiance follows an exponential decline with depth, with the derivative (e.g., the losses) contributing to local heating. In our previous studies of oceanic processes, a single exponential profile was used as the empirical form for the penetrative component of the solar irradiance (Chen et al. 1994a, Murtugudde et al. 1996, 1998, Murtugudde and Busalacchi 1999, Murtugudde et al. 1999), namely, $I(z) = I(0) \times \gamma \times \exp(-z/h_\gamma)$. Here, $I(z)$ is the downwelling penetrating solar irradiance at a given depth $z$ (positive downward). The value $h_\gamma$ represents the e-folding depth scale for the penetration of irradiance over the penetrative wave bands (380 - 700 nm); it is the inverse of the spectrally averaged diffuse attenuation coefficient ($K_D$). Strictly, it is more accurate to use a full spectral approach (e.g., Lewis et al. 1990, McClain et al. 1999), but the increase in accuracy is offset by the significantly increased computational load imposed by hyperspectral decomposition of the absorption of solar irradiance.
when implemented in a 3-dimensional OGCM context.

Based on Chen et al. (1994a), our previous studies employed $\gamma = 0.33$ and $h_\gamma = 17 \text{ m}$. Here, $\gamma = 0.47$ for all experiments as explained above. Our control runs were done with the $h_\gamma = 17 \text{ m}$ and these runs are compared with runs where $h_\gamma$ is computed from annual mean CZCS data (Fig. 1). To estimate $h_\gamma$ from the retrieved pigment concentration, we take $h_\gamma = 1/K_D$ where we assume $K_D$ can be represented as in Morel (1988) as $K_D = 0.027 + 0.0518 \times CHL^{0.428}$. In this formulation, the coefficient $0.027 \text{ m}^{-1}$ is that of pure ocean water ($h_\gamma = 37 \text{ m}$), and CHL is the CZCS derived pigment concentration. This formulation was originally developed to model the integral photon flux over the same wavebands as we are interested in here. We have assumed that this formulation is sufficiently accurate for simulations of the penetration of energy given the relatively small spectral variations over the relatively clear-water open ocean regions of interest.

The annual mean $h_\gamma$ is as low as $\sim 5 \text{ m}$ in regions of elevated biological activity such as the coastal and open ocean upwelling regions, and above $30 \text{ m}$ in the oligotrophic subtropical gyres. The constant value of $h_\gamma = 17 \text{ m}$ is a good approximation for the global mean of Fig. 1, but clearly not representative of any particular region or basin. The differences between CZCS derived attenuation depths and the constant value of $h_\gamma = 17 \text{ m}$ are shown in Fig. 2. The CZCS values are deeper than $17 \text{ m}$ in the Pacific nearly everywhere except in the eastern coastal regions. In the Atlantic, the mouths of the Amazon, Congo, Niger and Orinoco are clearly seen as regions with attenuation depths shallower than $17 \text{ m}$ as are the upwelling regions of Guinea and Angola domes. The Indian Ocean also has attenuation depths shallower than $17 \text{ m}$ in biologically active regions such as the coastal regions and the Arabian Sea.

The simulations with variable attenuation depth are referred to as PACVAR, INDVAR, and ATLVAR for the Pacific, Indian and the Atlantic Oceans, respectively.

3 Results

3.1 Effects of neglecting penetrative radiation

Chen et al. (1994a) compared the SST and its seasonal cycle along the equator with and without penetrative radiation to argue that when penetrative radiation was neglected there was excess warming during the weak wind conditions of the boreal spring months. Schneider and Zhu
(1998) compared a coupled model simulation with and without penetrative radiation. As stated earlier they used a constant attenuation depth of 15 m with a $\gamma = 1$. In the equatorial Pacific, they found that the penetrative radiation led to deeper mixed layers and reduced sensitivity of SSTs to upwelling. Our results are similar in that the subsurface stratification does weaken leading to deeper mixed layers and a more realistic cold tongue except in the far east near the coast (Fig. 3). The differences in SST and net heat fluxes between the control run and a simulation neglecting the penetrative radiation show that the surface fluxes tend to damp the SST differences. As expected, the effects of penetrative radiation are maximum in the upwelling season (Jul-Sep).

The minimum depth of the mixed layer in Schneider and Zhu (1998) was effectively 30 m (top two levels), and thus the wind-forcing was independent of MLD for mixed layer shallower than 30 m. This is crucial in the eastern equatorial Pacific and Atlantic Oceans where the MLDs are shallower than 30 m except when the trades are at their maximum during the boreal summer months. Since our model is forced with specified winds, the warming of the water column below the mixed layer leads to a weakening of the stratification and a rapid deepening of the mixed layer and reduced surface divergence. The ensuing weaker upwelling leads to warming of the SSTs despite the loss of radiation to the subsurface when penetrative radiation is accounted for. Thus the response to biological heat trapping is stronger in our model. For example, even though a constant attenuation depth is employed in the control run, the effects of penetrative radiation are maximum in the upwelling season when the phytoplankton concentrations and their effect on heat trapping are maximum.

In a coupled climate model, the SST changes in the cold-tongue region can potentially lead to a positive feedback since warmer eastern Pacific would result in a reduced east-west SST gradient and weaker winds. A reduction in upwelling would reduce the nutrient supply and diminish the biological response as is the case during warm ENSO events. Employing observed surface chlorophyll distributions to derive the attenuation depths would presumably account for the feedbacks that are involved in producing the observed ecosystem response. Note however that as pointed out by Schneider and Zhu (1998), the strong coupling between the equatorial zonal winds and off-equatorial SST is not represented here and will alter the feedbacks found in our simulations. Nonetheless, our simulations highlight the need for considering finer vertical resolution with penetration of momentum and solar irradiance to accurately capture all the
feedbacks in regions like the eastern equatorial Pacific.

3.2 Effects of variable attenuation depths

The results are presented as annual mean and seasonal differences between the simulations with spatially variable attenuation depths (Fig. 1) and control runs with constant attenuation depth of $h_\gamma = 17$ m. The value of $\gamma$ is set to 0.47 for all runs based on Frouin et al. (1989).

3.2.1 Pacific Ocean simulations: The annual mean differences between model and Levitus SSTs are shown in Fig. 4. As is the case for most state of the art OGCMs (Stockdale et al. 1994), the cold tongue is colder than observed by nearly 2.5°C. A band of colder than observed SSTs also extends across the basin around 15°N which maybe related inadequacies of the model or errors in the winds, radiation or cloudiness. Since the corrective fluxes required to correct these SSTs errors are well within the errors in the surface fluxes used to force the model, it is not possible to unambiguously place the blame on model deficiencies. The outcropping and thinning of the shallow isopycnals in the upwelling region during the boreal summer leads to excess supply of cold water near the surface during the following boreal spring since most models fail to restratify the water column. The excessive cooling in the cold tongue region is generally alleviated by reducing the Hellerman and Rosenstein (1983) winds by an arbitrary amount or the vertical mixing is tuned to improve simulated circulation in the region (for e.g., Yu and Schopf 1997).

The annual mean differences in surface temperatures and currents between PACVAR (spatially variable $h_\gamma$) and the control run ($h_\gamma = 17$ m) for the Pacific are shown in Fig. 5. The SST warming is as high as 1°C for the equatorial Pacific cold-tongue region. It is quite evident that the large region of colder than observed SSTs in the eastern equatorial and coastal region is greatly alleviated by employing the correct attenuation depths. During the boreal spring months when the MLDs are at their minimum (Chen et al. 1994a, Murtugudde et al. 1996) and entrainment is also at its minimum due to weakening of the trades, the subsurface warming is maximum. The restratification of the water column below the mixed layer is thus simply achieved by the subsurface heat source provided by the penetrating radiation. It is important to note that no fine tuning of the vertical mixing needs to be resorted to when the vertical distribution of solar energy is properly represented. It is thus clear from Fig. 5 that the colder
than observed cold tongue that plagues most OGCMs (Stockdale et al. 1994) may simply be related to the errors in the formulation of penetrative radiation.

The surface and subsurface current differences show that the equatorial divergence and the upwelling are diminished and the South Equatorial Current (SEC) is weakened. In the Pacific, the available radiation below the mixed layer is higher for PACVAR which clearly leads to warmer subsurface temperatures as seen in Fig. 5. Note also that a deeper mixed layer results in weaker surface currents and reduced equatorial divergence, which further enhances the surface warming and the weakening of stratification. The off-equatorial SST errors are clearly not affected as much since there is no wind-feedback in our model (Schneider and Zhu 1998).

The subduction of saline central Pacific waters around 160°W is enhanced in Fig. 5 which may contribute to maintaining the barrier layer structure in the western Pacific warm pool region (Lukas and Lindstrom 1991, Murtugudde and Busalacchi 1998). The role of the barrier layer in the supply of nutrients to the euphotic zone in the warm pool has been conjectured to be important (Mackey et al. 1995, McClain et al. 1999, Murtugudde et al. 1999). More studies with coupled bio-physical models are needed to understand whether radiative trapping due to biology in the east and associated dynamical changes really contribute to the barrier layer structure in the west. The middle panel of Fig. 5 shows that most of the warming occurs below the mixed layer and is caused by dynamical changes associated with penetrative radiation rather than direct warming due to the radiation itself. That is, the slight warming below the mixed layer by the penetrative radiation and the consequent weakening of the stratification leads to dynamics feedbacks described above that further enhance the subsurface warming. The advective time-scales thus dominate the heating time-scale leading to larger effects on the thermal structure as noted by Schneider and Zhu (1998).

A vertical profile of the radiation for the control run and PACVAR can be instructive in understanding the subsurface effects of penetrative radiation. The top panel of Fig. 6 shows the profiles at 129°W on the equator. For a given depth, PACVAR has higher penetrative radiation than the control run at all depths. This leads to more warming at that depth and hence weaker stratification. Since the wind-forcing is the same in both runs, the available TKE leads to deeper MLDs in PACVAR, and thus the surface currents weaken and the surface divergence diminishes. The warming shifts the temperature profile nearly uniformly down to 100 m. Note that the control run employed a shallower attenuation depth in the cold tongue region than that
computed from CZCS (Fig. 2). Most models typically employ a number in this range (for e.g., Schneider and Zhu (1998) used 15 m and Nakamoto et al. (2000) used 23 m).

The bottom panel of Fig. 5 shows the annual mean MLD differences for the tropical Pacific. As mentioned above, the MLDs are deeper for PACVAR in the cold tongue region. Surprisingly, the MLDs are also deeper in the relatively weak wind regions of the western Pacific warm pool in the ITCZ and SPCZ regions. The inadequacy of the formulation of turbulent kinetic energy generation is often blamed for shallower than observed mixed layer depths in these regions (for e.g., Murtugudde et al. 1996) while proper representation of penetrative radiation may be the major cause of the model deficiency.

The accompanying SST differences in these regions are however much smaller than in the cold tongue due to smaller available surface heat fluxes and also due to the negligible surface effect of subsurface changes. These small SST differences could still be significant since the mean SSTs in these regions are near the threshold of atmospheric convection. Lewis et al. (1990) estimated the effects of penetrating solar radiation on the heat budget of the equatorial Pacific. Using transparency of the ocean computed from CZCS and mixed layer depths estimated from ship observations, they concluded that most of the net surface heat flux is lost below the mixed layer as penetrative radiation. This is significant since the heat balance and the mechanisms for SST regulation in the western Pacific warm pool are still an open question (for e.g., Ramanathan and Collins 1991, Fu et al. 1992).

Annual mean surface heat flux and the penetrative radiation for the western Pacific are shown in Fig. 7 for the control run and for PACVAR. It is evident that the loss of penetrative radiation is significantly higher in the warm pool region for PACVAR and most of the net flux is indeed lost to the subsurface. The loss of radiation to the subsurface removes the heat from the layer that interacts strongly with the atmosphere and points to the need for revisiting the detailed heat budget calculations of the warm pool (Gent 1991) with accurate representation of the penetrative radiation losses.

While the deepening of the mixed layer due to penetration of solar radiation to greater depths was noted by Denman (1973), our result is rather counter-intuitive. That is, a greater loss of radiation from the mixed layer leads to warmer SSTs because of weaker stratification, deeper mixed layers, and reduced upwelling. The deepening of the mixed layer due to weakened stratification is captured in the level model in Schneider and Zhu (1998) by employing a
Richardson number based mixing (Pacanowski and Philander 1981) but the minimum depth is
effectively equal to the depth of top two levels (30 m). Their Fig. 9 clearly shows deepening
of the MLDS due to penetrative radiation even though the differences are less than 5 m and
the annual mean SSTs are colder over most of the eastern equatorial Pacific despite the large
increase in subsurface heating. This could simply be due to a lack of vertical resolution and the
fact that the minimum MLD was 30 m whereas the MLD variability in this region is between
5 m and 30 m. As stated earlier, it is also possible that the feedbacks between the equatorial
zonal winds and the off-equatorial SSTs may explain the differences. It is thus important to
use a coupled model not only with variable attenuation depths, but also with realistic variable
mixed layer depths in order to allow feedbacks between mixed layer variations, surface fluxes,
and subsurface stratification.

3.2.2 Atlantic Ocean simulations: The annual mean differences in surface tempera-
tures and currents for the Atlantic Ocean are shown in Fig. 8. There are some similarities in
the SST and current changes in the Pacific (Fig. 5) and the Atlantic Oceans since both of them
have seasonal surface phytoplankton blooms caused by equatorial upwelling. Note however that
the Atlantic is also affected by the radiation attenuation at the mouths of Congo and Niger
rivers in the east that could be due to suspended sediments and dissolved organic matter. The
Atlantic Ocean displays slightly stronger subsurface warming, especially near the eastern coast.
Westward surface currents are weaker to the east of 25°W, but stronger to the west of 25°W.
Similarly, upwelling is reduced around 10°W, but the mean downwelling is enhanced to the west
of 30°W.

The reasons for a weaker SEC in the western equatorial Atlantic are less straightforward.
The eastern equatorial Atlantic has fairly shallow attenuation depths due to Congo and Niger
river plumes which trap radiation, that serve to shallow the mixed layer (Fig. 8). This leads to
strengthening of the local eastward surface currents and the shallow downwelling near the coast
(Philander and Pacanowski 1986), leading to subsurface warming despite lowered penetrative
radiation. An increase in the penetrative losses in the western half of the equatorial region leads
to substantial subsurface warming there leading to deepening of the mixed layer similar to the
Pacific cold tongue region. The currents on the equator to the west of 25°W tend to be weakly
eastward associated with a minimum in the SEC (Philander and Pacanowski 1986), which are
weakened by deeper MLDs. Thus the weaker eastward currents and increased downwelling lead to local deepening of the thermocline to the west of 25°W, which enhances the subsurface warming in the west (Fig. 8, middle panel). Thus the surface and the subsurface temperature differences are warmer in the east and west due to different processes. The differences in the structure of the Equatorial Undercurrent (EUC) and the thermocline slope in each of the oceans (Philander and Chao 1991) is clearly manifest in the way they are affected by enhanced heating due to penetrative solar radiation. The change in thermocline slope centered around 25°W seen in the Atlantic (Fig. 8) is related to the seasonal response to winds with a small fetch (Philander and Chao 1991).

The bottom panel of Fig. 8 shows that the MLDs are shallower in the Guinea and Angola domes in addition to the mouths of the Amazon, Orinoco, Congo and Niger rivers. Shallower attenuation depths in ATLVAR (Fig. 2) increase the amount of radiation trapped near the surface and lead to shallower mixed layers. Deeper mixed layers in the western equatorial region just outside of the Amazon river discharge are caused by increased downwelling mentioned above. The interactions of radiation trapping and ocean circulation are clearly more complex in the tropical Atlantic not only due to biological activity, but also due to turbidity associated with riverine inputs.

The profiles of radiation at 5°E on the equator (bottom panel of Fig. 6) show that ATLVAR traps more radiation near the surface which is quite the opposite of the Pacific (top panel of Fig. 6). The subsurface warming is also different with less warming immediately below the mixed layer in the Atlantic. Note, however, that the subsurface warming in the Atlantic is confined largely to the east and west away from the main central equatorial upwelling region whereas in the Pacific the maximum subsurface warming is around the region of maximum equatorial upwelling (around 120°W). The differences between the control run and ATLVAR in the profiles of radiation are much higher than in the Pacific (top panel of Fig. 6) with the mixed layer being shallower in ATLVAR compared to the control run.

3.2.3 Indian Ocean simulations: The warming in the Indian Ocean due to biological heat trapping (Fig. 9) occurs away from the equator, but also in the regions that are known to be biologically active (Banse and McClain 1986, Murtugudde et al. 1999), namely, the Arabian Sea, around Sri Lanka, and in the southern tropical Indian Ocean. The local upwelling
and the associated ecosystem response at the southern tip of India are shown in Murtugudde et al. (1999). This is the region where a monsoon onset vortex forms annually (Rao and Sivakumar 1999) and an accurate simulation of SST in the region is important. The proximity of the thermocline to the surface due to large scale wind-stress curl and the associated Ekman pumping in the southern tropical Indian Ocean (Reverdin et al. 1986) does not always manifest itself in SSTs (Murtugudde and Busalacchi 1999). However, the intermittent entrainment of nutrients is seen as a region of surface phytoplankton bloom (Murtugudde et al. 1999) and this clearly affects surface heat trapping due to attenuation of incident radiation (top panel of Fig. 9). The warming in the Indian Ocean is typically less than 0.5°C in the annual mean, but higher in some seasons as shown later. As in the western Pacific warm pool region, the mean SSTs in the Indian Ocean are high and even small SST increases can be climatically significant.

The subsurface temperature differences in the Arabian Sea are minimal as can be inferred from Fig. 6 (middle panel). The MLDs are shallower for INDVAR (bottom panel of Fig. 9) as in the Atlantic and thus more radiation is trapped near the surface. But the differences in MLDs and radiation profiles with depth in the Arabian Sea are not as large as in the Atlantic. The MLDs are typically much deeper (\( \sim 60 \text{m} \)) than the attenuation depths in the Arabian Sea and thus the differences in the radiation escaping the mixed layer depth is minimal between the control run and INDVAR. The river mouths in the Bay of Bengal are demarcated by shallower MLDs and slightly warmer SSTs. The effects of freshwater input and surface heat trapping in the Bay of Bengal appear to be minimal compared to that in the Atlantic Ocean which is consistent with recent observations and modeling studies (Sengupta 2000, personal communications; Howden and Murtugudde 2000, Han et al. 2000).

Surface current changes include a weaker eastward equatorial Wyrtki Jet, stronger northward Somali Jet, and a stronger SEC in the southern tropical Indian Ocean. Slightly stronger alongshore currents along the coast of Sumatra and the associated cooling of SSTs can potentially be important since this region has a warm pool like characteristic with warm SSTs and a barrier layer structure (Murtugudde and Busalacchi 1999). Note also that this is the region of significant upwelling and surface cooling during anomalous Indian Ocean events (Murtugudde et al. 1999, Murtugudde et al. 2000).

The equatorial Indian Ocean is dominated by the strong semiannual Wyrtki Jets (Wyrtki 1973) which produce a convergence and downwelling in the east leading to a deeper mixed layer.
The attenuation depths in the control run (17 m) are typically shallower than in INDVAR except close to the coasts and in the Arabian Sea (bottom panel of Fig. 2). The MLDS respond by deepening in the equatorial and southern tropical Indian Ocean similar to the Pacific. Note however that the SST differences on the equator are near zero or slightly negative because in the Indian Ocean, while deeper MLDS lead to weaker surface currents, the mean currents are eastward. The weaker Wyrtki Jets and diminished downwelling in the east lead to cooling of the surface layer. But the temperatures are warmer in the thermocline region because the west to east slope of the thermocline is relaxed due to weaker eastward surface currents. The subsurface current structure along the equator is more complex in the Indian Ocean compared to the Pacific and Atlantic Oceans. The differences in the currents also display complex east-west structure below the mixed layer (middle panel of Fig. 9).

The surface and subsurface currents in the Indian Ocean display seasonal reversals due to the monsoon wind variations from the southwest to the northeast monsoon. Thus, annual mean differences can hide much larger seasonal effects in the Indian Ocean. Nakamoto et al. (2000) report from their OGCM experiments that the heating rate in the upper ocean in the Arabian Sea displays seasonality due to variability in surface chlorophyll concentrations. Nakamoto et al. (2000) used an isopycnal model with $h_r = 23$ m for the control run and the monthly mean CZCS derived attenuation depths for the simulation. The surface heat fluxes were specified and surface salinity was relaxed towards Levitus annual mean values. The seasonal cycle of the solar radiation is characterized by minima during southwest and northeast monsoons and maxima during inter-monsoon periods (see Fig. 3 of Nakamoto et al. 2000). The seasonality of the surface chlorophyll concentrations shows a strong boreal summer peak in the Somali current region and two weaker peaks in the Arabian Sea during the two monsoons.

Our results for seasonal differences in SST due to biological heat trapping are shown in Fig. 10 which are nearly identical to the results of Nakamoto et al. (2000). Maximum warming of about 0.6°C occurs during March-April with a slight cooling during July-August. A slightly smaller warming is seen during October through January. The small differences between Fig. 10 and Fig. 2 of Nakamoto et al. (2000) can easily be explained by the differences in the constant attenuation depth (our 17 m vs. their 23 m) and the differences in surface heat flux formulations. The more important thing is to note that the seasonality and the magnitude of warming can be simulated fairly accurately even by using annual mean attenuation depths. This
is important for computational efficiency in coupled climate models.

4 Simulations with basin mean attenuation depths

The previous sections have demonstrated that penetration of shortwave radiation can indeed affect the upper ocean circulation, especially SSTs. A recommendation can be made at this stage that coupled climate models or modeling studies that aim to understand processes that control SSTs should use accurate attenuation depth distributions. It is worth exploring however, if basin mean values of Fig. 1 can be used instead, to further simplify the formulation of penetrative short wave radiation. This would be more accurate than the global mean number of 17 m used in the control runs. For this purpose, we carried out climatological simulations in each basin with mean values for the domains (30°S–30° for the Pacific, 40°S–40°N for the Atlantic, and 30°S–26°N for the Indian Ocean, respectively) of 25 m, 20 m, and 21.5 m, for the Pacific, Atlantic, and Indian Oceans respectively. The simulations PACVAR, ATLVAR, and INVAR are compared to simulations with these basin mean attenuation depths (called runs PACB1, ATL1, and INDB1) in Fig. 11 in terms of annual mean SSTs and surface currents.

In the Pacific Ocean, the largest SST differences in Fig. 5 were seen in the cold-tongue region where the attenuation depth from Fig. 1 is about 20 m. Increasing the attenuation depth to 25 m hardly has any effect (top panel of Fig. 11) on SSTs or subsurface temperatures (not shown). There is a cooling of nearly 0.5°C off the coast of Central America in the eastern Pacific ITCZ region which can be significant since this is a region of high mean SSTs. The rest of the tropics is not significantly affected by employing a basin mean attenuation depth. The sensitivity to changes in attenuation depths is much higher when the attenuation depth is close to the thermocline region as was the case when \( h_\gamma = 17 \) m was changed to \( h_\gamma = 20 \) m near 129°W on the equator (top panel of Fig. 6). The seasonal evolution of the differences in temperature vs. depth at 129°W on the equator between PACVAR (variable \( h_\gamma \)) and the control run \( (h_\gamma = 17 \) m) and between PACVAR and PACB1 \( (h_\gamma = 25 \) m) are shown in Fig. 12. It should be kept in mind that these differences are between equilibrium solutions with the two attenuation depths and not the evolution of the differences. These differences shown at one location also include the effects of the three-dimensional circulation changes. Thus the dynamical effects from regions east of 129°W where the \( h_\gamma \) are over 23 m affect the location
shown in Figs. 6 and 12.

At 129°W, the mixed layer shoals to about 15 m in the control run during March-April whereas it only shallows to about 19 m in PACVAR (Fig. 12. When the two runs are spun up, they are both started from the same initial conditions based on Levitus (1994). The tendency of the model is to generate excess shoaling of the thermocline in the east. In the control run, the heat trapping is below the mixed layer but above the main thermocline during early part of the calendar year whereas for PACVAR, it is in the main thermocline. Thus the stratification is weaker for PACVAR during April-June leading to deeper mixed layers and the feedback described previously. While the exact sensitivity to changes in the attenuation depths in terms of the amplitude of the subsurface and SST warming are clearly dependent on the vertical resolution of the model, the feedbacks should be acting in the same sense for all models that tend to produce colder than observed cold tongue.

The weakened stratification in April-May leads to subsurface warming of over 1.5°C in PACVAR which further affects its evolution during the fall months and leads to SST warming of over 1°C in late September. The differences between PACVAR and PACB1 show that the weakening of the stratification and the associated subsurface warming are far less dramatic since the heating below the mixed layer is not altered significantly by changing $h_\gamma$ from 20 m to 25 m. There is still some subsurface warming for PACB1 compared to PACVAR and it does lead to deeper MLDs in the fall. However, the water being entrained into the mixed layer is not altered much, leading to similar net surface heat fluxes. The slightly deeper PACB1 thus warms a little less than PACVAR.

The Indian Ocean (bottom panel of Fig. 11) is also nearly unaffected by using a basin mean attenuation depth albeit seasonal SST differences can be slightly larger. The largest differences in SSTs in Fig. 11 are seen in the Atlantic Ocean, again at the mouths of the Congo and Niger Rivers. The attenuation depths from Fig. 1 in this region are $\sim 5$ m while the basin mean is 20 m. Not surprisingly, this degrades the model performance in terms of SST errors. Annual mean differences between model and Levitus SSTs for ATLVAR and ATLB1 shown in Fig. 13, clearly show that the model SSTs are much closer to Levitus SSTs in ATLVAR than in ATLB1. This is not surprising since our control run with $h_\gamma = 17$ m was already larger than the local attenuation depths in the Congo and Niger regions (Fig. 2). Note that the surface heat fluxes are computed by the atmospheric mixed layer model and no corrective fluxes or feedbacks to
observations are employed in any of the simulations.

It is evident from these comparisons that while the SST differences for the Pacific and Indian Oceans are apparently small in the annual mean, seasonal differences need to be investigated for their climatic impacts. In the Atlantic, the differences are unacceptably large even in the annual mean. Thus, it is best to use annual mean spatially variable attenuation depths for the most accurate SST simulations.

5 Interannual simulations in the Tropical Pacific

It is well known that the response of the ecosystem in the Pacific to ENSO events is fairly dramatic (Barber and Chavez 1983) with diminished (enhanced) chlorophyll concentrations in the cold tongue during warm (cold) ENSO events. This clearly affects the oceanic transparency to solar irradiance. The ocean color data from SeaWiFS (McClain et al. 1998) clearly demonstrates this effect for the El Niño-La Niña events of 1997-98 in terms of attenuation depths (not shown). Since the SeaWiFS time-series is not sufficiently long for forming monthly climatologies, we attempt to understand the potential impact of interannual variability of the ecosystem response on radiation attenuation, using our simulations with constant and spatially varying annual mean attenuation depths. It should remembered that the same interannual winds are forcing all simulations, hence the model SSTs reproduce observed SSTs in the NINO3 and NINO4 regions with high fidelity and the differences would be much higher in a coupled system if wind-SST feedbacks were to be allowed.

Since the cold tongue in the eastern equatorial Pacific shows the largest SST differences in Fig. 5, we will focus on this region and the most common ENSO index, namely, NINO3 (SST anomalies averaged over 150°W– 90°W, 5°S–5°N). During warm ENSO events, the thermocline is anomalously deep, upwelling is reduced, and biological activity is diminished (Barber and Chavez 1983). This would clearly result in an attenuation depth that would be larger than climatology, i.e., incident radiation would be able to penetrate much deeper. The opposite would be true during a cold ENSO event with enhanced upwelling, increased surface chlorophyll concentrations, and reduced attenuation depth. As seen in Fig. 1, the control run has shallower attenuation depths in the cold tongue than PACVAR and the run PACB1 with basin mean attenuation depth of 25 m has a deeper attenuation depth than PACVAR. We contrast these
simulations for the variability and evolution of NINO3 for insights into the effects of ENSO events on ecosystem response and hence penetration of solar radiation.

The differences in SST and SST anomalies in the NINO3 and NINO4 regions are shown in Fig. 14. The subsurface differences are larger (not shown) but the response for the cold and warm ENSO events can be understood in terms of differences for climatological runs shown in Figs. 5 and 11. During warm ENSO events the attenuation depth should be typically deeper than climatological values. Thus the SST differences would be like the top panel of Fig. 11, i.e., reduced impacts on SSTs. During cold ENSO events, the attenuation depths are shallower than climatological values, the SST differences would be similar to Fig. 5, i.e., warmer SSTs for PACVAR compared to the control run.

This is confirmed by Fig. 14. A few other features are immediately obvious. The differences in SSTs are somewhat smaller in the NINO4 region. The differences in SSTs and SST anomalies are smaller in both regions for PACVAR and PACB1 (C-B1) compared to the differences between PACVAR and the control run (C-B). Thus increasing the attenuation depth beyond the climatological mean values has less of an impact since most of the radiation is trapped at the climatological attenuation depths. This implies that during a warm ENSO event, deepening of the attenuation depths does not affect the upper ocean circulation significantly. However, details of the exact response of the coupled system to SST changes of this magnitude should be investigated in coupled models, especially for the effects of such SST differences on the east-west SST gradient and potential positive or negative feedbacks.

The other characteristic evident in Fig. 14 is the correlation of the differences between PACVAR and the control run (C-B) with NINO3 index itself. During warm ENSO events (e.g., 1982-83, 1986-87, 1997-98) the differences tend to be minimum and during cold events (e.g., 1984, 1988) they tend to be maximum. Similar behavior is also seen for the differences between PACVAR and PACB1 (C-B1) in the NINO3 region. This is not surprising since the deepening of the thermocline and the associated increase in the MLDs and attenuation depths would make it less sensitive to going from \( h_{\gamma} = 17 \) m in the control run to a deeper value in PACVAR. On the other hand, during cold ENSO events, higher surface chlorophyll concentrations and reduced oceanic transparencies with shallower MLDs would increase near surface heat trapping. This would mean SST differences between PACVAR and the control run would be in the same sense as seen in Fig. 5. Thus the error of using climatological attenuation depths would be larger
during cold ENSO events than during warm ENSO events. Using climatological attenuation depths results in warmer than observed SSTs in the eastern equatorial region during cold ENSO events which may reduce the east-west SST gradient, weaken the trade winds, and provide a negative feedbacks. On the other hand, in the real world, a shallowing of the attenuation depth due to enhanced surface chlorophyll levels may trap more heat in the mixed layer and also provide the same negative feedback. These feedbacks can only be investigated in fully coupled ocean-atmosphere and ocean-ecosystem models.

In summary, we can assume that since $h_{\gamma} = 17 \, m$ is shallower than climatological attenuation depth in the cold tongue, the control run is more realistic for cold ENSO events, and $h_{\gamma} = 25 \, m$ is deeper than the climatological value and thus more realistic for warm ENSO events. The SST differences between these simulations can be considered SST errors caused by using climatological attenuation depths. It can be inferred from Fig. 14 that the SST errors in the NINO3 and NINO4 regions are small ($\sim 0.2^{\circ}$C) during warm ENSO events and larger ($\sim 0.4^{\circ}$C) during cold ENSO events.

The remaining question of course is just how much smaller does the attenuation depth get in the cold tongue region during La Niña events? This question can be answered with some confidence based on the SeaWiFS data for the La Niña of 1998. This was one of the strongest known cold ENSO events in the last 50 years (McPhaden 1999) with one of the highest equatorial surface chlorophyll concentrations (Murtugudde et al. 1999) observed. The attenuation depths for the boreal summer months of 1998 reached a minimum of about 15 $m$ (not shown) and the SST errors in Fig. 14 were only slightly higher. Similarly, for the peak of the warm event during 1997-98, the attenuation depths were as deep as 22 $m$ in the cold tongue, which is only slightly deeper than the climatological values seen in Fig. 1. It should also be noted that the errors in SSTs will be much larger for cold events if the basin mean of 25 $m$ is employed.

Thus the conclusions drawn above regarding the SST errors associated with using climatological values will likely hold in general. Moreover, the goal here is to recommend ways to improve SST simulations in coupled ENSO prediction models which attempt to make 6 month to 12 month forecasts. Resorting to interannually varying attenuation depths will require estimating attenuation depths in the forecast mode which may not add substantially to the forecast skill as can be inferred from Fig. 14. A significant improvement in the initial conditions may be achievable on the other hand, by using accurate representation of penetrative radiation.
6 Concluding Remarks

An OGCM for the tropics was employed to study the effects of penetrative shortwave radiation on the upper ocean dynamics and thermodynamics. The ocean model explicitly represents a bulk mixed layer and is coupled to an advective atmospheric mixed layer. The SSTs are allowed to evolve freely with the only external quantities specified being those that are not directly controlled by SSTs, namely, winds, solar radiation, and cloudiness. This allows us to study the processes that affect SSTs when subsurface penetration of incident solar radiation is ignored or otherwise modified. The main goal was to explore the possibility of recommending a computationally economic, yet reasonably accurate formulation of penetrative solar irradiance for coupled climate models.

Our previous studies have resorted to a single exponential profile of radiation in the water column with a single value of attenuation depth for the entire tropics. Ignoring the penetration of shortwave radiation entirely produces results similar to the coupled model experiments reported previously. However, the processes involved in generating the surface effects are remarkably different because a variable depth mixed layer is represented explicitly. The sensitivity of the equatorial Pacific cold tongue to equatorial divergence and upwelling is reduced due to weakening of the subsurface stratification and deeper mixed layers for the same wind-forcing. This could be important in coupled models in determining the coupled feedbacks and the seasonality of the heat trapping effects even though a constant attenuation depth is employed.

Our main simulations replaced the constant attenuation depth with spatially variable annual mean attenuation depths derived from ocean color data. The region of most interest, namely, the cold-tongue in the eastern equatorial Pacific has climatological attenuation depths of $\sim 20$ m, which is deeper than the constant value of 17 m used in our control runs. This clearly determines the differences presented throughout the paper. However, 17 m is a typical number used in most state of the art models. We also present sensitivity of the upper ocean circulation to a different constant in each of the tropical oceans. Model simulations with ocean color derived attenuation depths are compared with control runs with constant attenuation depth in each of the tropical oceans separately.

The cold tongue in the eastern equatorial Pacific is warmer when realistic attenuation depths
are used indicating that the colder than observed cold tongue that plagues most of the present day OGCMs may simply be related to inaccurate representation of the radiation profile. The heat budget of the western Pacific may be crucially dependent on penetration of solar radiation since most of the net surface flux is lost to the subsurface. The interannual simulations are more erroneous in terms of NINO3 and NINO4 indices during cold events when the biological activity is enhanced and the heat trapping due to surface chlorophyll is increased. While the potential feedbacks need to be investigated in coupled climate models, it is likely that the errors induced by annual mean climatological attenuation depths instead of monthly mean attenuation depths are significantly smaller than SST errors when a single non-representative constant attenuation depth is used. When a representative constant attenuation depth such as the basin mean of the variable depths is used, the errors are greatly reduced but may still cause large wind feedbacks in the eastern ITCZ region. During cold ENSO events even the equatorial region will be affected since $h_\gamma$ in the cold-tongue region is much shallower than the basin mean.

The effects of using realistic attenuation depths in the Atlantic are somewhat similar to the Pacific but the details are quite different in terms of the processes involved due to the basic differences in the circulations of the two oceans. The equatorial region is dominated in the Atlantic Ocean by the discharges from the Congo and Niger rivers. It is shown that the model performance in terms of SST simulations compared to Levitus data is indeed better when variable attenuation depths are used. Interannual variability of the Atlantic and Indian Oceans are not presented here since they are inherently less dramatic than the ENSO variability in the Pacific.

The Indian Ocean is characterized by a lack of equatorial upwelling in the mean with the largest surface chlorophyll variability in the Arabian Sea and in the southern tropical Indian Ocean. The SST differences between the control run and the simulations with variable attenuation depths are considerably smaller in the annual mean than the Pacific and Atlantic Oceans. However, much of the Indian Ocean is dominated by relatively high mean SSTs and thus even small SST errors can be climatically important. It is demonstrated that even the annual mean spatially variable attenuation depths capture the seasonality of the biological heat trapping quite accurately in the Arabian Sea, since the mixed layer depths are always much deeper than the attenuation depths. The effects of variable attenuation depths are not as large in the Bay of Bengal as in the eastern Atlantic even though the Bay also receives significant riverine inputs.
The reasons for this are being investigated further.

It should be reemphasized that the goal of this paper is to recommend the simplest possible formulation for penetrative radiation for the coupled ENSO prediction models without adding computational penalties. It is evident from this study that a single exponential formulation with annual mean attenuation depths derived from ocean color can improve SST simulations and is probably quite adequate for coupled climate models. These studies will be re-evaluated when the SeaWiFS and other ocean color data provide a sufficiently long time-series to capture the climatological and interannual variability in the attenuation depths. Effects of radiation attenuation are also being studied with this OGCM coupled to a nine-component ecosystem model, which will be reported elsewhere.

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CAPTIONS

Figure 1: Attenuation depths derived from annual mean CZCS pigment data based on Morel (1988). The attenuation is higher and hence attenuation depths are lower in biologically active regions. The oligotrophic subtropical regions have the largest attenuation depths in the tropics.

Figure 2: Differences in attenuation depths derived from CZCS data and the constant depth (17 m) assumed for the control run.

Figure 3: Seasonal cycle of the differences in SST between the control and a simulation where subsurface penetration of shortwave radiation is not allowed. The contours show differences in surface heat fluxes.

Figure 4: Annual mean differences between model and Levitus SSTs for the control run ($h_\gamma = 17$ m). Like most state of the art models forced with observed winds, the cold tongue is colder than observed. Significant warming in this region is achieved by using spatially variable attenuation depths.

Figure 5: Annual mean differences in surface currents (vectors) and temperatures (colors) for variable and constant attenuation depths (top), differences of temperatures (color) and zonal/vertical currents along the equator (center), and mixed layer depth differences for the tropical Pacific.

Figure 6: Profiles of radiation (left) and temperatures with depth at locations shown. The dashed lines are annual mean mixed layer depths for constant (blue) and variable (red) attenuation depths. $HP$ values indicate attenuation depths at those locations.

Figure 7: Annual mean net surface heat fluxes and penetrative radiation losses in the western equatorial Pacific for the control run and for the simulation with variable attenuation depths. Nearly all net downward flux is lost to the subsurface as penetrative radiation in the bottom panel which is significant for the heat balance of this important region of warm SSTs and atmospheric convection.
**Figure 8:** Annual mean differences in surface currents (vectors) and temperatures (colors) for variable and constant attenuation depths (top), differences of temperatures (color) and zonal/vertical currents along the equator (center), and mixed layer depth differences for the tropical Atlantic.

**Figure 9:** Annual mean differences in surface currents (vectors) and temperatures (colors) for variable and constant attenuation depths (top), differences of temperatures (color) and zonal/vertical currents along the equator (center), and mixed layer depth differences for the tropical Indian Ocean.

**Figure 10:** Seasonal differences in SST along 20°N in the Arabian Sea between simulations with variable and constant attenuation depths. Even though annual mean attenuation depths are used, the seasonality of the biological heat trapping can be reproduced.

**Figure 11:** Annual mean differences in surface currents (vectors) and temperatures (colors) between simulations with variable and basin mean attenuation depths for the Pacific (top), Atlantic (middle), and Indian Oceans.

**Figure 12:** Temperature differences at 129°W on the equator between the simulation with spatially variable attenuation depths and the control run (top panel) and between the simulation with spatially variable attenuation depths and the basin mean attenuation depth. The white (black) line in each panel denotes the mixed layer depth for the variable (constant) attenuation depth run.

**Figure 13:** Annual mean differences between model and Levitus SSTs for simulations with basin mean and variable attenuation depths. Model SST errors are lower in the eastern equatorial Atlantic for variable attenuation depths.

**Figure 14:** Differences in NINO3 and NINO4 SSTs and SST anomalies between simulations with variable and constant attenuation depths (C-B), and variable and basin mean attenuation depths (C-B1). Note that B (B1) has an attenuation depth shallower (deeper) than the variable attenuation depth run (C).
Pacific Ocean Attenuation Depth Differences

Atlantic Ocean Attenuation Depth Differences

Indian Ocean Attenuation Depth Differences
Averaged between 2°S and 2°N
Pacific Ocean, SP - NoSP

Month -20 -20 -10 0 10 20 30 40 50 60 70 80 90 100 110
-1 -0.8 -0.6 -0.4 -0.2 0.0 0.2 0.4 0.6 0.8 1.0

Longitude

°C
Pacific Ocean SST Differences

- Levitus

B - Levitus
Indian Ocean SST and Surface Current Differences

Indian Ocean Equatorial Differences

Indian Ocean Mixed-Layer Depth Differences