A Calculation of Ocean Heat Storage and Effective Ocean Surface Layer Depths for the Northern Hemisphere

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ABSTRACT

In the hierarchy of simple ocean formulations available for coupling to atmospheric GCMs, a scheme whereby ocean surface-layer depths vary geographically and seasonally is deemed better than a fixed depth layer at all locations and seasons, but still is less sophisticated than dynamic ocean models. Yet such simple ocean formulations are useful for basic sensitivity studies. Here, a calculation of varying surface layer depths is done by first performing an ocean heat storage calculation using gridded, long-term mean mixed-layer depths and sea surface temperatures with a parameterized temperature structure beneath the mixed layer derived from weather ship data. Heat storage values in the midlatitudes are larger in the Atlantic than in the Pacific, which is in qualitative agreement with the weather ship data. Variants of the basic calculation show that neither mixed layer data nor SST data alone are sufficient to compute heat storage adequately. Using the results from the parameterized heat storage calculations, effective ocean surface-layer depths are computed. These are found to be deeper in the Atlantic than in the Pacific, with a strong semiannual monsoon signal apparent in the Indian Ocean. Since these calculations exclude the effects of vertical and horizontal motion further analysis as to the viability of these calculations can be done with the specified depths coupled to an atmospheric GCM.

1. Introduction

A number of recent climate modeling efforts have attempted to address the question of air-sea interaction and the role of the oceans in climate and climate change by using a hierarchy of ocean formulations coupled to atmospheric general circulation models (GCMs). The types of studies which have been done using these ocean parameterizations are reviewed, for example, by Meehl (1984). At one end of the spectrum are the energy balance or "swamp" oceans which have no heat storage capability and no dynamics (e.g., Manabe and Wetherald, 1975, 1980; Washington and Meehl, 1983). These must be run with an atmospheric model forced with annual mean conditions. Next in the hierarchy are the simple slab mixed layer oceans. Typically these specify an ocean surface layer of some fixed depth which acts as a heat reservoir and moisture source but does not include ocean dynamics (e.g., Manabe and Stouffer, 1980; Spelman and Manabe, 1984; Washington and Meehl, 1984). This type of ocean parameterization allows the atmospheric GCM to be run with a full seasonal cycle. The sea surface temperatures can respond to a change in external forcing in perturbation experiments only via the surface energy balance. Finally, there are fully computed ocean models of either the ocean surface layer (e.g., Pollard et al., 1983; see also a review of upper ocean models by Kraus, 1977) or the full ocean (e.g. Manabe et al., 1979; Washington et al., 1980; Spelman and Manabe, 1984), which can be coupled to atmospheric GCMs. These formulations are the most realistic in that ocean heat storage and dynamics are included, but they are also the most complicated and expensive.

One of the reasons for having such a hierarchy of ocean formulations available for coupling to atmospheric GCMs is that by adding progressively more detail to the ocean part of the simulation a better understanding may be obtained concerning the role of certain ocean processes in the climate system. However, as with any simplified parameterization, certain anomalies in the simulation can adversely affect the results of a given coupled model simulation. For example, Meehl and Washington (1984) show a sea surface temperature (SST) difference map, computed minus observed, for their simple 50 m fixed depth mixed-layer ocean model coupled to the NCAR Community Climate Model (CCM) (Washington and Meehl, 1984). Sea surface temperature anomalies of up to 6°C are sufficiently large by observed standards to be associated with significant regional climate anomalies. Meehl and Washington (1984) speculate as to the causes of the SST anomalies in their simple
mixed-layer model, and conclude that since heat transport and other effects are excluded, a mixed layer that remains at a fixed depth year round certainly could be expected to produce significant SST anomalies. For example, considering only the depth of the mixed layer which acts as a simple heat reservoir and moisture source, if the real ocean mixed layer were very shallow the deeper fixed-depth mixed layer could be expected to be cooler, and vice versa if the real ocean mixed layer were deeper than the fixed depth mixed layer. If this were the case, it could be possible to specify mixed layer depths as a function of time and space and thereby reduce the SST anomalies associated with this effect. Clearly many other ocean processes are playing a role in causing these anomalies. However, by specifying seasonally and geographically varying mixed layer depths and thereby reducing this source of discrepancy, two purposes would be served. First, the computed SST from the simple mixed layer should be closer to observed values. Therefore, results from a $2 \times CO_2$ or other perturbation experiment do not have to be interpreted in light of large SST anomalies in the control simulation. Second, areas where ocean heat transport or other processes not necessarily dependent on mixed layer depth are important would more accurately be delineated. Of course, this type of ocean parameterization presumes that the control simulation or any perturbation in the climate only affects SST via the surface energy balance. In essence, this method aims to specify observed SST by simply stipulating mixed layer depth. In the real ocean, of course, SSTs are the end product of many interacting processes, including mixed-layer effects. Clearly, any changes in ocean dynamics resulting from a climate change are not accounted for with the simple mixed layer technique. However, much can be learned about climate sensitivity with this type of coupled model where the ocean acts only as a heat reservoir and moisture source, and results from such experiments point to parameters and processes which can be studied in more realistic detail at the next level of sophistication in the hierarchy—the fully coupled ocean–atmosphere GCM.

Having recognized the feasibility of such an approach, the question becomes one of what depths to use for the specified mixed layer. Manabe and Stouffer (1980) calculate an “effective depth” of the upper ocean which is derived from an integration of the annual range of temperature as a function of depth. They arrive at annual zonal means in this way and use the global average of those means, 68 m, for the fixed depth of their simple mixed-layer ocean at all locations and seasons. Hansen et al. (1984) derive values of mixed layer depth as a function of season and location from a temperature-dependent criterion, specify those depths for their simple mixed layer ocean, and limit the deepest depth to 65 m for coupling to an atmospheric GCM. Van den Dool and Horel (1984) compute “active layer depths” for the upper ocean in the Pacific from energy balance considerations. Their values typically range from about 25 to 50 m.

The purpose of this paper is to formulate a technique for computing effective mixed-layer depths that vary in time and space, which could be used in a specified depth, simple mixed layer ocean formulation suitable for coupling to an atmospheric GCM. This will be done by first taking a set of observed mixed layer depths and SSTs, computing seasonal ocean heat storage, and comparing those values to calculations done with sets of hydrographic data (Oort and Vonder Haar, 1976; Lamb and Bunker, 1982). The present calculation will involve parameterizing the sub-mixed layer temperature structure which will be derived from ocean weather ship data. The results from the calculation then will be used to compute the equivalent seasonally changing ocean surface-layer depth, which will be referred to as an “effective ocean surface layer depth.” For illustrative purposes these ocean depths will be computed for the Northern Hemisphere, but the technique also could be applied to the Southern Hemisphere.

In Section 2 the basic heat storage calculation will be reviewed, and Section 3 will pose various methods of computing heat storage from observed mixed-layer depths and SST data. Section 4 will show computed heat storage results for the Northern Hemisphere and point to the necessity of accounting for the temperature profile below the mixed layer. Section 5 will show heat storage results for individual ocean basins. Finally, Section 6 will show the calculation of the effective mixed layer depths based on the previous heat storage computation. Zonal means will be compiled for the entire Northern Hemisphere as well as for individual ocean basins.

2. Ocean heat storage

The surface energy balance (e.g., Sellers, 1965) is defined as follows:

$$[R - R] + [(Q + q)(1 - \alpha)] = H + LE + G,$$

(1)

where $R$ is incoming infrared radiation, $R$ is upward infrared from the surface, $Q + q$ is the sum of direct and diffuse solar radiation, $\alpha$ is albedo, $H$ is sensible heat, $LE$ is latent heat, and $G$ is heat flow through the surface.

The sea surface temperature is a function of the three terms on the right side of Eq. (1), as well as being related to the upward infrared flux as follows:

$$R = \epsilon \sigma T^4,$$

(2)

where $\epsilon$ is surface emissivity, $\sigma$ the Stefan–Boltzmann constant, and $T$ the sea surface temperature.

In the ocean, $G$ has two major components:

$$G = Q_l + Q_r,$$

(3)
where $Q_r$ is rate of change of local ocean heat content which will be termed simply “heat storage” in discussions to follow and $Q_v$ is heat transported by ocean currents and other processes.

The quantity $Q_s$, heat storage, is defined as

$$Q_s(h, t) = \frac{\partial}{\partial t} \int_0^h c_p \rho T(z) \, dz,$$

where $Q_s$ is the rate of change of heat storage in a column of depth $h$ and unit surface area, $c_p$ the specific heat of sea water, $\rho$ the water density and $T(z)$ is the temperature profile.

Typical values for the seasonal cycle of zonal mean ocean heat storage compared to atmospheric heat storage have been computed by Oort and Vonder Haar (1976) averaged over the entire latitude circles for the Northern Hemisphere and reproduced in Fig. 1. The ocean heat storage maximum of roughly 100 W m$^{-2}$ is about five times the atmospheric value. The ocean maximum and minimum occur in mid-latitudes near 35°N with storage during summer and heat loss during winter. The Oort and Vonder Haar (1976) results will be discussed further in subsequent sections.

The quantity $G$, net surface flux [Eq. (3)], has been computed as a residual from the surface energy balance [Eq. (1)] and published for various ocean regions (Bunker and Worthington, 1976; Hastenrath and Lamb, 1978; Hastenrath and Lamb, 1979). Unfortunately there is no way to distinguish the relative magnitudes of ocean heat storage $Q_s$, ocean heat transport $Q_v$, and other neglected terms from these data.

The best method of obtaining the actual ocean heat storage $Q_s$ is by compiling large amounts of hydrographic data and performing the required integrations on all the soundings [Eq. (4)]. These calculations have been performed by Lamb and Bunker (1982) for the North Atlantic, Behringer and Stommel (1981) for the tropical Atlantic (but not shown in their paper), Oort and Vonder Haar (1976) for the Northern Hemisphere, Merle (1980) for the equatorial Atlantic, Fritz (1958) for the Northern Hemisphere from limited Pacific data, and Bryan and Schroeder (1960) from limited data for the North Atlantic. As mentioned earlier, the only Northern Hemisphere calculation with actual data from all oceans is the one by Oort and Vonder Haar (1976), but no individual ocean basin results are shown. Levitus (1984) performs heat content calculations as a function of latitude and ocean basin, but gives heat storage (rate of change of heat content) for only hemispheric and global means.

All of the calculations cited above were performed with different data sets and carried out in slightly different ways. For example, Oort and Vonder Haar (1976) limited their Northern Hemisphere integrations to a depth of 275 meters, while Lamb and Bunker (1982) extended their Atlantic calculations to 500 meters. As a result, there is agreement in the general character of the annual cycle of ocean heat storage (e.g., Fig. 1b), but some discrepancies in terms of magnitude exist with disagreements of the order of 30%.

Global ocean general circulation models have computed ocean heat storage (Bryan and Lewis, 1979; Manabe et al., 1979; Meehl et al., 1982). However, inherent model constraints, such as the lack of vertical resolution, limit the accuracy of these results.

3. Ocean heat storage computed from mixed layer depth and SST data

The ocean mixed layer is generally defined as the upper layer of the ocean where the water is well-mixed and nearly isothermal conditions are maintained (e.g., Knauss, 1978). Beneath the mixed layer the temperature falls rapidly with depth in the seasonal thermocline (if it exists). Below the seasonal thermocline the water temperature decreases more gradually with depth in the permanent thermocline and varies little throughout the year. Ideally, any integration of ocean heat storage should extend below the depth of any seasonal temperature fluctuation. To illustrate the annual cycle of the upper ocean, the long-term means (16 years, 1955–70) of vertical temperature profiles for ocean weather station VICTOR in the Pacific at 34°N, 164°E are shown in Fig. 2 from Tabata (1976) (first published by Ballis, 1973b). A shallow mixed layer forms in spring and summer, and deepens during fall and winter. Ekman divergence
and upwelling, internal waves, velocity shear in the vertical, the depth of penetration of radiation, wind stirring and convection due to temperature and salinity variations are all thought to affect the depth of the mixed layer.

The availability of various data sets of monthly mean mixed-layer depths (MLDs) provides the attractive possibility of using MLD and SST to compute ocean heat storage. However, as alluded to earlier, there is disagreement over exactly how the depth of the mixed layer should be defined. Baten (1972) gives MLDs for the North Pacific and uses a temperature gradient-related criterion to define mixed-layer depth. Colborn (1975) defines mixed-layer depths for the Indian Ocean both in terms of temperature gradient in the thermocline and the depth at which the temperature is 1°C less than the SST. Lamb (1984) takes the level at which the temperature is 1°C less than the SST to be the depth of the mixed layer based on a compilation of individual soundings for the Atlantic. The Northern Hemisphere atlases of Robinson (1976) and Robinson et al. (1979) similarly define the top of the thermocline, a crude estimate of mixed-layer depth, as the depth at which the temperature is 1.1°C less than the SST. Levitus (1982) gives mixed-layer depths for both temperature- and density-related criteria. He notes that the salinity stratification differences between Pacific and Atlantic limit mixing in many regions of the Pacific to about 150 m, while some areas in the Atlantic mix to depths of 300 m or more. Since the calculation of ocean heat storage in this paper uses the monthly mean MLDs of Robinson (1976) and Robinson et al. (1979), their definition will be used to define the depth of the upper well-mixed layer. The monthly mean observed SSTs used here are from Alexander and Mobley (1976). All data are digitized on 5° latitude-by-longitude grids extending from 2.5°S to 57.5°N.

The Robinson MLD data limit the mixed-layer depths to 120 meters in the Pacific. Therefore, for consistency all MLDs have been limited to 120 m for the present calculation, but the integrated depth will vary by ocean basin as will be seen shortly.

February and August MLDs from Baten (1972) and Robinson (1976) can be compared in Fig. 3. The MLD criteria are different for the two data sets, but a rough idea can be obtained concerning the consistency of such mixed layer data and the 120 m limitation of Robinson (1976). In general, midlatitude MLDs are of the order of 10 m during August in both data sets, whereas tropical MLDs remain deeper in both seasons. During winter, Baten (1972) shows MLDs extending from 140 to 180 m in areas of the northern Pacific where the Robinson (1976) data are limited to 120 m. However, discrepancies of the order of 20% or more exist in some areas. This could be due to different averaging periods or methods of analysis in addition to the different MLD criteria.

The annual cycle of zonal mean MLDs from the digitized data of Robinson (1976) and Robinson et al. (1979) are shown in Fig. 4 for the entire Northern Hemisphere and for individual ocean basins. As can be expected from Fig. 3, relatively deep MLDs occur year round in the tropics, with very shallow summer and deep winter MLDs present in midlatitudes. A strong semiannual signal is apparent in the Indian Ocean with deep MLDs occurring in summer and winter with shallowing in the transitional seasons. This is due in part to the twice-annual monsoon wind forcing with strong summer and winter monsoon winds deepening the mixed layer in those seasons as noted, for example, by McPhaden (1982).

Calculation of ocean heat storage as defined by Eq. (4) stipulates that integration must be carried out to the depth $h$ where no seasonal temperature change takes place. If only MLD and SST are known and it is assumed the MLD is isothermal at the value of the SST, there is no implicit knowledge of the submixed-layer temperature structure. Since it is necessary to integrate to the same depth every month, assumptions must be made concerning the subsurface temperature...
when only SST and MLD are known. Figure 5 schematically illustrates a temperature–depth profile. Here $D_0$ is the depth at which little or no seasonal change takes place, $T_0$ is the temperature at that depth, $T_1$ is the temperature of the mixed layer and $D_1$ is the mixed layer depth. Total integration would sum the area to the left of the curve of the actual sounding and determine the amount of heat present in the column of unit area and depth $D_0$ for a particular month. To simulate that integration knowing only SST ($T_1$) and MLD ($D_1$), the area

$$(T_1)(D_1)(\rho c_p)$$

which is single cross-hatched in Fig. 5 must be added to the area

$$(T_0)(D_0 - D_1)(\rho c_p)$$

which is double cross-hatched in Fig. 5. If only these two areas are computed, the additional heat present in the thermocline is not taken into account. To estimate this remaining heat, the area

$$C(T_1 - T_0)(D_0 - D_1)(\rho c_p)$$

can be computed (light and dark stippling in Fig. 5 if $C$, for example, is taken to be 0.5, or just dark stippling if $C$ is 0.25) where $C$ is an empirically derived constant. Therefore, the heat content $H$ in a column of unit area for one month is given by the sum of Eqs. (5), (6) and (7) when the submixed-layer temperature profile is taken into account. It also may be possible to estimate the thermocline profile with some type of similarity curve or exponential function. For our purposes here, however, a simple linear calculation is adequate considering the uncertainties of such estimations.

The rate of change of the heat stored in the column from one month to the next is given by

$$\frac{\Delta H}{\Delta t} = \frac{H_i - H_{i-1}}{\Delta t},$$

where $H_i$ is the heat stored in month $i$, $H_{i-1}$ is the heat stored in the preceding month, and $\Delta t$ is 30 days. Note that SST and MLD values to be used in this paper are monthly means taken to be centered at mid-month. Therefore, when differences are taken from month to month, the heat storage values are computed for the first of the month, and will be plotted in that way in the figures to follow.

FIG. 3. (a) February mean depths to the top of the thermocline, defined as the depth at which the temperature is 1.1°C less than the SST. Contour interval is 50 ft (15 m). Stippled areas are depths greater than 400 ft or 120 m. Cross-hatched areas are depths less than 200 ft or 60 m (Robinson, 1976). (b) February mean mixed-layer depths defined by a threshold temperature gradient in the thermocline. Contour interval is 20 meters. Stippled areas are depths greater than 120 m. Cross-hatched areas are depths less than 60 m (Bathen, 1972). (c) As in (a) except for August. (d) As in (b) except for August.
A method of calculating the rate of change of heat storage that does not include the subsurface temperature profile is given by

$$\frac{\Delta H}{\Delta t} = \left[ (T_i - T_{i-1}) \frac{(D_i + D_{i-1})}{2} \right] \rho c_p \Delta t,$$

(9)

where $T_i$ and $D_i$ are the SST and MLD of month $i$ respectively, and $T_{i-1}$ and $D_{i-1}$ are the SST and MLD from the preceding month. The quantity $\Delta t$ is thirty days as before. This method has the inherent weaknesses of not summing over the same column depth each month and having the sign of heat storage depend on the SST and column depth change.

The observed annual cycle of mean soundings in Fig. 2 from weather ship VICTOR at 34°N, 164°E was used to calculate the annual cycle of the rate of ocean heat storage using Eq. (9) and the results are plotted in Fig. 6a. The most interesting feature of this calculation is that the heat storage changes sign from positive (heat gain) to negative (heat loss) in August–September, two months sooner than shown by the full integration to be described shortly. Since the sign of heat storage in this method is dependent on SST as noted before, the SST at this location is decreasing even though the real ocean shows positive heat storage. The process most likely responsible for this phenomenon at this location is entrainment (for example, see discussion by Elsberry and Camp, 1978). For comparison, a total integration to 150 m for ship VICTOR is plotted as the solid line in Fig. 6 and calculated as follows. Considering the salinity-based depths of Levitus (1982), mixing is assumed to be limited to near 150 m and $D_0$ is taken to be that depth. The subsurface temperature $T_0$ is the SST during the month of deepest mixed layer or lowest SST minus 1.1°C (the Robinson MLD criterion).
The heat content of a column $D_0$ for month $i$ is given by adding Eqs. (5), (6) and (7):

$$H_i = (T_i D_i) + T_0 (D_0 - D_i) + [C(T_i - T_0)(D_0 - D_i)](\rho c_p),$$

(11)

where $T_i$ and $D_i$ are defined in the same way as in Eq. (9). If the third term is defined

$$A_i = C(T_i - T_0)(D_0 - D_i),$$

(12)

then the rate of change of heat content from month

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Fig. 5. Schematic of temperature vs. depth profile showing a typical profile (heavy line) and areas used to calculate heat stored in column of depth $D_0$.

Therefore, the equation for the integrated heat storage is given by

$$\frac{dH}{dt} = \sum_{k=1}^{k_{\text{max}}} (T^k_i - T_{i-1}^k) \Delta z_k \rho c_p \Delta t,$$

(10)

where $k_{\text{max}}$ is the number of depth intervals and $\Delta z_k$ is the depth of the interval which is taken to be 10 m. Equation 10 is used to compute storage for six weather ships: PAPA (50°N, 145°W), NOVEMBER (30°N, 140°W) and VICTOR (Fig. 6) in the Pacific, and CHARLIE (52°N, 35°W), DELTA (44°N, 41°W) and ECHO (35°N, 48°W) in the Atlantic. The Pacific weather ship data were taken from published values of Ballis (1973a,b,c) reproduced by Tabata (1976). The Atlantic data are interpolated from standard levels from the National Oceanographic Data Center (NODC) Nansen cast file for one-degree squares at the weather ship locations. These are monthly means roughly for the period 1965–73. These data are checked for consistency with Levitus (1982). Assuming that little mixing takes place below the salinity stratification levels noted by Levitus (1982), the value of $k_{\text{max}}$ in Eq. (10) is taken to be 15 (a depth of 150 m) for the Pacific ships and 30 (a depth of 300 m) for the Atlantic ships. These also are standard levels at which the weather ship data are roughly 1.1°C less than the SST during the coldest month in the Nansen cast data. These integrations are plotted as the solid lines in Figs. 6 through 12.

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Fig. 6. Annual cycle of ocean heat storage (W m$^{-2}$) calculated for ocean weather ship VICTOR, 34°N, 164°E, in the Pacific: (a) for a full integration to 150 m (solid), an SST dependent calculation (dashed), and using mixed-layer depths only (dash-dot); (b) the full integration (solid line) shown in part (a) compared to the parameterized heat storage calculation (dashed).
Fig. 7. Annual cycle of ocean heat storage calculated for ocean weather ship PAPA, 50°N, 145°W, in the Pacific of the full integration to 150 m (solid) and parameterized heat storage calculation (dashed).

$i - 1$ to month $i$ is given by combining Eqs. (8) and (11) to form an equation for heat storage:

$$\frac{\Delta H}{\Delta t} = [(T_i D_i) - (T_{i-1} D_{i-1}) - T_0 (D_i - D_{i-1}) + (A_i) - (A_{i-1})] \frac{\partial C}{\partial t}. \quad (13)$$

Results from the calculation of heat storage for the ship data using Eq. (10) are then substituted on the left-hand side of Eq. (13) to solve for $A_i$. Then Eq. (12) is used to solve for $C$. This is the constant that determines the temperature profile below the mixed layer. The MLDs ($D_i$) are taken as the nearest 10 m level in the ship data where the temperature is less than the SST by 1.1°C (again, the MLD criterion of Robinson, 1976). These values of $C$ are then averaged for the months when mixed layer changes are most rapid, April and November. At those times of year the temperature structure below the thermocline has a large effect on heat storage as seen in Fig. 6a where the dash-dot curve shows computed heat storage taking only the mixed layer into account (i.e., $C = 0$). Using the average value of $C$ determined in this way, 0.18, heat storage is recalculated for the six weather ships using Eq. (13). The results for this parameterized heat storage calculation are shown in Figs. 6–12 as dashed lines. It can be seen that

Fig. 8. As in Fig. 7 except for ship NOVEMBER, 30°N, 140°W, in the Pacific.

Fig. 9. As in Fig. 7 except for ship CHARLIE, 52°N, 35°W, in the Atlantic.

Fig. 10. As in Fig. 7 except for ship DELTA, 44°N, 41°W, in the Atlantic.
generally good agreement exists for both magnitude and phase at most times of year to within about 20% which, as was indicated in Fig. 3, is probably within the levels of accuracy of the observations. The agreement between the parameterized heat storage calculation is better for ships PAPA and NOVEMBER in the Pacific (Figs. 7 and 8), and CHARLIE and ECHO in the Atlantic (Figs. 9 and 11). There is less correspondence at VICTOR in the Pacific and DELTA in the Atlantic. These two ships are located near the Kuroshio and Gulf Stream currents respectively. Therefore, it appears that this parameterization works better for central ocean basin conditions.

4. Northern Hemisphere results

Annual cycles of zonal means of heat storage from 2.5°S to 57.5°N computed for the mixed layer only and for the parameterization developed in the previous section are shown in Fig. 12 along with Oort and Vonder Haar's (1976) values (Fig. 1) redrawn to correspond to zonal means taken over oceans only. Negative values imply heat loss, positive values mean heat gain. In Fig. 12a, where heat present in the thermocline is not taken into account, winter heat losses are a little more than half of the Oort and Vonder Haar (1976) results in Fig. 12c, and the summertime heat gain is weak, with even some slight heat loss during early summer near 45°–50°N. This is similar to the results in Fig. 6a for data from ship VICTOR. The smaller values are to be expected especially during the summer since changes in the heat content below the mixed layer are not taken into account when the mixed layer is very shallow.

Comparing the parameterized heat storage calculation in Fig. 12b to Oort and Vonder Haar's (1976) values in Fig. 12c, the following features appear in

FIG. 12. Annual cycle of zonal means of ocean heat storage (W m\(^{-2}\)) using (a) only mixed-layer depth, (b) parameterized heat storage and (c) values computed by integrating individual soundings to 275 m from Oort and Vonder Haar (1976), converted from their original values to apply to averages taken over ocean areas only.
common: 1) a large midlatitude summertime heat gain and wintertime heat loss; 2) a delay of the change in sign from positive to negative heat storage until the end of September probably associated with entrainment; 3) a tropical heat gain and loss leading the midlatitude maxima by about three to four months; and 4) evidence of a semiannual signal near the equator with heat gain in early spring and fall and heat loss during winter and summer. Maximum and minimum midlatitude values in Fig. 12b are of the order of 15% larger than the Oort and Vonder Haar values, which is again likely to be within the accuracy of the observations. The greatest discrepancy occurs in August–September near 10°–15°N where Oort and Vonder Haar (1976) show large negative values of heat storage. The parameterized calculation does not show a comparable feature, but does show values near zero at that latitude and time of year. This is a region of large heat convergence documented by Oort and Vonder Haar (1976). The parameterized heat storage technique was seen earlier to perform more poorly at the two weather ships located in regions of large ocean heat advection. If heat convergence or divergence is important at that location and time of year, it is likely that the parameterized heat storage method will not be able to reproduce that feature in Fig. 12c. Also the tropics have been shown to be regions where fluctuations of mixed-layer depths are forced by ocean dynamics, and entrainment is not a major factor (e.g. Merle, 1980; McPhaden, 1982). Since the present value of $C$ is tuned to midlatitude midocean temperature profiles where entrainment is thought to be important, it can be expected to be less realistic in the deep tropics.

5. Individual ocean basin results

Figure 13a shows the annual cycle of monthly zonal mean heat storage for the Atlantic Ocean computed from the parameterized method compared to bimonthly values derived by Lamb and Bunker (1982) for the Atlantic (Fig. 13b). There is basic agreement concerning patterns and magnitudes of heat storage in midlatitudes with maxima and minima in summer and winter of roughly +200 and −200 W m$^{-2}$ respectively. However, there is less agreement in the subtropics and tropics, particularly in spring and fall. Detailed heat storage calculations by Merle (1980) for the equatorial Atlantic show how varying conditions between the east and west sides of the basin produce significantly different heat storage results. In his Fig. 9b for the area 0–6°N, there is positive heat storage in February–March and September–November, and negative values for April–August and December–January. This closely resembles the computed pattern in Fig. 13a at those latitudes. However, Merle (1980) shows a much different pattern in the western Atlantic with positive heat storage at 0–6°N during June–November and negative for December–May. He notes that the deeper and more seasonally variable mixed-layer depths in the western Atlantic compared to the eastern Atlantic are associated with seasonal uplifting of the thermocline due to the dynamical response of the ocean to seasonally variable windforcing, and the effects of entrainment are of secondary importance. As noted earlier, the present value of $C$, which determines the temperature profile beneath the thermocline, is derived from midlatitude weather ships where entrainment mixing is important. Since the seasonal vertical movement of the thermocline is less in the eastern than western Atlantic, it can be
expected that errors in parameterizing tropical thermocline structure would be minimized there, and thus there is better agreement between the present heat storage calculation for the Atlantic near 0–6°N and Merle's (1980) values for the eastern Atlantic at those same latitudes.

Figure 14a shows the annual cycle of zonal mean heat storage computed for the Pacific, and 14b is for the Indian Ocean. As seen earlier for the weather ship data, the Pacific values of heat storage are of less magnitude than for the Atlantic as shown in Fig. 13, but the midlatitude pattern is comparable. The Indian Ocean (Fig. 14b) exhibits a large semianual signal with heat loss during summer and winter and heat gain during spring and fall associated with the twice-

annual monsoon regime in the atmosphere (Ramage, 1971). This pattern of heat storage is in qualitative agreement with values computed for Gan Island by McPhaden (1982).

If it is assumed that there is no net gain or loss of heat over the year in a long-term mean, the sum of the rates of heat storage at each latitude should be zero for the year. Table 1 shows the net residual for the parameterized heat storage calculation for each ocean basin. In spite of the simplicity of the calculation, many of the residual values for all three basins in Table 1 are near zero. In the Atlantic midlatitudes the negative residuals indicate a net loss of heat suggesting that various processes mentioned earlier are not accounted for properly.

6. Effective mixed-layer depths

It was seen in the previous section that by taking only the near-isothermal mixed layer depth into account, heat storage is not satisfactorily approximated. Therefore, a specified effective mixed-layer depth should take into account a layer of water somewhat deeper than the upper surface layer in order to specify SST's more accurately on the basis of adequately simulated heat storage. To do this, a seasonally varying, effective mixed layer depth \( h_e \) can be computed that is a reflection of the month-to-month variations of heat content in the seasonally changing part of the temperature profile. The expression for the heat content in a column \( D_0 \) for month \( i \), Eq. (11) can be rewritten:

\[
H_i = (T_0D_0 + (T_i - T_0)D_i) + \left( C(D_0 - D_i)(T_i - T_0) \right) \rho c_p.
\]

If \( T_0 \leq T_i \) year round, \( T_0D_0 \) is the same from month to month, and the seasonally varying component of the heat content can be written

\[
H_i = (T_i - T_0)D_i + \left( C(D_0 - D_i)(T_i - T_0) \right) \rho c_p.
\]

| Table 1. Residuals of annual heat storage for individual ocean basins (bars indicate incomplete data). |
|-------------------------------------------------|------------------|------------------|
| Latitudinal Band | Atlantic | Pacific | Indian |
| 60°N | - | - | - |
| 55° | - | - | - |
| 50° | - | 0.00 | - |
| 45° | -7.28 | -0.03 | - |
| 40° | -3.50 | -0.33 | - |
| 35° | -2.87 | 0.00 | - |
| 30° | -0.01 | -0.85 | - |
| 25° | 0.00 | -0.44 | - |
| 20° | 0.00 | 0.00 | 0.00 |
| 15° | 0.00 | 0.00 | 0.00 |
| 10° | 0.00 | -0.13 | 0.00 |
| 5° | -0.36 | 0.00 | 0.00 |
| 0° | 1.09 | 0.07 | 0.20 |

Fig. 14. Annual cycle of zonal means of parameterized ocean heat storage (W m\(^{-2}\)) for (a) Pacific and (b) Indian Ocean.
An alternative expression for the seasonally varying heat content is

\[ H_i^t = [(T_i - T_0)(D_0)]\rho c_p, \]

(16)

where \( D_e \) is an effective ocean surface-layer depth that accounts for the upper mixed layer and the effective subsurface depth to which heat storage is important. Substituting Eq. (16) into Eq. (15) and canceling terms, one obtains

\[ D_e = D_i + C(D_0 - D_i). \]

(17)

This equation takes into account the mixed layer depth \( D_i \) as well as the parameterized temperature structure beneath the mixed layer that depends on the constant \( C \). Recall that \( C \) was determined from heat storage calculations with the weather ship data.

Zonal mean effective mixed layer depths computed from Eq. (17) are shown in Fig. 15. Northern Hemisphere values (Fig. 15a) vary from less than 40 m in midlatitudes in summer to around 130 m in winter. The tropics remain near 80 m year round. For the individual basins, Atlantic midlatitude depths vary

![Diagram](image_url)

**Fig. 15.** Annual cycle of zonal means of effective ocean surface-layer depths (m) for (a) Northern Hemisphere, (b) Atlantic, (c) Pacific and (d) Indian Ocean.
from a minimum of 70 m in summer to 140 m in winter, with tropical values remaining near 90-100 m year round. In the Pacific, the effective depths are shallower than in the Atlantic and range from less than 40 m in summer to about 120 m in winter in midlatitudes, and remain at roughly 80 m year round in the tropics. This corresponds to the shallower MLDs seen in Fig. 4 for the Pacific, and heat storage values of smaller magnitude for the Pacific weather ships as well as for the zonal mean Pacific heat storage values. The semiannual monsoon forcing is evident in the Indian Ocean (Fig. 15d) with effective depths of greater than 80 m in summer and winter monsoon seasons, and depths of roughly 60 m during the transition seasons.

In Fig. 16 the annual mean effective ocean surface-layer depths have been compiled from the zonal means in Fig. 15 for the Northern Hemisphere and the individual ocean basins. This shows more clearly the differences between the ocean basins concerning the effective depth to which ocean heat storage is involved. The Atlantic has the largest annual mean values ranging from 80 to 100 m in the tropics to over 110 m at higher latitudes, followed by the Pacific with annual mean values near 75-85 m at all latitudes, and the Indian Ocean with depths of around 70 m. These depths are deeper than those calculated by Manabe and Stouffer (1980) and van den Dool and Horel (1984). The former used a purely temperature-dependent method which accounted for annual fluctuations of temperature with depth, and the latter computed an “active layer” determined by surface energy balance. Here, heat storage is first calculated and then the depths to which the fluctuations in heat storage are important are computed. The fact that effective ocean surface-layer depths computed in this manner are deeper than the Manabe and Stouffer (1980) and van den Dool and Horel (1984) results suggests that even small temperature fluctuations at lower depths are important in terms of heat storage and SST.

It also is useful to note that Meehl and Washington (1984) concluded that the fixed-depth 50 m layer of Washington and Meehl (1984) probably was either too shallow or too deep at various locations at different times of year. The surface-layer depths calculated here vary from less than 40 m to more than twice the 50 m fixed depth of Washington and Meehl (1984). This suggests that at least for their model these effective depths could improve the simulated SST patterns when coupled to their atmospheric GCM.

7. Conclusions

The problem is posed of how to prescribe effective ocean surface-layer depths, which vary by location and season, for coupling to an atmospheric GCM. This type of formulation is likely to be better than one which fixes the same mixed layer depth for all locations and seasons. It is noted that prescribing depths for such a simple ocean parameterization still has substantial flaws compared to a fully computed ocean GCM, but is consistent with the use of hierarchies of simple oceans coupled to GCMs for sensitivity studies. Such a formulation reduces errors in the basic SST state, but clearly does not come as close to simulating the real ocean as an oceanic GCM.

The method chosen for computing effective ocean surface-layer depths involves first calculating ocean heat storage using readily available gridded data sets of mixed layer depth (MLD) and sea surface temperature (SST). Using weather ship data, a parameterization for the temperature profile beneath the mixed layer is formulated in terms of the seasonal cycle of heat storage, and then this method is generalized for use with the Northern Hemisphere MLDs and SSTs. Comparing the annual cycle of zonal mean heat storage computed in this manner with values calculated from hydrographic data by Oort and Vonder Haar (1976) and others, it is noted that:

1) Use of mixed-layer depth and SST alone is not sufficient to simulate the annual cycle of zonal mean heat storage, especially during summer when the thermocline changes rapidly.

2) SST evolution is a poor indicator of ocean heat storage, especially in the fall season, partially because of entrainment.

3) The parameterized heat storage calculation is comparable in magnitude and phase with the Oort and Vonder Haar (1976) results at midlatitudes. However, there is less agreement in the tropics. This
is due possibly to the weaknesses of the parameterized heat storage method in properly simulating heat storage in areas where active heat transport or minimal entrainment is taking place.

4) The amplitude of Atlantic heat storage at mid-latitudes is larger than values for the Pacific. This is consistent with the results from the weather ship data and is due to the deeper mixing in the Atlantic indicated by the salinity stratification differences between the two basins noted by Levitus (1982) and others. Atlantic values are comparable to those from at least one other source computed from hydrographic data. The Indian Ocean heat storage reflects the semianual monsoon-forcing signal.

5) Using a formula for the seasonally changing heat content derived from the heat storage calculation, values of effective ocean surface-layer depths are computed. As expected from the heat storage results, deepest values are seen for the Atlantic, shallowest for the Indian Ocean.

Such seasonally and geographically dependent values for an effective ocean surface-layer depth can be used coupled to an atmospheric GCM. One clear weakness with depths computed in this way is that they exclude the effects of vertical motion, advection, and other processes that are not present in a specified-depth ocean formulation. This type of error could be evaluated further by coupling the specified depths to an atmospheric GCM and analyzing the resulting SSTs.

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