An Attempt to Estimate the Thermal Resistance of the Upper Ocean to Climatic Change

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(Manuscript received 1 July 1983, in final form 20 February 1984)

ABSTRACT

An attempt is made to estimate the thermal inertia of the upper ocean, relevant to climatic change. This is done by assuming that the annual variation in sea surface temperature (SST) can, to a first-order approximation, be described by a simple energy-balance equation. From the observed climatological annual variation in SST and in absorbed solar radiation we can estimate then a typical value of the heat capacity (C) of the active layer of the ocean. Also we can estimate how fast the SST is damped towards an equilibrium value (damping coefficient b). Within the same theoretical framework the decay time of SST anomalies allows us to estimate the seasonality of C/b.

The method is first tested on SST at six ocean weather ships and two coastal stations. The calculated depth of the active layer looks reasonable though somewhat small and it is encouraging that the seasonality in C/b, derived from daily SST data at one station, is similar to the observed seasonality in mixed layer depth. One of the problems seems to be that we need rather precise observations concerning solar radiation reaching the earth's surface. At many places such knowledge is not available. The spatial distribution of calculated active layer depth over the North Pacific is very similar to that of observed annual mean mixed layer depth but the mixed layer seems to be twice as deep as the active layer. Also the effective mixed layer defined and used by Manabe and Stouffer is substantially deeper than our calculated active layer.

The results are discussed in the context of both the surface energy balance and the vertically averaged energy balance. One of the interesting findings of this study is that the layer of the ocean involved in the annual cycle should be taken as 20–50 m rather than the more customary 60–80 m. Another conclusion is that the SST seems to damp (towards equilibrium) at least five times faster than the vertically integrated energy content of the climate system as a whole (including the ocean!).

1. Introduction

In the past decade many studies have appeared in which thermodynamical models were employed to study the earth's climate and its temperature in particular. The simplest model of all is a globally and vertically averaged energy-balance model of the type

$$C\frac{dT}{dt} = Q(1 - \alpha) - (a + bT), \qquad (1)$$

where T is a temperature typically of the surface but CT is representative of the energy in a column extending from the top of the atmosphere down to the depth in the ocean where energy exchange is small, $Q(1 - \alpha)$ is the absorbed solar radiation, α is the albedo, $a + bT$ is a parameterized form of the outgoing infrared radiation at the top of the atmosphere (the only damping) and C the thermal inertia of the system.

If Q is taken to be a constant, say the annually averaged incoming radiation, then a stationary solution to (1) can be obtained. For properly chosen values of α, $\partial\alpha/\partial T$, a and b, this value of T (T_{stat}) can be brought as close as we want to the observed annually and globally averaged surface air temperature (Schneider and Dickinson, 1974; North *et al.*, 1981). Note that T_{stat} does not depend on C and that the static perturbation of climate due to a CO_2-increase or a change in Q depends only on $\partial\alpha/\partial T$ and b. In view of this it is not surprising that the latter two quantitites have been studied in considerable detail, starting from Budyko (1969) and Sellers (1969) up to Simmonds and Chidzey (1982).

The value of C becomes vital if time dependent changes of climate are considered. Fortunately nature provides us daily and yearly with quite impressive climatic excursions. The annual (and to a lesser extent the daily) march of atmospheric and oceanic variables are in fact by far the best documented evidence of climatic change on earth, as it involves a considerable part of the upper ocean and the continental surfaces. The amplitude of the annual temperature wave and its lag with respect to solar forcing inform us indirectly about the thermal inertia C and the damping coefficient b as well. To some extent this idea was used already in an analysis by McEwen (1938).

For most time scales it seems that the largest con-

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tribution to the thermal inertia of the climate system resides in the ocean. The problem therefore is how to determine what part of the ocean is involved. In their simulation of the annual cycle Sellers (1973) and North and Coakley (1979) took a seasonally independent C which was derived from estimates of the mixed layer depth (h_m). The numerical (h_m) varied from 40 m at the pole to 90 m at the equator in Sellers, while North and Coakley used 75 m, uniform with latitude. Thompson and Schneider (1979) allowed for variation of the mixed layer depth both with latitude and season. Their annual mean value of h_m varied from about 10 m at the pole to 40 m at the equator with a second maximum of almost 100 m at 55°S. Their annual variation was restricted to the midlatitude and had an amplitude of about one-third of the annual mean value.

One reason for these rather different choices of h_m may be that the concept of a mixed layer is somewhat ambiguous. Different criteria can be used to derive h_m from vertical profiles of temperatures and/or density (Bathen, 1972; Levitus, 1982). Manabe and Stouffer (1980) constructed an effective mixed layer depth (h_{em}) appropriate to the annual cycle problem in a coupled atmosphere ocean model. In this model the ocean is represented by the energy balance of a slab of water of depth h_{em}. They defined h_{em} to be a measure of the depth of the upper layer effectively involved in the annual cycle in observed sea water temperature. More precisely

$$h_{em} \equiv \int_0^\infty B(z)dz/B(0),$$

where $B(z)$ is the amplitude of the annual cycle in sea water temperature at depth z. The global mean value of h_{em}, 68 m, was subsequently used in their General Circulation Model. The definition of h_{em} does not take into account the change of phase of the annual cycle with z and therefore the modeled temperature, using h_{em}, represents water somewhat below the surface rather than SST.

By their very nature, the incorporation of oceanic inertia in vertically averaged models is far from straightforward. Here we will focus the discussion on the incorporation of oceanic inertia in a coupled atmosphere–ocean model and more in particular on those models where the ocean is represented by a supposedly well mixed "slab." Our proposed solution is based on the following considerations. 1) The atmosphere feels only sea surface temperature (SST) and therefore the equation describing the slab has to represent SST. 2) Because there is considerable mixing, observed SST presumably represents a layer of some depth. 3) But since the mixing is not perfect the water temperature averaged over that layer does not equal SST. These three considerations will naturally lead to a definition of the active layer (h); observed SST behaves as if it were taken from a well-mixed slab of h meters.

To that end we first simplify the energy-balance equation of the upper ocean to an equation similar to (1). We then invert the problem and calculate the constants C and b needed to explain the observed annual cycle in SST forced by the sun; C can be converted into the depth of a slab of seawater (h). To test the usefulness of this method we first provide a few examples based on the data at Ocean Weather Stations A, B, D, J, N and P, at the light-vessel *Texel* (off the Dutch coast) and at the Scripps Pier in La Jolla, California. The method will be outlined in Section 2. In Section 2 we will also discuss a method to extract the seasonality in C/b from SST data. Data and results will be shown in Section 3; the results include calculated active layer depths over the entire North Pacific. More detailed results on the seasonality of the active layer depth at the Scripps Pier will be discussed in Section 4. The implications for both surface and vertically averaged energy balance equations are discussed in Section 5.

2. An energy-balance equation for the upper ocean

For the upper layer of the ocean we assume that the energy balance at a given location can be written, to a first-order approximation, as

$$C\frac{\partial T}{\partial t} = F - (a + bT), \tag{2}$$

where T is, from now on, the ocean's surface temperature, F the solar radiation absorbed by the ocean and $(a + bT)$ a parameterized form of all damping processes. In (2) C is the system's heat capacity given by

$$C = \int_0^h \rho c_p dz = \rho c_p h \approx 4.2 \times 10^6 h \quad [\mathrm{J\ m^{-2}\ K^{-1}}], \tag{2a}$$

where h is the depth of the active layer, ρ is water density and c_p is specific heat. Several remarks have to be made. 1) In a loose sense, h should be the mixed layer depth. Mixed layer and active layer coincide only when (a) the mixing is perfect down to depth h, (b) there is no contact with water below h and (c) h is invariant with season. These conditions are never fulfilled although it becomes close in shallow seas. Note again that T is SST and not the temperature vertically averaged over h. 2) Eq. (2) is constructed such that the annual cycle is forced by solar radiation; anything else damps. The damping $(a + bT)$ includes upwelling, horizontal mixing by eddies of all sizes and the exchange of heat at the surface by (net) infrared radiation and fluxes of sensible and latent heat. Although this assumption is crude it is likely that for the annual cycle experiment these processes are, in an overall sense, of a damping nature. After all, in the absence of solar forcing none of these processes would force an annual cycle in SST. 3) Eqs. (1) and (2) look very similar but

(1) is vertically integrated and (2) is only for the ocean's active layer. As a result C and especially a and b assume different numerical values in (1) and (2). 4) Obviously (2) is only an approximation to reality. The most prominent element missing in (2) is an explicit reference to an atmospheric temperature (T_a) that varies along with T over the annual cycle. In the Appendix we discuss an atmosphere–ocean system in which T_a is a variable as well. 5) Finally, note how (2) differs from the more common local heat budget equation

$$C\frac{\partial T}{\partial t} = Q, \tag{2b}$$

where Q is the heat flux into the ocean composed of shortwave and net longwave radiation and sensible and latent heat fluxes. Gill and Niiler (1972) and Leetmaa (1983) have shown that one can simulate the annual cycle in T quite successfully with (2b), at least away from boundary currents and upwelling areas. However, the calculation of Q in (2b) involves knowledge of T (that we want to solve for) so (2b) is a diagnosis of data and empirical expressions (to estimate Q) rather than a model to simulate T in response to the ultimate (solar) forcing. The latter is the aim of (2).

When we assume for a while that C and b are independent of season, (2) can be written as

$$C\frac{\partial(T-T_0)}{\partial t} = (F-F_0) - b(T-T_0), \tag{3}$$

where T_0 and F_0 are yearly averaged quantities related by

$$F_0 = a + bT_0.$$

The temperature response to the first harmonic in $F - F_0$, say $A \sin\omega t$, with ω being the annual frequency, is given by

$$T - T_0 = B\sin(\omega t - \Delta) + C_1 \exp(-bt/C), \tag{4}$$

where

$$\tan\Delta = C\omega/b, \tag{4a}$$

$$B = A[b^2 + (\omega C)^2]^{-1/2} = A\cos(\Delta)/b, \tag{4b}$$

and C_1 is determined by T ($t = 0$). For long enough times the second term in (4) vanishes, so the solution to (3) is a temperature wave lagging the solar forcing by a delay Δ ($0 \leqslant \Delta < \pi/2$). The features of the response are transparent: The delay Δ is large for large C and small b while the amplitude B is large for small C and b.

Our strategy is as follows. From climatological observations at a given location we determine B and Δ. We also need to know the amplitude of the annual cycle in solar forcing (A). Where direct pyrheliometer measurements are absent we will estimate A from insolation at the edge of the atmosphere and reasonable assumptions about the attenuation of the solar beam in the atmosphere. From A, Δ and B we calculate b, using (4b) and h, using (4a) and (2a). We essentially propose here to invert the energy balance problem which traditionally starts from choosing reasonable values of b, A and h and then solves for B and Δ.

This method can yield a reasonable result only when the annual cycle in the solar forcing is large, that is in middle and high latitudes. Within the tropical belt A is fairly small and variation of SST at the annual frequency (if there is any) must have been caused by the dynamics of the ocean–atmosphere system. A good example of this seems to be the eastern equatorial Pacific (Horel, 1982; Leetmaa, 1983). Here our Eq. (3) fails badly because we have assumed that the sun forces and anything else damps. So, at best we can expect reasonable values of C and b poleward of about 25°. To the north of 25°N the annual cycle in SST represents the overwhelming part of the variance in climatological mean SST (Chiu and Newell, 1983). Problems in the midlatitudes arise only 1) at some coastal areas where continental air flowing over the ocean contributes substantially to the forcing of an annual cycle in T and 2) in all areas where Δ is 80 days or more; here small errors in Δ lead to large errors in h and especially b.

Note that h and b derived via (4a) and (4b) are uniform with season, or more precisely, we calculate a number representative for the whole year. It is reasonable to expect that h does vary with season since mixed layers are deeper in winter than in summer. It is more difficult to anticipate the seasonality of b.

There is at least one way to extract some of the seasonality in C and b from observations of the SST. The second term in (4) indicates that the initial anomaly decays with an e-folding time scale C/b. If we extend (2) with a stochastic forcing term (say wind blowing across the ocean), anomalies in SST will have a chracteristic time-scale described by the autocorrelation function

$$\rho(\tau) = \exp(-b\tau/C). \tag{5}$$

For the Scripps Pier in La Jolla we have daily SST data at our disposal, so it is possible to evaluate $\rho(\tau)$ from data and to determine, as a function of time of the year, at what lag $\rho(\tau)$ has decayed to e^{-1}; this yields sesonality in C/b. In this context it is worth mentioning that (3) is essentially Frankignoul and Hasselmann's (1977) equation where the linear feedback coefficient is given (here) by C/b and F includes the random forcing necessary to create SST anomalies.

3. Data and results

a. Pilot study

We use in this pilot study data at eight stations. In Table 1 we list the locations, the size of the records and the nature of the data used (daily, monthly). The SST data of the six ocean weather ships are those dis-

TABLE 1. The stations used, their position, the period of their record and the type of data used.

N	Name	Location	Period	Type data
1	Weather ship A	62°N, 33°W	1948–71	Monthly mean climatological SST
2	B	57°N, 51°W	1948–72	Monthly mean climatological SST
3	D	44°N, 41°W	1950–72	Monthly mean climatological SST
4	J	53°N, 20°W	1951–71	Monthly mean climatological SST
5	N	30°N, 140°W	1948–72	Monthly mean climatological SST
6	P	50°N, 145°W	1950–72	Monthly mean climatological SST
7	Lightvessel *Texel*	53°N, 5°E	1890–75	Monthly mean climatological SST
8	Scripps Pier, La Jolla	33°N, 118°W	1920–82	Daily SST data

cussed by Esbensen and Reynolds (1981). For the last two stations in Table 1, near-coast stations, we have data concerning the solar radiation reaching the surface not too far away from the sites where SST is measured. The Scripps solar data are taken from Fig. 2 in Tont (1981). For these two stations we assumed a surface albedo of 10% to arrive at absorbed solar radiation. For the weather ships we took the solar radiation at the outer edge of the atmosphere (Sellers, 1965) and assumed that 48% (global figure) is absorbed by the surface at the positions of weather ships.

Table 2 gives the amplitude of the annual variation in SST at each of the stations, the delay of the maximum and minimum SST with respect to the 21st of June and December, the amplitude of the annual variation in absorbed solar radiation and the resulting estimates of b and h. The delays are determined by adapting a parabola to the monthly mean temperatures of the three warmest and coldest months. The average of the winter and summer delay is used in calculating b and h. A check on the Scripps Pier data indicates that delays determined from monthly means are equal, within a few days, to delays determined from 10-day running mean climatology.

The results, calculated damping coefficients (b) and active layer depths (h), are interesting but not uniformly encouraging. To mention a few positive aspects: 1) the method yields a reasonable order of magnitude for h, 20–40 m. 2) the method senses that the mixed layer in the middle of the ocean is deeper than in shallow coastal waters, although the difference is less than we had expected. 3) the North Sea case verifies quite favorably against the depth of the sea at the position of Light-vessel *Texel*. The sea is about 25 meters deep rather uniformly to the west, south and north, but to the east and bottom slopes upward towards the beach. So 16 meters seems satisfactory. One reason that this case verifies so well is that h has to be nearly constant with season in a shallow sea. Seasonality in C makes the analysis of (2) much more complex.

Other aspects are rather disturbing, especially the fact that it is next to impossible to determine h and b at the weather ships D, N and P. For large delays any small error in the delay will lead to huge errors in $\tan\Delta$ and therefore h and b. So the estimates for these three stations are probably useless. Simple theory prohibits delays to be larger than 90° and the fact that the observed delays are close to or even larger than a season indicates that either the situation is more complex than described by (2), or the data input is inaccurate.

Of particular interest is the huge damping coefficient b suggested by the data of most of the stations. In vertically averaged energy-balance climate models (including an atmosphere) b is usually taken to be 1.8 $\pm$ 0.4 W m^{-2} K^{-1}, whereas here we find as a typical value in (2) 10 W m^{-2} K^{-1}, which is probably an underestimate (see Appendix) of the true value of b. The ocean as a subsystem seems to be much more stable, equating large damping to stability here for convenience. It is hard to say whether $6 < b < 14$ is reasonable for midlatitudes because there is little to compare with. After considerable engineering on the surface energy-balance Haney (1971) arrives at an expression for the net downward energy flux through the atmosphere ocean interface:

$$Q = b(T_a^* - \text{SST}),$$

where T_a^* is an atmospheric equilibrium temperature. For the purpose of a comparison, it is unfortunate that the solar radiation is included in Q. For zonally averaged climatological conditions, b was determined to be about 33 W m^{-2} K^{-1} with only weak latitude de-

TABLE 2. Amplitude of the annual variation in monthly mean SST, the delay of the extremes with respect to 21 Dec (w) and 21 Jun (s), estimated (N = 1, 6) and measured (N = 7, 8) amplitude of the annual variation in absorbed solar radiation, the damping coefficient b and the mixed layer depth h.

N	SST (K)	Delay w (days)	Delay s (days)	Solar (W m^{-2})	b (W m^{-2} K^{-1})	h (m)
1	3.6	76	55	105	13.1	36
2	5.2	73	64	99	7.9	24
3	5.5	98	63	81	2.8	19
4	2.9	73	58	97	14.3	39
5	2.5	85	88	64	2.0	31
6	4.3	84	72	90	5.1	27
7	6.3	65	62	90	6.6	16
8	3.3	45	57	61	11.8	18

pendence. At best our results are not inconsistent with those of Haney.

Although 18 meters, the calculated active layer depth at the Scripps Pier, is more than the actual depth of the water there (local depth: 5 meters) it still may be, in some sense, a physically correct value. Due to the absence of a continental shelf near California, water representative of a much deeper sea continually mixed with coastal water.

b. *The Pacific Ocean*

We will now apply the method to the entire Pacific basin north of 25°S, with the exception of the equatorial strip (10°S $\leqslant \phi \leqslant$ 10°N) where the annual cycle in T is small and, if there is any, it is not forced locally by solar heating. This allows us to make a comparison with earlier estimates of the mixed layer depth by Bathen (1972), Levitus (1982) and Manabe and Stouffer (1980).

The climatological monthly mean values of SST on a 5° × 5° latitude–longitude grid were derived from two separate data sets. Between 25°S and 25°N the means were computed using data from 1946 to 1976. This dataset is described by Horel (1982). North of 20°N means were derived from data between 1946 and 1966. They have been used in many studies of the North Pacific (e.g., Douglas *et al.,* 1982). Since the data sets overlap along 20°N and 25°N, the means from the separate analysis at these latitudes were averaged together. Harmonic analysis of the SST data yields values of B and Δ at each grid point.

The solar data are estimates given by

$$F = Q^*(1 - \alpha)(1 - a),$$

where $Q^*(\phi, t)$ is the solar radiation at the edge of the atmosphere (Sellers, 1965), α is planetary albedo and a is absorption in the atmosphere; α and a are assumed to be spatially uniform with magnitudes of 0.25 and 0.2 respectively. Harmonic analysis of $F(\phi, t)$ yields a value of A at each grid point.

Figures 1 and 2 show amplitude (B) and phase (Δ) of the annual cycle in observed SST. In Fig. 1 one can observe how in both hemispheres B increases polewards. Two large zonal asymmetries in the spatial distribution of B are the maximum of 10°S off the coast of South America and the maximum east of Japan at 40°N. The phase lag behind the sun (Fig. 2) is typically 70 days. In the western subtropical seas of both hemispheres the delays are very short (30–40 days); in contrast Δ is large in the eastern subtropical Pacific. The latter is probably associated with persistent stratus during early summer in those areas. As noted before for $\Delta > 80$ days one cannot expect useful estimates of h and particularly b.

Figures 3 and 4 show the damping coefficient (b) and active layer depth (h) that result pointwise from observed values of A, B and Δ. Just as in the pilot study, a typical value of b is 10 W m^{-2} K^{-1}. The damping is largest (smallest) where Δ is smallest (largest), i.e., in the western (eastern) subtropical oceans. It is likely that evaporation in the western subtropical oceans contributes to the large b, thereby causing near-continental values of Δ. The active layer depth (Fig. 4) ranges typically from 20–40 m. Because the active layer should be related to the mixed layer it is useful to compare Fig. 4 with Bathen's (1972) estimates of the annual mean mixed layer depth (h_m) over the North Pacific which is redrawn in our Fig. 5. The spatial distributions of h and of h_m are quite similar. Both maps are characterized by three belts of high (ϕ > 40°N), low (20° < ϕ < 40°N) and high values (10° < ϕ < 20°N), each of the belts being inclined from WSW to ENE. There is, however, a rather large discrepancy in the depth of our active layer and Bathen's

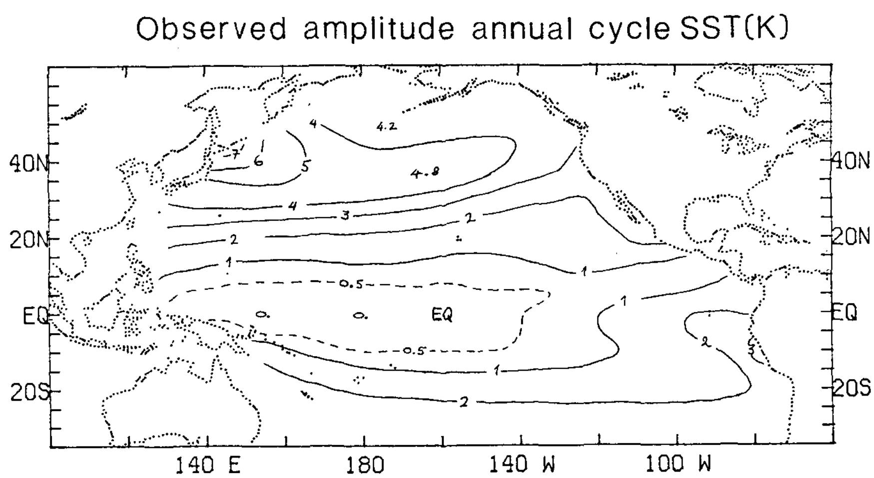

FIG. 1. The amplitude of the annual cycle (adapted to 12 long-term monthly means) in SST over the Pacific, in K.

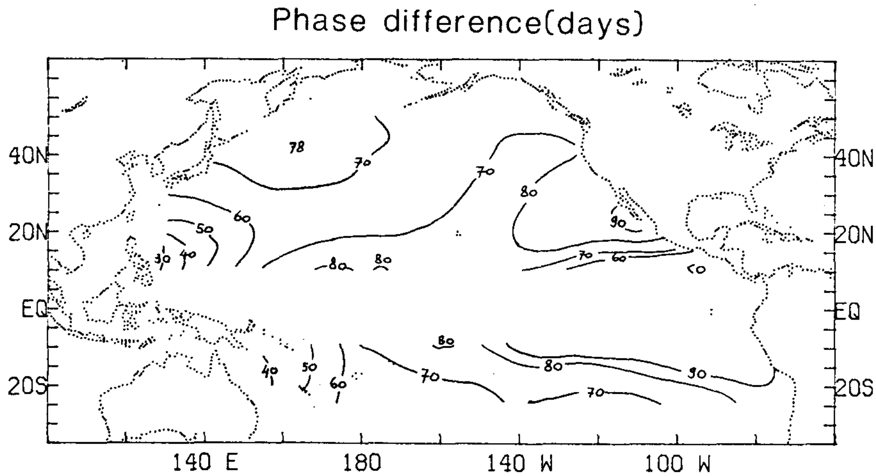

FIG. 2. The phase of the annual cycle in SST expressed in days behind June 21. The strip 10°S–10°N has been omitted.

mixed layer. There are several possible reasons for the difference between h and h_m. (1) As argued in the Appendix, our estimate of h is likely to be an underestimate, by 10% or so. (2) Bathen's values are very large; at the sites denoted by single and double asterisks Levitus (1982) derives much smaller values of annual mean h_m. (The difference between Bathen and Levitus is entirely due to the ambiguity in defining h_m.)

Another comparison can be made. Fig. 6 shows the active layer depth calculated from zonally averaged SST. Although our calculations are only for the Pacific we compare them here with Manabe and Stouffer's effective mixed layer depth (h_{em}) which is based on water temperature zonally averaged over all oceans. In order to estimate h and b between 10°S and 10°N we have used the semi-annual cycle in radiation and SST. At 10°N and 10°S both the estimates based on semi-annual and annual cycles are plotted and they are not too different. With the exception of the equator and 10°S–20°S h_{em} turns out to be considerably larger than h. Moreover the dependence on latitude is dissimilar. Disregarding the large h_{em} values from 5°N–15°N where the annual cycle is small anyway it turns out that h_{em} is 1.5 to 2 times h. It may be that h, based only on Pacific data, does not represent h determined from zonally averaged SST data, but the main reason for the difference is that h_{em} represents subsurface temperatures and h SST.

4. Seasonality in mixed layer depth

The delays separately for summer and winter in Table 2 give an indication of seasonality. At 6 out of 8 stations the delay is largest in winter. This can be understood via (4a) in terms of a difference in C/b in summer and winter. A deep mixed layer in winter

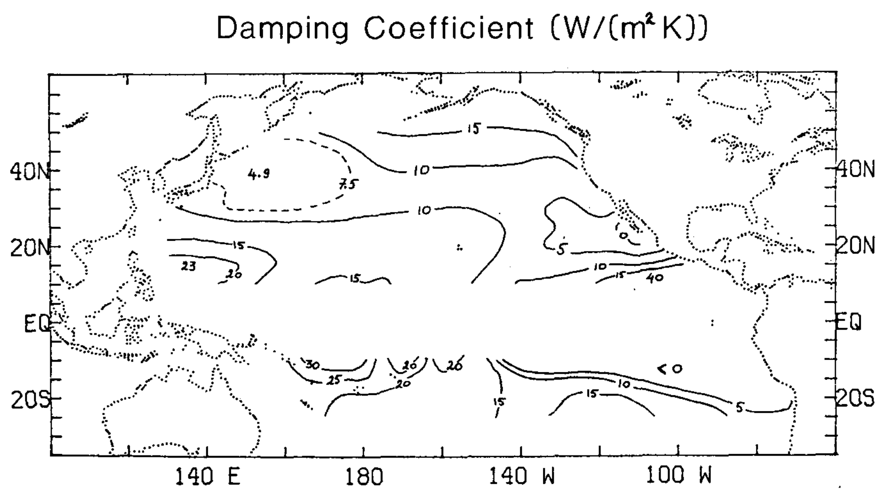

FIG. 3. Calculated values of the damping coefficient b, in W m^{-2} K^{-1}.

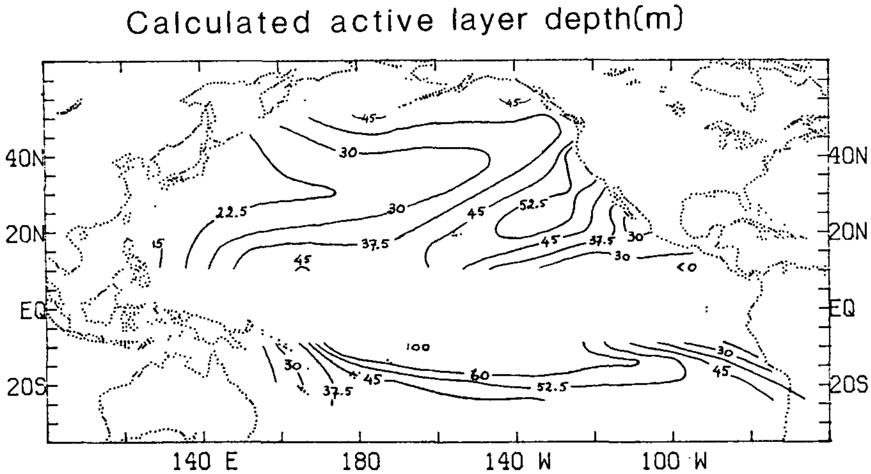

FIG. 4. Calculated values of active layer depth h, in m.

leads to a larger delay than a shallow one in summer. The fact that at all stations summer SST is more above the yearly mean than winter SST is below, fits in with that.

For one station (Scripps Pier) we have used daily data over many years to determine the damping timescale of anomalies in SST, as a function of the time of the year. This was done by calculating the autocorrelation function $\rho(i, \tau)$ which is defined here by

$$\rho(i,\tau)=\frac{\frac{1}{61}\sum_j T(i,j)T(i+\tau,j)-\frac{1}{61^2}\sum_j T(i,j)\sum_j T(i+\tau,j)}{\left\langle\left\{\frac{1}{61}\sum_j T(i,j)^2-\left[\frac{1}{61}\sum_j T(i,j)\right]^2\right\}\left\{\frac{1}{61}\sum_j T(i+\tau,j)^2-\left[\frac{1}{61}\sum_j T(i+\tau,j)\right]^2\right\}\right\rangle^{1/2}}, \quad (6)$$

where $T(i, j)$ is the SST at day i ($i = 1{,}365$) in year j and τ is the lag in days. The summation over j is from 2 to 62 (or the years 1921 to 1981). Data for 1920 and 1982 were used only for those calculations where data of adjacent Decembers and Januaries are needed (large lags).

Before $\rho(i, \tau)$ was determined the annual cycle and the trend were taken out of the data. This was done as follows. For each day i, a climatological mean value $\bar{T}(i)$ was determined. The $\bar{T}(i)$ values were smoothed by applying a running 10-day mean, yielding $\bar{\bar{T}}(i)$. Trends were determined by calculating 63 yearly means $[T(j)]$. Smoothing of these yearly means by a 10-year running mean yielded $[\tilde{T}(j)]$. For the first and the last five years $[\tilde{T}(j)]$ equals $[\tilde{T}(5)]$ and $[\tilde{T}(59)]$, respectively. The trend turns out to be very small. For all i and j, $T(i, j)$ was replaced by

$$T(i,j)-\bar{\bar{T}}(i)-([\tilde{T}(j)]-[\bar{T}]),$$

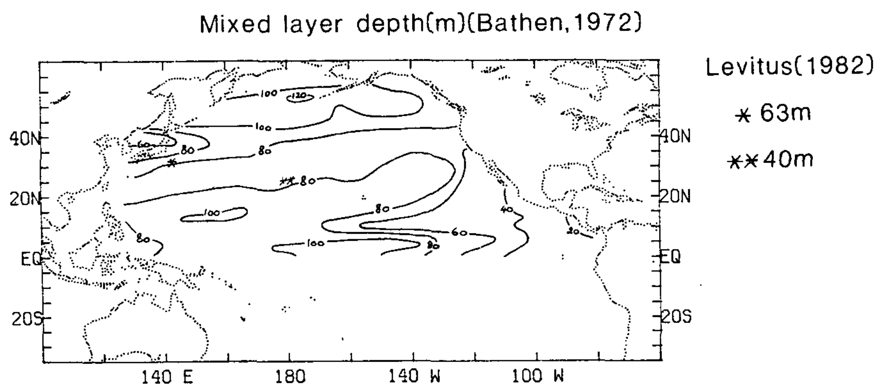

FIG. 5. Observed annual mean mixed layer depth in meters according to Bathen (1972). At the sites denoted by single and double asterisks values by Levitus (1982) are given for comparison.

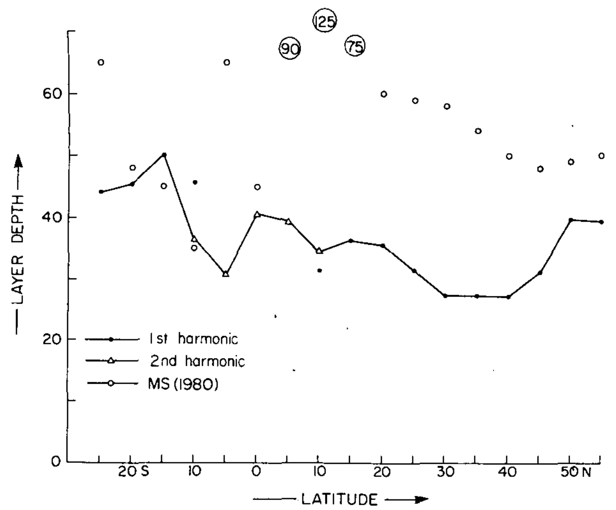

FIG. 6. Active layer depth (h) in the Pacific as a function of latitude based on zonally averaged SST. In the extratropics the annual cycle has been used to estimate h (solid circles) but in the strip 10°S $\leqslant \phi \leqslant$ 10°N we used the semi-annual cycle (delta). For comparison the open circles are effective mixed layer depths as defined by Manabe and Stouffer (1980); they were calculated from seawater temperature averaged zonally across all oceans.

where $[\bar{T}]$ is the grand mean over all i and j, being 16.88°C.

The value of $\rho(i, \tau)$ was calculated for each day of the year and for lags up to 3 months. Fig. 7 shows $\rho(i, \tau)$ as a function of the lag (horizontal axis) and time of the year (vertical axis) averaged by the calendar month. For lag 1 day the correlation varies typically from over 0.9 in January to about 0.8 in September. With increasing lag the correlations decrease rather smoothly. For a SST-anomaly present in January the e-folding time is about 65 days; in September it is only 6 or 7 days, a huge difference.

According to the procedure outlined in Section 2 we will interpret the e-folding time as C/b. Hence Fig. 7 is just as well a graph of C/b against season. The horizontal axis in the upper part of Fig. 7 is a calibration of C (or h) assuming a seasonally uniform b = 11.8 W m^{-2} K^{-1} (see Section 3). The depth of the active layer (according to the e-folding 1 line) varies from a few meters in August–September to about 20 meters in January. Because the calculated correlation coefficients are lowered by noise in the data (due to sampling and instrumental errors) these active layer depths are conservative estimates. The e-folding 29 line indicates that SST anomalies averaged over 29 days have a somewhat longer lifetime resulting in active layer depths of 4 m in summer to well beyond 25 m in winter. The yearly averaged value of 18 m discussed in Section 3 is in reasonable agreement with this. Observed climatological mixed layer depths just off the La Jolla coast vary from about 45 m in January to 0–10 m from May to October (California Cooperative Oceanic Fisheries, 1971); our calculations agree quite well with that. Of course, b is not absolutely constant throughout the year. The other extreme would be to assume that C is uniform with season. In that case the Scripps Pier data suggest b to vary from 5 (January) to about 100 W m^{-2} K^{-1} (August). In view of the similarity of calculated active and measured mixed layer depth variations we suspect that the truth is nearer to constant b and varying C.

Constancy of b with season for the La Jolla area is striking. The energy-balance in the California current is highly nonlocal as has been described in great detail by Nelson and Husby (1983). Therefore upwelling and advection could be forcing mechanisms rather than damping mechanisms and (2) may easily fail. However, after subtracting the yearly averaged cooling by upwelling and advection these two processes effectively warm in winter and cool in summer or, in other words, they behave as damping mechanisms on the annual temperature wave forced by the sun. Nevertheless it may very well be a coincidence that the net damping can be represented by a seasonally uniform coefficient b.

A deep (shallow) mixed layer in winter (summer) seems perfectly reasonable. However there is an apparent paradox in the Scripps data. In Section 3 we derived that the delay in summer is larger than in winter, which would indicate a deeper mixed layer in summer. This problem can be solved by reconsidering the solar data (Tont, 1981). Due to status and fog in June the maximum in solar forcing is not reached at

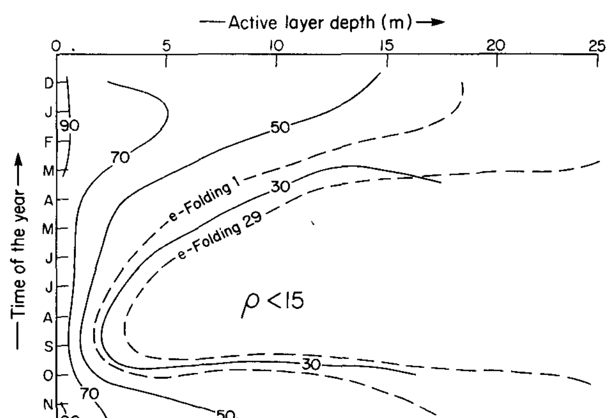

FIG. 7. Isolines of the autocorrelation (in %) of SST anomalies at the Scripps Pier in La Jolla (California) as a function of the lag (lower horizontal axis; days) and the time of the year (vertical axis, months). Anomalies are defined here as deviations from a long-term mean appropriate for the time of the year. The dashed isolines indicate the time over which daily SST (1) and 29-day mean SST (29) anomalies decay to e^{-1}. The upper horizontal axis converts the e-folding time into an active layer depth using the simple theory explained in the text.

June 21 but later in July. So we should reduce the summer delay to about 35 days. In view of this we can recalculate the b and h estimates at the Scripps Pier. The results are $b = 14.2$ W m^{-2} K^{-1} and $h = 15$ m, respectively. Because there are no solar data over the oceans we are not in the position to recalculate b and h at those sites, but we suspect that the large delays in the summertime at ship N (and the east Pacific in general) is erroneous due to the same status problem. So assuming, by default, that the solar forcing is minimum (maximum) at the 21st of December (June) may lead to serious errors.

5. Conclusions and discussion

Employing a simple energy-balance equation we have tried to calculate two aspects of the resistance of the ocean to forced climatic change. The climatological response of the ocean to the annual variation in solar forcing allows us to estimate typical values of the oceanic inertia (C) as well as the net damping (coefficient b) towards equilibrium. The oceanic inertia can be converted into the depth of a slab of water; this defines the active layer depth (h). The SST behaves as if it were taken from a perfectly mixed layer of depth h. In addition, the decay time of SST anomalies can be used for an independent estimate h/b. For most of the midlatitudes and subtropical areas we find active layer depths varying from 15 to 40 meters, which seems reasonable though somewhat small. The spatial distribution of calculated active layer depths over the Pacific is fairly similar to that of observed mixed layer depth. The damping coefficient suggested by the data is typically 10 W m^{-2} K^{-1}. These b and h estimates are typical values for the year as a whole. Assuming b to be uniform with season, the decay time of SST anomalies at the Scripps Pier indicates h to vary from 3 m in late summer to about 30 m in January. Again, this is in reasonable agreement with observations of the mixed layer throughout the year at that site.

As a bonus we have also derived the net damping of SST towards equilibrium. Compressed into one damping coefficient the damping $b \approx 10$ W m^{-2} K^{-1} seems large compared to the radiative damping $b = 2.1$ W m^{-2} K^{-1} used in North and Coakley's model. Let us compare the equations used, for a correct interpretation of the damping. While we use the surface energy balance

$$C_o \frac{\partial T_o}{\partial t} = F_o - (a + bT_o), \tag{7}$$

zonally and vertically averaged energy-balance models are formulated as

$$\frac{\partial E}{\partial t} = F_a + F_o - (a' + b'T_{as}) - D\nabla^2 T_{as}. \tag{8}$$

The suffix a and o are for atmosphere and ocean, T_{as} is the surface air temperature, E is the energy content of the column extending from the depth of the active layer to the top of the atmosphere and $D\nabla^2 T_{as}$ is a parameterized form of meridional heat transport by the ocean and the atmosphere. The reasons why b' in (8) is smaller than b in (7) are: 1) The inclusion of the oceanic part of $D\nabla^2 T_{as}$ in b and 2) the inclusion of heat exchange between the atmosphere and the ocean in b. Indeed the effect of horizontal heat transfer seems to be at least a doubling of the effective damping (Budyko, 1969). The vertical heat transfer between atmosphere and ocean plays an important role in the energy-balance of the two subsystems but, obviously, not in (8). So it is likely that ocean and atmosphere damp quickly towards their respective equilibria while the system as a whole damps much slower towards equilibrium (with outer space). A further technical reason for a difference in b and b' is that (7) is applied locally whereas (8) is a zonal average.

There are, however, additional fundamental differences in b and b'. In (8) the time derivative operates on E while T_{as} appears in the damping. For their convenience many authors have equated E to $(C_a + C_o)T_{as}$ but it is far from obvious that the time variation of the atmosphere's energy content (latent heat, internal, potential and kinetic energy) can be represented by T_{as} alone. In (8) a damping of the form $a'' + b''E$ is more appropriate and b'' may be very different from b'.

Our results concerning active layer depth and damping coefficient are relevant to various groups of models.

1) Atmospheric models designed for very long-term integration, such as Manabe and Stouffer's model. In these models the ocean is usually represented by an energy balance equation for the (effective) mixed layer. Since the atmosphere feels only SST it seems straightforward that the thermal inertia of the ocean, as derived by our empirical method, is the one needed to integrate such a coupled atmosphere-ocean model successfully through the annual cycle. The surface energy balance then reads $\rho c_p h \partial/\partial t$ SST $= Q$ where h is the active layer and Q consists of all heat exchanges at the surface, each of which can be evaluated. There is no application for b here. Even though h seems to be a good choice for the annual cycle there is no guarantee that other transient climate problems can be handled. After all, h may be time-scale dependent.

2) Stationary atmospheric models (monthly or seasonal time-scale) forced by heat sources and sinks related to SST anomalies. Skillful atmospheric predictions can only be made if SST anomalies can be projected skillfully into the future. A physically sound way to do this is to decay SST anomalies with the time scale C/b. C/b may depend on season and position.

3) Vertically averaged energy balance models for climate sensitivity studies. It has become customary to approximate $\partial E/\partial t$ (in Eq. (8)) by $C\partial T_{as}/\partial t$ (see North and Coakley, 1979, for example). Since C is

dominated by C_o it is straightforward to use the active layer depth as a measure of thermal inertia at least in oceanic areas. Rather than advocating the use of h too much in this context we want to express some reservation about equating $\partial E/\partial T$ to $C\partial T_{as}/\partial T$. Since the energy of the atmosphere is more than just sensible heat it is questionable whether the atmospheric part in $\partial E/\partial t$ can be written as $C_a\partial T_{as}/\partial t$. Moreover, to the extent that $\partial E/\partial t$ can be written as $C_o\partial T_o/\partial t + C_a\partial T_{as}/\partial t$ it is not clear at all that $C_a\partial T_{as}/\partial t$ can be neglected let alone that one can write $C_o\partial T_{as}/\partial t$. It may turn out to be virtually impossible to use oceanic inertia straightforward as a measure of C in a vertically averaged energy balance model.

Three final remarks are appropriate.

1) The essence of this paper is that mixed layer depths as determined in one way or another from oceanic temperature profiles may not be directly applicable to an interactive ocean–atmosphere model. Instead, we advocate as a strategy to invert the problem and to derive the ocean's inertia (active layer) from observed behavior of SST over the annual cycle (or any other well documented climatic change). This paper gives an example of such an approach. By simplifying the ocean's energy balance to its bare principle (2), the inversion of the problem becomes very easy. All we have done here is to show that there are seasonally uniform but location dependent choices for C and b for which the annual cycle (first harmonic only) in SST can exactly be reproduced. More in general, one might want to invert more complete energy balances such as the one described in the Appendix. The very best (and feasible) approach would be to run an Atmospheric Circulation Model with prescribed normal seasonally varying SST over many annual cycles. This yields the best possible, model consistent, climatology of the energy flux (EF_n) at the ocean–atmosphere interface. Equating $C\partial \mathrm{SST}/\partial t = EF_n$ at each location yields C as a function of space and season. Consequently using this C in an interactive mode (SST variable) guarantees the best possible reproduction of the annual variation in SST, at least in coupled models where the ocean is represented by a slab and where horizontal transports (in the ocean) and upwelling are not explicitly included.

2) The whole discussion in this paper is concerned with the climatological annual cycle of SST. It is intuitively acceptable that the annual cycle in incoming solar radiation forces the annual cycle in SST whereas upwelling, advection and evaporation act as a damping. However, at other time scales, especially short ones, any of the "damping" processes may act as a forcing; this invalidates (2) for those time scales. For the period 1934–50 we investigated for the Scripps Pier how well observed monthly mean anomalies in incoming solar radiation ($\hat{Q}$) are related to observed monthly mean SST anomalies ($\widehat{\mathrm{SST}}$). Fig. 8 displays the series of $\hat{Q}$ and SST for January and June. The lack of correlation is very clear. In no part of the year is there any trustworthy (statistical) correlation between (i) $\hat{Q}(j)$ and $\widehat{\mathrm{SST}}(j)$, (where j is the month of the year), (ii) $\hat{Q}(j-1)$ and $\widehat{\mathrm{SST}}(j)$, (iii) $\hat{Q}(j) + \hat{Q}(j+1)$ and $\widehat{\mathrm{SST}}(j) + \widehat{\mathrm{SST}}(j+1)$ and (iv), (model inspired) $\widehat{\mathrm{SST}}(j+1) - \widehat{\mathrm{SST}}(j)$ and $\hat{Q}(j) + \hat{Q}(j+1)$. Of course, this does not mean that incoming radiation has no influence on SST. Keeping anything else the same $\widehat{\mathrm{SST}}$ and $\hat{Q}$ would be positively correlated. However, "anything else" is not the same every year. For example, in Fig. 8 the El Niño case of 1940–41 shows up beautifully. Above normal storminess over the East Pacific may have caused both strong advection of warm water (high SST) and a lot of clouds (low Q).

3) It is possible to extend the analysis to the vertically integrated energy content of the atmosphere to derive atmospheric inertia (C) and damping (b) from the ob-

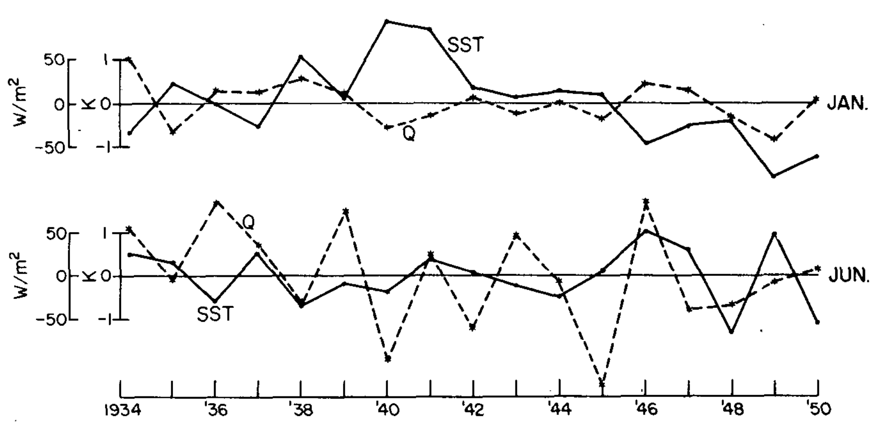

FIG. 8. Synchronous anomalies in monthly mean SST and incoming solar radiation at the Scripps Pier in La Jolla for January and June from 1934 to 1950. Anomalies are deviations from the 1934–50 mean. The units are K and W m^{-2}. Observations of solar radiation stopped in 1950 are less reliable prior to 1934.

served annual cycle in the energy content of the atmosphere. This problem is an order of magnitude more complicated than its oceanic counterpart because atmospheric energy is not just CT but CT plus latent energy (kinetic energy can be safely neglected). Moreover the atmosphere is not very well mixed. Nevertheless, a "back of the envelope" calculation with the climatology of Bismarck, North Dakota gave the following result: $b = 4.8$ W m^{-2} K^{-1} and $C = 1.4 \times 10^7$ J m^{-2} K^{-1} (a mixed atmosphere of 1400 mb thickness). These values look reliable. On a globe without oceans C should be at most 1.0×10^7 J m^{-2} K^{-1} (1000 mb) and Bismarck (a very continental site) is approaching that.

Acknowledgments. Thanks are due to Mr. S. Tont and Dr. A. Frantzen for providing solar radiation data. The North Pacific SST data were provided by Mr. D. Cayan. The discussions with Drs. G. North, G. Vallis, and Mr.'s D. Cayan and D. Blaauboer were greatly appreciated. We thank J. F. Garrett for suggesting the term active layer. Mrs. G. Johnston typed the manuscript.

APPENDIX

Ocean–Atmosphere System Including T_a

Here we address the question whether systematic errors are made in estimating h and b from an equation as simple as (3). Suppose that the "real world" atmosphere–ocean system were given by

$$C_a \frac{\partial T_a}{\partial t} = \gamma\sigma T_o^4 - 2\gamma\sigma T_a^4 + Q(1-\alpha)a + c(T_o - T_a) - D_a\nabla^2 T_a, \quad \text{(A1)}$$

$$C_o \frac{\partial T_o}{\partial t} = \gamma\sigma T_a^4 - \sigma T_o^4 + Q(1-\alpha)(1-a) - c(T_o - T_a) - D_o\nabla^2 T_o, \quad \text{(A2)}$$

where C_a and C_o are atmospheric and oceanic heat capacities, Q is solar radiation, α is planetary albedo, a is atmospheric absorbtion, γ is the efficiency of the (grey) atmosphere to emit/absorb infrared radiation, σ is the Stefan-Boltzmann constant, c is a coefficient for air–sea heat exchange, D_a and D_o are diffusion parameters. In contrast to (3), (A2) is linked explicitly to the variable T_a via radiation and air–sea heat exchange. (A1)–(A2) are somewhat in the spirit of Hoffert *et al.* (1980).

Annual cycles in T_o and T_a can be forced by a periodic variation of Q. All other parameters are kept constant over the annual cycle. While maintaining accuracy we can simplify (A1)–(A2) by linearizing to obtain

$$C_a \frac{\partial T_a}{\partial t} = (\gamma u + c)T_o - (2\gamma v + c + e)T_a + Q(1-\alpha)a, \quad \text{(A3)}$$

$$C_o \frac{\partial T_o}{\partial t} = (\gamma v + c)T_a - (u + c + d)T_o + Q(1-\alpha)(1-a), \quad \text{(A4)}$$

where T_a, T_o and Q are redefined as deviations from their annual mean ($\bar{T}_a$, etc). The coefficients u and v are determined from $4\sigma\bar{T}_o^3$ and $4\sigma\bar{T}_a^3$ respectively, that is $u = 4.6$ to 6.2 W m^{-2} K^{-1} [$\bar{T}_o = 273$ to 300 K (polar to equatorial seas)] and $v = 3.3$ W m^{-2} K^{-1} for $\bar{T}_a = 245$ K. Without rigorous defense the diffusion terms are simplified to eT_a and dT_o respectively.

Equation (A4) differs from (3) only in the presence of the term $(\gamma v + c)T_a$. Note also that the damping of T_o in (A4) is the net effect of outgoing infrared (u), air-sea heat exchange (c) and horizontal mixing (d); $b = u + c + d$. Our strategy will be to calculate annual cycles in T_o and T_a via (A3)–(A4) by assuming values for the mixed layer depth and the damping coefficient. Then we derive estimates of h and b by inserting the amplitude and the phase of the annual in T_o (B and Δ) as well as the solar forcing (A) into (4b), (4a) and (2). By comparing the assumed and estimated values of h and b we can check whether (3) is a valid approximation to (A4).

In principle (A3)–(A4) can be solved analytically, but since the final expressions are awkward we resort to numerical integration. Starting from arbitrary initial conditions it takes only a few model years for T_o and T_a to converge to a repeating annual cycle (within 0.01 K). The time step is 1 day. The system is forced by $Q(1-\alpha) = 100 \sin\omega t$ where ω is the annual frequency.

The results are given in Table A1. In the third row, denoted by an asterisk, we have the basic experiment in which all parameters of (A3)–(A4) are believed to have credible values. Except for the ones explicitly given in Table A1, the constants are $a = 0.2$, $C_a = 10^7$ J m^{-2} K^{-1} (assuming an atmosphere of 10 km height and density 1), and u, v, d and e are 5, 3.3, 4 and 2 W m^{-2} K^{-1}. As can be seen from the basic experiment, both b and h are underestimated using (3), h by about 10% and b by about 50%. Experiments 1 to 5 are conducted to investigate how sensitive (3) is to changes in the role of the atmosphere in the oceanic energy budget. The role of the atmosphere is changed by changing the emissivity and/or air–sea heat exchange.[2] It turns out that (3) is rather robust in estimating h; all errors made by neglecting the annual cycle in T_a are apparently combined into the estimate of b. Experiments 6 and 7 are rather "wild" deviations from the real world. For a mixed layer of 1000 m the theory breaks down; $\Delta > 90° \rightarrow b < 0$. For a mixed layer of 20 cm the neglect of T_a is, indeed, serious.

It seems that the systematic error in h is small. This

[2] Changing γ and c leads strictly speaking to other $\bar{T}_o$ and $\bar{T}_a$ and therefore u and v change as well. However, we neglect that effect here.

TABLE A1. The response of the model described by (A3)–(A4) to periodic solar forcing at annual frequency. From left to right: assumed emissivity γ, air-sea heat exchange coefficient c (W m^{-2} K^{-1}), oceanic damping b ($\equiv u + d + c$) (W m^{-2} K^{-1}) and mixed layer depth h (m); calculated response in T_a and T_o [amplitude (K), phase (days behind the sun)] and b and h calculated from (4a), (4b) and (2a).

	Assumed constants				T_a		T_o		Calculated	
Experiment	γ	c	b	h	Amp	Phase	Amp	Phase	b	h
1	0.75	1	10	40	3.1	42	2.5	79	6.6	36.9
2	0.75	5	14	40	2.8	51	2.6	77	6.9	35.7
3*	0.85	3	12	40	2.8	48	2.6	78	6.5	36.2
4	0.95	1	10	40	2.9	44	2.6	79	6.2	36.8
5	0.95	5	14	40	2.6	53	2.6	78	6.5	35.8
6	0.95	5	14	1000	1.5	12	0.1	91	−6.5	847
7	0.95	5	14	0.2	9.6	16	11.2	8	7.1	1.2

can be understood by rearranging (4a) and (4b) which yields

$$h = \frac{\sin(\Delta)A}{\rho C_p B\omega}.$$

Even though A is too small (we neglect downward infrared radiation as a forcing of the annual cycle in T_o) and Δ is too large (the net forcing including downward infrared radiation peaks later than June 21), h is still about right. However, the errors in A and Δ are additive in calculating b [since $b = A\cos(\Delta)/B$]. As an example: the net forcing (= solar + downward infrared + sensible heat exchanges) in experiment 3 peaks on June 29 (instead of 21) with an amplitude of 91.7 W m^{-2} (instead of 80.8 W m^{-2}). These differences lead to a considerable error in b.

The point of this Appendix is that neglecting the annual cycle in T_a leads to a small systematic error in h (-10%) and a sizeable systematic error in b (-50%). This conclusion holds within the context of (A3)–(A4). Relative to the real world, the role of atmospheric moisture is missing altogether from (A3)–(A4). It is not clear whether this changes our arguments. However, relying upon bulk aerodynamical formulas such as those used by Leetmaa (1983), the ocean–atmosphere latent heat exchange is only to a limited extent determined by the temperature of the atmosphere. In areas where evaporation is large (the tropics and subtropics) the air–sea temperature difference is small, and therefore T_a is somewhat redundant. Therefore it seems that (3) should yield a relatively accurate estimate of the active layer depth.

REFERENCES

Bathen, K. H., 1972: On the seasonal changes in the depth of the mixed layer in the North Pacific Ocean. *J. Geophys. Res.*, **77**, 7138–7150.

Budyko, M. I., 1969: The effect of solar radiation variations on the climate of the earth. *Tellus*, **21**, 611–619.

California Cooperative Oceanic Fisheries Investigations, 1971: CalCOFI Atlas 15, Scripps Institution of Oceanography, La Jolla, p. 188.

Chiu, L. S., and R. E. Newell, 1983: Variations of zonal mean sea surface temperature and large-scale air-sea interaction. *Quart. J. Roy. Meteor. Soc.*, **109**, 153–158.

Douglas, A. V., D. R. Cayan and J. Namias, 1982: Large-scale changes in North Pacific and North American weather patterns in recent decades. *Mon. Wea. Rev.*, **110**, 1851–1862.

Esbensen, K. E., and R. W. Reynolds, 1981: Estimating monthly averaged air-sea transfer of heat and momentum using the bulk aerodynamic method. *J. Phys. Oceanogr.*, **11**, 457–465.

Frankignoul, C., and K. Hasselmann, 1977: Stochastic climate models: Part 2. Application to sea-surface temperature anomalies and thermocline variability. *Tellus*, **29**, 289–305.

Gill, A. E., and P. P. Niiler, 1972: The theory of the seasonal variability in the ocean. *Deep-Sea Res.*, **20**, 141–177.

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

Hoffert, M. I., A. J. Callegari and C-T. Hsieh, 1980: The role of deep sea heat storage in the secular response to climatic forcing. *J. Geophys. Res.*, **85**, 6667–6680.

Horel, J., 1982: On the annual cycle of the tropical Pacific atmosphere and ocean. *Mon. Wea. Rev.*, **110**, 1863–1878.

Leetmaa, A., 1983: The role of local heating in producing temperature variations in the offshore waters in the eastern tropical Pacific. *J. Phys. Oceanogr.*, **13**, 467–473.

Levitus, S., 1982: *Climatological Atlas of the World Ocean.* NOAA, Prof. Pap. No. 13, 174.

Manabe, S., and R. J. Stouffer, 1980: Sensitivity of a global climate model to an increase of CO_2 concentration in the atmosphere. *J. Geophys. Res.*, **85**, 5529–5554.

McEwen, G. F., 1938: Some energy relations between the sea surface and the atmosphere. *J. Mar. Res.*, **1**, 217–238.

Nelson, C. S., and D. M. Husby, 1983: Climatology of surface heat fluxes over the California current region. NOAA Tech. Rep. NMFS SSRF-763, 155.

North, G. R., and J. A. Coakley, 1979: Difference between seasonal and mean annual energy balance model calculations of climate and climate sensitivity. *J. Atmos. Sci.*, **36**, 1189–1204.

——, R. F. Cahalan and J. A. Coakley, 1981: Energy balance climate models. *Rev. Geophys. Space Phys.*, **19**, 91–121.

Schneider, S. H., and R. E. Dickinson, 1974: Climate modeling. *Rev. Geophys. Space. Phys.*, **12**, 447–493.

Sellers, W. D., 1965: *Physical Climatology.* University of Chicago Press, 272 pp.

——, 1969: A global climate model based on the energy balance of the earth-atmosphere system. *J. Appl. Meteor.*, **8**, 392–400.

——, 1973: A new global climate model. *J. Appl. Meteor.*, **12**, 241–254.

Simmonds, I., and C. Chidsey, 1982: The parameterization of longwave flux in energy balance climate models. *J. Atmos. Sci.*, **39**, 2144–2151.

Thompson, S. L., and S. H. Schneider, 1979: A seasonal zonal energy balance climate model with an interactive lower layer. *J. Geophys. Res.*, **84**, 2401–2414.

Tont, S. A., 1981: Upwelling: effects on air temperature and solar irradiance. *Coastal Estuarine Sci.*, **1**, 57–62.