MONITORING OCEAN SURFACE LAYER PROCESSES

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ABSTRACT

The monitoring of several upper ocean processes over large spatial scales on monthly and longer time scales is discussed. The surface forcing functions of wind stress, heat flux, and freshwater flux are treated in terms of the potential roles that in situ observations, satellites, and atmospheric general circulation models might play in producing the most useful fields on a regular basis globally. Thermocline mixing and daily cycling are presented as examples of phenomena that, once understood from intensive regional studies, could be observed from a much wider perspective using relatively easily measured data. Specifically, episodic cooling, a symptom of intense thermocline mixing in the North Pacific during autumn, appears to occur over large regions of the Southern Hemisphere, depending on the annual cycle of wind forcing. Also, the daily cycle of equatorial oceanic turbulence seen in the short Tropic Heat study of 1984 appears to persist throughout the year over 20° of longitude. A similar cycle is not evident during heating season in the Sargasso Sea.

INTRODUCTION

Only a few of the parameters important to ocean surface layer processes are currently being monitored. One example is the vertical temperature profile from ship of opportunity XBT casts. Most of the others are summarized in the Climate Diagnostics Bulletin (CDB), distributed monthly by NOAA’s Climate Analysis Center (CAC). In addition, the Australian Bureau of Meteorology’s National Climate Center issues a Southern Hemisphere Climate Monitoring Bulletin (SHCMB) each month.

A large international effort is underway to take large numbers of XBT casts in most of the world’s oceans, excepting the Arctic and Southern Oceans. From these casts, important parameters such as thermocline depth and heat content can be computed on a quasi-synoptic basis. In the Anomaly Dynamics Study, for example, such data and winds were used to test the Ekman pumping mechanism in the North Pacific (White et al., 1980). In a study of thermal variability in the Pacific from 20°S to 50°N, White et al. (1985) used 85000 XBT observations. Recently, the Tropical Oceans Global Atmosphere (TOGA) program has provided the impetus for a rapid expansion of the XBT network in all three equatorial oceans. In the Pacific an ambitious program calls
for sampling with 1 to 2° latitude and 10 to 20° longitude spacing on a monthly to seasonal time scale. An extensive program is planned initially for the Indian Ocean in order to establish its data base. An Atlantic XBT network is well on its way to becoming operational.

The oceanographic data in the CDB are particularly useful for following the generation and eastward propagation of Kelvin waves along the Pacific equator. These data include pan-Pacific sea level anomalies and Equatorial Pacific thermocline depth (depth of the 20°C isotherm from the XBT program), surface wind psuedo-stress vectors, and zonal wind indices. Interest in Equatorial Pacific wave processes stems from the focus of the TOGA program on ENSO (El Niño, Southern Oscillation) phenomena. Global fields in the CDB appear to be less concerned with specific processes. However, they do include several air-sea interaction parameters; namely, sea level pressure, sea surface temperature (SST), 850mb vector wind, and outgoing longwave radiation. The SHCMB deals primarily with meteorological variables, but it does include SST anomalies.

Of course equatorial waves and the surface forcing are not the only important processes that govern the upper ocean’s structure. However, studies of other processes, such as mixing, diurnal cycling, Langmuir circulation, entrainment, and advection, have largely been confined to local, small scale, short duration, observational programs, such as JASIN, MILE, STREX, FASINEX, and Tropic Heat. In order to assess the overall impact of a particular process on the state of the upper ocean it is not enough to observe it in one of these programs. It is also necessary to determine both its frequency of occurrence and its global distribution. Such assessments are essential to the success of large scale oceanographic and climatic programs like WOCE (World Ocean Circulation Experiment) and TOGA.

Realistically it will not be possible to mount intensive process oriented experiments covering all seasons in more than a few regions of the world’s oceans. Therefore, indirect methods, through indices or signatures involving more easily measured parameters, need to be developed and tested for monitoring important surface layer processes globally. In the case of the surface forcing functions and waves in the Equatorial Pacific, the situation appears to be relatively well in hand. The former will be discussed further in the next section. Generally, other processes are not being looked at in this way, but the following attempts to do so for a particular thermocline mixing process and for an aspect of diurnal cycling.

SURFACE FORCING FUNCTIONS

The ocean is forced by the surface fluxes of momentum (wind stress), \( \bar{F}_0 \), heat, \( Q_0 \), and freshwater, \( F_0 \). Global wind stress measurements are showing a great deal of promise, because of the advent of satellite techniques for both scalar (altimetry, radiometry) and vector (scatterometry) wind estimates, and because of the long term efforts to establish the parameterization:
\[ \tilde{r}_0 = \rho_a C_D U \tilde{U} \] 

where the drag coefficient \( C_D \) is a function of measurement height and stability, \( \rho_a \) is the air density and \( \tilde{U} \) is the wind velocity, with \( U \) its magnitude.

The surface heat flux into the ocean, \( Q_0 \), is the balance between net solar heating \( SW \), net longwave radiation \( LW \), the latent heat flux \( H_L \), and the sensible heat flux \( H_S \):

\[ Q_0 = SW + LW - H_L - H_S \] 

The bulk transfer coefficients, \( C_E \) and \( C_T \), in the formulae

\[ H_L = \rho_w \Lambda E_0 = \Lambda C_E U (q_0 - q_a) \]

\[ H_S = \rho_a c_p C_T U (T_0 - T_a), \]

where \( \Lambda \) is the latent heat of evaporation, \( \rho_w \) is the density of seawater, \( q_0 \) and \( q_a \) (\( T_0 \) and \( T_a \)) are the air humidities (temperatures) at the surface and measurement height, respectively, and \( c_p \) is the specific heat of air, are less well established than \( C_D \). Nevertheless, suitable measurements of the mean quantities in (3) and (4) would give global estimates of \( H_L \) and \( H_S \). The prospect of monitoring \( Q_0 \) globally has improved greatly, because of the recent development of techniques for estimating \( SW \) from satellite imagery (Gautier et al., 1980). Should present efforts to retrieve \( LW \) also prove fruitful, then it will become feasible to produce fields of \( Q_0 \) routinely.

The difference between the rates of evaporation, \( E_0 \), and precipitation, \( P_0 \), is

\[ F_0 = E_0 - P_0 \]

Although (3) parameterizes \( E_0 \), measurements of \( P_0 \), either from satellite or in situ, are unproven and may preclude using (5). However, the net divergence of moisture in a vertical column of the atmosphere is a direct measure of \( F_0 \), that it may be possible to extract from an atmospheric general circulation model, such as those used at the major forecast centers, such as ECMWF (European Center for Medium Range Weather Forecasting) and NMC (the U.S. National Meteorological Center).

The problem of producing global fields of the forcing functions for WOCE is addressed in Large (1985). What emerges there and here is that should the following become available:

1) Standard bulk parameterizations (1), (3) and (4) with established uncertainties
2) Satellite SW and LW,

3) Satellite and in situ SST ($T_0$ and $q_0$),

4) Atmospheric GCM assimilation of bulk parameters, especially $U, q_a$, and $T_a$, and computation of $F_0$,

then useful estimates of the forcing functions could be produced routinely over all the world's oceans. Of course the accuracy would likely be better in the more well travelled and sampled northern hemisphere.

Should such an effort be undertaken, there would need to be a significant effort put into data communication and management. Figure 1 shows the present scheme used by NOAA to produce monthly SST maps. It uses data from voluntary observing ships (VOS) and select buoys as ground truth for satellite radiometric fields. Also shown are other potential sources of data that could be incorporated for a better product. The improvements possible include smaller spatial and shorter temporal resolution, greater accuracy, and a more geographically uniform reliability.

![Diagram](image)

**Fig. 1** Satellite and in-situ data flow of measurements relevant to producing sea surface temperature fields. Solid lines represent the flow used by CAC, and the dashed lines are potential augmentations.
Figure 2 shows the hypothetical data flow needed to produce surface wind stress fields. The most straightforward path is to produce the fields from satellite data alone. It is being pursued by NASA. Should this approach prove to be too unreliable it may be necessary to gather more of the available surface wind data together to produce the stress field as is being done for TOGA. Figure 3 illustrates the possible role of an atmospheric general circulation model. It is now being explored for the limited case of incorporating near–real–time satellite wind data (from ERS–1) into the surface analysis now performed at ECMWF before forecast runs. Such a scheme may exclude a great deal of data that arrives too late. The alternative that would make maximum use of all data is to run a delayed analysis. Since the result would not be of use to forecasts there are considerable political and economic problems with implementing such an operation. Therefore, it is first essential to establish that such an effort is needed and worthwhile.

Fig. 2 Hypothetical data flow into a wind analysis center, where time binned wind fields as well as the wind stress field could be produced.
THERMOCLINE MIXING

One of the important discoveries of the Storm Transfer and Response Experiment (STREX) was that the episodic nature of the autumn SST cooling is indicative of intense vertical mixing in the seasonal thermocline, whereby cool thermocline water is exchanged with warm mixed layer water, without the mixed layer depth necessarily increasing as required for entrainment (Large et al., 1986). Furthermore, subsequent horizontal advection in the now locally warmer thermocline could be responsible for the ultimate removal of summertime mixed layer heat from the water column. This heat loss was estimated to be sufficient to account for the annual average imbalance between $Q_0$ and local heat storage observed by Tabata (1965).

The STREX scenario appears to be common over a large area of the Gulf of Alaska and is a major component of the annual heat budget. The question to be addressed here is: Are there any other ocean regions where similar physics are important and what are

Fig. 3 Hypothetical data flow for assimilation and flux field production by an atmospheric general circulation model. Potential venues are the European Center for Medium-Range Weather Forecasting (ECMWF) and the U.S. National Meteorological Center (NMC).
the necessary conditions? First, it is necessary to develop quantitative measures of the episodic behavior of the autumn SST decrease in STREX. These measures can then be computed for all the 1979 southern hemisphere SST records from the FGGE drifting buoy array. Where these values are similar to those found in STREX it is probable that similar thermocline mixing also occurred.

The episodic nature of the STREX fall cooling is illustrated in Fig. 4, where the fraction of the net cooling over 50 days is plotted cumulatively against that fraction of the 50 days over which the cooling took place. Because of the brief intense cooling events, 80% of the ~3.0°C takes place in just 20% of the time and cooling occurs only 50% of the time. Solar heating and some negative $H_L$ are occasionally sufficient (12% of the time) to produce some heating. A measure of the departure from uniform cooling is the area between the solid and dashed curves, $A = 0.60$.

Plots similar to Fig. 4 were produced for 120 FGGE buoys with a complete SST

![Graph](image)

**Fig. 4** Normalized cumulative sea surface temperature cooling curve from nine drifting buoys near Ocean Weather Station "PAPA" over 50 fall days of both 1980 and 1981.
record over the 60 autumn days from 16 April to 15 June, 1979. An example of the SST data from buoy 54602 is shown in Fig. 5 along with the sea level pressure at the buoy. Episodic cooling is evident and as in STREX it appears to be associated with some, but not all, of the storm conditions (low pressure). The SST records were smoothed by averaging over each 48-h period and excluding any period with fewer than two SST observations. The results from buoy 54602 (Fig. 6) are very similar to the STREX experience, implying that in this area (40–45°S, 160–170°W) of the South Pacific Ocean, the mixed layer physics, including thermocline mixing, during the fall season is similar to that found in the North Pacific. The probability distribution (histogram) of “amount of cooling over two days” is very distinct for regions of episodic cooling. It is very non-Gaussian with a large standard deviation. There are also too many cases of values very much smaller (more negative) than the mean amount of cooling.

From the time series and histograms it was possible to associate some of the buoys with episodic cooling and these are plotted with solid circles and trajectories in Fig. 7. Other buoys do not seem to experience this mode of cooling and they are shown as open circles and dashed trajectories. Others, where it is unclear, have been omitted. At present this selection process is rather subjective and efforts are underway to find quantitative criteria by which to judge whether or not buoy SST time series is indicative of episodic cooling.

![Graphs of SST and SLP](image)

**Fig. 5** Sea surface temperature (SST) and sea level pressure (SLP) at FGGE buoy 54602 (40–45° S, 160–170° W) from 16 April to 15 June 1979.
Fig. 6 Normalized cumulative sea surface temperature cooling curve at FGGE buoy 54602, from 16 April to 15 June 1979.

Figure 7 suggests that thermocline mixing occurred in two distinct regions: one in the South Pacific and the other in the South Atlantic. They both extend over most of the basins in longitude and are both bounded approximately by 35 and 45° south latitude. There does not appear to be a corresponding region in the Indian Ocean, nor is there any evidence of episodic cooling anywhere around the Southern Ocean, which includes the southern latitudes corresponding to the STREX area (50°N).

The geographic distribution found in Fig. 7 appears to be governed by the summer and fall surface wind forcing. In order to have intense thermocline mixing cause episodic cooling, the summer winds need to have been weak enough for a shallow warm mixed layer to form over a strongly stratified seasonal thermocline, as is the case in the Gulf of Alaska. From Fig. 8a it is remarkable how closely the episodic cooling regions of both the Atlantic and Pacific correlate with the latitudes of low (<5 m/s) mean 1979 summer winds. Farther south the winds were very much stronger, probably preventing the establishment of a seasonal mixed layer and thermocline. Secondly, the autumn
winds must be very strong to generate sufficient inertial shear across the thermocline for it to go unstable and provide the energy for mixing. Although there is a large area of weak summer winds over the Indian Ocean (Fig. 8a), the autumn winds in this area appear too weak (Fig. 8b) for thermocline mixing. In contrast the areas of episodic cooling in Fig. 7 experienced an intensification from low to moderate (5–10 m/s) in the mean wind from summer to fall.

Fig. 7 Trajectories of 98 FGGE drifting buoys over the Austral autumn of 1979. Solid circles indicate the 16 April positions of buoys clearly displaying episodic cooling and the solid lines show their 60 day movement. Open circles and dashed trajectories represent buoys that do not seem to have experienced such cooling.

THE DIURNAL CYCLE

Renewed interest in the diurnal cycle of the upper ocean has been spawned by the Tropic Heat observation that near equatorial (Pacific) values of turbulent kinetic energy dissipation rate (and hence vertical mixing) are much larger at night than during the day (Chereskin et al., 1986). Also, the mid–latitude diurnal cycle has recently been intensely sampled and modelled (Price et al., 1986).

The upper ocean heat budgets during the day and at night are, respectively,

\[ H_i^D = SW - L^D - A^D - R^D \], and \( (6) \)
Fig. 8 Mean geostrophic surface wind speed (m/s) in the southern hemisphere, a: January–February (summer) 1979 and b: April–May (autumn) 1979.

\[ H_t^N = - (L^N + A^N + R^N) \],

where \( H_t \) is the average rate of change in heat content; \( -L = LW - H_L - H_S \); \( A \) and \( R \) are, respectively, the net advection and turbulent cooling; and superscripts D and N refer to day and to night, respectively. Subtraction of (7) from (6) yields an expression for the night-to-day bias in the sum of the loss terms,

\[ B = \Delta (L + A + R) = (H_t^D - H_t^N) - SW \].

Imawaki et al. (1987) demonstrate that in the eastern Equatorial Pacific there is no night-to-day bias in either \( L \) or \( A \), and so they were able to reduce (8) to a measure of the bias in the turbulent cooling.
\[ \Delta R = (H_t^D - H_t^N) - SW \]  

\[ \Delta H_t = (H_t^D - H_t^N) = \frac{\rho w c_w}{\Delta t} \int_{-h}^{0} (2T(2) - T(1) - T(3)) \, dz \]  

where \( c_w \) is the water's specific heat, \( \Delta t = 12 \) hours, \( h \) is depth, and the ocean temperatures at sunrise, sunset, and sunrise the next day are \( T(1), T(2), \) and \( T(3), \) respectively. Thus, using only thermistor chain data and SW values from satellites, they were able to extend the conclusions of the Tropic Heat observations of dissipation rate made at 140°W. Their results from 125°W (Fig. 9) and 140°W indicate that the bias persists throughout the year at both longitudes. It is also evident at 1°S at 140°W.

**Fig. 9** Monthly mean estimates of \( \Delta H_t \) (open circles), SW (solid line) and \( \Delta R \) (crosses) from mooring data on the equator at 125° W. Abscissa is the month of 1984 and 1985 (from Imawaki et al., 1987.)

Observations of the diurnal cycle (34°N, 70°W) in the Sargasso Sea have been reported by Price et al. (1987). From April through August of 1983 there were also three thermistor chain drifting buoys in the vicinity. It is evident that days of strong solar heating and little wind display large diurnal variations in surface temperature. Observations of daytime temperature changes and of SW give the sum of the daytime loss terms in (6), \( (L^D + A^D + R^D) \). Since there is no solar heating at night the measured heat content change must reflect the sum of the nighttime loss \( (L^N + A^N + R^N) \). In contrast to the Equatorial Pacific, the tendency during the spring and summer in the Sargasso Sea appears to be for \( B \) in (8) to be negative.
Figure 10 shows the average profile from the drifting thermistor chains of

\[ \Delta T_i(z) = (\Delta t^D)^{-1} (T(2) - T(1)) - (\Delta t^N)^{-1} (T(3) - T(2)) \]  

(11)

where \( \Delta t^D \) and \( \Delta t^N \) are the length of day and night, respectively. Only 120 buoy days, where the diurnal SST warming at a buoy was greater than the overall average (0.37\(^\circ\)C), are included in the Fig. 10 average. The average SW over these days was likely greater than the average SW at LOTUS over all the days (435 W/m\(^2\)). Even so, when this SW value is used in (8) along with \( \Delta H_i = 300 \) W/m\(^2\) from the profile in Fig. 10 to \( h = 10 \) m, a negative value of \( B = -150 \) W/m\(^2\), results. Further work is needed to determine if this result is significant relative to the uncertainties and if so, to isolate which of the loss terms is responsible for the bias.

![Graph](image)

**Fig. 10** Average profile from drifting thermistor chains of \( \Delta T_i \) (11) over a total of 120 buoy days when the diurnal sea surface temperature warming was greater than its overall average (0.37\(^\circ\)C).

DISCUSSION

The advent of WOCE and TOGA has meant that more attention is being paid to the ocean basin and global scales than ever before. These programs require ocean data over very large scales, relatively frequently. It seems likely that fields of the
surface forcing functions will be produced routinely. However, a great deal of effort is still needed to obtain the necessary accuracies and coverage. Central to this activity are satellite observations whose value, although probably considerable, has yet to be fully demonstrated; however, their range of applications extends beyond the forcing functions.

Satellite altimetry promises to provide routine monitoring of global sea level; relative on its own and absolute in areas with sufficient supplementary data. Simulation experiments with eddy resolving ocean general circulation models indicate that TOPEX–like altimetric measurements will be extremely powerful in constraining models to behave properly. Even the lowest layers of a quasigeostrophic model feel the effects of the altimetric data, if properly assimilated, within a month or so (W. R. Holland, personal communication, 1987). Altimetry will also be able to track features of the turbulent mesoscale eddy field and of planetary waves.

Monitoring the upper ocean is particularly challenging, because of the numerous processes at work there, and its relatively rapid response to atmospheric forcing, which necessitates frequent sampling. Results from the foregoing examples of equatorial waves, thermocline mixing, and the daily turbulence cycle are encouraging and rewarding. It would be exciting and important if other upper ocean processes could be assessed globally using relatively easily obtained data. Some possible examples are subduction, Ekman advection, and Langmuir circulation.

REFERENCES


