The response of the upper ocean to solar heating. II: The wind-driven current

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(Received 28 February 1985; revised 30 July 1985)

SUMMARY

The current profile generated by a steady wind stress is disturbed by the diurnal variation of mixed layer depth forced by solar heating. Momentum diffused deep at night is abandoned to rotate inertially during the day when the mixed layer is shallow and then re-entrained next night when it deepens. The resulting variation of current profile has been calculated with a one-dimensional model in which power supply to turbulence determines the profile of eddy viscosity. The resulting variations of current velocity at fixed depths are so complicated that it is not surprising that current meter measurements have seldom yielded the classical Ekman solution. However, the progressive vector diagrams do exhibit an Ekman-like response (albeit with superimposed inertial disturbances) suggesting that the model might be tested by tracking drifters designed to follow the flow at fixed depths. The inertial rotation of the current in the diurnal thermocline leads to a diurnal jet, the dynamical equivalent of the nocturnal jet in the atmospheric boundary layer over land. The role of inertial currents in deepening the mixed layer is clarified, leading to proposals for improving the turbulence parametrizations used in models of the upper ocean. The model predicts that the diurnal thermocline contains two layers of persistent vigorous turbulence separated by a thicker band of patchy turbulence in otherwise laminar flow.

1. INTRODUCTION

Ekman’s (1905) theory for the current profile generated by a steady wind stress, assuming a constant eddy viscosity $K$ independent of depth, yielded the famous spiral which has a scale depth $L_E = \pi (2K/f)^{1/2}$, where $f$ is the Coriolis frequency. The steady current has 4% of its surface speed at the Ekman depth $L_E$, which has a value of 89 m for $f = 10^{-4}$s$^{-1}$ and $K = 0.04$ m$^2$s$^{-1}$. The validity of Ekman’s theory has been confirmed by measurements of the current profile in a vigorously and (compared with $L_E$) deeply convecting layer under drifting ice by Hunkins (1966) and McPhee (1980). But attempts to observe an Ekman spiral in the open ocean have not met with much success. The problem must be that the conditions assumed in Ekman’s theory are not met. The wind stress is not sufficiently steady, or the eddy viscosity is either not steady or not independent of depth. Csanady and Shaw (1980) have investigated the effect of change in wind stress. This paper shows that changes in the profile of $K$ induced by the daily cycle of solar heating lead to significant deviations from Ekman’s spiral at all latitudes and seasons.

It has long been recognized (e.g. Richardson 1920) that turbulence is weakened (reducing $K$) in a stable density gradient. More recently it was discovered (Woods 1969) that the density gradient in the seasonal thermocline is sufficient to quench the turbulence generated by the wind stress, leaving laminar flow and a negligible $K$. The Ekman current is therefore constrained to flow above the seasonal thermocline, which lies on average at a depth (<50 m) too shallow to permit the development of a classical Ekman spiral. Gonella (1971) published an analytical solution for the Ekman problem (steady wind stress) with $K$ equal to a constant value above a fixed depth $H$ and zero below. Our model was tested by demonstrating that it yields Gonella’s result under those conditions. The work presented in this paper may be thought of as the logical extension of Gonella’s with the values of $K$ and mixed layer depth $H$ varied in a way consistent with the diurnal cycle of solar heating. In our study the temporal variation of $H$ is prescribed; changes in the current profile do not influence $H$, although they do influence $K$.

It is appropriate to refer briefly to more ambitious studies by other authors who have developed models in which the current, density and turbulent viscosity and diffusivity...
profiles all interact in response to surface momentum and buoyancy fluxes. The fully-coupled problem is nonlinear and cannot be solved analytically. It requires a computer model. Munk and Anderson (1948) pioneered the subject using an analogue computer to calculate the current and density profile using parametrizations of the eddy viscosity and (density) diffusivity in terms of Richardson number which had been deduced by Taylor (1931) from Jacobsen's (1913) measurements of salinity and current measurements in the Kattegatt pycnocline. A number of similar parametrizations were proposed (see Bowden 1962), but they all had to be abandoned when it was discovered that turbulence in the ocean is quenched when the Richardson number exceeds a critical value close to unity (Woods 1969). Taylor's interpretation of the Kattegatt data had been wrong. Subsequent parametrizations have assumed quenching at a critical Richardson number (e.g. Mellor and Durbin 1975). Following procedures used in modelling the atmospheric boundary layer (Mellor and Yamada 1974) the tendency in recent years has been to use so-called 'higher-order closure' parametrizations with constants determined in the laboratory rather than from field data (see the review by Mellor and Yamada (1982)). The models of Worthem and Mellor (1980) and Klein and Coantic (1981) exemplify the application of such parametrization to models of the wind-generated current profile and the density profile in the upper ocean. Such studies have focused almost exclusively on predicting the changing temperature profile, the surface temperature and the depth of the mixed layer. Of particular significance for the present work was the emphasis by Simpson and Dickey (1981) on the need for care in parametrizing the solar heating profile in such models, but they, too, concentrated on the temperature profile, rather than the current profile. It is surprising that these models, which it is claimed do such a good job of simulating the atmospheric and presumably therefore the oceanic boundary layer, have not been used to examine the response of the Ekman current profile to the diurnal cycle of heating and cooling. We considered using a higher-order closure model, but concluded that our objectives could be achieved more economically and just as effectively by the simpler approach described below.

2. Theory

The aim of our study was to determine the response of the Ekman current to the diurnal changes in the mixed layer calculated in part I (Woods and Barkmann 1986). The principal simplification in the present study is that the mixed layer depth is not affected by the turbulence in the diurnal thermocline generated by Reynolds stresses acting on the wind-driven current. The point will be discussed later.

For a steady wind stress, surface heat flux and cloud cover, the model predicts a significant diurnal variation of mixed layer depth in response to the diurnal variation of solar elevation. An example, based on Bunker's (1976) monthly-mean surface meteorology at 41°N 27°W, is shown in Fig. 1. That diurnal variation of mixed layer depth $H$ and turbulent power input $E$ was used as prescribed mixed layer properties for our study of the Ekman current.

The value of $K$ in the seasonal thermocline ($z > H_{max}$) can drop to the value of the molecular viscosity; essentially Gonella's (1971) assumption. The rare billow turbulence mixing events in the seasonal thermocline due to internal wave-induced shear instability (Woods 1968) yield a mean energy dissipation rate ($\varepsilon$) that is a thousand times smaller than that in the mixed layer (Osborn 1980) and make a negligible contribution to the vertical flux of momentum.

The value of $K$ depends on turbulence sustained by the Reynolds stresses associated with the current shear; and is given by the Richardson law (Hinze 1977):
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\[ K = 0.1 \epsilon^{1/3} h^{4/3} \]  
(1)

where the power input to the turbulence is calculated from the current shear

\[ \epsilon = K (\partial V / \partial z)^2 \]  
(2)

and the vertical scale of the turbulence is put equal to the Ozmidov (1965) scale:

\[ h = (\epsilon / N^3)^{1/2} \]  
(3)

or to the mixed layer depth \( H \), whichever is the lesser. By combining (1) with (2) and (3) we obtain

\[ K_i = 0.1 K_{i-1} \{ (\partial V_{i-1} / \partial z) / \sqrt{N_{i-1}} \}^2 = 0.1 K_{i-1} (Ri)_{i-1}, \]  
(4)

\( i \) denoting the time step. The Brunt–Väisälä frequency \( (N) \) in the diurnal thermocline increases during the day owing to the absorption of solar energy. The change was calculated by the Woods–Barkmann model on the assumption of no turbulent heat flux in the diurnal thermocline. The diurnal thermocline mean value \( \overline{N} \) used in Eq. (3) was based on vertically averaging the profile in the depth range \( H < z < H_{\text{max}} \) at every time step, so that \( \overline{N} \) varies with time but not with depth. We shall see later (Fig. 4) that in practice the value of \( K \) in the diurnal thermocline is seldom significantly greater than zero. Computation with \( K = 0 \) in the diurnal thermocline gives essentially the same current profile.

3. THE MODEL

The coordinate system has \( x, y, z \) positive to the west, north and downwards respectively. The one-dimensional Reynolds equation (5) was integrated numerically using a Crank–Nicolson scheme with 0.1 m depth interval and 5-minute time step:

\[ \mathbf{v} = \mathbf{u} + i \nu = - \frac{i}{f} \left[ \frac{\partial \mathbf{v}}{\partial t} - \frac{\partial}{\partial z} \left( (K + \nu) \frac{\partial \mathbf{v}}{\partial z} \right) \right] \]  
(5)

where \( \nu \) is the molecular viscosity. The surface boundary condition is given by

wind stress = \( \tau = - \rho (K + \nu) \partial \mathbf{v} / \partial z \) = constant.  
(6)

Figure 1. Diurnal variation of mixed layer depth in May at 41°N 27°W calculated from Bunker's climatological mean surface meteorology by Woods and Barkmann (1986). The model described in this paper assumed a sawtooth variation between the same diurnal minima and maxima.
The bottom boundary condition is given by
\[ \mathbf{V}(t, z_{\text{max}}) = 0; \quad z_{\text{max}} = 100 \text{ m}. \]  
(7)
The initial conditions are an ocean at rest
\[ \mathbf{V}(t \leq t_0) = 0. \]  
(8)
The vertical profile of Brunt–Väisälä frequency is given by
\[ N = \begin{cases} \frac{N(t)}{z}, & H < z \leq H_{\text{max}} \text{ (diurnal thermocline)} \\ N_0 \exp\left(\frac{(H_{\text{max}} - z)/10}{m}\right), & z > H_{\text{max}} \text{ (seasonal thermocline).} \end{cases} \]  
(9)
The transitions at depths \( H \) and \( H_{\text{max}} \) were smoothed over a vertical distance of 2 m, in order to overcome errors encountered in testing the model (see Strass 1983).

The vertical scale of mixing is given by
\[ \begin{align*}
\text{mixed layer (} z \leq H \text{):} & \quad h = H \\
\text{thermoclines (} z > H \text{):} & \quad h = (\varepsilon/N)^{-3/2} \text{ or } H, \text{ whichever is smaller.} 
\end{align*} \]  
(10)
The boundaries of the diurnal thermocline are prescribed with \( H_{\text{max}} \) constant and \( H \) following a sawtooth variation.

(a) Model tests with no diurnal variation

The model was tested by calculating the current profile in three cases for which we have analytical solutions, namely:

(1) Ekman's (1905) problem: \( K \) constant and uniform down to more than twice the scale depth;

(2) Krauss's (1973) problem: \( K \) constant and uniform between the surface (\( z = 0 \)) and a solid bottom (\( z = Z \)), with a no-slip boundary condition;

(3) Gonella's (1971) problem: \( K \) constant and uniform in a mixed layer of constant depth \( H_{\text{g}} \) overlying a thermocline in which \( K = 0 \), with a free-slip boundary condition.

The Ekman transport for all the above cases should be directed 90° to the right of the wind stress (in the northern hemisphere) and have a magnitude \( \tau/f \).

The initial condition in all tests was \( \mathbf{V} = 0 \) as indicated in (8), an ocean at rest, so the results exhibit a transient oscillation of the form described by Fredholm's theory (given in Ekman (1905)). The results are illustrated in Fig. 2 for test (1), Fig. 3 for test (2). Table 1 shows the Ekman transport for various values of constant thermocline depth

<table>
<thead>
<tr>
<th>Depth of mixed layer ( H ) (m)</th>
<th>Model-calculated mass transport divided by ( \tau/f )</th>
<th>Deviation to the right of the wind (deg.)</th>
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<tbody>
<tr>
<td>70</td>
<td>1.000</td>
<td>89.54</td>
</tr>
<tr>
<td>50</td>
<td>1.000</td>
<td>89.58</td>
</tr>
<tr>
<td>30</td>
<td>1.000</td>
<td>89.64</td>
</tr>
<tr>
<td>10</td>
<td>1.003</td>
<td>89.61</td>
</tr>
<tr>
<td>5</td>
<td>1.007</td>
<td>89.50</td>
</tr>
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in test (3). Further illustrations are available in Strass (1983). We conclude that the model accurately reproduces current profiles determined analytically for a steady profile of $K$.

(b) Standard conditions for runs with diurnal variation

The results to be presented in the next section were obtained from four-day integrations of the model with a steady wind stress equivalent to a $10\,\text{ms}^{-1}$ south wind and a mixed layer depth varying linearly between 50 m at 06 h and 20 m at 12 h. The stratification in the diurnal thermocline ($H < z < 50\,\text{m}$) was equivalent to a temperature gradient increasing linearly with time during the forenoon from 0 to $10\,\text{mK/m}$. The variations of $H$ and $\bar{N}$ are shown in Fig. 4.

4. Results

The model was integrated at $41.7^\circ\text{N}$ (inertial period 18 h) for four days under the standard conditions for the diurnal variation of $N(z)$ described above.

(a) Turbulence

The variations in the profiles of Brunt–Väisälä frequency, power input to turbulent kinetic energy, length scale of the turbulence, eddy viscosity and absolute current are illustrated in Fig. 4. Remember that, according to Eqs. (1), (2) and (3), the values of $h$, $\varepsilon$ and $K$ depend on the current shear, but that $h$ is prescribed in the mixed layer. The strong shear at the top and bottom of the diurnal thermocline gives local maxima in the power supply to the turbulence, as predicted by Kitaigorodskii (1979), who assumed that the maximum possible fraction (0.15 according to Ellison (1957)) of the turbulent power supply from inertial current shear at that level would be consumed by turbulent entrainment of water from the seasonal thermocline, i.e. by increasing $H_{\text{max}}$, (which is not allowed to occur in our model integrations). Pollard et al. (1973) have also argued that the seasonal thermocline may be rapidly entrained into the mixed layer when changes in the wind stress vector generate vigorous inertial motion, but their model did not address the problem of changes in $\varepsilon$ in the diurnal thermocline. The results of our model show that turbulence is generated vigorously by the vertical shear in the inertial jet at the base of the diurnal thermocline. The shear is large provided the top of the thermocline ($H_{\text{max}}$) is shallower than the Ekman depth ($L_{E}$). In winter when $H_{\text{max}} \geq L_{E}$ the diurnal
Figure 4. Profiles over a period of 4 days showing: (a) Brunt–Väisälä frequency; (b) energy dissipation rate; (c) turbulence length scale; (d) eddy viscosity; (e) absolute velocity. The numbers in (e) indicate local time.
thermocline has no shear maximum at the base of the diurnal thermocline, and no
turbulence will be generated to affect entrainment of the seasonal thermocline. This
result provides the best rationale for assuming that the entrainment parameter decreases
exponentially as $H/L_E$ (Wells 1979). It suggests a better parametrization, namely to take
account of the effect of the inertial jet on entrainment by reducing the entrainment
parameter as $H_{\text{max}}/L_E$. However, it will still be necessary to take account of the increasing
fraction of turbulence consumed by viscous heating as $H$ increases, i.e. a decrease of the
slab flux Richardson number ($R_f$) as the mixed layer deepens. The two effects can be
combined in a parametrization of the form

$$m = R_f \cdot \exp(-H_{\text{max}}/L_E) \cdot \exp(-H/H_0)$$  \hspace{1cm} (11)

where $H_0$ is a scale length to be determined empirically.

The inertial current shear maxima at the top and bottom of the diurnal thermocline
produce layers of vigorous turbulence a few metres thick that persist for several hours.
Such layers have been noted in flow visualization studies (Woods and Fosberry 1966).
The power input to the turbulence in the remainder of the diurnal thermocline is very
much weaker, and decreases as solar heating stratifies the water. Eventually the power
supply falls below the critical value necessary to sustain continuous turbulence. In other
words, continuous turbulence would be impossible with the calculated power input and
density gradient because the energy-containing eddies (i.e. at the Ozmidov scale) have
too low a Reynolds number (Stillinger et al. 1983). We know from flow visualization
studies that when the mean power supply falls below that critical value the character of
the flow changes from continuously to intermittently turbulent, with short-lived billow
turbulence events in a generally laminar flow. The power supply to the individual mixing
events remains large enough to give a relatively high Reynolds number for the energy-
containing eddies (of order 100 to 1000).

The model results are consistent with the conceptual model derived from flow
visualization studies. In principle it should be possible to use the model to predict the
timing of the transition from continuous to intermittent regimes, given the initial current
profile and temporal variation of wind stress and surface meteorology.

(b) The diurnal jet

The absolute velocity profiles in Fig. 4 suggest the formation of a current speed
maximum every day with a near duplication of the profile at 18 h on the first and fourth
day. The recurrence is the result of beating between the 18 h inertial period and the 24 h
solar heating cycle. The development of the current maximum in the diurnal thermocline
is revealed in greater detail in Fig. 5. The top panels show profiles of Brunt–Väisälä
frequency (which is prescribed) and the absolute current calculated by the model at 90-
minute intervals. The maximum speed occurs in the diurnal thermocline during the late
afternoon and early evening. The bottom panels show the variation of the current vector
with depth at the same 90-minute intervals. It is seen that the current profile varies
relatively little in the top 20 m, which is always mixed (according to the standard
conditions used in the model runs reported here). Inside the diurnal thermocline, the
current rotates inertially with little change in the current shear. There is a large current
shear at the top and bottom of the diurnal thermocline. This diurnal jet is not a precisely
repeated feature. Even if the wind stress is constant, as in our model runs, there is the
regular beating between the inertial and solar periods referred to above. The jet is
triggered each day by the forenoon rise of the diurnal thermocline. The initial details of
the current profile also depend on the time of day at which the model integration is
started.
Figure 5. Profiles of: (a) Brunt–Väisälä frequency; (b) the corresponding absolute velocity at local times in the first day since the onset of wind. The profiles of the current vector, (c), are taken out of the period of the jet occurrence.
There is an obvious dynamical similarity between the diurnal jet that our model predicts to occur in the diurnal thermocline and the nocturnal jet observed in the atmospheric boundary layer on clear nights (Thorpe and Guymers 1977). In both cases, the Ekman profile is disturbed by stratification established radiatively following the daily quenching of buoyant convection. In the ocean the rising sun is responsible for the quenching of convection and the water column is stabilized by absorption of solar radiation. In the atmosphere (over land), convection is strong during the day while the sun warms the ground, but dies as the sun goes down, leaving long-wave radiation to dominate the vertical heat flux and stratify the air column. In both cases, ageostrophic flow arising because of the surface wind stress is linked to the boundary during some part of the diurnal cycle and isolated from it (when $K$ becomes negligible) at another. During the latter interval of time the flow rotates inertially, giving rise to the jet. The inertially-rotated momentum is subsequently re-entrained into the mixed layer when the convection is re-established, deepening the latter. It then imposes a shock to the mixed layer, so that the velocity profile can never settle down to a regular diurnal cycle, even with a constant wind stress.

(c) General remarks about the velocity profile

We remarked earlier on the failure of experiments designed to detect the Ekman spiral in the ocean. That is no longer surprising in view of the results presented above, which show that solar heating so changes the profile of $K$ that the wind-driven current cannot settle down to a steady state even if the wind stress is constant.

Figure 6 shows the temporal variation over four days of the current vector at constant depth, such as might be recorded by a moored current meter (assuming no internal wave undulations). The penetration of momentum into the seasonal thermocline (the top of which lies nominally at a depth of 50 m) arises from the smoothing of the $N$ profile described earlier (see Fig. 4).

It is common practice for oceanographers to present current meter records as progressive vector diagrams, in which the displacement vector $X = ix + jy$ is given by

$$X(t) = \int_{t_0}^{t_0+\tau} V \, dt.$$  

Progressive vector diagrams for four days are given in Fig. 7. The displacement vectors describe the trajectories of drifters constrained to follow the motion at a fixed depth. Such an experiment was performed by Saunders (1980) during the JASIN expedition (Charnock and Pollard 1983). Alas, the results do not provide a sensitive test of our model because the drifters could only be followed during the night, the acoustic tracking system did not work during the day owing to sound refraction by the diurnal thermocline—the 'afternoon effect' of naval acoustics. Note that Fig. 7 represents only a 'Lagrangian' view of the wind-driven current for particles constrained to stay at constant depth. In practice particles will be mixed vertically giving them a mean displacement 90° to the right of the wind stress.

(d) Ekman transport

Earlier it was demonstrated that the model accurately calculates the Ekman transport when the $K$ profile is kept constant. It is less easy to specify the equivalent test when there is a diurnal thermocline in which the motion rotates inertially before being re-entrained into the mixed layer during the night. Vertically integrating the current profile at any instant will clearly not yield a transport with magnitude equal to $|\tau|/f$ and directed
Figure 6. Hodographs taken from different depths. The dots are time marks, the number to the left of the slash is the day, the number to the right denotes local time.
90° to the right of the wind stress. It is necessary to calculate an average over an appropriate interval. The inertial or diurnal periods seem appropriate candidates. The results of calculating the mean transport are given in Table 2.

The results are based on a regular cycle of mixed layer depth. In spring, it decreases systematically (see Woods and Barkmann 1986) and some of the momentum subducted into the diurnal thermocline each morning is not re-entrained next night. This represents a leakage of momentum from the Ekman current, and a source of inertial wave energy in the thermocline.

| TABLE 2. Ekman transport averaged over diurnal and inertial periods compared with $|r|/f$. |
|---|---|---|---|
| Latitude (deg.) | Inertial period (days) | Average for the last inertial period of a 4-day run | Average for the last diurnal period of a 4-day run |
| | | Mass transport, divided by $|r|/f$ | Deviation to the right of wind (deg.) | Mass transport, divided by $|r|/f$ | Deviation to the right of wind (deg.) |
| 85.7 | 0.50 | 0.926 | 89.64 | 1.015 | 89.43 |
| 41.7 | 0.75 | 0.975 | 84.26 | 0.913 | 73.90 |
| 29.9 | 1.00 | 0.905 | 85.53 | 0.905 | 85.53 |
| 23.5 | 1.25 | 0.928 | 88.21 | 0.906 | 88.91 |
| 19.4 | 1.50 | 0.977 | 83.82 | 1.127 | 72.31 |
| Mean | | 0.962 | 86.17 | 0.973 | 82.02 |
| r.m.s. deviation | | 0.047 | 2.34 | 0.098 | 8.29 |

Difference from mass transport under stationary mixing depth of 30 m

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<td>0.038</td>
<td>3.47</td>
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5. DISCUSSION

The aim of this paper has been to clarify the response of the Ekman current to solar heating, as a phenomenon in physical oceanography. It has not been our intention to simulate the actual current profile at any particular location and occasion. The model used in our study is unlikely to be able to produce accurate simulations, not because its ingredients are wrong in principle, but rather because we have deliberately simplified those details for the sake of clarity, and because the model leaves out processes that are likely to affect the current profile in the ocean. The model simplifications have been discussed in detail earlier. They include the neglect of: (1) the current shear on the variation of the mixed layer depth; (2) turbulent entrainment of the seasonal thermocline. We acknowledge these deficiencies, but do not expect them to have a large influence on our results. Davis et al. (1981) had success with a similar model in simulating observations of mixed layer depth and temperature in response to changes in the wind.

The principal cause of uncertainty lies in the profile of $K$, which depends on the density profile. We have used the Woods–Barkmann model results. They have drawn attention to the absence of a data set that can test models of the diurnal thermocline, in the sense of discriminating decisively between the quite different parametrization schemes used by different authors. Budgets of the diurnal heating cycle cannot be closed with sufficient accuracy to permit such tests because of errors in the estimates in surface fluxes and of horizontal flux divergence (see, for example, Stommel et al. 1969; Howe and Tait 1969).

Nevertheless, it would be reassuring to be able to refer to a data set that confirmed some of the more striking results of our modelling study, in particular the occurrence of a diurnal jet. Price, Weller and Pinkel (1985) have measured the temperature and current vector profile from the drifting research platform 'Flip'. They too found that the change in heat content of the top 50 m of the ocean was much larger than expected from the surface fluxes "presumably because of advection", and significant contribution of salinity on some days. Equally, "the low frequency motion (relative to the drifting platform) plus tides and inertial motions largely obscured the diurnal cycle of wind-driven velocity". Nevertheless, the current in the diurnal thermocline was observed to rotate independently of that in the mixed layer, as predicted by our model. (However, the observed rotation was not purely inertial.) Also, the current in the mixed layer was observed to have a maximum speed around noon as it appears in our model results, the slippery seas effect identified by Woods and Houghton (1969). As with most case studies, too many factors were acting at once.

We are not aware of any other data that can be used to test our model predictions. The Lagrangian technique of Saunders (1980) is particularly attractive and we hope it will be repeated (with day and night tracking) in the future.

6. CONCLUSION

It has been shown with the help of a simple one-dimensional model that diurnal variation of the mixing caused solely by solar heating produces a jet in the diurnal thermocline when there is a steady wind stress. This is the dynamical equivalent of the nocturnal jet that occurs in atmospheric boundary layers. We, therefore, name it the diurnal jet of the upper ocean. It arises from inertial rotation of the current vector in the diurnal thermocline while the mixed layer becomes shallower following the quenching of convection by solar heating. The re-entrainment of the rotated momentum when the mixed layer deepens after sunset leads to a disturbance which prevents the current profile
from settling down to the steady form predicted by Gonella's (1971) theory. There is some recent experimental evidence of the diurnal jet, but field data are not yet adequate to provide a test of the assumptions in the model. The inertial shear maxima at the top and bottom of the diurnal thermocline lead to layers of vigorous turbulence that persist for several hours. Elsewhere in the diurnal thermocline the turbulence weakens during the day as solar heating stratifies the water; by the afternoon the mean power input to the turbulence drops below the critical value, opening up regions of laminar flow. These results are consistent with flow visualization observations and microstructure profiles showing layers of persistent turbulence in a largely laminar flow diurnal thermocline.

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