Prescriptions for Heat Flux and Entrainment Rates in the Upper Ocean during Convection

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ABSTRACT

A detailed investigation of the upper ocean during convection reveals

- the vertical structure of potential temperature, $\theta$, to be steady in time, and
- the current shear to vanish in the bulk of the mixed layer.

These imply that a "slab"-type model may be an adequate representation of the convective ocean boundary layer (OBL). In contrast, when convection is not the dominant forcing mechanism, the OBL is stratified and can support a significant current shear. This indicates the inadequacy of "slab" models for the nonconvective OBL.

Two independent estimates of the vertical heat flux profile in the convective OBL were made. The first estimate results from heat conservation and the steadiness of the vertical structure of potential temperature. The second estimate is based on the turbulent kinetic energy (TKE) balance and the vertical profiles of TKE dissipation rate. The estimates are consistent and suggest that the nondimensional vertical heat flux due to turbulence has a linear depth dependence of the form $1 + a_0(z/D)$, where $z$ is the depth, $D$ is the mixed layer depth, and $a_0$ is a constant with a mean value of 1.13, consistent with numerical and laboratory results and with observations in the convective atmospheric boundary layer. An estimate of the entrainment rate, derived from observed quantities, is $-1 \times 10^{-4} \text{ m s}^{-1}$. This is within a factor of 2 of estimates derived from alternative formulations.

1. Introduction

Experiments in the convective ocean boundary layer (OBL) indicate that the vertical structures of the time-averaged potential temperature, $\theta$, and TKE dissipation rate, $\epsilon$, in the convective OBL are similar to those in the convective atmospheric boundary layer (ABL) over land. Both systems include four distinct regimes (proceeding from the ocean's surface down and from the land's surface up):

1) A "cool skin" (a few millimeters thick) near the surface of the ocean (Khunzhua et al. 1977; Paulson and Simpson 1981) analogous to the "hot skin" (or microlayer) near the land surface (Stull 1988). Physical processes are governed by molecular diffusion; large vertical gradients of potential temperature are typical ($\partial \theta / \partial z \sim -300 \text{ K m}^{-1}$ for the OBL and $\partial \theta / \partial z \sim -1 \times 10^4 \text{ K m}^{-1}$ for the ABL).

2) A superadiabatic ocean surface layer (OSL) of $O(10-20 \text{ m})$ in which $\partial \theta / \partial z \sim -2 \times 10^{-4} \text{ K m}^{-1}$ (Anis and Moum 1992) and a superadiabatic atmospheric surface layer (ASL) of $O(10-100 \text{ m})$ in which $\partial \theta / \partial z \sim -1 \times (10^{-2} - 10^{-1}) \text{ K m}^{-1}$. In some cases $\epsilon$ in the OSL followed Monin–Obukhov similarity scaling (Soloviev et al. 1988; Anis and Moum 1994a), similar to what is observed in the ASL (Driedonks and Tennekes 1984; Stull 1988). However, an increasing number of field studies shows that mixing can be much more vigorous in the OSL with values of $\epsilon$ much larger than expected from Monin–Obukhov similarity scaling (Anis and Moum 1992, 1994a; Agrawal et al. 1992). The enhanced dissipation rates may result from several mechanisms unique to the OSL, such as surface wave breaking and surface wave stress (Anis and Moum 1994a), or Langmuir circulation.

3) A well-mixed layer (ML) in which $\partial \theta / \partial z \sim 0$ and $\epsilon$ scales well on average with the surface buoyancy flux, $J_b^\theta$. Mean values of $\epsilon / J_b^\theta$ are similar in both the ML of the convective OBL ($\epsilon / J_b^\theta = 0.44 \sim 0.87$: Shay and Gregg 1986; Lombardo and Gregg 1989; Anis and Moum 1992) and the convective ABL ($\epsilon / J_b^\theta = 0.64$: Caughey and Palmer 1979). Thicknesses of the ML are $O(50-150 \text{ m})$ and $O(1000-2000 \text{ m})$ for the OBL and the ABL, respectively.

4) A stable thermocline in the OBL of $O(100-200 \text{ m})$ and a stable inversion layer in the ABL of $O(100-500 \text{ m})$ where $\partial \theta / \partial z > 0$. Turbulence is intermittent.
and is generated by shear, Kelvin–Helmholtz instabilities, internal waves, and overshooting thermals (e.g., Gargett 1989 for the OBL and Stull 1988 for the ABL). This regime is often referred to as the entrainment zone.

The simplest representation of a convective BL is as a uniform “slab.” In ABL models, mean profiles of $\theta$, humidity (and other scalars), and wind speed are assumed to be constant with height with quasi discontinuities at the top of the ML (e.g., Stull 1988). Similarly, in OBL “slab” models, it is assumed that mean temperature, salinity, and horizontal currents are vertically homogeneous inside the “slab” with a discontinuity across the base of the ML (e.g., Niler and Kraus 1977). Since the vertical structure of $\theta$ is assumed to remain constant with time, the heat flux is a linear function of the vertical coordinate $z$, decreasing from its effective surface value, $w^i\theta^i(0)$ (defined just beneath the oceanic “cool skin”, or just above the atmospheric microlayer), to its value near the top/base of the ML, $w^i\theta^i(D)$. (Turbulence fluxes of quantities such as humidity and pollutants in the ABL and salinity in the OBL will similarly be linear functions of $z$.) Linear heat flux profiles were observed in laboratory studies (Willis and Deardorff 1974; Deardorff et al. 1980) and field experiments in the convective ABL (e.g., Zhou et al. 1985; Young 1988) and in numerical simulations (e.g., Deardorff 1974; André et al. 1978). André and Lacarrère (1985), using third-order numerical simulations of a buoyancy driven OBL, have similarly demonstrated the linear decrease of buoyancy flux with depth. A field experiment in the convective aquatic boundary layer of a freshwater reservoir inferred a linear heat flux profile with depth (Imberger 1985; in this instance the system was assumed to be purely 1D and the heat flux was estimated directly from the time rate of change of temperature of the water column).

Closure of the ML equations requires knowledge of the entrainment processes at the top/base of the ML. Unfortunately, relatively little is known about the details of entrainment in the convective ABL (and even less so in the convective OBL), mainly due to the difficulty of making accurate measurements of the highly intermittent turbulence in the entrainment zone. Generally, three distinct processes are responsible for TKE production and entrainment in the convective ABL (Driedonks and Tennekes 1984): (a) convection due to surface heat/buoyancy flux, (b) surface wind stress, and (c) shear production due to wind shear at the top of the ML. Opposing the entrainment-producing processes in the ABL are the stable inversion layer at the top of the ML and viscous dissipation of TKE.

A widely used closure scheme of the ML equations of “slab” models is by parameterization of the entrainment rate, $w_e$. Due to the complicated nature of entrainment processes, no general parameterization exists that is valid for all cases (see Deardorff 1983 for a thorough treatment and comparison of entrainment formulations). Another closure method relates the heat/buoyancy flux at the top/base of the ML to the heat/buoyancy flux at the surface, such that $-w^i\theta^i(0)/w^i\theta^i(D) = A$, where $A$ is a constant. Most of the suggested values for $A$ in convective boundary layers, from theoretical, experimental, and laboratory studies are between 0.1 and 0.3, although values ranging from 0 to 1 have appeared (Stull 1976). This spread of values may be just another manifestation of the complicated processes involved.

In this study we examine the details of the vertical structure of $\theta$, $\epsilon$, and shear of the horizontal current as revealed by a large number of profiles obtained in the nighttime convective OBL. We focus our attention on the time period when convective forcing was relatively steady (the period during which the ML rapidly deepened is discussed by Anis and Moum 1994b). Throughout this quasi-steady convective forcing period (lasting 11.5–13 h) the vertical structure of $\theta$ was observed to remain essentially constant. Two conclusions are drawn from the steadiness of the vertical structure of $\theta$: 1) a slab-type behavior of the convective OBL and 2) a linear depth dependence of the heat flux. The first conclusion is further supported by the vanishing vertical shear of the horizontal currents in the bulk of the ML.

Using a simplified TKE balance at the base of the ML ($z = -D$), we have estimated $w^i\theta^i(-D)$. Based on these estimates the ratio $-w^i\theta^i(0)/w^i\theta^i(-D)$ and the slope of the linear heat flux profile were calculated and found to be similar to values from laboratory, numerical, and ABL field studies. As mentioned above, the knowledge of $w^i\theta^i(-D)$ provides also a closure condition for the ML equations and enables one to estimate the entrainment rate. A second estimate of the heat flux profile was made using $\epsilon$ profiles and the assumption of a TKE balance between buoyancy flux and dissipation, $\nu_s(z) \sim \epsilon(z)$. This estimate suggests, independently from the estimate above, a linear heat flux profile with a slope that is consistent with the slope calculated from the ratio $-w^i\theta^i(0)/w^i\theta^i(-D)$.

Experimental details, meteorological background conditions, and observations are described in section 2, with a specific focus on the vertical structure of $\theta$, $\epsilon$, and shear. Two independent estimates of the vertical heat flux profile in the nighttime convective OBL are made in section 3 and compared to each other and to results from laboratory, field, and numerical studies. An estimate of the entrainment rate is made in section 4 and compared to several entrainment formulations. Summary and conclusions are presented in section 5.

2. Observations

a. Experimental details

A large dataset was collected in the upper OBL of the Pacific Ocean between 13 and 20 March 1987 as the R/V Wecoma progressed along 140°W at a mean speed of ~2.5 m s$^{-1}$ from 17°N to 6°N (Park et al.
of salinity at the ocean’s surface was less than 10%; \( J_0^* > 0 \) for convective forcing. The Monin–Obukhov length, \( L \), which during convective conditions defines the depth at which the wind stress and the surface buoyancy flux are of equal importance (such that for \( z \gg L \) wind stress is the main TKE source, while for \( z < L \) buoyancy production is the main TKE source; \( z > 0 \), upward and \( L < 0 \) during convection, as defined in Table 1), had mean nighttime values between \(-10.6 \) and \(-17.5 \) m (Table 1). These values are similar to those observed in other diurnal oceanic convective boundary layers (e.g., Shay and Gregg 1986; Lombardo and Gregg 1989). Sea state conditions were moderate throughout the experiment with swells with significant heights between 1.8 and 2.7 m and wind waves with significant heights between 0.6 and 0.9 m (Table 1).

c. The vertical structure of potential temperature, TKE dissipation, and velocity shear

Net heat loss from the ocean’s surface resulted in destabilizing convective conditions during each of the six nights of the experiment. During the initial phase of the convective period (about 2–3 h), when \( J_0^* \) increased nearly linearly to its nighttime quasi-steady value, the stable daytime stratification between the ocean’s surface and the seasonal pycnocline was eroded. The rate of ML deepening was between 8–19 m h\(^{-1} \) (Anis and Moum 1994b) and final nighttime ML depths, limited by the top of the seasonal pycnocline, were between 50.6 to 82.2 m (Table 2; ML depth, \( D \), was estimated from individual profiles as the depth at which the density exceeded the surface value by \( \sim 0.005 \) kg m\(^{-3} \), and then averaged for the night). During the major part of each night (lasting 11.5–13 h), \( J_0^* \) varied little and \( D \) traced the undulations in the depth of the seasonal pycnocline. We refer to this part of the night as the quasi-steady forcing phase; this phase is the focus of our analysis.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>1</th>
<th>2</th>
<th>3</th>
<th>4</th>
<th>5</th>
<th>6</th>
<th>Mean</th>
</tr>
</thead>
<tbody>
<tr>
<td>( r_0 ) (N m(^{-2} ))</td>
<td>0.08</td>
<td>0.10</td>
<td>0.08</td>
<td>0.07</td>
<td>0.07</td>
<td>0.11</td>
<td>0.11</td>
</tr>
<tr>
<td>( T_{air} - SST ) (K)</td>
<td>-1.84</td>
<td>-1.23</td>
<td>-1.53</td>
<td>-0.88</td>
<td>-0.73</td>
<td>-0.86</td>
<td>-0.86</td>
</tr>
<tr>
<td>( J_0^* ) (W m(^{-2} ))</td>
<td>212</td>
<td>211</td>
<td>171</td>
<td>166</td>
<td>205</td>
<td>199</td>
<td>189</td>
</tr>
<tr>
<td>( 10^3 J_0^* ) (m(^2) s(^{-1} ))</td>
<td>1.61</td>
<td>1.66</td>
<td>1.37</td>
<td>1.31</td>
<td>1.54</td>
<td>1.61</td>
<td>1.48</td>
</tr>
<tr>
<td>( L ) (m)</td>
<td>-10.6</td>
<td>-14.2</td>
<td>-11.7</td>
<td>-10.8</td>
<td>-17.3</td>
<td>-17.5</td>
<td>-13.9</td>
</tr>
<tr>
<td>( H_{sw} ) (swell m)</td>
<td>2.4</td>
<td>2.7</td>
<td>2.0</td>
<td>1.8</td>
<td>2.2</td>
<td>2.1</td>
<td>2.1</td>
</tr>
<tr>
<td>( H_w ) (wind waves m)</td>
<td>0.8</td>
<td>0.9</td>
<td>0.7</td>
<td>0.6</td>
<td>0.9</td>
<td>0.9</td>
<td>0.8</td>
</tr>
</tbody>
</table>

Table 1. Nighttime average values of meteorological parameters. The surface wind stress, \( r_0 \); the difference between the air temperature (measured from the R/V Wecoma mast at 8-m height) and the sea surface temperature (SST; measured from the ship’s hull-mounted Doppler thermistor at 5-m depth), \( T_{air} - SST \); the net surface heat flux, \( J_0^* \); the net surface buoyancy flux, \( J_0^* \) (positive for convective conditions); the Monin–Obukhov length scale, \( L = -u^* \sigma /e J_0^* \) (where \( u^* = \sqrt{\tau_{air} / \rho} \) is the ocean surface friction velocity and \( L < 0 \) during convection); the significant wave height, \( H_w \), for swell and wind waves (from R/V Wecoma ship’s officers’ log). Averages of the above quantities were calculated for the time period during which \( J_0^* \) was quasi-steady (starting about 2 h after \( J_0^* \) changed sign from negative to positive and ending about 2 h before \( J_0^* \) changed sign back).
Investigation of the vertical temperature structure, as a function of time, was carried out by first referencing individual profiles of $\theta(z)$ to the average value of $\theta$ in the ML (in the depth range $-D < z < 2L$) and then averaging over periods of 2 hours each during the quasi-steady convective forcing phase of each night. As an example, the envelopes of the 95% bootstrap confidence interval of 2-h averages of $\theta$, as functions of the scaled depth $z^* = z/D$ (calculated in depth bins of 0.04D), are presented in Fig. 1 for five sequential periods during night 4. (We used $z^*$ instead of $z$ since $D$ varied during the night as it followed the undulations of the top of the seasonal pycnocline, probably due to a combination of internal wave activity and local variability; this, of course, independent of mixed layer physics and hence we scale this variablity out of the problem.) The variability in $\theta(z)$ is restricted to $O(0.001)$ K in averaged profiles (Fig. 1). From this, we infer that the vertical structure of $\theta(z)$ does not change significantly during the quasi-steady convective forcing phase. Three distinct regimes are noticed from the profiles of $\theta$: a superadiabatic OBL, where $\partial \theta/\partial z < 0$, in the upper 20%–40% of the OBL (see also Anis and Moum 1992); a relatively well-mixed layer beneath it in which $\partial \theta/\partial z \sim 0$ (the lower part of the ML often had a small positive temperature gradient similar in magnitude to that of the superadiabatic OBL; we attribute this to entrainment of cooler water below the ML base into the ML); a stable layer with a sharp positive gradient of $\theta$ capping the OBL from below and coinciding with the top of the seasonal pycnocline. The vertical structure of $\theta(z)$ was found to be constant within 95% confidence limits throughout the quasi-steady convective forcing phase of each of the six nights of the experiment (beginning and ending profiles from the quasi-steady forcing phase of each night are shown in Figs. 2a–f).

The vertical structure of $\epsilon$ was nondimensionalized as $\epsilon^*(z^*) = \epsilon(z^*)/\epsilon^*_B$ for individual profiles and ensemble averaged in depth bins of 0.04D for each night (as we did for the profiles of $\theta$). The results are shown in Fig. 3 for each of the 6 nights of the experiment. A similar computation was made by ensemble averaging the individual profiles of all nights (Fig. 4). A consistent feature of the averaged profiles was a rapid and roughly exponential decrease of values of $\epsilon^*(z^*)$ in the upper 20%–40% ($\sim 1–2L$) of the convective OBL (Anis and Moum 1994a). In the lower half of the ML, roughly between $z^* = 0.5$ and $z^* = 0.9$, values of $\epsilon^*(z^*)$ decreased linearly with $z^*$. Before dropping rapidly by several factors of ten, a slight increase in $\epsilon^*(z^*)$ was observed near the base of the ML between $-z^* = 0.95$ and $-z^* = 1.05$. This is an indication of enhanced mixing in the entrainment zone, possibly a result of the enhanced shear levels observed near the base of the ML.

Horizontal current velocity measurements from the R/V Wiecoa hull-mounted ADCP were estimated every 4 m vertically, although they are independent approximately only every 12 m and interpolated in time to match the RSVP casts used in our analysis. The velocity shear was computed from the slopes of linear fits, over a 12-m depth interval, to the $U$ and $V$ velocity components, such that $S^2 = \left(\partial U/\partial z\right)^2 + \left(\partial V/\partial z\right)^2$. Analysis of nighttime profiles showed that the estimates of $S^2$ in the bulk of the ML were equivalent, on average, to the detection limit of $O[1–2 (\times 10^{-6} \text{s}^{-2})]$ in estimating $S^2$ from the ADCP velocity measurements (Fig. 5). That is, the ML shear is not distinguishable from 0 s$^{-1}$. In the lower 20% of the ML a relatively sharp increase in shear was observed with a maximum of about $S^2 \sim 6 \times 10^{-6} \text{s}^{-2}$ near the base of the ML.

To determine if the low shear values were characteristics for the upper OBL during nighttime only, we plotted the time series of hourly averages of $S^2$ estimated between 19.1 m and 31.1 m and centered at 25.1 m (19.1 m is the shallowest depth at which $U$ and $V$ could be estimated). A clear distinction between nighttime and daytime emerges from inspection of the time series of the squared buoyancy frequency, $N^2$ (Fig. 6b) and $S^2$ (Fig. 6c). Larger values of $S^2$ during daytime, compared to nighttime, are closely associated with the larger values of $N^2$ resulting from the increasing stability due to daytime heating.

To illustrate the difference between daytime and nighttime shear, two datasets, one for daytime and one for nighttime, were constructed from near surface values of $S^2$ (centered at a depth of 25.1 m). The resulting distributions of $S^2$ are presented in Fig. 7. Nighttime was characterized by a large number (almost 40%) of estimates of $S^2$ below the equivalent detection limit ($\sim 1 \times 10^{-6} \text{s}^{-2}$), while during daytime almost none (less than 1%) of the $S^2$ were smaller than the detection limit. The mean daytime value of $S^2$ was larger by a factor of about...
4 compared to the mean nighttime value. A standard Kolmogorov–Smirnov test (Press et al. 1992), applied to the nighttime and daytime datasets, showed that the distributions of the two datasets were significantly different at a confidence level better than 99.999%. An independent Student’s t-test and F-test confirmed, respectively, that the variance and mean of the two datasets were significantly different, again at a confidence level better than 99.999%; however, this is to be expected for substantially different distributions.

3. Estimates of the vertical heat flux profile during convection

a. Estimate from potential temperature profiles and heat conservation

Using Reynolds time averaging, the equation for the conservation of heat in the OBL can be written as

$$\frac{\partial \theta}{\partial t} + U_i \frac{\partial \theta}{\partial x_i} = \nu \frac{\partial^2 \theta}{\partial x_i \partial x_i} - \frac{\partial (u' \theta')}{\partial x_i} - \frac{\partial I}{\partial z}.$$  \hspace{1cm} (1)

where $\theta$, $U_i$, and $I$ are the time-averaged potential temperature, current velocity, and penetrating solar radiation, respectively; primed variables are turbulence components; $\nu$ is the thermal molecular diffusivity; and $z$ is positive upward. The mean molecular heat conduction term [first term on the rhs of (1)] is much smaller than any other term in the ML, away from the surface, and hence can be neglected. Also, during nighttime $I = 0$ and the last term on the rhs of (1) vanishes.

To isolate the effect of convection, we assume a purely 1D process such that the time rate of change of $\theta$ is solely due to the vertical turbulence heat flux. Equation (1) then simplifies to

$$\frac{\partial \theta}{\partial t} = - \frac{\partial (u' \theta')}{\partial z}.$$  \hspace{1cm} (2)
Integrating (2) vertically to the surface \((z = 0)\) and rearranging terms results in an equation for the vertical heat flux due to turbulence at depth \(z\),

\[
\overline{w'\theta'}(z) = \overline{w'\theta'}(0) + \int_{0}^{z} \frac{\partial \theta}{\partial t} \, dz'.
\]

Equation (3) then permits an estimate of the vertical heat flux at some depth \(z\) by vertically integrating the time rate of change of temperature between the surface and \(z\) and using the surface heat flux, \(J_o^q = \rho C_p w'\theta'(0)\) (estimated from shipboard meteorological measurements; Table 1). In practice this is almost always impossible to do using oceanic data because of the relatively small temperature changes resulting from convection compared to those due to lateral temperature variability. For example, using a difference form of (2) and assuming a kinematic heat flux difference between the surface and the bottom of the ML of \(\Delta w'\theta' \sim 5 \times 10^{-5} \text{ K m s}^{-1}\) (based on a surface heat flux of \(\sim 200 \text{ W m}^{-2}\) and a negligible heat flux at the base of the ML), a time period of \(\Delta t \sim 10 \text{ h}\) and a mean ML depth \(D = 65 \text{ m}\) results in a cooling of \(|\Delta \theta| = |\Delta (w'\theta') \Delta t / \Delta z| < 0.03 \text{ K}\). For comparison, sea surface temperature measurements indicated a warming each night of up to a few tenths of K (Anis and Moum 1994b; Fig. 1b). This is not only larger in magnitude than expected from convective cooling it is of
the wrong sign. Rather, the sea surface warming was associated with our southward passage and was further complicated by a series of sharp fronts.

As shown above, even if convection is solely responsible for the vertical structure it is still impractical to use (3) in oceanic applications. A way to alleviate this problem is by taking the vertical derivative of (1) (assuming a 1D process, neglecting the molecular term, and recalling that $I = 0$ during nighttime), which results in

$$
\frac{\partial}{\partial t} \left( \frac{\partial \theta}{\partial z} \right) = - \frac{\partial}{\partial z} \left( \frac{\partial (w' \theta')}{\partial z} \right),
$$

where the order of the $z$ and $t$ derivatives was changed.

From (4), the vertical structure of the turbulence heat flux can be found from the time rate of change of the temperature gradient, $\partial \theta / \partial z$. This can be accomplished more readily, since now we are not concerned with the absolute temperature differences between two profiles of $\theta(z)$, which are separated in both time and space, but only with the change in their relative vertical structure as a function of time. Because the latter does not change significantly throughout the quasi-steady convective forcing phase of each of the six nights (Fig. 2), we have $\partial (\partial \theta / \partial z) / \partial t \sim 0$. By (4), this implies that the vertical heat flux due to turbulence in the convective OBL is linear in $z$; that is,

$$
\overline{w' \theta'}(z) = \frac{\overline{w' \theta'}(0) + [\overline{w' \theta'}(0) - \overline{w' \theta'}(-D)]}{D} \frac{z}{D},
$$

or in dimensionless form

$$
\overline{w' \theta'^*}(z^*) = 1 + [1 - \overline{w' \theta'^*}(-1)]z^* = 1 + a_b z^*,
$$

where $\overline{w' \theta'}(0)$ and $\overline{w' \theta'}(-D)$ are the heat fluxes at the surface and the base of the ML, respectively, $\overline{w' \theta'^*}(z^*) = \overline{w' \theta'}(z) / \overline{w' \theta'}(0)$, and $a_b = [1 - \overline{w' \theta'^*}(-1)]$. Field experiments in the convective ABL (e.g., Zhou et al. 1985; Young 1988); numerical simulations (e.g., Deardorff 1974; André et al. 1978 for a convective ABL; André and Lacarrèrè 1985 for a convective OBL); laboratory experiments (e.g., Willis and Deardorff 1974); and a field experiment in the convective aquatic boundary layer of a freshwater reservoir (Imberger 1985) all indicate a linear dependence on $z$ of the turbulence heat/buoyancy flux.

To assess the expected uncertainty in the inference of a linear heat flux due to the possible variability with time of $\partial \theta / \partial z$, we use (4) to write the uncertainty in the vertical heat flux divergence as $\Delta (\partial (w' \theta') / \partial z^*)$.
\[ D(\Delta z^*)/\Delta t \Delta(\partial \theta/\partial z^*) \]. Suppose \( \theta \) is homogeneous rather than superadiabatic toward the sea surface; this represents an upper bound to the variability between profiles made over the course of a single night (Fig. 2). Using mean values from our experiment (\( D \sim 65 \) m, \( \Delta z^* \sim 0.5 \), \( \Delta t \sim 8 \) h, and \( \Delta \partial \theta/\partial z^* \sim 0.01 \) K), \( \Delta(\partial (w^* \theta^*)/\partial z^*)/\Delta z^* \sim 1 \times 10^{-5} \) K m s\(^{-1}\). Since this is only about 20% of the total change in the heat flux in the
ML, assuming that it decreases from its surface value \((\sim 5 \times 10^{-5} \text{ K m s}^{-1})\) to some small value near the base of the ML, we conclude that the expected changes in \(\partial \theta / \partial z\) are small enough not to upset the inference of a linear heat flux profile.

An estimate of the upper limit on the magnitude of \(\overline{w'\theta'}(-D)\) [or \(\overline{w'\theta'}^\ast(-1)\)] can be made assuming a simplified form of the steady state TKE equation near the base of the ML in which the mechanical production of TKE, \(P\), is balanced by TKE dissipation and by buoyancy destruction, \(J_b\); namely, \(P = -J_b + \varepsilon\). The small, observed increase in \(\varepsilon\) at \(-z/D \sim 1\) (Figs. 3 and 4) and the increase in shear magnitude near the base of the ML (Fig. 5) seem to support such a balance during our experiment. Evidence for a balance of this type at the base of the oceanic ML is provided by the numerical simulations of a buoyancy-driven OBL by André and Lacarrère (1985), while observational evidence for a similar balance at the top of the convective ABL comes from a field study discussed by Zhou et al. (1985). Using the simplified TKE balance, defining a flux Richardson number as \(R_f = -J_b/P\), and using a critical value for \(R_f\) of 0.15 (Osborn 1980) results in an upper bound on the magnitude of the buoyancy flux given by

\[
|J_b(-D)| \leq 0.2 \varepsilon(-D). 
\]

Neglecting salinity effects, the heat and buoyancy fluxes are related through \(J_b = g\alpha \overline{w'\theta'}\) \([g is the gravitational acceleration and \alpha is the thermal expansion coefficient of seawater with \(\varepsilon = 2.9 - 3.2 \times 10^{-4} \text{ K}^{-1}\) for our experiment]) so that

\[
|\overline{w'\theta'}(-D)| = \frac{|J_b(-D)|}{g\alpha} \leq 0.2 \varepsilon(-D),
\]

or in dimensionless form

\[
|\overline{w'\theta'}^\ast(-1)| \leq 0.2 \varepsilon^\ast(-1),
\]

where \(\varepsilon^\ast(-1) = \varepsilon(-D)/J_f^\ast\). Nighttime means of \(\varepsilon^\ast(-1)\) varied between 0.36 and 0.90 with an average value of 0.63 (Table 2). Substitution of these values

\begin{table}[h]
\centering
\begin{tabular}{|c|c|c|c|c|c|}
\hline
 & 1 & 2 & 3 & 4 & 5 & 6 & Mean \\ \hline
\(D\) (m) & 64.4 & 68.5 & 50.6 & 67.5 & 64.3 & 82.2 & 65.1 \\ \(10^5 \varepsilon(-D)\) (m$^2$ s$^{-3}$) & 0.56 & 1.14 & 0.79 & 0.62 & 1.36 & 1.15 & 0.63 \\ \(\varepsilon^\ast(-1)\) & 0.36 & 0.65 & 0.61 & 0.47 & 0.90 & 0.67 & 0.13 \\ \(\varepsilon^\ast(z^\ast)\) & 0.28 & 0.44 & 0.54 & 0.69 & 0.42 & 0.53 & 0.58 & 0.68 \\ \(10^5 \overline{w'\theta'}^\ast(-1)\) & 0.07 & 0.13 & 0.12 & 0.09 & 0.18 & 0.13 & 0.13 \\ \(a_n\) & 1.07 & 1.13 & 1.12 & 1.09 & 1.18 & 1.13 & 1.13 \\ \(b_n\) & 0.30 \(\pm\) 0.10 & 1.26 \(\pm\) 0.13 & 1.50 \(\pm\) 0.14 & 1.25 \(\pm\) 0.08 & 0.32 \(\pm\) 0.13 & 0.57 \(\pm\) 0.07 & 0.93 \(\pm\) 0.05 \\ \(b_n\) & 0.66 \(\pm\) 0.07 & 1.64 \(\pm\) 0.09 & 1.99 \(\pm\) 0.10 & 1.49 \(\pm\) 0.06 & 1.10 \(\pm\) 0.10 & 1.15 \(\pm\) 0.04 & 1.40 \(\pm\) 0.04 \\ \hline
\end{tabular}
\caption{Nighttime average values of mixed layer-related quantities: \(D\) is the depth of the ML; \(\varepsilon(-D)\) is the TKE dissipation rate at the base of the ML (calculated over the depth range \(-D < z < -D + 2.5 \text{ m}\)); \(\varepsilon^\ast(-1)\) is the dimensionless TKE dissipation rate (calculated for each profile over the scaled depth range \(0.98 < -z/D < 1.02\) and then ensemble averaged for a particular night; values in parentheses are the 95% bootstrap confidence limits); \(10^5 \overline{w'\theta'}^\ast(-1)\) is the upper bound on the nondimensional vertical heat flux near the base of the ML (calculated as \(\overline{w'\theta'}^\ast(-1)\) \(\leq 0.2 \varepsilon^\ast(-1),\) using the mean values of \(\varepsilon^\ast(-1)\)); \(a_n\) and \(b_n\) are the values resulting from a least-squares fit of the form \(b_n + a_n z^\ast\) to \(\varepsilon^\ast(z^\ast)\) (where \(\varepsilon^\ast(z^\ast) = c(z^\ast)/J_f^\ast\) and \(z^\ast = z/D\) in the lower half of the ML (Figs. 3 and 4). The \(\pm\) values are the probable uncertainties in \(b_n\) and \(a_n\), resulting from the fit to individual nightly mean profiles and to the overall mean profile. The latter, written in parentheses, are error estimates as determined from two graphically fitted straight lines in the 95% bootstrap confidence envelope (see Fig. 4).}
\end{table}
into (7a) resulted in estimates of the upper limit on the magnitude of $w^r \theta^r \epsilon(z^r)$, ranging between 0.07 and 0.18 for the individual nights with an overall mean of 0.13 (Table 2). Using (5a) and the mean value $|w^r \theta^r \epsilon(z^r)| \leq 0.13$, we suggest that a reasonable approximation for the nondimensional heat flux profile in the convective OBL during our experiment may be given by

$$w^r \theta^r \epsilon(z^r) = 1 + 1.13z^r. \quad (8)$$

Note that the heat flux at the base of the ML is negative (i.e., flows out of the ML).

Figures 1 and 2 indicate a maximum in $\theta$ near the middle of the ML, similar to the minimum in $\theta$ observed near the middle of the ML of the convective ABL (e.g., Stull 1988). This is simply due to cooling from above, by heat loss from the ocean surface, and entrainment of cool water from below, and results in cooler temperatures in those regions and a maximum in $\theta$ near the middle of the ML.

The steadiness of the vertical structure of $\theta$ also implies that throughout the depth of the convective OBL the change in temperature, as a function of time, has to be uniform. This suggests that, at least in a time-averaged sense, the convective OBL cools like a slab. From dimensional analysis an “overturn,” in which a convective eddy communicates from top to bottom of the ML, has a timescale of $T_c = (D^2/J_0^b)^{1/3}$, which is on the order of one hour for $D = 65$ m and $J_0^b = 1.5 \times 10^{-7}$ m$^2$ s$^{-3}$. If, during an overturn, the OBL is completely mixed, then potential temperature changes averaged over time periods larger than $T_c$ will be consistent with the picture of the ML cooling as a uniform slab. However, one has to remember that on timescales smaller than $T_c$ the convective OBL is highly intermittent (consecutive profiles of $\theta$, separated by 6–7 minutes, reveal considerable small-scale variability and complexity; Anis and Moom 1992). The slab-type behavior of the ML is further given credence by the fact that the shear magnitude in the bulk of the ML was at the noise level but increased relatively sharply near the ML base (Fig. 5).

b. Estimate from dissipation profiles and TKE balance

In the convective OBL, away from the ocean’s surface and away from the ML base, TKE production is mainly by buoyancy forcing. For a steady-state, neglecting turbulence transport, and since the background shear is indistinguishable from zero, neglecting also mechanical production, we expect a balance between TKE dissipation and buoyancy production such that

$$\epsilon(z) \sim \epsilon_b(z). \quad (9)$$

Neglecting salinity effects (9) can be rewritten as

$$\epsilon(z) \sim g \alpha w^r \theta^r(z), \quad (10)$$

where we use $J_0^b = g \alpha w^r \theta^r(0)$.

Equation (10) permits an independent estimate of the depth-dependent heat flux from the dissipation profiles. Within the ML (away from the surface layer and the ML base) we fitted a linear profile, by least squares, of the form

$$\epsilon(z^r) = b + az^r. \quad (12)$$

The coefficients $a$ and $b$ are equivalent via (11) to those in (5a) and (8). That is, if (11) is true, we expect $b \sim 1$, $a \sim a_b$.

Least-square fits to the observed $\epsilon(z^r)$ profiles, performed in the lower half of the convective OBL (Figs. 3, 4), resulted in values of $a$, between 0.30 and 1.50 for the individual nights, and an overall mean of $a = 0.93$ (Table 2). The uncertainty in the mean value of $a$, was estimated in two ways: the first, $0.0 = 0.05$, is the probable uncertainty (Press et al. 1992) in the least-square fit to the mean $\epsilon(z^r)$ profile for all nights. The second uncertainty estimate, given by the range 0.73–1.23, was determined by a graphical fit of two straight lines in the 95% confidence envelope (Fig. 4). This resulted in one fit with a minimum slope of 0.73 and a second fit with a maximum slope of 1.23. Of the two estimates, the second represents a better way to estimate the probable uncertainty in $a$, since it takes into account the spread of values of $\epsilon$, around its mean, at each depth bin. Based on this estimate we conclude that $a$, is consistent, at the 95% confidence level, with the average value $a_b$ of 1.13.

Estimates of $b$, ranged from 0.66 to 1.99 for the individual nights with an overall mean of 1.40 (Table 2). The uncertainty was estimated in a similar way as for $a$, and the uncertainty interval for the overall mean value of $b$, (estimated from the two graphically fitted straight lines; Fig. 4) was 1.25–1.60. Although this interval has values slightly larger than the expected value of 1, we believe that it is not significantly different from 1; for example, estimates of $\epsilon$, which are systematically too large by 25%, could account for this departure without causing a change in the slope of $\epsilon(z^r)$ [correction of an error producing dissipation values too large by 25% will also slightly reduce the values of $a_b$, through $\epsilon(z^r)$, such that the overall mean value of $a_b$ will be 1.10 instead of 1.13]. An uncertainty analysis by Oakey (1982), who employed a profiler similar to our RSVP, suggests that uncertainties in estimates of $\epsilon$ may be a combination of random errors (maximum of ±27%), systematic errors (maximum of ±32%), and errors due to the assumption of isotropic turbulence (maximum of ±50%). An error may also be introduced by the obvious difficulty in estimating the precise depth of the base of the ML, $D$. However, we estimate this error to be ±10%, at most, for individual profiles and less so when averaged over many profiles.
c. Additional considerations and comparison to atmospheric and numerical results

Physical processes in the convective OBL, neglected in our simplified TKE balance, may also account for some of the differences between the two independent estimates of the heat flux profiles. Of these the most important process is the additional production of TKE by either direct wind stress or by surface waves. However, the enhanced TKE dissipation observed in the upper 30%-40% of the OBL (Anis and Moum 1994) may effectively balance this additional TKE production. Numerical studies of the convective ABL (André et al. 1978) and the convective OBL (André and Lacarrère 1985) indicate turbulence transport of TKE from near the surface into the ML. This transport causes a net convergence, effectively production of TKE, in the lower half of the OBL, which has to be balanced by TKE dissipation. The additional dissipation resulting from this balance, which unfortunately cannot be estimated from our measurements, may be another reason why our estimates of $b_i$ were slightly larger than unity.

In spite of the simplifying assumptions we have made, the fact that the vertical structure of $\theta$ was constant throughout each night and that $\epsilon$ decreased linearly with depth in the lower half of the ML are solid and consistent results, both leading to the same conclusion of a linear heat flux profile. Moreover, the suggested slope of the heat flux profile (i.e., $a_h = 1.13$) is in general agreement with the range of 1.1-1.3 suggested by Stull (1976) based on a large number of published values. Simulations of the Wangara experiment by Deardorff (1974) and André et al. (1978), using different numerical schemes, resulted in $a_h = 1.13$ and $a_h = 1.15$, respectively. A numerical study of a buoyancy-driven OBL (André and Lacarrère 1985) resulted in $a_h = 1.16$. When an additional turbulence-generating velocity shear, $\Delta U = 0.1$ m s$^{-1}$ (about twice the magnitude of the mean value of $\Delta U$ observed at the base of the convective OBL; see next section), was imposed at the bottom of the OBL the slope increased to $a_h = 1.20$, due to the enhanced entrainment.

4. Entrainment rates

The deepening rate of the ML due solely to vertical mixing processes is given by $w_0$, the entrainment rate. If, at the bottom of the ML, all mixing processes are associated with the entrainment of the deeper and denser fluid into the ML, then (e.g., Stull 1988)

$$w_0 \Delta \rho = -w^2 \rho' (-D),$$

(13)

where $\Delta \rho$ is the density jump at $z = -D$ (we used the density flux representation, instead of heat flux, for the purpose of comparison below). From (8), with aid of the relation $J_0 = g \alpha w' \theta'$, we have $-w^2 \rho' (-D) = 0.13 w \rho' (0) = -0.13 \rho J_0^2 / g$. Using the representa-tive values $\Delta \rho = 0.16$ kg m$^{-3}$ (estimated as $\Delta \rho = |\partial \rho / \partial z| \Delta z$, where $|\partial \rho / \partial z| = 3.2 \times 10^{-2}$ kg m$^{-4}$ was calculated for the depth range $-D < z < -D + 2.5$ m and averaged for all nights, and $\Delta z = 5$ m), $\rho = 1023$ kg m$^{-3}$, and $J_0^2 = 1.5 \times 10^{-7}$ m$^3$ s$^{-3}$ results in an estimate of $w_0 \sim -1.3 \times 10^{-5}$ m s$^{-1}$. This rate translates to a deepening of the ML of about 0.6 m during a period of 12 h. Unfortunately, such small entrainment rates are close to impossible to verify experimentally in the ocean because of the much larger changes due to lateral variability and waves in the stratified interior. However, since from (13) the entrainment rate is inversely proportional to $\Delta \rho$, measurements at sites where the latter is much smaller than we have observed may provide a better opportunity for experimental verification.

Entrainment during the quasi-steady convective forcing phase differs from the rapid deepening (8-19 m h$^{-1}$; section 2c) of the ML observed at the beginning of each night in several aspects: 1) during the rapid deepening period (2-3 h) $J_0^2$ was not steady but rather increased roughly linearly with time to its quasi-steady nighttime value; 2) although ML deepening was initiated by the onset of destabilizing convective conditions, additional forcing was required from energy balance considerations; 3) during the major part of each night, when $J_0^2 \sim$ const, the ML base was deep and far enough from the surface such that only convection was effectively driving the entrainment (this is further evidenced from the fact that $\epsilon \sim J_0^2$ away from the surface). The rapid ML deepening phase is discussed in detail in Anis and Moum (1994).

Deardorff (1983) derived an entrainment rate based on the TKE equation and second-moment equations for the buoyancy flux and the density fluctuation variance in the entrainment zone of the ML. The TKE equation was closed utilizing various empirical closure functions and constants. Formulation of the normalized entrainment rate, $w_0 / u^2$ or $w_0 / w^2$, was presented (Fig. 3 of Deardorff 1983) as a function of three Richardson numbers: $R_i = c_i^2 / u^2$; $R_i^* = c_i^2 / w^2$ ($w^2 = (J_0^2 D)^1/3$ is the convective velocity scale); and $R_i = c_i^2 / (\Delta U)^2 (\Delta U$ is the magnitude of the velocity jump at the base of the ML). These three Richardson numbers represent, respectively, the importance of the three mechanisms contributing to entrainment, namely, surface wind stress, surface buoyancy flux, and shear across the ML base, relative to the retarding stable density gradient at the ML base, represented by the velocity scale $c_i = (g / \rho \Delta \rho D)^{1/3}$. In the general case, when both the surface stress and convection are important, the formulation for $w_0$ and $R_i^*$ is used with $w^2$ replaced by $w_0' = (w_0^2 + \tau_i u^2)^{1/2}$ and $R_i^*$ by $R_i^* = (R_i^* + n_i \tau_i^{1/3})^{1/2}$, where the coefficient $\eta_i < 1.8$.

From the mean values $J_0^2 = 1.5 \times 10^{-7}$ m$^3$ s$^{-3}$, $D = 65.1$ m, $\tau_0 = 0.09$ N m$^{-2}$ (Table 1), and $\Delta \rho = 0.16$
kg m$^{-2}$, and the definitions of $w_0$, $u_0$, $c_i$, Ri$_e$, and Ri$_s$, it is found that $w_0 \approx w_0'$, to about 5%, and Ri$_e \approx$ Ri$_s$, to about 10%. This indicates that for our experiment Deardorff's entrainment formulation for a convectively driven ML can be used to a good approximation. Using that formulation in conjunction with the mean values $w_0 = 2.1 \times 10^{-2}$ m s$^{-1}$, Ri$_e$ = 221, and Ri$_s$ = 71, results in $w_0 \sim 2.5 \times 10^{-3}$ m s$^{-1}$ (Ri$_e$ was estimated using $(\Delta U)^2 = S^2(\Delta z)^2 = 1.4 \times 10^{-3}$ m$^2$ s$^{-2}$, where $S^2 = 5.6 \times 10^{-5}$ s$^{-2}$ was calculated for the depth range $-D < z < -D + 2.5$ m and averaged for all nights, and $\Delta z = 5$ m).

From dimensional analysis, it is argued that the dimensionless entrainment rate, $w_e / u'$, can be expressed as $w_e / u' = f(Ri)$ (e.g., Turner 1973), where $(Ri_0) = g \beta \Delta T u' / \nu^2$ and $l$ and $u'$ are the turbulence length and velocity scales, respectively. The turbulence velocity scale usually assumes the velocity scales of the forcing mechanism, for example, $u' = w_0$ or $u' = u_0$ for a BL driven by, respectively, convection or surface stress. The length scale is commonly taken to be the ML depth, such that $l = D$. In the respective cases of convective- and stress-driven entrainment, one then has $w_e / w_0 = f(Ri_e)$ and $w_e / u_0 = f(Ri)$, where Ri$_e$ and Ri$_s$, were defined above. Laboratory studies of the entrainment zone of a convectively ML by Deardorff et al. (1980) showed that the dimensionless entrainment rate could be represented fairly well by the relation $w_e / w_0 = 0.25(Ri_e)^{-1}$. Using this relation and the relevant mean numerical values from our experiment, a value of $w_e \sim -2.4 \times 10^{-5}$ m s$^{-1}$ is found. Note, however, that the relation $w_e / w_0 = 0.25(Ri_e)^{-1}$ results directly from the definition of $w_0$ and Eq. (13), when $-\overline{w' \rho' / D} = -0.25 \rho J / g$ is used instead of $-\overline{w' \rho' / D} = -0.15 \rho J / g$, the mean estimate for our experiment.

In a formulation by Stull (1976), the potential energy change due to turbulent entrainment was related to the TKE energy integrated over the depth of the ML. Applying this to an idealized slab-type ML model, and making some simplifications and approximations, an entrainment rate equation was derived. Values of the coefficients in Stull's equation were then determined by testing the theoretical equation against a large number of laboratory and field experiments. Using Stull's entrainment parameterization for the convective case (presented graphically in Fig. 7 of Deardorff 1983) and the estimates from our study for $w_0$, Ri$_e$, and Ri$_s$, results in $w_e \sim -1.2 \times 10^{-3}$ m s$^{-1}$. In spite of the lack of consensus on the various entrainment formulations, the values of $w_e$ calculated from (13) and from all of the entrainment formulations discussed above are consistent to within a factor of 2.

5. Summary and conclusions

An open-ocean site experiment was conducted in the nighttime convective OBL when the surface buoyancy flux was quasi-steady. The vertical structure of potential temperature, TKE dissipation rate, and shear of the horizontal velocity were investigated using a large number of vertical profiles. The major results of this study are the following.

- The nighttime convective OBL cooled as a uniform slab during the quasi-steady forcing phase. This is inferred from the steadiness of the vertical structure of potential temperature, that is, $\delta \theta / \delta z / \delta t = 0$, where $\theta$ was averaged over time periods larger than $T_c$, the convective timescale. The slab-type behavior is further supported by the vanishing shear of horizontal velocity in the bulk of the ML and the large increase in shear near the base of the ML. In contrast, during daytime, when the upper OBL was slightly stably stratified due to solar heating, and mixing was chiefly driven by the surface wind stress, a clear increase in shear in the upper part of the OBL was observed.

The clear distinction observed between the nighttime well-mixed layer, which was unable to support a significant shear, and the daytime stratified upper OBL with a noticeable shear, supports André and Lacarrère's (1985) results from a third-order numerical model. Their results indicate that it is not always possible to describe the OBL as a well-mixed layer as is frequently done in most of the simple parameterized models. This was particularly true for the case of a stress-driven OBL, since the mechanically induced turbulence was found to be relatively inefficient in redistributing the momentum introduced at the surface. However, in the case of buoyancy-driven turbulence, heat and momentum mixing was found to be much more efficient and to lead to a simplified vertical structure. The evidence from our study is consistent with André and Lacarrère's conclusions and provides experimental justification for the adequacy of the use of slab models for the OBL during convection. Conversely, our results demonstrate the inadequacy of such models during periods when the OBL is driven by surface stress, though exceptions may occur in extreme cases such as storms when exceptional high surface stresses induce vigorous mixing and a truly well-mixed layer may result.

- Two independent estimates of the turbulence heat flux profile suggest that the heat flux in the convective OBL decreases linearly with depth. The first method of estimation of the heat flux profile was based on heat conservation and the steadiness of the vertical structure of potential temperature. From this estimate, the dimensionless linear heat flux is represented by $w \theta^*(z^*) = 1 + a_0 z^*$. Values of the constant $a_0$, estimated from the ratio $-\overline{[w \theta'(-D)]/[w \theta'(0)]}$, were between 1.07 and 1.18 with a mean of 1.13. The second method of estimation of the heat flux profile was based on profiles of TKE dissipation rate and a simplified TKE balance between the buoyancy flux and dissipation in the bulk of the ML, away from the surface of the ocean and the
base of the ML. This estimate suggests, similarly, a linear heat flux profile where \( w^* \theta^* (z^*) = b_\theta + a_\theta z^* \). Estimates of \( b_\theta \) ranged from 0.66 to 1.99 with an overall mean of 1.40, and estimates of \( a_\theta \) were between 0.30 and 1.50, with an overall mean of 0.93. As discussed in section 3, the estimates of \( b_\theta \) are not significantly different from 1, while the estimates of \( a_\theta \) are not significantly different from those of \( a_\theta \). The linear heat flux profile suggested by the two methods is in good agreement with results from laboratory studies, numerical models (for both the OBL and the ABL), and field studies in the convective ABL.

- Estimates of the ratio \( -[w^* \theta^* (-D)]/[w^* \theta^* (0)] \) ranged between 0.07 and 0.18 with a mean of 0.13, and are in good agreement with most of the values quoted for the convective ABL (e.g., Stull 1976). Knowledge of the heat flux at the base of the ML, \( w^* \theta^* (-D) \), is of crucial importance in slab models since it provides a closure condition for the ML equations and permits an estimate of the entrainment rate.

- The entrainment rate, estimated from the equation for the transport across an interface (13) and the mean heat flux and density jump at the base of the ML, was found to be \( \sim -1.3 \times 10^{-5} \text{ m s}^{-1} \). Entrainment rates calculated from different formulations, using the relevant numerical values from our experiment, were found to be within a factor of 2 of our estimate. This rather good agreement may be related to the fact that the simplifying assumptions underlying some of the entrainment formulations, such as an idealized slab-type mixed layer, seem to hold for convective ocean boundary layers.

Although there appear to be clear similarities between the mixed layers of both the ocean and atmosphere during convection, there is a growing body of evidence that the upper part of the OBL, or the surface layer, is distinctly different (e.g., Agrawal et al. 1992; Anis and Moun 1994a). Specifically, mixing is enhanced in the OSL over predictions for the ASL and the superadiabatic potential temperature gradient is comparatively smaller (Anis and Moun 1994a). Both the short timescale variability (i.e., less than the convective timescale) of the vertical heat flux and its relation to the unique processes in the upper OBL such as wave–turbulence interactions and Langmuir circulation will have to await future availability of direct heat flux measurements.

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