Conditional Sampling of Updrafts and DOWndrafts in the Marine Atmospheric Boundary Layer

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ABSTRACT

The properties of updrafts and downdrafts in the lower third of the marine atmospheric boundary layer (MABL) over the central Pacific Ocean are investigated using a conditional sampling technique. When the drafts are classified according to their heat and moisture content, the properties of the major classes (moist updrafts and dry downdrafts) are in agreement with a parcel displacement model of vertical mixing. The minor class events appear to be the result of the reversal of motion of the major class events. Drafts that consume turbulent kinetic energy (TKE), although small in number in the lower third of the MABL, have spatial scales comparable with drafts that produce TKE. At the lowest level, the area occupied by positively buoyant downdrafts exceeds the area occupied by negatively buoyant updrafts by a factor of 2. Updrafts and downdrafts produce a large fraction of the total fluxes of heat, moisture, and momentum, increasing from 75% at 0.07 z; to 85% at 0.32 z. Of the draft contribution to the fluxes, 95% is due to the mean properties of the events and only 5% is due to the correlated fluctuations within the events. A convective mass flux parameterization, based on the mean conditions within updrafts and downdrafts, is obtained for the lower third of the mixed layer.

1. Introduction

Convective elements are largely responsible for vertical transport within the mixed layer and for entrainment at the capping inversion. Although the existence of "plumes" and "thermals" in the surface and mixed layers has been recognized since the 1950s, the techniques for probing their characteristics in a systematic way have been developed only recently. Two approaches are currently popular for the investigation of flux-producing events in turbulent flow. One is conditional sampling, which originated in the studies of intermittency in laboratory boundary layers (e.g., Kovasznay et al., 1970; see Khalsa, 1980) for a more extensive review of early applications of conditional sampling techniques). In studies of atmospheric boundary layers, Warner and Telford (1967) and Frisch and Businger (1973) employed methods similar to the formalized conditional sampling methods now in use (Lenschow and Stephens, 1980; Greenhut and Khalsa, 1982). The second approach is the use of a joint probability distribution (JPD), which displays the two fluctuating variables that make up a flux. The JPD has been used in the study of eddy structure by Holland (1973) and more recently by Mahrt and Paumier (1982, 1984), Wilczak and Businger (1983) and Grossman (1984).

The two approaches have a number of contrasting strengths and weaknesses. In conditional sampling, the choice of discrete events preserves the temporal relationship between the variables, while in a JPD this information is lost. On the other hand, event selection in conditional sampling depends upon a set of specified criteria which may reflect the prejudices of the experimenter. In addition, different criteria may yield different results. In contrast, a JPD is strictly objective.

Once the criteria for conditional sampling are established and an indicator function is generated, conditional averages and conditional variances and covariances can be computed for any set of variables. Each variable can be independently averaged in conditional sampling and this is its main advantage over the JPD. Holland (1973) and Grossman (1984) used conditional mean distributions (CMDs) to find averages of other variables associated with portions of a JPD. Since averages produced in this way are functions of the two variables of the JPD, a degree of independence is lost in studying the properties of additional variables. As an example, consider the case of a CMD for temperature based on a humidity–vertical velocity JPD. Suppose the CMD gives an average temperature perturbation of zero in the moist updraft quadrant. Conditional sampling of the same data using an indicator function based on vertical velocity might reveal
a balance between warm/moist updrafts and cool/moist 
updrafts both having large temperature perturbations. 
In addition, using conditional sampling, the average 
size, area occupied, and perturbations in vertical ve-
locity, temperature and moisture can be obtained for 
the two classes of drafts. This information is not avail-
able from a JPD.

In this paper the conditional sampling technique 
developed by Greenhut and Khalsa (1982) (hereafter re-
ferred to as GK) is applied to aircraft data taken within 
the lower third of the marine atmospheric boundary 
layer (MABL). In this region there is considerable vari-
ation of draft properties with height (Lenschow and 
Stephens, 1980). The present work and GK differ from 
other conditional sampling studies in the explicit treat-
ment of downdrafts as distinct from the environment. 
Statistics are obtained on the size and area occupied 
by events and on perturbation quantities and contri-
butions to variance and covariance due to updrafts and 
downdrafts as defined by an indicator function based 
on fluctuating vertical velocity.

When drafts are classified according to their tem-
perature and moisture content, the behavior of the most 
common classes is consistent with a parcel displacement 
model of vertical motion in the presence of the 
observed profiles of wind, temperature and humidity. 
The remaining events are at least partly due to reversals 
of motion of events in the major classes. Our expla-
nation of the origin of the thermodynamically classified 
events is supported by the along-wind momentum per-
turbations within each class. The thermodynamically 
classified data are also used to examine the role played 
by drafts in the production and consumption of tur-
bulent kinetic energy in the mixed layer.

Mean event properties account for most of the flux 
produced by updrafts and downdrafts and therefore 
conditional averages can be used in a parameterization 
similar to the convective mass flux parameterization 
introduced by Betts (1975, 1976). Extending the results 
of GK, we find that, along with heat and moisture, the 
flux of along-wind momentum can be parameterized 
in this way as well.

2. Data and scaling parameters

The data used in this analysis were obtained from 
flights of a NOAA WP-3D research aircraft during the 
Equatorial Pacific Ocean Climate Studies (EPOCS) 
program. Data from two flight periods are used: 28 
February through 18 March 1979 (EPOCS I) and 14 
June through 2 July 1979 (EPOCS II). The flight paths 
for both periods were concentrated mainly in the cen-
tral equatorial Pacific between 150 and 160°W and 
between the equator and approximately 10°N. For 
each period there were also flights near the equator in 
the vicinity of 170°E and 110°W. The EPOCS flights 
are discussed in detail by Greenhut and Bean (1981).

Turbulence data were obtained from the NOAA gust 
probe system mounted on the aircraft (Bean et al., 
1976; Gilmer et al., 1978). The data were digitized at 
80 Hz after low-pass filtering with a four-pole Butte-
worth filter having a –3 dB point at 11.5 Hz. Intervals 
approximately 5 minutes (33 km) in length were used 
in the computation of fluxes by eddy correlation tech-
niques. During processing, the mean and a linear trend 
were removed from the time series of each variable 
and the sampling rate was reduced to 40 Hz.

In the North Pacific Experiment (NORPAX) data, 
GK found that the fluxes, and therefore the mixed layer 
scale parameters, could be rather different on either 
side of the near-equatorial convergence zone (CZ). 
The location of the CZ in the vicinity of the EPOCS flights 
was determined using visible satellite imagery. On four 
of the eight days chosen for study, data were obtained 
only one side of the CZ. On three of the remaining 
days it was found that since conditions were sufficiently 
similar on both sides of the CZ it was unnecessary to 
analyze the data separately. Only on Julian Day 064 
was it necessary to separate the observations into re-
gions north and south of the CZ. For this day the sur-
face fluxes were approximately three times as large to 
the north of the CZ as they were to the south of it.

Scaling parameters and surface fluxes for the result-
ing nine observation periods are given in Table 1. The 
runs labeled 0MC and 0CH both occurred on Julian 
Day 060 and are distinguished by their departure and 
arrival locations: Majuro Island to Christmas Island 
and Christmas Island to Hawaii.

The profiles of potential temperature, θ, obtained 
during descents to and ascents from the low-level tur-
bulence flight legs, were used to determine the height 
of the first inversion, z1, for each run. The inversion 
height was taken to be the height at which the θ-profiles 
for the descent and ascent changed slope. When the 
two heights were not the same, an average was taken 
to obtain z1 for the flight period.

Composite profiles of potential temperature θ, water 
vapor mixing ratio, q, and virtual potential tempera-
ture, θv, are shown in Fig. 1. Individual profiles were 
averaged over intervals of z/z1 of width 0.1 before com-
posing. Since the aircraft did not usually descend to 
the lowest level before beginning a flux run, the lowest 
two points in the composite profiles are averages over 
fewer observations than the remaining points and 
therefore have a greater uncertainty. The profiles in 
Fig. 1, typical of those observed over tropical oceans, 
show a slightly positive slope in potential temperature 
and a negative slope in mixing ratio through most of 
the mixed layer with steeper slopes above the inversion. 
Figure 1 also shows the composite profile of wind speed 
development from the mean. Because of a large variability 
in wind speed magnitude, the average wind speed in 
the layer 0.1 ≤ z/z1 ≤ 1.0 was removed from each profile 
before compositing. The mean boundary layer wind 
speed for all runs was 9.7 m s⁻¹. Figure 1 shows a po-
itive shear in wind speed within the boundary layer
which decreases but remains positive above the inversion.

The aircraft performed a stepping maneuver during the turbulence runs. The flight levels were 46, 91 and 152 m in EPOCS I and 85, 165 and 220 m above the surface in EPOCS II. For EPOCS I the average fluxes at the 46 m level were taken to be representative of the surface fluxes and are given in Table 1. For EPOCS II, the lowest level, 85 m, is probably above the constant flux layer so the surface fluxes were obtained by linear extrapolation of the average flux profiles to z = 0.

The scaling parameters in Table 1 were calculated using

\[ u^* = (\tau_0/\rho)^{1/2} = \left[ -\left(\overline{u'w'}\right)_0 \right]^{1/2} \]

\[ w^* = \sqrt{\frac{g^2 B_0}{\rho_c T_0}} T_0 \]

\[ q^* = \frac{E_0}{\rho L_c w^*} \]

\[ T^* = \frac{B_0}{\rho_c w^*} \]

where the subscript 0 denotes the surface value and g is the gravitational acceleration. The primed quantities are fluctuations about the mean of a run after a linear trend has been removed. The ratios \( z_i/(\overline{L}) \), where \( L \) is the Monin-Obukhov length, lie between 2 and 10 indicating slightly unstable conditions typical of the tropical marine boundary layer.

The composite normalized flux profiles are shown in Fig. 2. The points in these and in all subsequent profiles were obtained by first averaging all flight legs at a given height for a given flight period. These averages were normalized and then averaged over all flight periods for a given height interval. Only the seven EPOCS I flight periods contributed to the lowest level at \( z/z_l = 0.07 \). For \( z/z_l = 0.13 \) and 0.23, the points in the profiles are averages over all nine flight periods. At the highest level, \( z/z_l = 0.32 \), the average is made up of only the two flight periods of EPOCS II. As in GK, we find that normalization by mixed layer scaling parameters results in values at a given height \( z/z_l \) which are approximately the same from day to day as indicated by the small standard errors of the mean in Fig. 2. We are therefore confident that the normalized profiles we have obtained are good representations of the conditions in the experimental area. Standard errors are not shown in subsequent figures since they are as small, and more often smaller, than those in Fig. 2.

3. Selection criterion for conditional sampling

The parameter we use to characterize the flux-producing convective elements in the MABL is vertical velocity. Although both temperature and humidity have been used by previous workers to define thermals, we have selected vertical velocity because it allows the

![Fig. 1: Profiles of potential temperature (θ), virtual potential temperature (θ_v), water vapor mixing ratio (q) and wind speed deviation (V - ⟨V⟩), where ⟨V⟩ is the mean wind speed in the layer 0.1 < z/z_l < 1. The profiles are averages over the data taken during the descents and ascents for the nine observation periods in Table 1.](image-url)
study of transport carried out by coherent downdrafts as well as updrafts. Using vertical velocity to define events provides information on the origin and dynamics of coherent vertical structures that is not available when they are defined using temperature or humidity.

The criterion for an updraft (downdraft) is that \( w' \) be larger (smaller) than a threshold \( w_{th}^u \) (\( w_{th}^d \)) for a minimum time, \( t_{min} \). As discussed in GK, a minimum event time eliminates smaller spurious events and prevents larger events from being divided by small breaks. The length of the minimum event time was selected by inspecting a number of indicator function time series which were produced using the same threshold but with different values of \( t_{min} \). A value corresponding to three samples in the block averaged data, or 0.375 sec, gave optimum indicator function time series, retaining small \( w' \) events with well-defined associated \( T' \) and/or \( q' \) perturbations while rejecting still smaller events which appeared to be spurious. The corresponding minimum length of 41 m at an airspeed of 110 m s\(^{-1}\) is well below the dominant scales of flux producing events as shown, for example, by Mahrt and Paumier (1984).

During the time that \( w' \) is larger (smaller) than \( w_{th}^u \) (\( w_{th}^d \)), an indicator function, \( I(t) \), is set equal to +1 (−1). Otherwise \( I(t) \) is set equal to 0 and this portion of the time series is referred to as the "environment." The indicator function is used to perform conditional sampling on the other variables. For convenience \( I(t) \) is separated into two functions:

\[
I(t) = I^+(t) - I^-(t)
\]

where \( I^+(t)(I^-(t)) \) is +1 during updrafts (downdrafts) and zero otherwise.

Large-scale variations in the time series of \( w' \) can make \( I(t) \) unrepresentative of the updrafts and downdrafts on the scales of interest. For this reason a high-pass filter having a −3 dB point at 0.026 Hz was applied to the time series of \( w' \) and the other variables thereby eliminating scales larger than approximately 4 km. The time series was then smoothed by block averaging every five points producing an effective sampling rate of 8 Hz.

When thresholds are used, events are selected for which the indicator variable, in our case \( w' \), is well displaced from the mean. It is thus reasonable to base the choice of thresholds on the frequency distribution of \( w' \). We have set the threshold for updrafts (downdrafts) equal to the square root of one half the variance of all positive (negative) values of \( w' \) about \( w' = 0 \),\(^1\) Since the frequency distribution of \( w' \) is positively skewed, this method gives \( |w_{th}^u| > |w_{th}^d| \). As shown in GK and confirmed in this study, the consequence of this selection is that the mean vertical velocity in the environment is zero. A complete discussion of the methods used for selecting events is given by GK and the reader is referred there for details and references to previous work.

The determination of thresholds was done separately for each run at each flight altitude. The resulting profiles of thresholds, normalized by \( w_{th} \), are shown in Fig. 3. The ratio \( |w_{th}^u/w_{th}^d| \) is approximately constant with height and equal to 1.2. This is about 15% smaller than the ratio observed by GK using NORPAK data at 0.32z. Comparing the two cases, we see that the difference is mainly due to the presence of more vigorous downdrafts in the EPOCS data where \( |w_{th}^u/w_{th}^d| \) is 10−15% larger than in the NORPAK case. This is reflected in a smaller average coefficient of skewness for EPOCS, 0.57 at 0.32z, compared to 0.71 for NORPAK.

4. Updraft and downdraft statistics and conditionally sampled meteorological variables

a. Average size and frequency of events

The average event size is defined as

\[
d^z = \frac{V_a \sum t_i^z}{n^z}
\]

where \( V_a = 110 \text{ m s}^{-1} \) is the aircraft velocity, \( t_i^z \) is the period of time for the \( i \)th updraft (+) or downdraft (−) event and \( n^z \) is the total number of updrafts or downdrafts in the sample. The number of events per unit length is given by

\[
N^z = \frac{n^z}{t_{tot}}
\]

where \( t_{tot} \) is the total duration of the sample. The product \( d^zN^z \) gives the proportion of the time series occupied by updrafts and downdrafts and is also equal to the fractional area occupied by the events if it is assumed that they are randomly distributed (Zipser and LeMone, 1980).

The profiles of event size, number density and area occupied obtained from the EPOCS dataset are shown

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1 Although GK failed to mention the factor of one-half in the definition of thresholds, the method of selecting thresholds is the same in their paper as in the present work.
in Fig. 4. Updrafts and downdrafts have approximately the same average size at a given height, but the average number of downdrafts exceeds the average number of updrafts by about 33%. The fractional area occupied, $d^2/N^2$, shows the same excess for downdrafts. This excess compensates for the weaker vertical velocities in downdrafts. As shown in Fig. 5, the conditionally averaged vertical velocities in updrafts, $w^+$, exceeds that for downdrafts, $w^-$, by about 33% at all heights and therefore the area-weighted vertical velocities for updrafts and downdrafts, $d^2N^2w^+$, are nearly equal in magnitude (Fig. 5). Since the mean has been removed from the time series prior to analysis, the average vertical velocity for each flight is zero and therefore $d^2N^2w^+ = d^2N^2w^-$ implies that the contribution from the remainder of the time series (the environment) is zero. As discussed in the previous section, zero mean vertical velocity in the environment is a consequence of our choice of thresholds.

The average event size for both updrafts and downdrafts increases with height (Fig. 4). There is also a decrease with height in the number of events per unit length. For updrafts, this can be understood in terms of the coalescence of plumes with increasing height in the lower part of the boundary layer as previously observed by Frisch and Businger (1973), Kaimal et al. (1976) and Manton (1977). For downdrafts, the reverse appears to be the case, i.e., large downdrafts separate into more numerous but smaller downdrafts near the surface. Some of the small downdrafts are the result of reversed motions as discussed in the next section. However, for the most part, the nature of downdraft size and distribution near the surface is necessitated and perhaps caused by the presence of small, more vigorous upward-moving thermals that intermingle with and cause the breakup of the larger downward moving parcels.

The fractional area occupied by updrafts and downdrafts increases with height by about 12% between the lowest two levels (Fig. 4) and then remains approximately constant at about 0.15 for updrafts and 0.20 for downdrafts. The smaller area occupied at the lowest level may be partly due to our sampling method, which eliminates all events with lengths smaller than 41 m. Events of this size are likely to be more prevalent nearer the surface.

As shown in the previous section, the EPOCS data analyzed here contain downdrafts that are more vigorous than those observed by GK in the NORPAX data. In comparing results, we shall consider only the highest point in the EPOCS profiles since there the normalized height is approximately equal to that for the NORPAX flights. The normalized conditionally averaged vertical velocity for updrafts at the highest level in Fig. 5, $w^+/w^* = 1.14$, is equal to that observed by GK in the NORPAX data, but for downdrafts, $w^-/w^* = -0.84 \pm 0.06$ which is about 12% larger in magnitude than the NORPAX value, $-0.75 \pm 0.02$. This is compensated for by a correspondingly smaller fractional area occupied by downdrafts in the EPOCS data. For both EPOCS and NORPAX, the fractional area occupied by updrafts is about the same.

Differences on the order of 30% exist between the normalized event size and normalized event frequency at the highest level in Fig. 4 and those found by GK using NORPAX data. In the latter case, the scale height was the lifting condensation level (LCL) calculated from parameters observed at flight level. As pointed out by GK, when the boundary layer is not completely well-mixed, the LCL may be an overestimate of its depth by 25% or more. If the NORPAX scale heights are reduced by about 25%, then much better agreement is obtained between the EPOCS and NORPAX results for normalized event size and normalized event frequency except for the differences arising from different areas occupied by downdrafts as discussed above.

Lenschow and Stephens (1980) obtain profiles of $d^+/z_l$ and $N^2z_l$, which are similar in shape to those given in Fig. 4. Our mean event sizes are about a factor of 2 larger and our mean number of events per unit length are about a factor of 2 smaller than theirs. (The quantity 0.68 in Eq. (3) of Lenschow and Stephens (1980) should
be increased by a factor of two, as should the labels on the horizontal axis in the plot of $N^+z_i$ vs $z/z_i$ in their Fig. 3; Lenschow, private communication, 1984.] The fractional area occupied by thermals observed by Lenschow and Stephens is in approximate agreement with the fractional area for updrafts in Fig. 4. Lenschow and Stephens based their indicator function on fluctuations of the mixing ratio, $q'$, and used a threshold of one half the standard deviation of $q'$ with a minimum event size of 25 m. Although $w'$ and $q'$ are well-correlated in the MABL (the correlation coefficient varies from 0.47 at our lowest level to 0.36 at our highest), the distributions of $w'$ and $q'$ over an event are generally not the same. These differences may be the cause of the discrepancies between our statistics and those of Lenschow and Stephens for event size and number density.

b. Conditional averages and event contributions to the fluxes

The conditional average of a fluctuating variable $x'$ in updrafts (+) and downdrafts (−) is given by

$$x^\pm = \frac{\int_0^{t_{\text{tot}}} x'(t) I^\pm(t) dt}{\int_0^{t_{\text{tot}}} I^\pm(t) dt}$$

where $t_{\text{tot}}$ is the total duration of the sample and $I^+(t)$ and $I^-(t)$ are the indicator functions for updrafts and downdrafts, respectively, as defined in Section 3. The profiles of the conditional averages for vertical velocity, mixing ratio, virtual temperature and alongwind momentum are given in Fig. 5. In the height range of 0.07 ≤ $z/z_i$ ≤ 0.32, the conditionally averaged vertical velocities are fairly constant with height, increasing by only 19 and 12% from the lowest to the highest level for updrafts and downdrafts, respectively. As discussed previously, $w'/w_*$ at 0.32$z_i$ is in good agreement with the observations of GK, while $w'/w_*$ at the same height is about 12% larger. The profile of vertical velocity obtained by Lenschow and Stephens (1980) shows a decrease approaching the surface and at the top of the mixed layer. However, in the height range that corresponds to Fig. 5 their profile is constant. Their normalized thermal velocity is about one-half our value of $w'/w_*$.

In GK it was shown that within an updraft or downdraft, 95% of the contribution to the flux is due to the
mean structure of the event and only 5% is due to eddy
correlations within the event. Generalizing this, we can
approximate the contributions of updrafts and downdrafts
to the total normalized flux of \( x \) by

\[
P_{x}^{\pm} \approx \frac{d^{+}N_{x}^{\pm} x^{\pm}}{w_{x}} \tag{1}
\]

where the sum over products of event means [Eq. (A6)
in GK] has been replaced by the product of the condi-
tional averages. Since the area-weighted vertical ve-
clocities are approximately constant with height above
the lowest level, the height variation of the conditionally
averaged variables in the lower part of Fig. 5 is a good
approximation to the height variation of flux contribu-
tions by updraft and downdrafts. Using (1) and
comparing with the total normalized flux profiles in
Fig. 2, we can estimate the percent contribution to the
fluxes by updrafts and downdrafts. For example, at
0.32z_{i}, the contributions are approximately the same
for \( q' \), \( T_{v} \) and \( u' \) and are about 60% for updrafts, 20%
for downdrafts and, therefore 20% for the environment.
(Note that for this calculation we are using \( w_{x} \approx 2w_{x}
\) as determined by the entries in Table 1.) A more
accurate calculation of these proportions using the tech-
niques of GK [see their Eqs. (2) and (3)] gives similar
results: 65%, 20% and 15% at 0.32z_{i}, changing to 50%,
25% and 25% at 0.07z_{i} for updrafts, downdrafts and
the environment, respectively. There is good agreement
between these results and those of GK at 0.3z_{i}.

In Fig. 5, the virtual temperature excess for updrafts
decreases more rapidly with height near the surface
than the corresponding deficit for downdrafts. Since
the air just above the sea surface is nearly saturated
and at the same temperature as the sea surface, the
gradients of both mixing ratio and potential tem-
perature are probably negative in the layer just above
the surface. The temperature profiles in Fig. 1 do not show
a superadiabatic layer but it must be kept in mind that
the lowest points are averages over only a small number
of observations since the aircraft rarely ended its de-
cescent or began its ascents at the lowest level. In
addition, the aircraft was often not low enough to detect
the superadiabatic layer. A superadiabatic layer will
produce updrafts that are relatively warm and moist.
As these parcels continue up into the subadiabatic layer,
they become relatively cool once they pass the level at
which the ambient potential temperature exceeds the
potential temperature of the updraft. This scenario is
confirmed in Fig. 6a which shows warm/moist updrafts
decreasing in area occupied with increasing height while
cool/moist updrafts show a corresponding increase.
The change from warm to cool updrafts presumably
causes the profile of conditionally sampled virtual
temperature for updrafts in Fig. 5 to undergo a large
decline with height between 0.13z_{i} and 0.23z_{i}. For
downdrafts, the same large change in \( T_{v} \) is not observed.
Downdrafts must descend fully into the superadiabatic
layer before becoming relatively cool and therefore
warm/dry is the dominant class of downdrafts
throughout the layer of observation (Fig. 6a).

5. Thermodynamic classification of updrafts and
downdrafts

The conditional averages presented in the previous
section describe the mean properties of updrafts and
downdrafts but do not reveal the large and interrelated
variations that occur from event to event. Additional
information on the nature of updrafts and downdrafts
is obtained by classifying them according to their tem-
perature and humidity perturbations. The observed
properties will be shown to be consistent with a parcel
displacement model of convective motions.

The profiles of \( \theta \), \( q \) and \( V \) in Fig. 1 show a mixed
layer in which the gradients of \( \theta \) and \( V \) are everywhere
positive and the gradient of \( q \) is everywhere negative.
Thus a parcel displaced upward will be cool and moist
and have less horizontal momentum relative to its en-
vironment while a parcel displaced downward will ac-
quire the opposite characteristics. This is observed in
the mean event properties of Fig. 5. At the lowest levels,
updrafts tend to be warm instead of cool since in the
tropics the sea surface temperature usually exceeds the
air temperature and therefore the gradient must be su-
peradiabatic close to the surface.

At some level, parcels of relatively warm near-surface
air encounter an environment of the same \( \theta \) and be-
come neutrally buoyant with respect to temperature.
However, the humidity excess produces overall positive
buoyancy and the parcel will continue to rise, trans-
porting potentially cooler air upward and producing a
negative sensible heat flux. For most of our flights, the
sign of \( wT' \) changes from positive to negative between
0.07z_{i} and 0.13z_{i}.

To preserve mass balance, upward motion must be
compensated by downward motion which is often thought
to take the form of an overall subsiding en-
vironment that is warm and dry and has greater hor-
izontal momentum relative to updrafts. However condi-
tional sampling shows that much of the downward
compensating motion occurs within discrete events
with large perturbation quantities. Images of the con-
vective boundary layer produced with remote sensing
devices such as lidar (Melli et al., 1985), FM-CW radar
(Rowland and Arnold, 1975) and sodar (Taconet and
Weill, 1983) show coherent downdrafts extending from
the mixed-layer top to near the surface. Updrafts push-
ing into the stable layer that caps the mixed layer dis-
place relatively warm and dry air down into the mixed
layer. This air will continue downward if mixing gen-
ers negative buoyancy. Although we cannot deter-
mine the level of origin of the observed downdrafts or
whether they are vertically coherent, the magnitudes
of \( \theta' \), \( q' \) and \( u' \) suggest that they are produced at higher

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levels and are not simply locally compensating motions.

In order to explore the thermodynamic properties of updrafts and downdrafts, following GK we have separated them into four classes, warm/moist, warm/dry, cool/moist and cool/dry, depending on the signs of $\theta'$ and $q'$ averaged over an event. Conditional averages were computed individually for the eight classes of events. The results for area occupied ($dN$) and normalized $q'$, $T_v'$ and $u'$ are shown in Fig. 6. Moist updrafts and dry downdrafts, occupying the greatest area, will be referred to as major classes, while the remaining
classes (dry updrafts and moist downdrafts) will be referred to as minor. The major classes have conditionally averaged $q'$ that is consistent with unmodified parcel displacement in the presence of the composite $q$ profile in Fig. 1. Warm/moist updrafts ($W/M^*$) and cool/moist updrafts ($C/M^*$) occupy the same fractional area at the lowest level. Above this level, $C/M^*$ events increase and $W/M^*$ events decrease in area occupied. The $W/M^*$ events must originate in the near-surface superadiabatic layer and become relatively cool after rising a small distance into the subadiabatic layer. The $C/M^*$ events increase in frequency with height beginning at the top of the superadiabatic layer.

Warm/dry downdrafts ($W/D^*$) are more prevalent than cool/dry downdrafts ($C/D^*$) and increase in area occupied with height for the first three levels. Above the second level $C/D^*$ events decrease with height in area occupied. Downdrafts will not be relatively cool until they enter the superadiabatic layer and therefore are less common relative to their warm counterparts. Although the composite $\theta$ profile in Fig. 1 does not show a superadiabatic layer above the lowest level ($0.13z_c$), there were probably individual days with deep superadiabatic layers capable of producing $C/D^*$ events. For example, a positive heat flux of 5.6 W m$^{-2}$ was measured at 0.13$z_c$ on day 059 indicating the presence of a deep superadiabatic layer, and at this height cool/dry downdrafts occupied nearly as much area as warm/dry downdrafts. Cool/dry downdrafts occupied the smallest area relative to warm/dry downdrafts on day 165, which had the largest negative sensible heat flux observed at 0.32$z_c$. Presumably this day had a rather shallow superadiabatic layer near the surface. The magnitude and sign of the overall sensible heat flux is determined by the balance of drafts of different $\theta'$.

The normalized vertical velocity, $w'/w^*_s$, shows no pronounced variations with height for any of the classes and is therefore not included in Fig. 6. However, the mean value of $w'/w^*_s$, averaged over all levels, varies between classes. Moist updrafts have the largest value of $w'/w^*_s$ with warm events having a slightly larger value (1.16) than cool events (1.08) as is expected due to their greater buoyancy. Dry updrafts have the next greatest $w'/w^*_s$ (0.94). For downdrafts, all classes have approximately the same value of $w'/w^*_s$ (−0.77).

A parcel displaced vertically will also carry a perturbation in horizontal momentum in the presence of vertical wind shear. In Fig. 6d, moist updrafts have negative $u'$ and dry downdrafts have positive $u'$, all tending to decrease with height.

The minor classes, moist downdrafts and dry updrafts, account for only a small fraction of the area occupied by drafts (Fig. 6a). Cool/moist downdrafts ($C/M^*$) are an exception with their area occupied equal to about half the area occupied by cool/moist updrafts ($C/M^*$) at all levels. We believe that $C/M^*$ events originate as $C/M^*$ events which have been partly or wholly forced downward by downdrafts. The reversal of updraft motion by large downdrafts has been documented in lidar motion pictures (E. Eloranta, private communication, 1985). Both $C/M^*$ events and $C/M^+$ events have significantly negative $u'$, a property that a parcel displacement model predicts should occur only for updrafts. Furthermore, the increase with height of area occupied for $C/M^*$ events parallels the trend for $C/M^+$ events. One feature not shared by $C/M^*$ events and $C/M^+$ events is the sign of $T'_c$. Mixing in the process of reversal decreases $q'$ to the point that the negative temperature perturbation determines the sign of $T'_c$. Since $C/M^*$ events that remain positively buoyant are probably short-lived, they make up a small fraction of the total sample of $C/M^*$ events.

The remaining minor classes occupy a very small area (Fig. 6a). Warm/moist downdrafts and cool/dry updrafts produce negative buoyancy flux indicating that these are forced events. Warm/dry updrafts have positive buoyancy and positive $u'$ at the upper levels suggesting they might be reversed warm/dry downdrafts. Mixing with environmental air can produce a change in buoyancy while the parcel still retains its $u'$ signature. Detrainment, where part of a draft is separated, may also contribute to some of the reversed motions and account for the fact that the average size of minor events is smaller than that of their major counterparts.

6. Event contribution to the variances and covariances

In this section, we discuss the contributions of updrafts and downdrafts to the variances and covariances of the meteorological variables. In GK it is shown that the contributions to the flux of $x$ by updrafts (+) and downdrafts (−) is

$$F_x = \frac{1}{t_{tot}} \int_0^{t_{tot}} w'(t) x'(t) f^2(t) dt$$

(2)

where $t_{tot}$ is the total sample time. The results of GK and those in Section 4 indicate that between 75 and 85% of the total fluxes in the lower third of the mixed layer is produced by drafts. In GK it was also shown that of the flux produced by drafts, approximately 95% is due to the mean event properties, the remaining 5% being due to the eddy correlation within events. In this section, we examine the height dependence of the event contributions to the variances and the covariances.

In Table 2 the proportion of the variances and covariances of $w'$, $u'$, $q'$ and $T'_c$ due to downdrafts, the environment and updrafts is given. Since updrafts and downdrafts contain the larger excursions of $w'$, they produce over 75% of $\sigma_w^2$. For the other variables, the drafts produce less than half the variance. If fluctuations were distributed randomly through the time series, the variance contributions would vary as the area occupied,
TABLE 2. Percent contributions, averaged over the lower third of the mixed layer, to the variance and covariance of vertical wind, along-wind momentum, water vapor mixing ratio and buoyancy by downdrafts, the environment (envr), and updrafts.

<table>
<thead>
<tr>
<th></th>
<th>Variance</th>
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<td>Down</td>
<td>Envr</td>
<td>Up</td>
<td>Down</td>
<td>Envr</td>
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<tr>
<td>$q'$</td>
<td>20</td>
<td>55</td>
<td>25</td>
<td>25</td>
<td>20</td>
</tr>
<tr>
<td>$T_v$</td>
<td>18</td>
<td>58</td>
<td>24</td>
<td>22</td>
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</table>

i.e. approximately 20, 65, and 15% for downdrafts, environment, and updrafts, respectively. Table 2 shows that the environment produces 5 to 10% less variance in $u$, $q$ and $T_v$ than implied by the area occupied and that updrafts produce 10% more variance. Downdrafts occupy a greater area but produce less variance than updrafts.

Table 2 also shows that the fluctuations due to drafts are better correlated with $w'$ than are the fluctuations in the environment. Since conditionally averaged $u'$, $q'$ and $T_v$ are approximately twice as large in updrafts as in downdrafts, updrafts are responsible for a little over twice as much flux as downdrafts.

The contributions to the variances and covariances by the event means within each class are calculated by taking the products of conditionally sampled $w'$ with conditionally sampled $u'$, $q'$ and $T_v$. As discussed in GK, the difference between this method and an exact calculation is due to the contributions of eddy correlations within events. The exact calculation uses (2) for covariance and an analogous equation for variance. In Table 3 the contributions to the variances and covariances from the event means are expressed as a percentage of the total variance or covariance within each state. For example, 90% of the vertical velocity variance due to updrafts and downdrafts is produced by the event means. The event mean $w'$ for individual events in the environment is generally small producing a contribution to the environmental $a_w^2$ that is less than 20%. In contrast, the event means of $u'$, $q'$ and $T_v$ in the environment appear to be significantly different from zero since they are responsible for about half the variance contributed by that state. For updrafts and downdrafts, 70 to 80% of the draft variance of $u'$, $q'$ and $T_v$ is produced by their mean properties. Similarly, over 90% of the covariance between $w'$ and $u'$, $q'$ and $T_v$ within updrafts and downdrafts comes from the mean event properties.

Although many values in Table 2 are the same for all heights in the lower third of the mixed layer, there are some exceptions. For example, the proportion of the variance of $w'$ due to updrafts increases from 39% at 0.07 $z_i$ to 48% at 0.32 $z_i$. In addition, the percent contributions to the covariances by updrafts and downdrafts generally increase with height indicating that drafts play a more important role in producing the fluxes with increasing height while the fluxes themselves decrease in magnitude. The normalized standard deviation of $w'$, $\sigma_w/w_{w'}$, increases from 0.62 at 0.07 $z_i$ to 0.71 at 0.32 $z_i$, in parallel with the increase in the magnitudes of the thresholds $w_{th}$. (Fig. 3). The coefficient of kurtosis for $w'$ also increases between the lowest and highest levels, from 3.5 to 4.0. These observations are consistent with the increase in draft contribution to the total fluxes, from 75% at 0.07 $z_i$ to 85% at 0.32 $z_i$.

In terms of the thermodynamic classification of events, the major classes (moist updrafts and dry downdrafts) together produce 80 to 90% of the overall fluxes in agreement with the results of GK. The height variation of the contribution to the fluxes by a thermodynamic class can be inferred from the product of the area occupied and the perturbation quantities in Fig. 6 as discussed in section 4b [see (1)].

7. Partitioning of buoyancy flux

A common assumption in mixed layer models is that production (consumption) of turbulent kinetic energy (TKE) occurs only at levels where the buoyancy flux, $w\Theta_v$, is positive (negative). Stage and Businger (1981) present a model in which production and consumption occur at all levels, an assumption with considerable consequences to the energetics of the layer. Wilczak and Businger (1983, hereafter referred to as WB) explore this issue using tower data obtained in the dry convective boundary layer. We will interpret the present data in a manner similar to WB but with the additional advantage of having knowledge of the scale of the events involved.

In WB, the data are divided into four classes corresponding to the quadrants in a $w\Theta_v$ joint probability distribution. From our thermodynamically classified events, we form analogous classes based on the signs of $w'$ and $T_v$. Following the notation of WB we call updrafts with positive (negative) buoyancy $P^+$ ($N^+$) and downdrafts with negative (positive) buoyancy $P^-$ ($N^-$), where $P$ ($N$) refers to TKE production (consumption) and the superscript + (-) denotes updrafts (downdrafts). In Fig. 6c it can be seen that warm/moist, cool/
moist and warm/dry updrafts all have \( T' > 0 \) and therefore make up the \( P^* \) state. Cool/dry and cool/moist downdrafts and, at the lower three levels, warm/dry downdrafts all have \( T' < 0 \) and therefore make up the \( P^- \) state. Warm/moist downdrafts and warm/dry downdrafts at the highest level constitute the \( N^- \) state and cool/dry updrafts make up the \( N^+ \) state.

Flux intensity is defined as the ratio of the proportion of flux produced in a state to the fraction of time in that state. Wilczak and Businger find that near the surface the \( P^- \) state occupies a greater area but has a smaller flux intensity than the \( P^+ \) state. In data from our lowest level the same relationships are found. The \( P^+ \) events occur 17% of the record and contribute 37% of the buoyancy flux, and \( P^- \) events occupy 13% of the record and contribute approximately 57% of the flux. The flux intensities, 2.2 for \( P^- \) and 4.4 for \( P^+ \), thus differ by a factor of two, in agreement with WB.

A comparison of the properties of the \( N^+ \) and \( N^- \) states shows both similarities and differences between the two datasets. At the lowest level of our data, \( N^+ \) and \( N^- \) occupy fractional areas of approximately 0.6% and 1.2%, respectively, which are much smaller than in WB. However, the fractional areas are relatively invariant with height, which is also the case in WB. The contributions to the buoyancy flux are approximately \( -2\% \) and \( -4\% \) for \( N^+ \) and \( N^- \), respectively, and therefore the flux intensities are equal (both 0.3), in agreement with WB who obtain flux intensities of approximately 0.45 for both states. In WB, \( N^+ \) occupies a greater area and contributes more flux, whereas in our data \( N^- \) dominates the TKE consuming classes.

The agreement between the present analysis and the results of WB on the relative magnitudes of the flux intensities suggests that the energetics of the two boundary layers that produced the data are similar, at least near the surface where the comparison is being made. Many of the differences between the two sets of results are most likely due to the different methods of partitioning the data. Our events are those for which \( w' \) exceeds either a positive or negative threshold, and which have a minimum duration. The remaining 68% of the data that do not satisfy these criteria make up the environment. In WB, thermally direct states occupy 2.4 times the area occupied by our \( P^+ \) and \( P^- \) states and their flux intensities are a factor of two smaller. The \( N^+ \) and \( N^- \) states are most likely small in size with a small associated \( w' \) and therefore generally fall into our environmental classification. However, large events which satisfy the \( N^+ \) and \( N^- \) criteria do occur in our data and, although they occupy a much smaller area, have properties similar to those of WB.

Our results and those of WB support the hypothesis of Stage and Businger (1981) that consumption of TKE occurs all the way down to the surface. However, WB attribute the nonzero consumption (as well as a component of the production) near the surface to convectively driven processes that are constant with respect to height. Randall (1984) states that the results of WB contradict Stage and Businger since this background consumption and production is irrelevant to the TKE balance of the mixed layer and, moreover, because the consumption near the surface is dominated by cold air entrained into accelerating thermals rather than by warm downward moving air as would be expected if the process were related to the mixing of entrained inversion layer air down into the mixed layer.

Our minimum event length criterion does not allow us to detect the negatively buoyant air embedded in updrafts, as proposed by WB. However, we do find TKE consuming events which have spatial scales of the order of 0.12\( z_i \), comparable to the average value of 0.15\( z_i \) for the TKE producing elements. Our TKE consuming events, in contrast to WB, are more often positively buoyant downdrafts than negatively buoyant updrafts. Although \( N^- \) events occupy a much smaller part of the record than their counterparts in WB, their size and characteristics suggest that they are related to entrainment processes as proposed by Stage and Businger.

Randall attempts to reconcile results of WB with the more traditional partitioning model by differentiating between the scales of the principal or flux producing circulation and those of the "background" production and consumption. Our results indicate that this approach is not viable since we find that TKE consuming events have scales that are comparable to those of TKE producing events.

8. Convective mass flux parameterization

A parameterization of vertical energy fluxes in clouds was developed by Betts (1975) and later extended to dry convection in the subcloud layer (Betts, 1976). In the parameterization, the fluxes of dry and moist static energy are represented by the product of a convective mass flux, \( \omega^* \), and a perturbation quantity which is the difference between properties inside and outside convective elements. A profile of \( \omega^* \) was derived by Nicholls and LeMone (1980) using aircraft data obtained over the tropical Atlantic ocean during GATE. The dashed curve in Fig. 7 is from a smooth fit to their results which were derived from

\[
\bar{w'}(r) = \omega^*(x_c - x)
\]

where \( x_c \) is the quantity inside convective elements and \( x \) is the horizontal average over a flight leg. The dashed profile in Fig. 7 applies to both the fluxes of latent heat and sensible heat.

The mass flux parameterization is useful in describing convection in the cloud and subcloud layers and in interpreting flux profiles derived from budget studies (Betts, 1975, 1976; Ebessen, 1975). In GK it was shown that, except in the cases where downdrafts were
the dominant contributors to the flux, the derived values of $\omega^*$ for latent and sensible heat flux were in good agreement with the results of Nicholls and LeMone. In GK, $x_1$ was taken as the updraft conditional average and $x$ was set equal to the area weighted average of the downdraft and environmental conditional averages. In order to extend the parameterization to the cases where downdrafts dominate, an approach was then taken where downdrafts were treated on an equal footing with updrafts. The resulting flux parameterization, containing no free parameters, was given by

$$\bar{w}'x' = d^+w^+x^+ - d^-w^-x^- = d^+w'(x^+ - x^-)$$

(3)

where the environmental contribution is zero since its mean vertical velocity is zero. This parameterization was successful in all cases because, as we have seen in (1), nearly all of the flux produced by drafts results from their mean properties. In GK it was found that in order to get exact agreement with the measured fluxes, the right hand side of (3) had to be multiplied by 1.25. This compensates for the missing contributions from the small-scale eddy correlations within events and from the environment.

In the present case, we generalize (3) to

$$\bar{w}'x' = \omega_{**}^*(x^+ - x^-)$$

(4)

where the mass flux, $\omega_{**}^*$, is once again a free parameter. The profiles of $\omega_{**}^*$ for latent heat, sensible heat and alongwind momentum are shown in Fig. 7. To our knowledge, this is the first time that momentum flux has been successfully parameterized in this way. In all cases, there is a slight trend toward increasing values with height in the range $0.07 \leq z/z_i \leq 0.32$ in agreement with the profile of Nicholls and LeMone. Since $\omega_{**}^*$ is nearly constant with height in the lower third of the mixed layer, the flux profiles are almost entirely determined by the variations of the conditional averages in the difference $x^+ - x^-$. The ratios $\omega_{**}^+/\omega_{**}^*$ and $\omega_{**}^-/\omega_{**}^*$ are approximately constant with height and equal to 0.86 and 0.96, respectively. Although it appears that in our case the mass flux parameter for sensible heat is different from that for latent heat and momentum, this should be viewed with caution since the sensible heat fluxes are small and therefore subject to large uncertainty. The standard errors for the normalized sensible heat flux parameter are of the order of 0.03 which is three times larger than the standard errors for the latent heat and momentum flux parameters. The mass flux parameters for latent heat and momentum shown in Fig. 7 are about 15% smaller than those obtained by Nicholls and LeMone. This is to be expected since our updraft/downdraft difference is presumably larger than their convective element difference which is taken relative to the flight mean.

9. Summary and conclusions

The properties of updrafts and downdrafts in the lower third of the marine atmospheric boundary layer have been investigated using a conditional sampling technique. Our results in the lower third of the mixed layer are an extension of those from a previous experiment flown at a single height over the same region of the Pacific Ocean. While some draft properties such as vertical velocity and area occupied vary little over the height range of our observations, other properties such as draft size and buoyancy perturbation have strong height dependence.

The properties observed are in general agreement with the predictions of a parcel displacement model of vertical mixing. Decreasing mixing ratio with height produces updrafts that are mostly moist and downdrafts that are mostly dry. At the lowest level, updrafts are most often relatively warm indicating the presence of a superadiabatic layer below. A change in sign of the sensible heat flux from positive to negative generally occurs between $0.1z_i$ and $0.2z_i$, despite the positive slope of the $\theta$ profile over all levels flown. This is reflected in a decrease in the area occupied by warm updrafts which have risen out of the superadiabatic layer and an increase in the area occupied by cool updrafts. In the case of dry downdrafts, those that are warm occupy a greater area at all levels than those that are cool.

We have argued that the observed dry updrafts and moist downdrafts for the most part are due to reversed motions. In particular, there is strong evidence that cool/moist downdrafts result from the overturning of cool/moist updrafts. There is also evidence that warm/dry updrafts are the reversed motion of warm/dry downdrafts. The remaining classes, warm/moist downdrafts, cool/dry updrafts and, at the highest level, warm/dry downdrafts, all produce negative buoyancy flux and therefore consume turbulent kinetic energy. At all levels TKE consuming events occur in agreement
with the model of Stage and Businger (1981). At the lowest level these events are more often downdrafts than updrafts, which is opposite to the findings of Wilczak and Businger (1983) in a dry continental boundary layer. Because the scale of these events is comparable to the scale of TKE producing events, they are not exclusively background processes and may be linked to entrainment of inversion layer air, as suggested by the model of Stage and Businger (1981). These issues will be explored further in future work with data obtained through the entire depth of the mixed layer.

Updrafts and downdrafts together produce more variance in $u'$, $q'$ and $T'_e$ per unit area occupied than that produced by the environment, confirming that drafts produce the larger fluctuations in these variables. Downdrafts produce approximately 75% of the fluxes at the lowest level increasing to 85% in the upper levels. This is reflected in an increase in the coefficient of kurtosis for $w'$ with height.

The mean values of $u'$, $q'$ and $T'_e$ within updrafts and downdrafts account for 70% to 80% of the draft variance and over 90% of the draft covariance with $w'$. Thus, the variation of average draft properties, including area occupied, explains to a large degree the behavior of the fluxes with height. For this reason a flux parameterization based on mean event properties works well, including the case of along-wind momentum.

Conditional sampling is clearly an effective method for investigating the properties of the flux-producing updraft and downdraft events in the lower MABL. Future work will apply this technique to data from the upper portions of the mixed layer where updrafts and downdrafts play a critical role in the process of entrainment. When extended throughout the mixed layer and into the cloud layer, the observed draft properties will shed light on the mechanisms of cloud–subcloud layer interactions.

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