Quasi-Lagrangian Measurements in Convective Boundary Layer Plumes and Their Implications for the Calculation of CAPE

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ABSTRACT

Measurements were made to determine the level of origin of air parcels participating in natural convection. Lagrangian measurements of conservative variables are ideal for this purpose. A simple remotely piloted vehicle was developed to make in situ measurements of pressure, temperature, and humidity in the convective boundary layer. These quasi-Lagrangian measurements clearly show that convective plumes originate in the superadiabatic surface layer. The observed boundary layer plumes have virtual temperature excesses of about 0.4 K in a tropical region (Orlando, Florida) and of about 1.5 K in a desert region (Albuquerque, New Mexico). The water vapor contribution to parcel buoyancy was appreciable in Orlando but in Albuquerque was insignificant. These observations indicate that convective available potential energy should be determined by adiabatically lifting air parcels from the surface layer, at screen level.

1. Introduction

The subcloud layer is usually not well mixed and, as a consequence, the air parcel lifting level can have a large influence in the thermodynamic properties of modeled cumulus clouds and in calculations of convective available potential energy [CAPE; Williams and Rennó (1993)]. It is therefore important to determine the level of origin of air parcels that initiate natural convection. Lagrangian measurements in convective plumes are ideally suited for this purpose.

Conventional aircraft sampling does not allow for Lagrangian measurements of an individual convective plume but only provides a statistical picture of plumes. Furthermore, for aircraft flying at around 100 m s^{-1}, as currently used for atmospheric research, corrections are required to compensate for both compressional and viscous heating of the air and for effects of evaporative cooling if the sensor is wet by cloud or precipitation. The correction to the temperature measurements are comparable to the temperature perturbation being measured (i.e., the parcel buoyancy), that is, at least 1 K. Paluch (1979) was the first to show concrete evidence of the occurrence of penetrative downdrafts in cumulus clouds. One probable reason for the success of Paluch's observation was the fact that she used a sailplane to make the measurements. The slow speed of the sailplane (20–30 m s^{-1}) reduced the uncertainty of the temperature measurements to less than 0.5 K, even if the sensors were wet (Lawson and Cooper 1990).

Small aircraft have been explored for in situ meteorological measurements. Konrad et al. (1970) used a small radio-controlled airplane to make temperature and humidity soundings from the surface up to a height of 800 m. Bluestein (1990) suggested that a small aircraft flying patterns under a thunderstorm would be ideal for sampling the updraft air. Holland et al. (1992) and McGeer and Holland (1993) reported on the development of a small airplane for atmospheric measurements. They envision a fully autonomous airplane costing over $10,000 and capable of ranging over 7000 km in flights lasting up to 4 days.

We developed a simple remotely piloted vehicle (RPV) capable of making quasi-Lagrangian measurements of pressure, temperature, and humidity in convective boundary layer plumes. We refer to the measurements as quasi-Lagrangian because the RPV can stay within the plume, but it sinks at about 0.5–1.0 m s^{-1} relative to the plume. The slow speed of our RPV (10–20 m s^{-1}) virtually eliminated the need for corrections in the temperature measurements (Lawson and

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TABLE 1. Technical specifications for the RPV.

Dimensions	
Wingspan (m)	2.67
Length (m)	1.18
Wing area (m^2)	0.44
Weights	
Empty weight (kg)	2.00
Radiosonde (kg)	0.03
Nicad battery pack (kg)	0.50
Lithium battery pack (kg)	0.20
Gross weight (kg)	2.73
Performance	
Minimum speed (m s^{-1})	10
Maximum speed (m s^{-1})	50
Thermaling speed (m s^{-1})	15
Minimum sink speed (m s^{-1})	0.5
Lift/drag ratio	20
Turning radius (m)	50
Engine running time (min)	3

Cooper 1990). We used the RPV to make quasi-Lagrangian measurements in summer boundary layer plumes in a tropical environment (Orlando, Florida) and in a desert environment (Albuquerque, New Mexico). We also used both neutrally buoyant and tethered balloons with suspended radiosondes to make boundary layer observations. To the best of our knowledge those are the first quasi-Lagrangian measurements made in convective updrafts.

Our observations clearly show that the convective boundary layer plumes originate in the superadiabatic surface layer. These observations are in agreement with soundings made with radiosonde balloons entrained in updraft cores of thunderstorms (Davies-Jones 1974). Davies-Jones' observations show that the strong updraft air originates in the surface layer and ascends virtually undiluted. These results suggest that CAPE should be determined by adiabatically lifting air parcels from the surface layer, at the screen level. This procedure implies substantial values for CAPE, even for reversible ascent with no ice processes included. Our observations are discussed in sections 3 and 4. We discuss the advantages and disadvantages of each sounding method in section 5. In section 6 we discuss the implications of our observations to the computation of CAPE.

2. Description of the RPV and the balloon probes

We developed a very economical (costing less than \$1,000, excepting costs of the radiosonde equipment) reusable RPV for in situ boundary layer measurements. Our goal was to build a small sturdy off-the-shelf model sailplane (weighing less than 5 kg) carrying a standard Vaisala radiosonde to make high quality quasi-Lagrangian measurements of pressure, temperature, and humidity in boundary layer plumes. The radio-controlled RPV is flown within visual range by a pilot at the ground, and the measurements are telemetered to the radiosonde receiver at the ground station. The RPV is electrically powered, portable, and can be easily disassembled for transport. Table 1 displays the technical characteristics of the RPV, while the equipment necessary for its construction and operation is listed in Table 2. The technical data for the radiosonde sensors are shown in Table 3.

A similar Vaisala radiosonde was also used attached to a tethered balloon whose height was controlled with a hand-held winch with 1000 m of line. This assembly allowed for repeated soundings of the boundary layer. Finally, measurements were made with untethered neutrally buoyant balloons with a Vaisala radiosonde attached. During initial flight tests with the RPV, we experienced serious problems with the original Vaisala radiosonde battery, which produced excess heat. When in contact with the RPV radio control system within the cockpit, this heating led to the complete failure of the control system. The solution (suggested by K. Goss of Vaisala Inc.) was to use lithium batteries to power the onboard radiosonde.

On a typical mission we would wait for the first signs of a convective plume at the surface (wind gusts, soaring birds, rising dust) and then hand-launch the RPV in the plume's direction. Assisted by the electrical engine, we would direct the RPV to the area of best lift, turn off the engine, and begin circling within the plume (Fig. 1). The RPV was centered in the plume by visually optimizing its rate of climb. After reaching the limit distance for visual control (about 800 m for an overhead plume in the clear air of the New Mexico desert), we would fly the plane out of the plume and begin to spiral it downward.

3. The observations

The standard Vaisala radiosonde was used to make measurements of pressure, temperature, and humidity

TABLE 2. Equipment used in the construction and operation of the RPV.

Component	Cost (\$)	Remarks
Graupner ASW22 B270	220	Electric powered sailplane (GR4264)
Motor/prop	55	Direct drive speed 700 Turbo (GR1165)
Motor control	60	Power switch 25 (GR3879)
Battery charger	80	ac/dc variable rate charger (AST115)
Nicad battery pack	45	8 cell 1400 MA h battery pack (B081400R)
Lithium battery pack	10	2 cell, camera battery
Radio control	250	Futaba 4 channels (PCM)
Radiosonde	150	Vaisala RS80 radiosonde
Radiosonde receiver	10,000	Vaisala
Personal computer	1,000	To store the radiosonde data

in boundary layer plumes. We computed the RPV's height (altitude) by integrating the hydrostatic equation, which is a good approximation even in strong updraft cores (Davies-Jones 1974). Thus, the height z at a pressure level p, is given by

$$z = -R_d \int_{p_0}^{p} T_v d \ln p, \tag{1}$$

where R_d is the specific gas constant of the dry air, p_0 is the surface pressure [$p_0 \equiv p(z = 0)$], and T_v is the virtual temperature, given by

$$T_v = (1 + 0.608r)T, \tag{2}$$

where T (K) is the absolute temperature and r (kg kg^{-1}) is the water vapor mixing ratio. There is an inherent noisiness in the pressure measurements (Table 3) that leads to an uncertainty of about 10 m in the estimation of the RPV's height.

The water vapor mixing ratio was computed from the radiosonde relative humidity by

$$r = \mathrm{RH} r_{\mathrm{sat}}(p, T), \tag{3}$$

where

$$r_{\mathrm{sat}} = 0.622 \frac{e_{\mathrm{sat}}(T)}{p} \tag{4}$$

and e_{sat} (Pa) was computed by the Wexler empirical formula

$$e_{\mathrm{sat}}(T) = 611.2 \exp\left[\frac{17.67(T - 273.15)}{T - 29.65}\right]. \tag{5}$$

Finally, the potential temperature was computed by

$$\theta = T\left(\frac{100\,000}{p}\right)^{0.286}, \tag{6}$$

where p (Pa) is the pressure and the virtual potential temperature was computed by

$$\theta_v = T_v\left(\frac{100\,000}{p}\right)^{0.286}. \tag{7}$$

Both these quantities are conserved in dry-adiabatic parcel displacements.

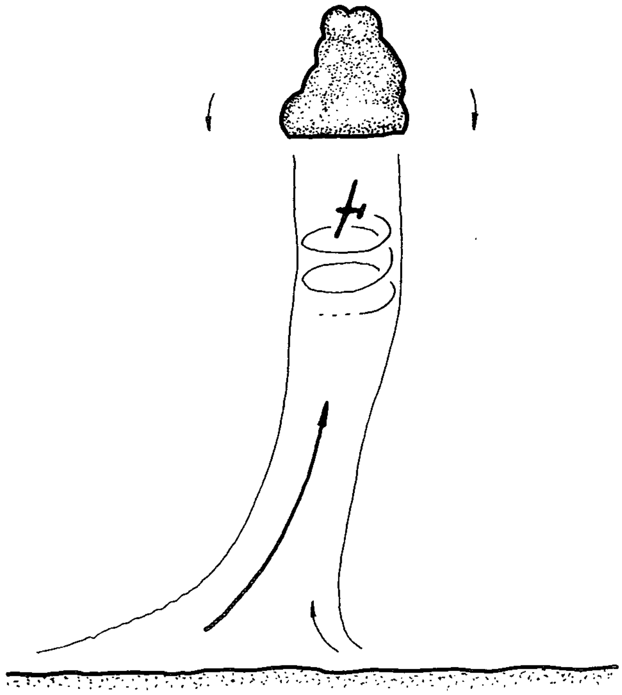

FIG. 1. Schematic illustration of a typical sounding of a boundary layer plume by the RPV.

a. Tropical plumes observed with the RPV

1) CASE 1

Figures 2a–d display the observations made with an RPV flight on 3 August 1992 near Orlando, Florida. The observations in ascent (inside the plume) are represented by dots, while observations on descent (outside of the plume) are represented by crosses. At the time of the flight, about 1100 local time, small cumulus clouds were present in the area. Figure 2a shows that, initially, the RPV climbed slowly, and then about 30 m above the ground got well centered in the plume and climbed steadily to about 360 m. Then, the RPV was flown out of the plume and down to the ground, sampling the ambient air. From Figure 2a, the average rate of climb for the RPV is estimated to be 2.0 m s^{-1}. Since, the RPV's sink speed in still air is about -1 m s^{-1}, the vertical velocity of the observed plume is 3.0 m s^{-1}. By the same reasoning, we estimated that

TABLE 3. Technical data for the radiosonde sensors (Vaisala 1992).

Parameter	Pressure (capacitive aneroid)	Temperature (capacitive bead)	Humidity (thin-film capacitor)
Measuring range	1060 to 3 hPa	+60° to −90°C	0%–100% RH
Resolution	0.1 hPa	0.1°C	—
Accuracy	0.5 hPa	0.2°C	2% RH
Lag	—	2.5 s (at 6 m s^{-1})	1.0 s (at 6 m s^{-1})

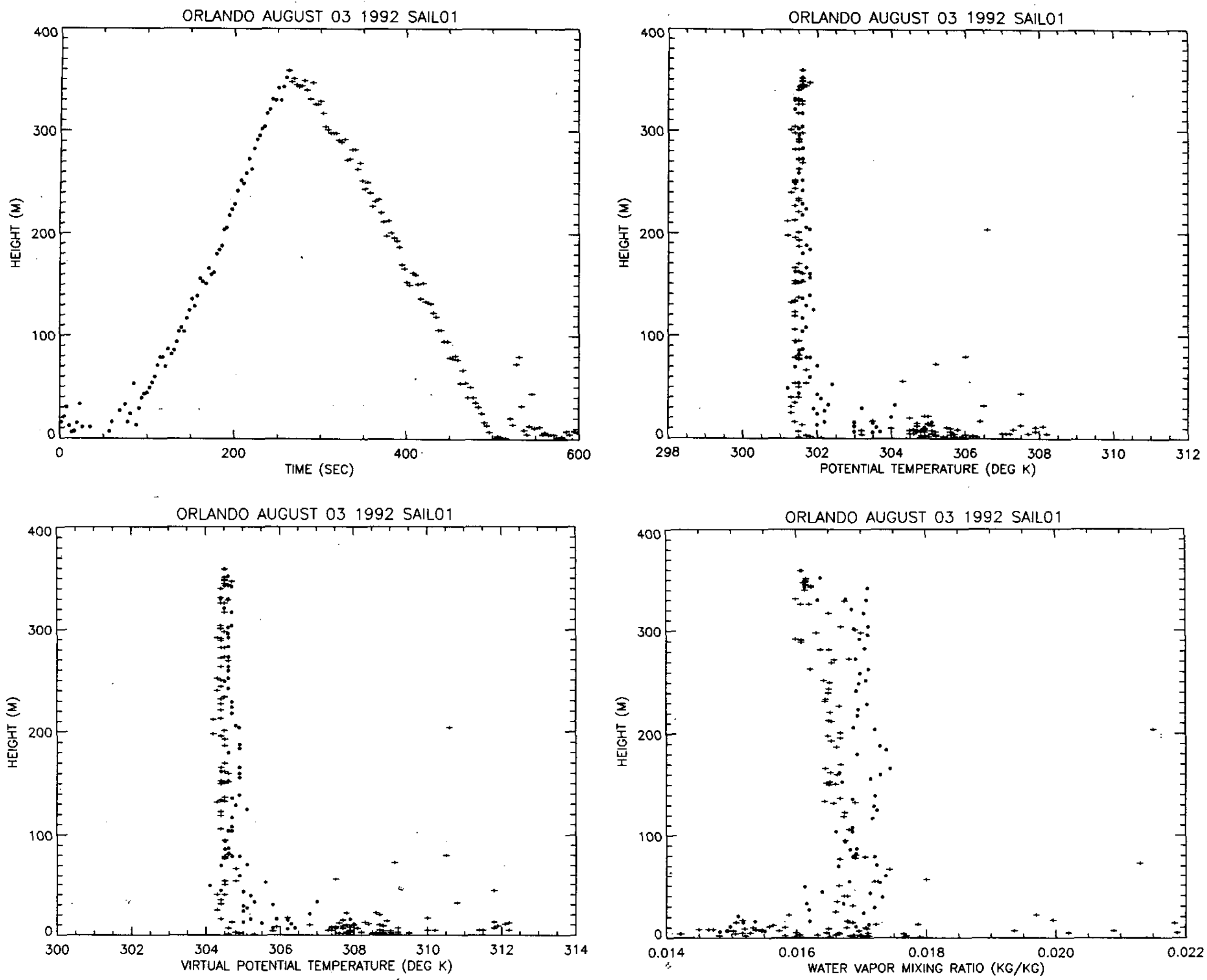

FIG. 2. Observations made with an RPV flight near Orlando, Florida, on 3 August 1992: (a) altitude, (b) potential temperature, (c) virtual potential temperature, (d) water vapor mixing ratio. The observations during ascent (inside the plume) are represented by dots; observations on descent (outside of the plume) are represented by crosses.

the air outside the plume was subsiding at -0.5 m s^{-1}. Figures 2b and 2c show that in the tropical plumes, the moisture excess contributes to about 50% of the plume's buoyancy. Figure 2c shows that except below 50 m, where the plume's buoyancy is as large as about 0.7 K, the buoyancy is only 0.4 K from 100 m to approximately 200 m. Above this level, the buoyancy is only about 0.2 K, and substantially smaller than the values observed over the desert near Albuquerque, New Mexico (1.5 K, see sections 3c and 3d). Finally, Fig. 2d shows that this tropical plume is always moister than the ambient air.

2) CASE 2

This case is concerned with the observations made with an RPV flight on 6 August 1992. At the time of the flight, 1030 local time, small cumulus clouds were present in the area. We do not show figures for this case. The average vertical velocity of the observed plume is estimated to be 3.9 m s^{-1}, and the ambient air is estimated to be subsiding at -0.4 m s^{-1}. The plume's buoyancy is not higher than about 0.2 K. Again, the plume's buoyancy is substantially smaller than the values observed in the desert plumes (1.5 K, see sections 3c and 3d). Finally, the moisture excess contributes to about 50% of the plume's buoyancy.

b. Tropical plumes observed with a tethered balloon—Case 3

Figures 3a–d display the observations made with a tethered balloon sounding on 25 July 1992. At the time of the sounding, 1130 local time, small cumulus clouds were present in the area. At around time 2740 s a convective plume rose through the altitude of the tethered balloon, causing a perturbation in temperature and moisture lasting about 40 s. At this time, vultures were soaring just overhead of the tethered balloon, and a rush of air was observed at the surface. A small cumulus cloud appeared above the balloon's location a short time later. Figure 3a shows that at the time that the plume was observed, the tethered balloon was at approximately 100 m above ground. The convective plume produced a temperature perturbation of 0.8 K (Fig. 3b) and a water vapor mixing ratio perturbation of 1.5 g kg^{-1} (Fig. 3d), which resulted in a virtual temperature perturbation of 1.1 K (Fig. 3c).

c. *Desert plumes observed with the RPV*

1) CASE 4

Figures 4a–d display the observations made with an RPV flight on 3 August 1993. At the time of the flight, about 1300 local time, small cumulus clouds were present in the area. Figure 4a shows that, initially, the RPV climbed slowly until an altitude of about 50 m above the ground. Then, the RPV became well centered in the plume and climbed steadily to about 320 m. It stayed at this height for about 150 s, perhaps due to a reduction in the plume's vertical motion, but more probably due to the RPV's wander from the center of the plume. The RPV then climbed another 150 m and again stopped climbing for another 150 s. Finally, the RPV climbed steadily again to about 800 m. During the RPV's ascent, we observed the formation of a small cumulus directly over the sampled plume. Since at this height the RPV was approaching the limit of visual range, the RPV was directed out of the updraft. Once outside the updraft, a spiral pattern was flown down to the surface. On descent, at about 650 m, the RPV briefly encountered another updraft. The average vertical velocity of

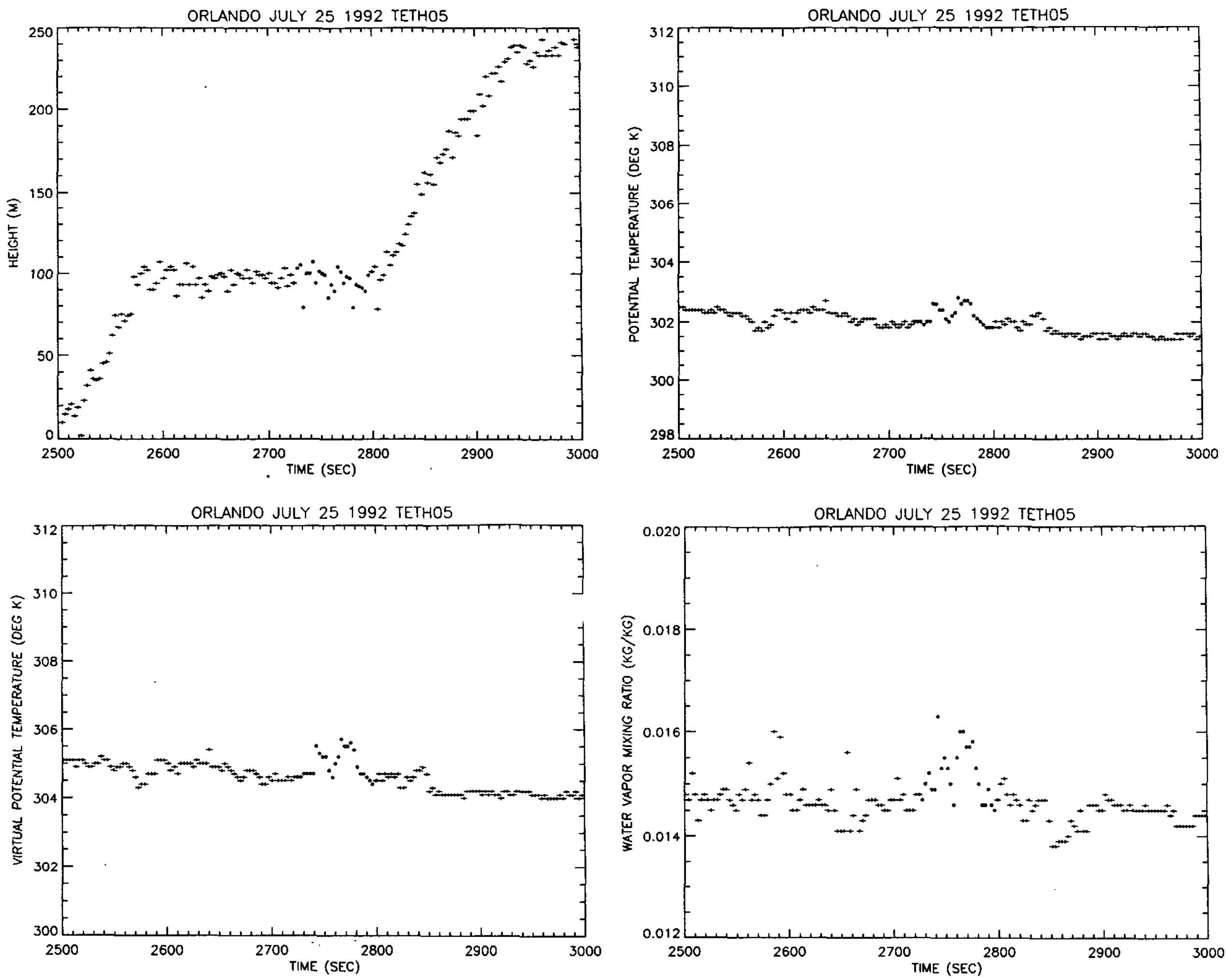

FIG. 3. Sounding made with a tethered balloon near Orlando, Florida, on 25 July 1992: (a) altitude, (b) potential temperature, (c) virtual potential temperature, (d) water vapor mixing ratio. The observations during the plume's passage at a height of about 100 m are represented by dots; observations outside of the plume are represented by crosses.

the observed plume was 3.5 m s^{-1}, and the air outside the plume was subsiding at -1.7 m s^{-1}.

Figure 4b shows that the plume's potential temperature, approximately 317 K, was about 6 K cooler than the maximum air temperature observed at the surface. Figures 4b and 4d reveal that the slightly larger buoyancy at about 100 m was partly due to the plume's larger moisture excess at low levels (i.e., buoyancy caused by a virtual temperature effect). Figures 4c and 4d show, respectively, that the plume's maximum buoyancy was about 1.5 K and that the plume's water vapor mixing ratio was about 5%–15% larger than the environment's. The plume's potential temperature excess, about 1.3 K, does not show any systematic change with height. A substantial reduction in the plume's buoyancy was observed at the heights where the RPV stopped climbing, at around 320 and 470 m as shown in Fig. 4c. The water vapor mixing ratio (shown in Fig. 4d) for the desert environment is only about half as large as values observed near Orlando.

2) Case 5

This case is concerned with the observations made with a second RPV flight on 3 August 1993. At the time of the flight, about 1400 local time, small cumulus clouds were present in the area. We do not show figures for this case. The vertical velocity of the plume is estimated to vary between 2.6 and 3.8 m s^{-1}. The air outside the plume is estimated to be subsiding at about -3.0 m s^{-1}. The virtual potential temperature presents a maximum value of 1.5 K near an altitude of 180 m. Interestingly, the profile of water vapor mixing ratio (not shown) shows that this desert plume is slightly

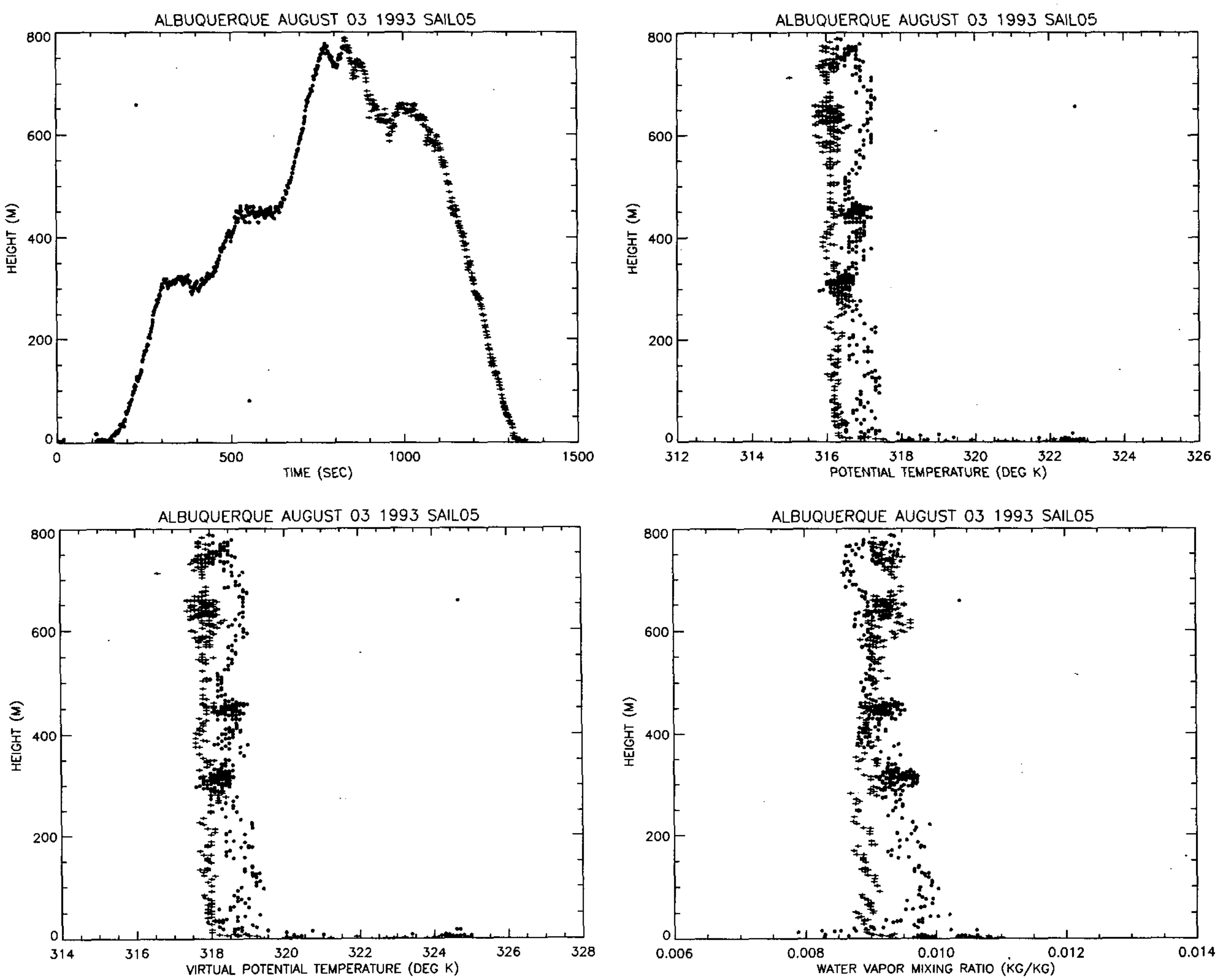

FIG. 4. Observations made with an RPV flight near Albuquerque, New Mexico, on 3 August 1993: (a) altitude, (b) potential temperature, (c) virtual potential temperature, (d) water vapor mixing ratio. The observations during ascent (inside the plume) are represented by dots; observations on descent (outside of the plume) are represented by crosses.

less moist than the ambient air (see section 4 for a discussion).

d. Desert plumes observed with a neutrally buoyant balloon—Case 6

Figures 5a–d display the observations made with a neutrally buoyant balloon on 4 August 1993. At the time of the balloon launch, about 1500 local time, small cumulus clouds were present in the area. Figure 5a shows that the balloon climbed steadily from 10 m above the ground to about 950 m. The balloon ascended at approximately 2.6 m s^{-1} to the 210-m level and then increased its rate of climb to about 6.3 m s^{-1} and stayed at this value until the 800-m level. A small cumulus cloud was observed to form just above the balloon. The balloon climbed to about 950 m above the ground and then descended to about 720 m at about -1.5 m s^{-1}. From this level, the balloon initially climbed unsteadily for about 150 m and then climbed steadily at 3.2 m s^{-1} to its maximum height, about 1050 m above the ground. From this level, the balloon descended at -1.0 m s^{-1} to about 825 m and then at about -2.0 m s^{-1}. At approximately 200 m, the balloon traversed a weak plume, climbing about 50 m, and then finally sank to the surface, approximately 5 km from the launching site.

Figure 5b shows that the plume's potential temperature, approximately 315.5 K, is about 6.5 K cooler than the maximum air temperature observed near the surface. Figure 5c shows that except below 150 m, where the plume's buoyancy is as high as about 2.5 K, the buoyancy is about 1.5 K from 150 m to approximately 600 m. Above this level, the buoyancy is about 1.0 K. The maximum updraft (6.3 m s^{-1}) occurred in the environment where the buoyancy was about 1.5 K. Finally, Fig. 5d again shows that the plume's water vapor mixing ratio might be smaller than the ambient mixing ratio (see section 4 for a discussion).

4. Summary of the observations of plume buoyancy

A comparison of thermal buoyancy and the buoyancy due to water vapor excess only (BWVE) is shown for all case studies in Table 4. In general, in the tropical plumes, the moisture excess contributes an appreciable fraction of the plume's buoyancy. In the desert plumes, the moisture contribution to buoyancy is of the order of 15% for the morning desert plumes and is negligible or even negative for afternoon desert plumes. Therefore, while the tropical plumes are often moist or "virtual plumes," the desert plumes are just thermal plumes or "warm plumes."

In an earlier discussion of the desert plumes, it was noted that the water vapor mixing ratio within the plumes often was less than the ambient values. We speculate that for the afternoon desert plumes, the water vapor mixing ratio might be smaller than the ambient mixing ratio because, in the early morning, "virtual plumes" moisten the boundary layer by the vertical transport of the water vapor condensed on the surface during the previous night. At the same time the surface layer is dried by the replacement of the moist air by dry air from the top of the convective layer, brought down by penetrative downdrafts. As the day progresses, the available moisture at the desert surface is depleted, and the surface becomes very hot and dry. With this condition plumes might become hotter and dryer than the ambient mixed layer air (Fig. 4d clearly shows that near the surface the air inside the convective plumes is dryer than the ambient air).

It is important to note that in spite of the plume's buoyancy, its vertical velocity is approximately constant, indicating that the plume is in dynamic equilibrium. Thus, our observations are in contradiction with results of calculations based on the inviscid parcel method, where buoyancy implies acceleration. Our observations lead to the following question: *Is the eddy viscosity parameter—necessary to balance the observed buoyancy—small enough so that it does not produce a complete mixing of the plume with the environment (this would be in contradiction with our observations, which do not show a substantial mixing)?* Assuming that buoyancy is balanced by the turbulent viscosity, the momentum equation for a rising air parcel can be written as $g\Delta T/T - \nu_{\rm turb}\nabla^2 v = 0$, where g is the gravity acceleration, ΔT is the plume's temperature excess over the environment, $\nu_{\rm turb}$ is the turbulent viscosity, and v is the updraft velocity. We take $\Delta T = 2$ K, $T = 300$ K, and $\nabla^2 v = 1 \times 10^{-3}$ m^{-1} s^{-1}, where $\nabla^2 v \sim \Delta v/l^2 \approx 10/100^2$, Δv is magnitude of the velocity difference between updrafts and downdrafts, and l is the characteristic plume's scale length. Thus, for balance, we must have $\nu_{\rm turb} = 65$ m^2 s^{-1}. The eddy viscosity based on scaling analysis of fully developed turbulence is given by $\nu_{\rm turb} \sim l_{\rm eddy}\Delta v_{\rm eddy}$. Thus, assuming $\Delta v_{\rm eddy} \sim \Delta v \approx 10$ m s^{-1} (i.e., turbulent eddies are as energetic as the plume), we find that eddies of the scale length of 10 m are enough to produce $\nu_{\rm turb} \approx 100$ m^2 s^{-1} and balance the buoyancy of 2 K. The turbulent mixing time $\tau_{\rm mix}$ can be written as $\tau_{\rm mix} \sim l^2/\kappa_{\rm turb}$, where the turbulent diffusivity is $\kappa_{\rm turb} \sim \nu_{\rm turb}$. In this case, the mixing timescale is $\tau_{\rm mix} \approx 100$ s. The convective timescale is given by $\tau_c \sim h/v$, where h is the plume's height. Thus, the convective timescale is $\tau_c \approx 100$ s. Since $\tau_{\rm mix} \approx \tau_c$, it is conceivable that the eddy viscosity is large enough to produce a balance between buoyant and viscous forces, and yet not produce a complete mixing of the plume with the environment.

5. The optimal probe for boundary layer measurements

Buoyancy in boundary layer plumes is obviously a delicate quantity (i. e., the measurements must be accurate to the order of 0.1 K for it to be assessed properly), with values of the order of 1 K. Slow-moving in situ probes are essential for obtaining accurate measurements. We argue that, for sampling of boundary layer plumes, the RPV has advantages over tethered and free neutrally buoyant balloons in measuring buoyancy. The RPV is capable of making quasi-Lagrangian measurements in the convective boundary layer plumes to sample the environmental air in the neighborhood of the sampled plume. Since the RPV is fully controllable, it is the best method for quasi-Lagrangian measurements in boundary layer plumes.

We argue that the untethered neutrally buoyant balloon with a radiosonde attached is the preferred method for sampling the near-surface air over an extended environment, where the boundary layer plumes originate. Since it is not controllable, the untethered neutrally buoyant balloon is less valuable for measuring buoyancy in boundary layer plumes. Bluestein (1990) reported on various attempts to sample thunderstorm updrafts with a radiosonde. However, in most attempts no

TABLE 4. Plume's maximum buoyancy due only to temperature excess (temperature buoyancy) and due only to water vapor excess (BWVE).

Case no.	Temperature buoyancy (K)	BWVE (K)	Environment/probe
1	0.2	0.2	Tropical/RPV
2	0.1	0.1	Tropical/RPV
3	0.8	0.3	Tropical/tether
4	1.3	0.2	Desert/RPV
5	1.5	0.0	Desert/RPV
6	1.4	0.0	Desert/neutral

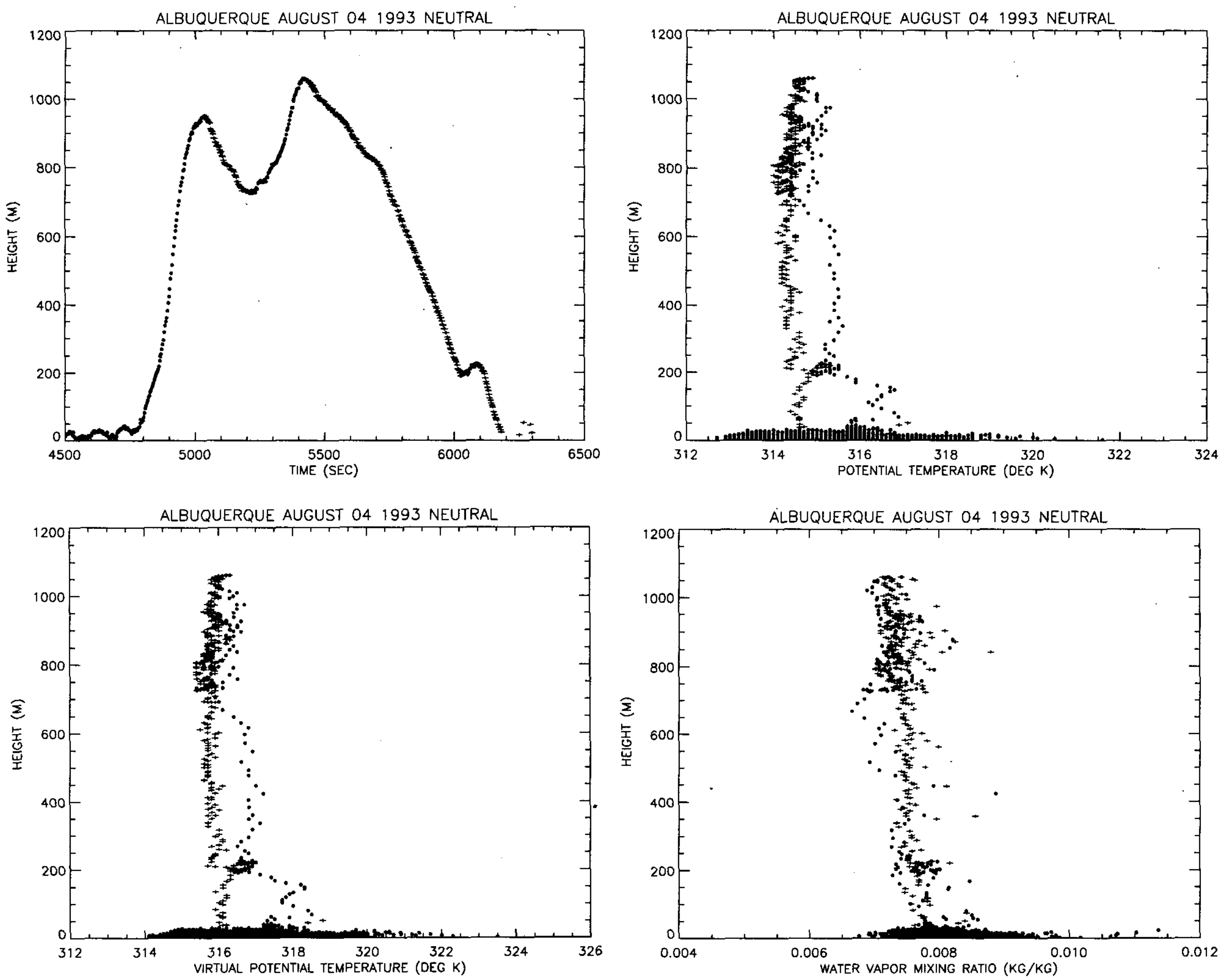

FIG. 5. Observations made with a neutrally buoyant balloon near Albuquerque, New Mexico, on 4 August 1993: (a) altitude, (b) potential temperature, (c) virtual potential temperature, (d) water vapor mixing ratio. The observations during ascent (inside the plume) are represented by dots; observations on descent (outside of the plume) are represented by crosses.

updraft data were collected because the balloon failed to enter the updraft. Bluestein suggested that an RPV flying patterns under the thunderstorm would be ideal for such measurements.

The tethered balloon allows for repeated soundings of the boundary layer and is well suited to sample low-level air closely in space and time. It might be the best sounding method to follow the diurnal evolution of the thermodynamic properties of the boundary layer.

6. Discussion

The air parcel lifting level has a considerable influence on calculations of CAPE (Williams and Rennó 1993). Consequently, it is very important to determine the origin level of air parcels participating in natural convection. Betts (1982) and Xu and Emanuel (1989) have argued that the tropical atmosphere is nearly neutral to adiabatic parcel ascent from the subcloud layer. However, in these studies, parcels were lifted from the top of the subcloud layer rather than the surface, as suggested by our observations. Even when the less unstable subcloud air was used, a significant positive buoyancy was observed. Xu and Emanuel (1989) attributed this observed departure from moist neutrality to instrument error. However, if this were the case, one would expect a random error around the mean, a result not found. The study of Williams and Rennó (1993) showed, in the same part of the world as studied by Xu and Emanuel (1989), that the CAPE values are substantial (of the order of 1000 J kg^{-1}) when the air parcels are lifted from the surface layer, even for reversible ascent with no ice processes included.

Our observation that the air participating in natural moist convection originates in the surface layer is also supported by Doppler radar, surface layer, aircraft, and radiosonde observations. The Convection Initiation and Downburst Experiment, (Wilson et al. 1988) showed that at least 80% of the storms forming on the Colorado plains were initiated along preexisting radar detected boundaries. Radar observations of summer convection in Florida suggest that the lifting of undilute surface layer parcels by mesoscale boundary layer features, such as sea breezes and gust fronts, play a very important role in the initiation of precipitating convection (D. Boccippio 1994, personal communication). Tower measurements in the surface layer during the passage of sea-breeze fronts clearly show signs of the convective updrafts at heights as low as 11 m (Chiba 1993); below this level the plume is not well defined. Furthermore, Marwitz (1973) and Fankhauser (1976) found from aircrafts flights under Colorado thunderstorms that updraft air originated near the surface. Observations made with a motorglider, over the Eyre Peninsula in South Australia, also lead to the conclusion that convective plumes originate in the surface layer (Williams and Hacker 1992). Davies-Jones (1974) studied soundings in which the radiosonde balloons where entrained in strong updraft cores of Oklahoma thunderstorms. All soundings indicated that the updraft air originates in the surface layer and ascends virtually undiluted. Soundings under thunderstorms in Texas also show evidence that the updraft air originates near the surface (Bluestein 1990). Finally, there is evidence over the ocean that convective plumes originate in the marine surface layer (Phelps and Pond 1971; Antonia et al. 1979).

The fact that soaring birds frequently are observed to soar from a few meters above the surface to cloud base (Scorer 1954) also supports the idea that the air participating in natural moist convection originates in the surface layer. Sailplane pilots frequently soar from around 100 m above the surface to cloud base (Reichmann 1978). Reichmann (1978) reports a flight in the 1965 German National Championship: *"Over flat, smooth country–ideal for landing–my course parallels a highway. At 600 feet, the air comes to life for the first time. I fly S-turns but find nothing worth circling in. By 450 feet I'm sure there must be a thermal developing somewhere nearby, but at my low altitude—I have already picked a field to land in—I won't be able to use it unless I can catch it right where it triggers. The landscape looks completely flat and featureless. On the downwind leg, at perhaps 350 feet, I decided—more so that I can say that I tried than with any hope for success—to make a slight jog over a little cairn, a few stones piled up with a stake at the top: a surveyor's monument, perhaps ten or fifteen feet tall. And, in fact, I feel a surge, I start a left circle, make sure after the first turn that I can still make it to my landing field—I still lose 30 to 50 feet of altitude initially, but then center the lift better, hold my altitude, until the lift finally improves and I get away in a strong 400 feet per minute thermal. . . . Such experiences are far from rare. I have made it a habit to fly over every possible triggering point before I have finally to commit myself to landing. . . ."* Reichmann concludes his observations by stating that the more uniform both the air and the surface, the smaller the irregularities need to be in order to work as triggering points for convective plumes. The fact that the triggering points are at the surface are further evidences that convective plumes originate in the superadiabatic surface layer.

The observation that the air participating in natural moist convection originates in the surface layer is not surprising. More than half of solar radiation energy absorbed at the surface is transferred to the atmosphere as both sensible and latent heat. This energy is vertically transported by the convective motions, and then radiated to space.

7. Plans for the future

The small size and low weight of the RPV makes it sufficiently durable that even severe atmospheric turbulence does not cause structural damage. Furthermore, the RPV is sufficiently inexpensive to allow for occasional losses. Therefore, it shows considerable promise for measurements in cumulus clouds and perhaps even in thunderstorms.

In the next phase, we plan to install a simple autopilot in our RPV, so that in-cloud flights will be possible. With this capability we will be able to make quasi-Lagrangian measurements beginning in the surface layer and terminating at cloud-top level in fair weather cumulus clouds. Later, we plan to include a GPS and a more stable pressure sensor to enable us to make detailed three-dimensional maps of our measurements.

In a later phase we will explore the possibility of constructing a simple "smart" autopilot to keep the RPV centered in some specified thermodynamic field. For example, if we choose the maximum virtual potential temperature as the target field, the RPV would be centered in a given boundary layer plume. Such an autopilot would be extremely useful for quasi-Lagrangian measurements within a cloud.

8. Summary

Observations using both RPV and balloon-borne sondes usually show that the air participating in natural moist convection originates in the superadiabatic surface layer. The observed convective boundary layer plumes have virtual temperatures in excess of 0.4 K in a tropical environment and 1.5 K in a desert environment. This temperature excess is approximately constant throughout the convective boundary layer. Furthermore, we discovered that while the tropical plumes are moist or "virtual plumes," the desert plumes are just "warm plumes."

Since these results have shown that the moist unstable air participating in boundary layer convection originates in the superadiabatic surface layer, they indicate that CAPE should be determined by adiabatically lifting air parcels from this level. This procedure has also been suggested by Doswell and Rasmussen (1994) in efforts to standardize the calculation of CAPE. The resulting values of CAPE are substantial, even if reversible ascent is assumed with no ice process included.

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