Vertical Velocity in Oceanic Convection off Tropical Australia

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(Manuscript received 21 September 1993, in final form 10 March 1994)

ABSTRACT

Time series of 1-Hz vertical velocity data collected during aircraft penetrations of oceanic cumulonimbus clouds over the western Pacific warm pool as part of the Equatorial Mesoscale Experiment (EMEX) are analyzed for updraft and downdraft events called cores. An updraft core is defined as occurring whenever the vertical velocity exceeds 1 m s^{-1} for at least 500 m. A downdraft core is defined analogously. Over 19 000 km of straight and level flight legs are used in the analysis. Five hundred eleven updraft cores and 253 downdraft cores are included in the dataset.

Core properties are summarized as distributions of average and maximum vertical velocity, diameter, and mass flux in four altitude intervals between 0.2 and 5.8 km. Distributions are approximately lognormal at all levels. Examination of the variation of the statistics with height suggests a maximum in vertical velocity between 2 and 3 km; slightly lower or equal vertical velocity is indicated at 5 km. Near the freezing level, virtual temperature deviations are found to be slightly positive for both updraft *and* downdraft cores. The excess in updraft cores is much smaller than that predicted by parcel theory.

Comparisons with other studies that use the same analysis technique reveal that EMEX cores have approximately the same strength as cores of other oceanic areas, despite warmer sea surface temperatures. Diameter and mass flux are greater than those in GATE but smaller than those in hurricane rainbands. Oceanic cores are much weaker and appear to be slightly smaller than those observed over land during the Thunderstorm Project.

The markedly weaker oceanic vertical velocities below 5.8 km (compared to the continental cores) cannot be attributed to smaller total convective available potential energy or to very high water loading. Rather, the authors suggest that water loading, although less than adiabatic, is more effective in reducing buoyancy of oceanic cores because of the smaller potential buoyancy below 5.8 km. Entrainment appears to be more effective in reducing buoyancy to well below adiabatic values in oceanic cores, a result consistent with the smaller oceanic core diameters in the lower cloud layer. It is speculated further that core diameters are related to boundary layer depth, which is clearly smaller over the oceans.

1. Introduction

The effects of cumulus convection must be accounted for in most large-scale numerical models, especially general circulation models, since it is of a scale too small to be explicitly resolved. One can deduce the effects of convection by assuming that the convection accounts for the residual in large-scale budgets of heat and moisture based on observational data. The apparent properties of the convection can then be found given a type of cloud model. Such an approach was successfully used by Arakawa and Schubert (1974) and subsequently improved upon by many investigators. The improvements have usually been based upon observations—for example, the introduction of the effects of downdrafts (e.g., Johnson 1976), the tilt of the convective elements from the vertical (e.g., Cheng 1989a,b), and attempts to separate out the effects of mesoscale from convective-scale motions.

A second approach for studying the effects of convection is to look at the convective elements themselves. LeMone and Zipser (1980) and Zipser and LeMone (1980) examined the properties of tropical oceanic convective updraft and downdraft cores in the Global Atmospheric Research Program's Atlantic Tropical Experiment (GATE), finding their diameters surprisingly small and their strengths surprisingly weak when compared to updraft and downdraft cores en-

* The National Center for Atmospheric Research is sponsored by the National Science Foundation.

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countered in the Thunderstorm Project (Byers and Braham 1949). The results were supported in later papers by Jorgensen et al. (1985) and Jorgensen and LeMone (1989) in studies of hurricane rainbands and subtropical oceanic systems, respectively.

This study uses similar methods to examine the properties of vertical drafts in the convective region of oceanic mesoscale convective systems (MCSs) observed during the Equatorial Mesoscale Experiment (EMEX) (Webster and Houze 1991), which was conducted during January and February 1987 in the oceanic regions of northern Australia, part of the western Pacific warm pool. The data used in this study are described in section 2. Section 3 details the results of this study and the comparisons of warm pool convection to the other areas previously studied. Section 4 presents some discussion of the results, and section 5 summarizes the results and conclusions of this study. LeMone et al. (1994) describe the effects of filtering on the statistics of convective cores presented herein.

TABLE 1. Summary of flight legs used in the study.

			Number of	
Height (m)	Number of legs	Length (km)	Updraft cores	Downdraft cores
5800	13	1069	35	8
5100	60	5818	159	67
4800	26	2093	48	15
4500	27	1935	69	54
3500	1	110	3	0
3200	43	2834	80	38
2600	2	220	0	0
2200	35	2444	51	36
1900	4	338	3	3
1300	15	937	34	25
700	2	211	9	2
500	7	605	15	7
300	1	52	2	0
200	5	506	3	4
Total	241	19 172	511	259

2. Data

The primary observation platforms for EMEX were three research aircraft; a National Oceanic and Atmospheric Administration P-3, the National Center for Atmospheric Research Electra, and the Australian Commonwealth Scientific and Industrial Research Organization F-27. Measurements of horizontal wind, vertical velocity, temperature, dewpoint, and aircraft attitude and position derived from inertial navigation equipment (INE) at 1 Hz are used in the analysis. Data from the Electra and F-27, which were recorded at 20 Hz and 50 Hz, respectively, are block averaged to produce a 1-Hz dataset. Only straight and level flights that penetrate convection are used here. Whether or not a flight leg sampled convection was subjectively determined from the P-3 horizontal radar composites of Houze et al. (1988). Since the aircraft sampled the same MCS in the same span of time, these composites were used in all cases. Most legs include environmental air as well as air beneath or in cloud; thus cores outside of clouds but associated with the convection would be included in the dataset. The criteria used to make this determination were a generally cellular structure of the echoes and peak reflectivity values of 30 dBZ or greater. Since most of the legs eliminated were in either clear air or stratiform cloud, only a few, very weak events were excluded by this procedure. A similar analysis by LeMone et al. (1994) using the whole dataset revealed statistics nearly identical to those reported here.

Table 1 summarizes the over 19 000 km of flight legs from all 10 EMEX missions used in the analysis. Legs from all three aircraft were typically 75 km in length. The P-3 sampled the convective region between 4.5 and 5.8 km using a "sawtooth" pattern designed for construction of pseudo-dual-Doppler wind fields in the area between adjacent legs. These legs typically were at 60° angles to the axis of the convective region. Both the F-27 and the Electra flew legs in the convective region approximately normal to the line. The Electra usually sampled regions between 2.0 and 4.0 km. The F-27 flew at levels from near the surface to 2 km. All 10 missions were in MCSs; no isolated clouds were sampled. The P-3 data from the first mission were not used due to instrumentation problems. The Electra did not participate in the 10th mission. The F-27 did not participate in missions 3 and 7. On these days, all sampling below 4 km was carried out by the Electra.

Vertical velocity w was computed using the method described in Jorgensen and LeMone (1989). During straight and level flight the vertical velocity is accurate to 1 m s^{-1}, but perturbations from the mean have an uncertainty of only 0.1 m s^{-1}. Fortunately, the average bias can be removed. This is done by subtracting out the average vertical velocity over all legs at the same height on the same day and assuming the real average vertical velocity is close to zero. This is different from previous studies, in which investigators subtracted out an offset equal to the average vertical velocity over each leg. In a short leg with a large excursion of w, the old method could have the effect of creating anomalous cores, as well as altering the vertical velocity by approximately 0.1 m s^{-1} (Zipser and LeMone 1980). The new method reduces the magnitude of this problem, since the effect of large excursions is reduced with a larger sample size.

The vertical velocity thus computed is used to describe events called *cores*. An updraft core is defined as a portion of a flight leg where the magnitude of vertical velocity continuously exceeds 1 m s^{-1} for at least 500 m (equivalent to 3–5 s of flight time). Downdraft cores are defined analogously. These definitions are the same as those used in LeMone and Zipser (1980), Jorgensen et al. (1985), and Jorgensen and LeMone (1989). This definition is somewhat arbitrary, but it separates meaningful convective events from more turbulent features. Several characteristics of the cores are calculated. These are the intercepted length of the core D (henceforth loosely referred to as "diameter"), equal to the product of the leg-averaged true airspeed and the time in the core; the arithmetic mean of all vertical velocities in the core $\bar{w}$; the maximum 1-s vertical velocity encountered in the core w_{max}; and mass flux per unit length normal to the flight track (called mass flux) in kilograms per meter per second. This is calculated as mass flux = $\rho \bar{w} D$, where ρ is the air density. Note that we avoid any assumptions about geometry in both our definition of D and mass flux.

Figure 1 is a time series of vertical velocity data from the P-3 for 16 January 1987 (EMEX 3) between 2148:05 and 2153:36 UTC. The average true airspeed during this leg was 147.8 m s^{-1}, giving a total leg length of 49.1 km, shorter than most of the legs during the project. A zero offset of 0.378 m s^{-1} has been removed. Small fluctuations around zero can be seen until approximately 2150:00 UTC, when the vertical velocity rapidly rises to peak values of just over 12 m s^{-1}. After these peaks, w quickly decreases to a minimum of -2.2 m s^{-1} at approximately 2150:55 UTC. Another small positive excursion follows with a peak w of 1.5 m s^{-1} near 2151:10 UTC. By 2151:20 UTC, the small fluctuations around zero have returned and remain until the end of the leg. Following the criteria shown above, the two updrafts are considered cores. The downdraft does not meet the size criterion for a core because the portion stronger than 1 m s^{-1} lasts only 3 s, equivalent to approximately 440 m. The first updraft has a D of 5.8 km, $\bar{w}$ of 5.2 m s^{-1}, w_{max} of 12.2 m s^{-1}, and mass flux of 21 000 kg m^{-1} s^{-1}. This was one of the largest, most intense cores observed in EMEX. The second updraft has a D of 1.2 km, $\bar{w}$ of 1.3 m s^{-1}, w_{max} of 1.5 m s^{-1}, and mass flux of 1100 kg m s^{-1}.

For reference, a typical cumulus cloud with a diameter of 1 km and a $\bar{w}$ of 1 m s^{-1} would have a mass flux of 1000 kg m^{-1} s^{-1}. At the other extreme, a cumulonimbus cloud with a diameter of 10 km and a $\bar{w}$ of 10 m s^{-1} would have a mass flux of around 100 000 kg m^{-1} s^{-1}. Most values of mass flux observed over the tropical oceans fall toward the lower end of this spectrum.

Frequency distributions of D, $\bar{w}$, w_{max}, and mass flux were completed for four height intervals of 0–700 m, 700–2500 m, 2500–4300 m, and 4300–5800 m. These levels are chosen to closely correspond with the intervals chosen by previous investigators. The median and 10% level (i.e., stronger than 90% of the cores in the sample) were examined for each parameter at each height level. The parameters in the sample could be biased too small, as described by Jorgensen et al. (1985), because the flight legs may not have penetrated the center of the cores. Following their reasoning, if we were to assume that the cores were circular in cross section with vertical velocity greatest at the center, random sampling of similar individuals would lead to an average D equal to $\pi/4$ times the actual core diameter, and similar biases would exist for $\bar{w}$, w_{max}, and mass flux. Furthermore, a brief (1 s) excursion in w below the 1 m s^{-1} threshold can have the effect of splitting a single large core into two smaller cores. This is observed occasionally in the data. Filtering can be used to reduce or eliminate this effect, but no filtering is done on the data in this study. As shown in Part II, filtering slightly reduces the 10% level of $\bar{w}$ and w_{max} and increases the 10% D. See Part II (immediately following) for more details.

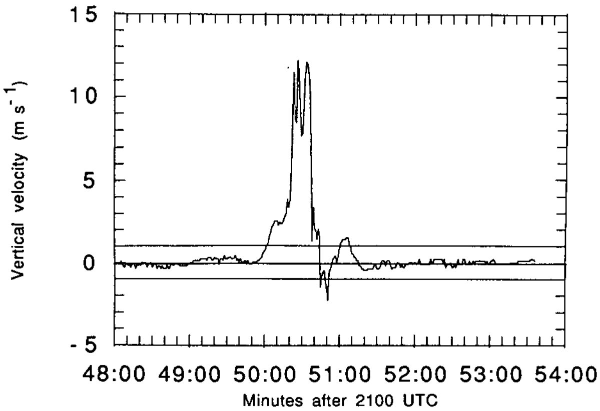

FIG. 1. Time series of vertical velocity (m s^{-1}) encountered by the P-3 during EMEX 3 from 2148:05 to 2153:36 UTC 16 January 1987 while flying near 11.6°S, 128.7°E. Flight level was 5100 m. Zero offset of 0.378 m s^{-1} has been removed from the vertical velocity.

3. Results

Using the criteria established above, the sample includes a total of 511 updraft and 253 downdraft cores. All parameters were approximately lognormally distributed, with many more small or weak cores than large or strong cores. Updraft cores were generally larger and stronger than downdraft cores and occurred about twice as frequently. For all days and altitudes combined, the median updraft core had a D of approximately 1.0 km, a $\bar{w}$ of 2.2 m s^{-1}, a w_{max} of 3.2 m s^{-1}, and a mass flux of 1800 kg m^{-1} s^{-1}. The median downdraft core had a D of about 750 m, a $\bar{w}$ of -1.5 m s^{-1}, a w_{max} of -2.0 m s^{-1}, and a mass flux of -990 kg m^{-1} s^{-1}.

Figure 2 shows the median and 10% values for D, $\bar{w}$, w_{max}, and mass flux as a function of height. The altitude of each height interval is computed as the frequency-weighted mean of the flight-leg heights. The 10% D for updrafts (Fig. 2a) shows a strong increase

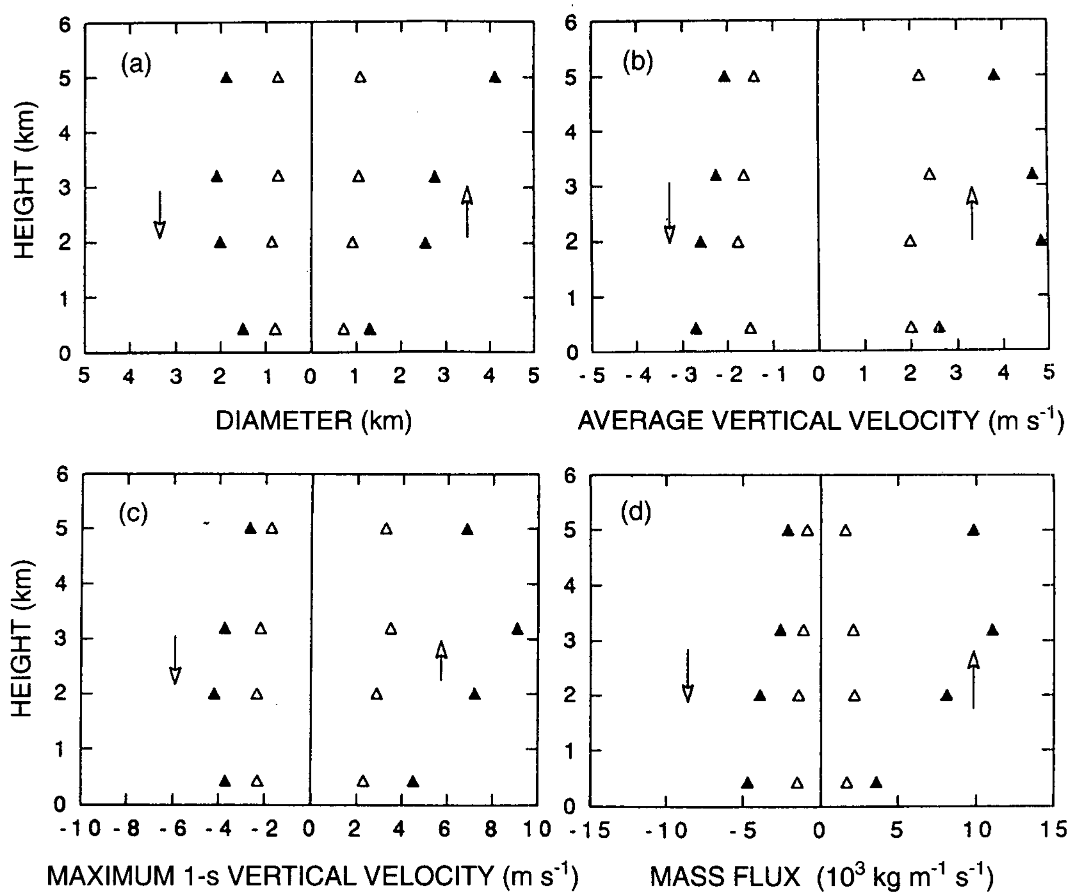

FIG. 2. Variations with altitude of median (open) and strongest 10% level (solid) statistics of (a) diameter or intercept length, (b) average vertical velocity, (c) maximum 1-s vertical velocity, and (d) mass flux per unit length normal to the flight track.

with height.[1] Other values for D show little variation with height. The 10% value of $\bar{w}$ (Fig. 2b) increased with height from ~0.3 km to between 2 and 3 km. Slightly lower values of $\bar{w}$ occur above that level. Similar tendencies were found for updraft w_{max} (Fig. 2c) and mass flux (Fig. 2d). The lower vertical velocities in the altitude range 4–5 km could be real, or they could result from an instrumentation bias between the different aircraft and/or a sampling bias introduced by the differing sampling strategies of the aircraft at different heights on different days.

For downdrafts, 10% and median values of $\bar{w}$ (Fig. 2b) and w_{max} (Fig. 2c) are nearly constant with height. Median mass flux for downdrafts (Fig. 2d) also showed little variation with height. Ten-percent level downdraft mass flux tends to increase downward reaching a maximum near the surface.

Figure 3 shows a comparison of the 10% level $\bar{w}$, for EMEX, GATE [GARP (Global Atmospheric Research Program)] Atlantic Tropical Experiment; LeMone and Zipser 1980), TAMEX (Taiwan Area Mesoscale Experiment; Jorgensen and LeMone 1989), hurricane rainbands (Jorgensen et al. 1985), and the Thunderstorm Project (Byers and Braham 1949). The 10%-level values of $\bar{w}$ for EMEX and the other oceanic cases are of approximately the same magnitude. As pointed out by the previous investigators, these oceanic cores are one-third to one-half as strong as those found during the Thunderstorm Project. Also apparent in Fig. 3 is the earlier discussed tendency for the intensity of the updrafts to change little or even decrease between 3 and 5 km for the oceanic cases. This is not apparent in the TAMEX case, but with only 92 cores at all levels, that dataset had the smallest sample. Comparison of

[1] The P-3 flew most of the higher-level legs, so it is natural to ask whether the 60° angle of its track to a convective band would lead to artificially large cores. If we assume as a worst case that the updraft is a sheet parallel to the band axis, the P-3 cores should be $(2/3)^{1/2} = 1.15 \times$ cores at lower levels. However, the cores at the highest levels are 1.5 times those at the next lower level.

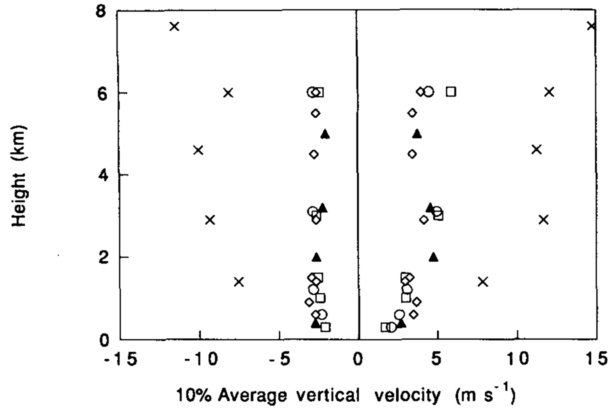

FIG. 3. Comparison of strongest 10% level average vertical velocity in updraft and downdraft cores for EMEX (triangles), GATE (circles), hurricanes rainbands (diamonds), TAMEX (squares), and the Thunderstorm Project (crosses). Adapted from Zipser and LeMone (1980).

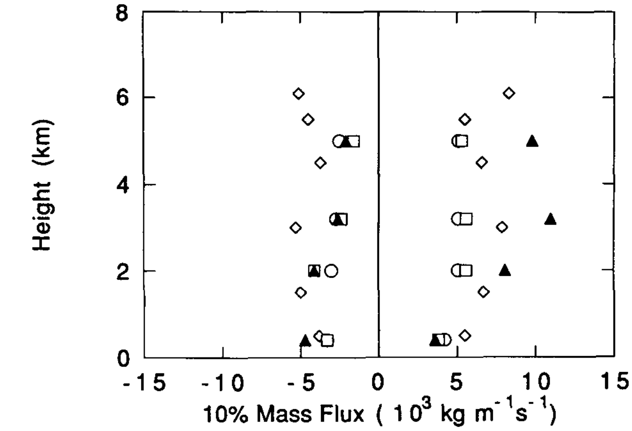

FIG. 5. As in Fig. 3 except for 10% level mass flux. Thunderstorm Project data not available.

10% w_{max} (Fig. 4) for the oceanic cases (Thunderstorm Project data not available) reveals results similar to those found for $\bar{w}$. At most levels, 10% updraft mass flux (Fig. 5) for EMEX was larger than the 10% mass fluxes found in the other oceanic cases (Thunderstorm Project data are not available). Ten-percent downdraft mass fluxes do not differ significantly.

Figure 6 shows 10% diameters for the oceanic cases and the Thunderstorm Project. Oceanic 10% updraft diameters range from 1.6 to 4.1 km in diameter, with hurricane rainband and EMEX diameters typically being the largest. The updraft cores from the Thunderstorm Project are the largest of all, being greater than 4.0 km at all levels. Oceanic 10% downdraft cores were between 1.0 and 2.0 km. Thunderstorm Project 10% downdraft cores were approximately twice as large as the oceanic cores.

Are the differences between continental and tropical oceanic convective cores real? Zipser and LeMone (1980) acknowledge two significant differences between the methods used in collecting the Thunderstorm Project data and the GATE data (which are similar to subsequent oceanic datasets). First, the oceanic data mostly represent cores that are embedded in mesoscale convective systems, while the majority of Thunderstorm Project cores were in relatively isolated large cumulus congestus or cumulonimbus. Second, while the aircraft used to collect oceanic core statistics had inertial navigation systems, the Thunderstorm Project aircraft did not. This necessitated using the aircraft response to the updrafts and downdrafts to determine core properties—a procedure that would be expected to filter the data somewhat.

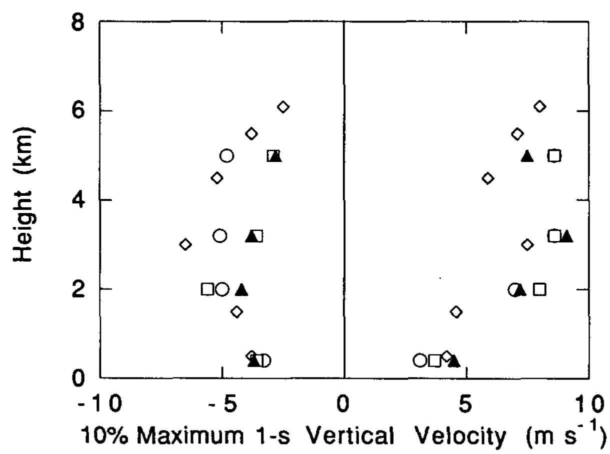

FIG. 4. As in Fig. 3 except for 10% level maximum 1-s vertical velocity. Thunderstorm Project data not available.

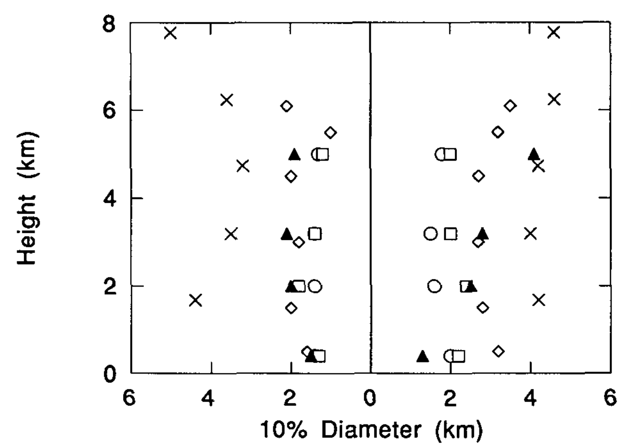

FIG. 6. As in Fig. 3 except for 10% diameter.

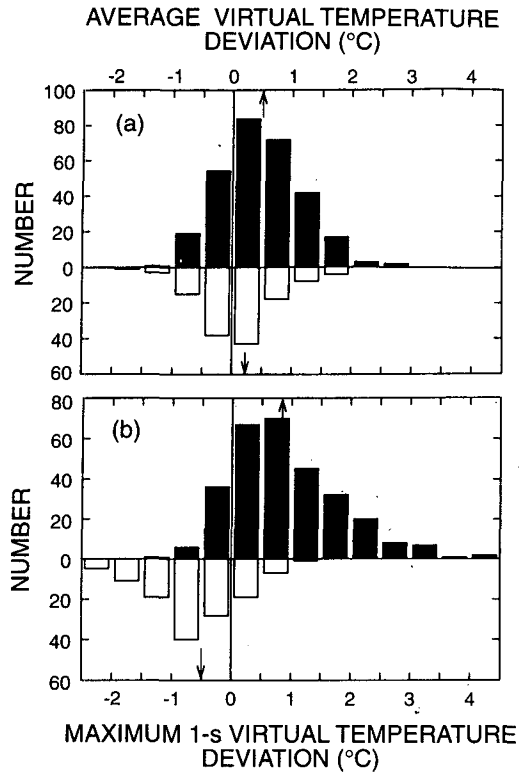

FIG. 7. Histogram of (a) average virtual temperature deviation for P-3 updraft cores (solid) and downdraft cores (open) and (b) maximum (minimum) 1-s virtual temperature for updraft (downdraft) cores, minus environmental value. Arrows denote the median of each distribution. Only flights between 4300 and 5800 m are included.

The answer is a qualified "yes." First, the differences are just too large to be accounted for by the differences in sampling strategy. While the Thunderstorm Project aircraft often took aim at isolated, obviously active clouds or radar targets, the aircraft used in the oceanic experiments flew long legs without deviating to penetrate or avoid convection; these penetrated not only convective bands with active convection often present but also surrounding stratiform clouds with some embedded convection. We recognize that this strategy probably led to an oversampling of weak cores and an undersampling of strong ones, relative to the Thunderstorm Project. However, in four oceanic experiments, if really strong cores existed some of them should have been sampled at some time. This did not happen: *extreme* 1-s *positive vertical velocities never exceeded* 17 m s^{-1}. Furthermore, if we look at the LeMone and Zipser (1980)/Zipser and LeMone (1980) dataset, comparison of the GATE cores in relatively isolated clouds (days 214, 254) to those in convective systems revealed no major systematic differences, suggesting that the sampling bias effect was small in comparison to the differences shown in Figs. 3 and 6.

Likewise, the effective filtering of the procedure used to estimate aircraft vertical velocity in the Thunderstorm Project would not explain the large differences. Filtering would only *reduce* the vertical velocity extrema, suggesting that the updrafts and downdrafts in the Thunderstorm Project would be even stronger than the data suggest. Filtering might exaggerate the diameter somewhat, but we are convinced that the Thunderstorm Project cores are significantly larger, at least near cloud base. It is shown in LeMone et al. (1994) that an 8-s filter increases core diameter significantly only through merging of adjacent events (for an aircraft traveling ~110–140 m s^{-1}), and this does not become important until well above cloud base. The calculations in Lenschow (1976) suggest that the procedure used filters out air motions ~10 s or less in duration. Converting to spatial dimensions, this is about equivalent to the filtering studied by LeMone et al., since the Thunderstorm Project aircraft flew at ~110 m s^{-1} (Braham 1993, personal communication),[2] slightly less than the speed of the aircraft sampling the oceanic cores. Finally, continental updraft diameters near cloud base comparable to those in the Thunderstorm Project have been measured using aircraft with inertial navigation equipment in both the National Hail Research Experiment (NHRE: Fankhauser et al. 1982) and the Cooperative Convective Precipitation Experiment (LeMone et al. 1988).

Deviations of virtual potential temperature (θ_v') were computed for updraft and downdraft cores encountered by the P-3 using data from the CO_2 radiometer and the General Eastern cooled-mirror dewpointer using a procedure outlined in Jorgensen and LeMone (1989). Since both the Rosemount temperature probe and the dewpointer are unreliable in warm clouds, the CO_2 radiometer temperature is substituted for the Rosemont temperature and saturation is assumed. A bias, calculated as the average difference between the Rosemount and CO_2 readings for the environmental portion of each leg, is added to the radiometer to account for the calibration drift. The average deviation (from the average *environmental* value) of θ_v in a core and the extreme 1-s deviation are defined analogously to $\bar{w}$ and w_{max}, respectively. The CO_2 instrument senses radiation in the 15-μm absorption band along a path estimated to be ~200 m in clear air and 33 m in stratus cloud (Albrecht et al. 1979). In cloud, some of the signal comes from the radiance emitted by liquid water. Thus, the brightness temperature measured is some average of

[2] Braham notes that these speeds were documented in Air Weather Service Tech. Rep. 105-39 (1949) and U.S. Weather Bureau Tech. Paper No. 7 (1948 and 1949).

the ambient moist air and the surrounding hydrometeors. Aircraft roll induced by turbulence in cores could lead to some bias in the measurement for a long pathlength. Given the short pathlength of the instrument, this effect should be minimal and no corrections for roll are made.

Both average (Fig. 7a) and 1-s extreme (Fig. 7b) deviations were approximately normally distributed. Consistent with the findings of Jorgensen and LeMone (1989), average temperature deviations for both updraft and downdraft cores near the freezing level are positive. The mean average temperature deviation for updraft cores is 0.55° ± 0.68°C and the mean maximum virtual temperature deviation is 0.98° ± 0.94°C. Table 2 shows linear correlation coefficients of average and maximum temperature deviation with $\bar{w}$, w_{max}, and D. Both average and extreme temperature deviations are well correlated with updraft $\bar{w}$ and w_{max}. A weaker correlation is also present for updraft D. Downdraft parameters show some correlation with the average θ_v deviation, but coefficients are near zero for the extreme temperature deviations.

In general, average core θ_v deviations for updrafts are much lower than is predicted by simple parcel theory. Examination of Fig. 7b indicates that some updraft cores do achieve 1-s maximum excesses approaching their undilute value of 4°–5°C. A sample sounding from EMEX 3 is shown in Fig. 8. The lower average values most likely reflect the effects of entrainment of environmental air into the convective core.

The mean average temperature deviation for downdraft cores is slightly positive: 0.17° ± 0.63°C, while the mean 1-s extreme negative virtual temperature deviation is −0.38° ± 0.73°C. Figure 7b also shows that downdraft virtual temperatures were occasionally 2°C lower than the environment, but only rarely. We follow Jorgensen and LeMone (1989) in hypothesizing that warm downdrafts exist mainly because of water loading. Overshooting downdrafts would also be warm.

4. Discussion

The vertical velocities of oceanic convective cores over the western Pacific warm pool have been shown to be approximately the same magnitude as those found in other tropical oceanic regions of the world. This finding is perhaps surprising in view of the fact that sea surface temperatures over this part of the warm pool during EMEX were, on average, >30°C (Griffith 1992), much higher than those in GATE (~27.5°C), TAMEX [~28°C for the MCS in Jorgensen et al. (1991), which accounted for a significant fraction of TAMEX cores, and 1°–2° cooler for the remainder of the TAMEX cores], and in some hurricanes (SSTs vary). These oceanic cores are much weaker than those found over continental regions. The question arises, What is responsible for limiting the updraft velocities over the ocean?

TABLE 2. Linear correlation coefficients of average and maximum θ_v deviation with w, w_{max} and D. Downdraft cores are in parentheses.

Parameter	Average θ_v deviation (°C)	Maximum θ_v deviation (°C)
$\bar{w}$	0.65 (0.31)	0.77 (0.04)
w_{max}	0.61 (0.30)	0.78 (0.04)
D	0.23 (0.12)	0.35 (0.03)

The first possibility is that the thermodynamic instability is less over the ocean than over the land. The reconstructed soundings in the inflow environment of the EMEX systems presented in Griffith (1992), along with results from TAMEX (Jorgensen and LeMone 1989) and GATE (Zipser and LeMone 1980), reveal that a significant amount of buoyancy can and does exist over the tropical oceans. Soundings from EMEX had an average convective available potential energy (CAPE), which is proportional to the positive area between the level of free convection and the equilibrium level on a thermodynamic diagram of 1765 J kg^{-1}. This is between the values found for severe (Bluestein and Jain 1985) and nonsevere (Bluestein et al. 1987) squall lines in Oklahoma, and only slightly less than the CAPE for the mean Thunderstorm Project sounding (Lucas et al. 1994).[3]

Examination of many soundings reveals that, though CAPE is often roughly the same magnitude for continental and oceanic regimes, the shape of the associated positive area can be different. Over the ocean, the positive area is generally "skinny." The soundings achieve only 4°–5°C of buoyancy, but this buoyancy is maintained through a large fraction of the troposphere. Over the United States, the positive area is often "fat"; an ascending parcel can achieve 8°–9°C or more of buoyancy, but the level of free convection is relatively high and the parcel equilibrium level relatively low compared to oceanic soundings. This is illustrated in Fig. 9, which shows the virtual temperature excesses predicted by simple parcel theory for an oceanic and a continental sounding. The oceanic sounding has a CAPE of 1750 J kg^{-1}, and the CAPE for the continental sounding is 1575 J kg^{-1}. Although the continental sounding has less CAPE than the oceanic case, it achieves a maximum virtual temperature excess of 11.3°C, compared to 4.5°C for the EMEX sounding.

[3] This is in contrast to the 3000 J kg^{-1} value erroneously reported in Zipser and LeMone (1980). See Lucas et al. (1994) for details.

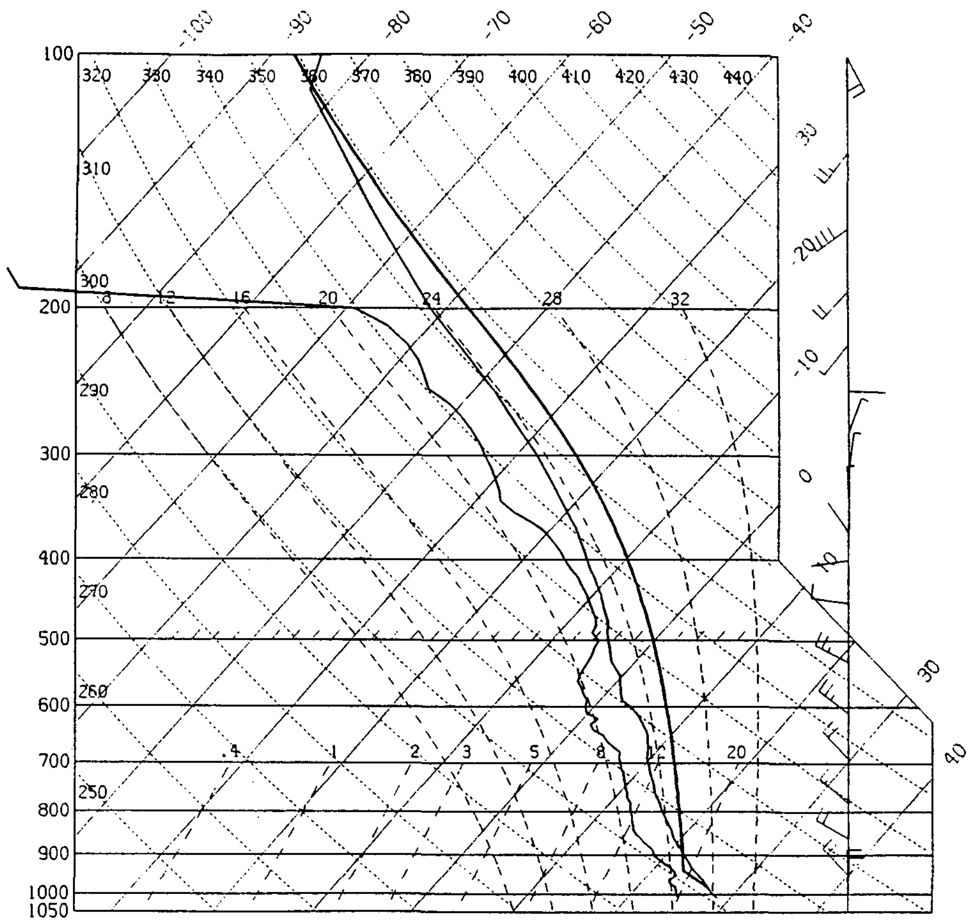

FIG. 8. EMEX 3 inflow environment sounding. Sounding has been reconstructed from low-level aircraft data and the 2300 UTC 16 January 1987 Darwin, Australia, sounding. Heavy line is parcel path for moist-adiabatic ascent of the lowest 50 mb.

This difference may be significant when considering the effects of water loading.

The second possibility is that water loading effects are greater for the oceanic environment than they were for the Thunderstorm Project. Based on the reconstructed EMEX soundings (Griffith 1992), the adiabatic liquid water content of a parcel ascending from cloud base to 5 km is nearly 10 g kg^{-1}, equivalent to 2.7° of negative buoyancy. However, the mean Thunderstorm Project sounding has a similar adiabatic liquid water content, about 9 g kg^{-1}. That is, cloud-base mixing ratios are in the 17–19 g kg^{-1} range for most tropical clouds (or clouds in maritime tropical air masses) whether over land *or* water, and the saturation mixing ratios at 5 km are about 9 g kg^{-1}. Experience, albeit with imperfect instrumentation, is that actual clouds rarely achieve the adiabatic values of 8–10 g kg^{-1}, and usually contain much less. Putting it another way, the question asked by Xu and Emanuel (1989), whether the tropical atmosphere is conditionally unstable, can be answered in the affirmative, because reversible ascent is indeed rare. We can speculate that given the known higher updraft velocities in clouds over land, water loading would be greater, not less, in land updrafts. The important point is that *any given amount of water loading is a smaller proportion* of the total buoyancy for the "fat" positive-area portions of the *continental soundings* compared with the usually "skinny" positive areas in *oceanic soundings.* This would lead to stronger updrafts for a given CAPE. Jorgensen and LeMone (1989), using a simple parcel model with a water loading term and observed liquid water contents included, reproduced an updraft vertical velocity profile for a "skinny" TAMEX sounding with a decrease in intensity between 3 and 5 km. Downdraft cores in TAMEX, like those in this study, had positive temperature deviations. Jorgensen and LeMone (1989) attributed this to water loading. The values predicted by this model for both updrafts and downdrafts are close to what is actually observed in the atmosphere, although

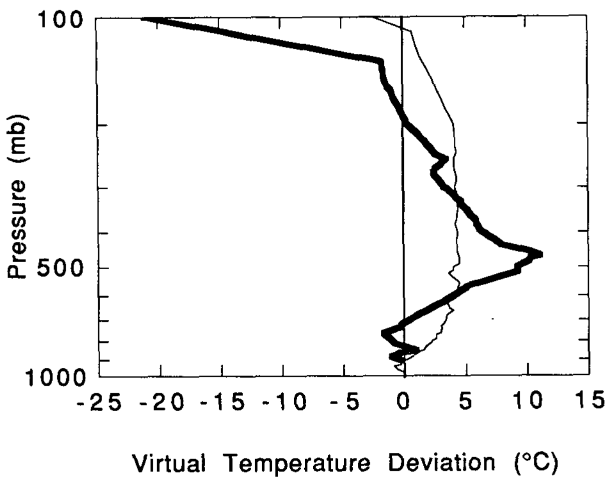

Fig. 9. Virtual temperature of air parcel minus environmental value predicted for moist-adiabatic ascent of the lowest 50 mb for and oceanic (thin line) and continental (thick line) sounding. Oceanic sounding is same as Fig. 8. Continental sounding is from Woodward, Oklahoma, at 0130 UTC 11 June 1985 just before the passage of a squall line.

the liquid water contents used are only a small fraction of the adiabatic values. A comprehensive analysis of continental and oceanic soundings should be done to examine this hypothesis more completely.

The third possibility is that entrainment is more effective for oceanic cores. As discussed earlier and shown in Fig. 6, continental updraft core diameters appear to be larger than those over the oceans, particularly near cloud base. This would reduce entrainment and leave the cores more buoyant. Table 2 indicates a tendency for the larger updraft cores to have a larger maximum temperature deviation. Hypothesized larger diameters over the continent would be consistent with larger typical convective boundary layer depths. We speculate that is true because boundary layer eddies and hence the diameter of initial cumuliform clouds (or the eddies swept up in the inflow of a convective band) roughly scale with the boundary layer depth (Kaimal et al. 1976; Grossman 1982; LeMone 1980).[4] The data support this conjecture. Over the ocean, boundary layer depths of this undisturbed air are the order of 500 to 600 m. Convective boundary layers over Florida are of the order of 1–2 km deep based on data gathered during the 1991 Convection and Precipitation/Electrification Experiment (CaPE) (J. C. Fankhauser 1993, personal communication), while those over the High Plains (e.g., NHRE) are 2 km and greater (Fankhauser and Wade 1982). The 10% cloud-base updraft diameters roughly scale with these depths, with the largest cores in cumulus congestus and cumulonimbus over the High Plains in NHRE ~3 km (Fig. 6.47, Fankhauser et al. 1982), intermediate values for congestus over Florida (from CAPE, ~2.5–3 km),[5] and the smallest values over the ocean (1.2 km from Fig. 6), in spite of the fact that the last sample includes only cumulonimbus.

Analysis of virtual temperature excesses at and near the freezing level indicated that while some areas of undilute air may exist within a core, average observed values of θ_v are generally substantially smaller than those predicted by parcel theory. We consider this to be strong evidence for considerable entrainment of environmental air into the oceanic cores at rather low levels in the clouds. However, entrainment is difficult to quantify for several reasons. First, the cores often ascend in an environment different from that observed by a sounding. Second, cores at a given level have different histories. It is difficult, if not impossible, to know the initial conditions associated with a particular core. Third, the original height of the air being entrained cannot be fully known because of uncertainties in the mechanism of entrainment. Entrainment through the cloud top would mix in air with characteristics different from air mixed in from the sides.

We might expect some effect of shear on convective core properties but could detect none. This is conceivably because the effects are too subtle to be detected using the sampling technique and analysis described here. Also, the range of shears encountered in EMEX was relatively small. The presence of deep, strong shear has been shown to be a necessary ingredient for producing strong, long-lived convection (Weisman and Klemp 1986). Forcing by the attendant gust fronts and mesoscale flows with associated pressure fields would be expected to play a role in generating or enhancing convective-scale vertical velocity. Finally, LeMone (1989) noted that cumulus congestus cloud diameter increased with vertical shear at cloud base in CCOPE, speculating that the shear enabled generation of gravity waves which obtained their energy from the boundary layer and imposed their scale on the boundary layer and hence cloud updrafts (Clark et al. 1986). Vertical wind shears in EMEX, like those of GATE (Barnes

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[4] Larger draft diameters that result from supercell dynamics are not relevant here, since none of the data include supercell penetrations. Some of the larger drafts, however, could have scales enlarged by interaction of the boundary layer (and hence broadening of boundary layer structures) as a result of gravity wave interaction with the boundary layer (see Clark et al. 1986). However, this effect should be a function of shear, and no shear effects are evident in the data here.

[5] Based on data from 3 days, from Barnes et al. (1992) and J. C. Fankhauser (1993, personal communication). Although the relative differences between Florida and High Plains updrafts are small and require more careful evaluation, they are supported by Fankhauser's impressions.

and Sieckman 1984), are typically smaller in magnitude than the shears found over Oklahoma (Bluestein et al. 1987). Vertical wind shears were sometimes very strong in EMEX, but in these cases the strong shear was confined to the lowest 1–2 km of the troposphere. Neither the strength nor orientation of the shear vector had a discernible effect on the properties of the convective cores.

5. Summary and conclusions

Aircraft data collected during penetrations of cumulonimbus clouds in oceanic MCSs off tropical Australia in January and February 1987 during the Equatorial Mesoscale Experiment have been analyzed to study the distribution of the mean vertical velocity $\bar{w}$, maximum vertical velocity w_{max}, intercepted width D, and mass flux of convective cores in tropical oceanic MCSs. Statistical analysis revealed that, despite the warm sea surface temperatures, cores in this region are similar in strength to cores in other oceanic regions but tend to be slightly larger than those found in other oceanic areas. These oceanic cores are one-third to one-half as strong and smaller than their continental counterparts. This result is surprising in view of evidence that there is little difference in CAPE between the land and ocean environments for which the aircraft penetration datasets were analyzed.

Calculation of virtual temperature excesses for EMEX cores near the freezing level showed that while some portions of an updraft core have excesses that approach values associated with undilute ascent, the average excess in a core was substantially below that predicted by parcel theory. Downdraft cores at these levels were on average slightly warmer than the environment, as Jorgensen and LeMone (1989) observed for TAMEX. As in that study, we attribute warm downdrafts mainly to water loading, although in some instances we probably sampled decelerating downdrafts.

We suggest that the vertical velocities of cores over water are far less than those over land due to a greater effectiveness of water loading and entrainment in reducing buoyancy in cores in the oceanic clouds. The actual water loading is believed to be subadiabatic, but any given loading tends to be a larger fraction of the available buoyancy for the oceanic environment. The smaller diameters observed for oceanic cores are believed to be the principal factor in the observed subadiabatic virtual temperature excesses, and we speculate that these are, in turn, related to the much reduced boundary layer depth over the ocean.

Part II of this paper (LeMone et al. 1994) describes in detail the effects of filtering on the convective core statistics. Filtering is seen to slightly decrease the median and 10% of $\bar{w}$ and w_{max} and slightly increase median and 10% of D.

Acknowledgments. The authors would like to thank Tae Chang of the Mesoscale and Microscale Meteorology Division of NCAR, who provided computer support that made this project possible. Special thanks are due Jorgen Jensen and Brian Ryan for providing the data from the F-27, and to Jim Fankhauser, Jim Dye, Dave Alexander, and two anonymous reviewers for providing useful comments on the manuscript. We also thank the flight crews of the P-3, Electra, and F-27 for aiding in the collection of the data. EZ and CL were supported by National Aeronautics and Space Administration Grant NAG5-1569 and National Science Foundation Grant ATM-9019757; ML by NCAR and NSF Grant ATM 9215507.

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