Flow and Mixing in New Mexico Mountain Cumuli

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ABSTRACT

Convection and cloud formation over mountains during weak winds and strong insolation were studied using an instrumented aircraft. Previous studies in cloudless situations had shown the existence of convergence over the mountain range at low levels, and divergence aloft. The present observations indicate that cumulus clouds and even thunderstorms in their early stages do not significantly affect the strength of the low level convergence. Furthermore, net divergence was found around clouds up to at least 6 km MSL. This favors the vertical mixing model of a cumulus cloud over the lateral entrainment model. In one case, the convergent low level heat island circulation over the mountain was observed to change to an almost non-divergent, asymmetric circulation as a thunderstorm over the mountain reached maturity.

1. Introduction

In the American southwest it is well known that summer thunderstorms generally form first over the mountains. This is the second of a series of papers reporting detailed observations of this phenomenon.

The first paper (Raymond and Wilkening, 1980) dealt with the case in which little moisture was present, allowing the precursor circulation to be isolated from the effects of latent heat release and rainfall. In that paper, we documented the production of a toroidal circulation due to the heat island effect of a mountain range illuminated by the sun. This circulation is convergent in lower levels and divergent aloft, and is superimposed on the ambient flow. The strength of this mesoscale flow is found to be a few meters per second in the horizontal wind, and on the order of a meter per second or less in the vertical component, averaged over the area of the mountain range—approximately 200 km².

Superimposed on the updraft region of the mesoscale flow (convective core—Braham and Dragini, 1960) are convective eddies with typical horizontal wavelengths of 3 km. These eddies exhibit vertical velocities of typically ±5 m s⁻¹, and are responsible for a large fraction of the vertical heat and moisture transport over the mountain range. The eddies and the mesoscale flow interact symbiotically, in that the heat dumped aloft by the eddies drives the mesoscale flow, which, in turn, removes excess heat and maintains a lapse rate suitable for the maintenance of convection. The net result is that low level air is efficiently transported aloft over mountain ranges, making these geological features excellent source regions for thunderstorms when moisture is present.

During the summer thunderstorm season of 1980, we staged another series of observations of this phenomenon. The purpose of these observations was twofold: First, we wanted to see what happened to the heat island circulation when moisture was added. Second, since the resulting storms tend to occur with great regularity in time and space, we wanted to use the mountain region as an open-air laboratory for the study of thunderstorm dynamics—particularly the early stage, which is difficult to observe in a less ordered environment.

A particular facet of storm behavior that interested us was the nature of the entrainment or mixing process. It has been known since the work of Stommel (1947), that cumulus clouds were not simply parcels of undiluted boundary-layer air adiabatically lifted without mixing. Observations of temperature and liquid water content of clouds clearly showed that mixing with air from above the boundary layer had to be occurring. In analogy with laboratory experiments, it was hypothesized that cumulus clouds accomplished this mixing by virtue of being entraining plumes (e.g., Morton, 1957; Squires and Turner, 1962).

In ordinary buoyant plumes, density mixes linearly. That is, if a parcel from the plume mixes with a parcel of surrounding fluid, the resulting mixture has a density intermediate between the densities of the two initial components. This is not necessarily the case when a cloud containing liquid water mixes with surrounding dry air. In this situation, the cloud water evaporates, and the resulting cooling can yield a mixture with density greater than that of either initial component.

Recognizing this, Squires (1958a) proposed that
clouds were subject to penetrative downdrafts of initially dry air from the cloud top environment. Observations of the structure of cumulus clouds (Squires, 1958b) suggested that this might be the primary mode of mixing in cumuli. It was found that liquid water content tended to be no higher on the average near the center of a cloud than at the edges. This is consistent with dry parcels penetrating downward at random through the cloud, but inconsistent with the idea that all dilution is due to mixing of air from the sides of the clouds as occurs in entraining plumes.

Based on simple buoyancy and thermodynamic arguments, Fraser (1968) built a consistent model of a cumulus cloud in which the primary dilution process is vertical mixing. Such a cloud must have inflow from the environment at both its top and bottom, and net outflow around the sides. This is in contrast to the entraining plume model of a cloud in which net flow around the sides of the cloud must be inward.

Given such contrasting theories on how cumulus clouds work, which theory is favored by observation? Surprisingly, pertinent evidence is rather sparse. Telford and Wagner (1974, 1980) presented a single observation of net outflow around the sides of a small cumulus cloud. The observation was obtained by use of wind finding equipment on an instrumented aircraft. Paluch (1979) determined, by means of thermodynamic conservation arguments, that cloud material at mid-levels in Colorado cumulus congestus clouds was a mixture of air from both above and below the observational level. Both of these observations clearly favor the vertical mixing model. However, direct measurements on fully developed thunderstorms generally show net inflow up to 6 or 7 km MSL (e.g., Byers and Braham, 1949). The nature of the dilution process has therefore not been well explored, and indeed, appears to vary with the situation.

The main emphasis of the present work has been to make convergence measurements by flying boxes around the convective core over a mountain range using an aircraft equipped to measure wind components. Where feasible, traverses were made through the convective core as well. This yielded information on storm interiors that was analyzed in the manner of Paluch (1979).

The order of presentation is as follows: Section 2 describes the observations taken, the analysis techniques used and summarizes the case studies made. A particular case study is examined in detail in Section 3. Supporting evidence from other case studies is presented in Section 4. Section 5 outlines conclusions reached and the Appendix presents an analysis of errors expected in convergence box measurements.

2. Observations and analysis

The primary tool used in this program was the instrumented Queen Air aircraft, N304D, operated by the National Center for Atmospheric Research (NCAR). The measurements of most interest obtained aboard the aircraft were the vertical and horizontal wind, temperature, humidity, cloud water content and aircraft position. Winds were obtained by subtracting the aircraft velocity, derived from an inertial platform, from the aircraft-relative wind measured by a gust probe system. A reverse flow thermometer was used for all temperature measurements, minimizing in-cloud wetting problems. Humidity measurements were made by two different dewpoint instruments as well as a Lyman–alpha hygrometer. Cloud water was obtained using a Johnson–Williams probe.

All data were recorded at a 20 Hz rate. For the purposes of this work, 1 s averages were taken. Additional measured parameters were radar altitude and surface temperature derived from a downward-lookig radiometer. Forward and side-looking time-lapse pictures were also taken from the aircraft.

On selected flights, radon monitoring equipment was carried aboard the aircraft. The results of these measurements will be reported elsewhere.

The observational area for these experiments was the Magdalena Mountains. This forested range has a maximum elevation of ~3.2 km MSL and is surrounded by grassland plains of approximate mean elevation 2 km MSL. The plains slope upward slightly to the west. In addition, there is some locally higher terrain in the form of low foothills just to the west of the mountains. The mountains are ~20 km long in the north–south direction and 10 km wide. The Langmuir Laboratory for Atmospheric Research operated by the New Mexico Institute of Mining and Technology, is located atop these mountains. A topographic map of the range is shown in Fig. 1.

Time-lapse movies of the clouds over the Magdalena Mountains were made from Socorro, which is 25 km east-northeast of the mountains. Estimates of cloud-top height were made using these movies, based on the assumption that the clouds remained directly over the mountains. For the case studies reported here, this is not a bad assumption, but the reported heights should be regarded as approximate. The nearest National Weather Service radiosonde station is Albuquerque, 120 km north-northeast of the site. Soundings for the period of interest were obtained.

The results of this paper rest, for the most part, on the accuracy of convergence measurements made using the Queen Air aircraft. We have therefore spent considerable effort to establish confidence in these measurements.

A convergence box is a flight along a closed path at a constant altitude around the object being studied. The horizontal divergence \( D \) is given by the ratio

\[
D = F/A,
\]

(1)
where

\[ F = \oint v_n ds \]  \hspace{1cm} (2)

is the line integral of the cross-track wind \( v_n \) around the flight path, and the line integral

\[ A = \oint x dy \]  \hspace{1cm} (3)

is the area inside the box. The positions east and north of a reference point are denoted respectively by \( x \) and \( y \). The cross-track wind is taken positive for outward flow.

The quantity \( F \), which we call the flux, clearly equals \( \bar{v}_n C \), where

\[ C = \oint ds \]

is the circumference of the boxed region and \( \bar{v}_n \) is the mean cross-track wind. Errors in the divergence calculation reduce essentially to errors in \( \bar{v}_n \). The Appendix discusses the various sources of errors in this quantity. We conclude that the mean cross-track wind can be measured to within \( \pm 0.2 \, \text{m s}^{-1} \), as long as the duration of the box measurement is much less than the 84 min precessional period of the inertial platform.

When the track of a box circuit is completely outside the region of significant divergence, the value of the flux, rather than the divergence itself, is the more interesting quantity. This is because the flux is not a function of the track location as long as the aircraft remains outside the convective core. For this reason, we present our results in terms of the flux rather than the divergence itself.

Thirteen flights were made from 24 July to 9 August 1980. Table 1 summarizes pertinent information about each of these flights, listing the number of box circuits and traverses, the type of convection occurring and the days on which radon measurements were made.

Times are reported in Mountain Standard Time (MST) (GMT – 7 h) except when otherwise noted. For flights 12 and 13, the primary mission was to investigate mesoscale phenomena other than those associated with the Magdalena Mountains. These flights will not be discussed further here. Those flights marked "strong winds" had winds in excess of 5 m s\(^{-1}\) below mountain top level. Mountain-wave activity was present in these cases, further complicating an already complex situation. During flights 6 and 8 there were mid-level clouds that reduced insolation, resulting in delayed convection.

Elimination of flights 6, 8, 12 and 13, as well as
Table 1. Summary of flights.

<table>
<thead>
<tr>
<th>Flight</th>
<th>Date</th>
<th>Time (MST)</th>
<th>Maneuvers</th>
<th>Clouds</th>
<th>Comments</th>
</tr>
</thead>
<tbody>
<tr>
<td>1*</td>
<td>24 July</td>
<td>0805–1127</td>
<td>5 boxes, 3 traverses</td>
<td>Rain shower</td>
<td>Radon</td>
</tr>
<tr>
<td>2*</td>
<td>25 July</td>
<td>0820–1056</td>
<td>3 boxes, 3 traverses</td>
<td>Small cumulus</td>
<td>No time-lapse movies</td>
</tr>
<tr>
<td>3*</td>
<td>26 July</td>
<td>0805–1110</td>
<td>6 boxes</td>
<td>Small cumulus</td>
<td>No time-lapse movies</td>
</tr>
<tr>
<td>4*</td>
<td>29 July</td>
<td>0834–1131</td>
<td>4 boxes, 4 traverses</td>
<td>Towering cumulus</td>
<td>Radon</td>
</tr>
<tr>
<td>5*</td>
<td>30 July</td>
<td>0841–1133</td>
<td>5 boxes, 3 traverses</td>
<td>Thunderstorm</td>
<td>Canceled due to no insolation</td>
</tr>
<tr>
<td>6</td>
<td>31 July</td>
<td>0915–0944</td>
<td>1 traverse</td>
<td>Lingering stratus</td>
<td>Radon</td>
</tr>
<tr>
<td>7*</td>
<td>1 August</td>
<td>0849–1132</td>
<td>5 boxes, 4 traverses</td>
<td>Small cumulus</td>
<td>No time-lapse movies</td>
</tr>
<tr>
<td>8</td>
<td>3 August</td>
<td>0900–1143</td>
<td>7 boxes</td>
<td>Cumulus congestus</td>
<td>Strong winds</td>
</tr>
<tr>
<td>9</td>
<td>4 August</td>
<td>0919–1231</td>
<td>7 boxes, 2 traverses</td>
<td>Thunderstorm</td>
<td>Strong winds, radon</td>
</tr>
<tr>
<td>10*</td>
<td>6 August</td>
<td>0858–1213</td>
<td>6 boxes, 1 traverse, spiral ascent</td>
<td>Thunderstorm</td>
<td>Strong winds, radon</td>
</tr>
<tr>
<td>11</td>
<td>7 August</td>
<td>0906–1148</td>
<td>4 boxes, 4 traverses</td>
<td>Thunderstorm</td>
<td>Primary mission elsewhere</td>
</tr>
<tr>
<td>12</td>
<td>8 August</td>
<td>0910–1116</td>
<td>3 traverses</td>
<td>Towering cumulus</td>
<td>Strong winds, primary mission elsewhere, radon</td>
</tr>
<tr>
<td>13</td>
<td>9 August</td>
<td>0931–1204</td>
<td>3 boxes, 3 traverses</td>
<td>Cumulus congestus</td>
<td>Strong winds, primary mission elsewhere, radon</td>
</tr>
</tbody>
</table>

* Analyzed for this paper.

those with strong winds leaves a sample of seven flights with weak winds and strong insolation, namely 1, 2, 3, 4, 5, 7 and 10. Of these cases, three produced a rainshower or thunderstorm, three resulted in only small cumulus clouds and one had towering cumulus with no precipitation.

3. Flight 10, 6 August 1980

On 6 August 1980 a thunderstorm developed over the Magdalena Mountains under nearly ideal conditions of strong insolation, weak winds and negligible rate of change in the large scale environment. Fig. 2 shows the evolution of cloud top height of the storm with time, as well as a chronology of Queen Air maneuvers and other events of interest. The storm formed directly over the mountains between 1030 and 1130 MST and drifted slowly to the west, decaying by 1300. Small precursor clouds were visible over the mountains intermittently from 0800 and continuously from 0835. Bases were 4.1 km MSL (625 mb) or ~1 km above the mountain ridge.

Starting at ~1050, a layer of stratus was observed to be projecting from the cloud on its north and west sides. Due to the flight path of the aircraft, this apparent detrainment could have begun as much as 10 min earlier and not have been seen. The height of this stratus was subsequently determined to be 6.4 km MSL. Precipitation was first observed at 1108 and an anvil cloud began to form at 1112. The anvil was
Fig. 3. Ambient sounding for 6 August 1980. Below 460 mb the sounding was obtained by the Queen Air in clear air in the vicinity of the cloud between 1140 and 1213 MST. Above 460 mb the Albuquerque sounding at 1200 GMT (0500 MST) is used. The wind components ($V_x = $ westerly, $V_y = $ southerly) have been subjectively smoothed. $Q$ is the mixing ratio, $\theta$ the potential temperature, $\theta_e$ the equivalent potential temperature and $\theta_e^*$ the saturated equivalent potential temperature. $\theta_e^*$ is computed exclusively from the Albuquerque sounding. The dashed lines represent $\theta_e$ values of air flowing into the storm at different times.

well above the detraining stratus, and in no way connected with it.

Fig. 3 shows a composite environmental sounding. The sounding below 460 mb was taken by the Queen Air between 1140 and 1213. Above this level, the Albuquerque 1200 GMT (0500 MST) sounding was used. Note that the soundings match reasonably well except for the $V_x$ (westerly) wind component. The Queen Air sounding was taken during descent south and east of the storm in clear air, while the storm was in progress.

a. Forcing

We now examine the way in which the storm was forced by the mountain range. Fig. 4 shows the evolution of the sub-cloud layer with time. Illustrated are Queen Air soundings centered at 0915 and 1159 respectively, and the Albuquerque sounding at 0500 (1200 GMT). Note that all these soundings converge near 650 mb, indicating a reasonably steady, homogeneous environment above the mixed layer.

By 1159 the mixed layer extended to ~690 mb. The temperature difference between the 0915 and 1159 soundings above 800 mb (i.e., the level of the plains) corresponds to an additional enthalpy per unit surface area added to the atmosphere of ~1.3 x 10^8 J m^-2. Spread uniformly over the 164 min interval between the soundings, this corresponds to an average surface heat flux of 135 W m^-2. (Comparing the early Albuquerque sounding to the Queen Air soundings at these levels is not fruitful because of the 120 km separation.)

The above calculation ignores the possible effect of advection on the temperature profile. It therefore yields an effective surface heating, useful as a lower boundary condition for numerical simulations.

Comparison of the mixing ratio profiles between 0915 and 1159 reveals a net increase in water vapor mass per unit surface area between 660 and 800 mb, i.e., the area to the left of the curve between these limits increased during the period. This could be due either to moist advection or an integrated latent heat flux from the surface amounting to 2.5 x 10^6 J m^-2 (1 mm layer of water evaporated). Over the period, this averages to a latent heat flux of 250 W m^-2, which is not unreasonable, in that significant rain fell on the two previous days, causing the ground to be moist.

We now discuss the first four convergence boxes, which as Fig. 2 shows, were made during the growing phase of the thunderstorm. These boxes were flown alternately at 626 and 730 mb to see if the previously observed pattern of inflow below the peaks and outflow above (Raymond and Wilkening, 1980) was repeated. As Fig. 5 shows, this was the case. Flux toward the mountain up to 65 km s^-1 was observed at 730
mb, while flux away from the mountain reached 29 km m s\(^{-1}\) at 626 mb. Also shown in Fig. 5 is the mean potential temperature over the box circuits. The 730 mb temperatures reflect the warming of the mixed layer with time. The boxes at 626 mb were above the top of the mixed layer and therefore do not exhibit a temperature increase.

Figs. 6–9 show the wind vectors along each box track as well as the vertical wind, potential temperature, mixing ratio and equivalent potential temperature traces. For clarity of presentation, these data have been low-pass filtered as described by Raymond and Wilkening (1980), with a cutoff wavelength of 1 km.

The effect of the foothills located on the west side of the mountains is reflected in the traces of potential temperature and mixing ratio—it is significantly warmer and dryer there. Since the terrain in this region is somewhat higher than the mean height of the surrounding plains, a potential temperature curve characteristic of the east sides of the 730 mb boxes has been added to Fig. 5. This curve is more likely to be representative of the potential temperature over the plains.

From boxes at just two levels, it is impossible to determine the vertical extent of the inflow and outflow regions. However, between convergence boxes, ascents and descents were made along the same track used for the boxes. The wind signal is too noisy to use as an indicator of inflow or outflow without a full box maneuver. However, the mixing ratio can be treated as a conservative tracer, and it should take on values characteristic of the convective core in regions of outflow.

![Fig. 4](image_url)  
**Fig. 4.** Evolution of the sounding in the sub-cloud layer with time on 6 August 1980 (flight 10). Variables as in Fig. 3.

![Fig. 5](image_url)  
**Fig. 5.** Flux and mean potential temperature for convergence boxes at two levels as a function of time on 6 August 1980 (flight 10). The lower potential temperature curve for the 730 mb box reflects the mean potential temperature to the east of the mountains.
Fig. 6. Flight path with horizontal wind vectors and traces of vertical wind $W$, potential temperature $\Theta$, mixing ratio $Q$, and equivalent potential temperature $\Theta_e$ for box 1 on 6 August 1980 (flight 10). All data have been smoothed with a low pass filter with a cutoff wavelength of 1 km. The cross indicates the location of Langmuir Laboratory (see Fig. 1) and the arrow is the direction of flight. The path length around the box is $S$, and is indicated by the numbers on the flight path as well as on the traces.

Fig. 10 shows the ambient profiles of mixing ratio at 0915 and 1145. Also shown is the mixing ratio taken during the ascents and descents between boxes. Note that from 630 to 690 mb, mixing ratios were generally much higher than ambient values. Below 690 mb, mixing ratios actually decreased from values attained at higher levels, and can be reconciled with ambient conditions there. We suggest that inflow was occurring below 690 mb and outflow between 690 and 630 mb. Note that mixing ratios exceeding am-

Fig. 7. As in Fig. 6, except for box 2, 6 August 1980.
bient occurred on all sides of the storm in this layer, indicating outflow from all sides. It is interesting that the observed cloud base (coincident with the 626 mb boxes) was almost exactly at this upper limit to the outflow. The outward flux measured at this level almost certainly under-represents the flux in the main body of the outflow layer.

Fig. 11 shows ambient potential temperature and potential temperature near the storm in the manner of Fig. 10. The outflow air was nearly 1°C cooler than ambient at 0956, but warmed to near ambient by 1100.

Between 0956 and 1100 the tops of the clouds went from <5 km MSL to >10 km. Why did this rapid
growth occur? It is reasonable to assume that air flowing out of the convective core is characteristic of air entering cloud base, since the core is typically well mixed. With the above assumption, we find that the flow out of the core at 670 mb had an equivalent potential temperature $\theta_e$ of 70.2°C at 0956, while the value at 1100 was 73.2°C. About a third of this increase was due to the rise in temperature, and two-thirds to the increase in mixing ratio.

The dashed lines in Fig. 3 represent adiabatic ascent at these two equivalent potential temperatures. It is obvious that the increase in $\theta_e$ results in significantly more moist convective instability as well as a lowering of the level of free convection from 560...
to 610 mb. This makes sense in light of the development of the cloud system over this interval.

b. Detrainment layer

An ascent, spiraling around the cloud, was made from 1057 to 1125. During this ascent, precipitation began falling from the cloud. It was found that the air near the storm was on the average only 0.3 g kg\(^{-1}\) moister and 0.2\(^\circ\)C warmer than the environment.

A convergence box was made at 470 mb, at the elevation of the detrainning stratus cloud. As Fig. 12 shows, substantial divergence was occurring at this level, resulting in an outward flux of 77 km m s\(^{-1}\)
and justifying our description of the stratus as detraining. No inversion is evident at the level of detraining. However, the detraining appears to be occurring near the level of minimum $\theta_e$ in the environment. The storm top was in excess of 12 km MSL during this box.

c. Storm phase

From 1150 to 1210 a convergence box (box 6) and a traverse were flown at 730 mb, or well below cloud base (see Fig. 13). By this time, the storm appeared to be developing a region of preferred inflow on the south and east sides. This was visually evidenced by the formation of an extensive region of flat, precipitation-free cloud base on that side of the storm. Precipitation was displaced to the west.

The vertical wind and thermodynamic traces in box 6 (Fig. 13) reflect this asymmetry. The updrafts on the east side were quite strong (up to 5 m s$^{-1}$) and had horizontal scales significantly greater than those evident earlier. The eastern segment of this box track was just under the edge of the cloud. By contrast, the vertical wind fluctuations were quite weak, west of the storm, with possibly a mean downward component. The values of mixing ratio and potential temperature there are characteristic of environmental air at 670 mb, indicating that this air had been forced down by 60 mb.

We hypothesize that the change from essentially symmetric inflow recorded in box 4 to the asymmetric flow of box 6 is the result of internal storm dynamics dominating the bouyant flow associated with the heat island. The net flux for box 6 was only 14 km m s$^{-1}$ inward, with strong inflow on the south and east sides counterbalanced by flow away from the storm on the west side. It is interesting to note that the low level winds, though weak, were from the southeast, on the inflow side of the storm.

As Fig. 5 shows, the mean potential temperature over the box 6 track was $\sim 0.6^\circ$C warmer than the environment at that level. This anomaly is also present in the traverse following box 6 (see Fig. 13). The heat island is thus still present at this time despite shadowing effects of the cloud.

No evidence of an evaporatively forced downdraft and outflow was found. However, such a flow could have occurred below the level of box 6.

4. Supporting studies

We now present a number of additional case studies as well as a composite measurement that illuminates the nature of mountain-induced convection.

a. Flight 3, 26 July 1980

On this day, the atmosphere was drier and more stable than average. Only small clouds formed, with insignificant effect on the dynamics of the heat island. Six box circuits were flown below mountain top level, alternating between 776 and 732 mb.

Unfortunately, the environment appeared to be changing rather rapidly during this flight. As Fig. 14
rather than time change in the mixing ratio, as the soundings between the boxes to the north of the mountains consistently show mixing ratios as high or higher than the early sounding.

The most interesting feature of Fig. 16 is the observed warming of the convective boundary layer. The increase of potential temperature with time is not smooth, but occurs primarily between ascending and descending soundings. This is probably an artifact due to lag in the reverse flow thermometer. A lag of only a few seconds in the rather rapid descents is sufficient to account for the observed effect.

The vertical lines with time labels represent estimates of the potential temperature of the mixed layer at various times. Ascending and descending values were averaged in each case to minimize the lag effect. The "mixed" layer was clearly not completely mixed in this case, and this represents a source of error. Nevertheless, the intersections of these lines with the 0817 sounding yield a rough estimate of the evolution of the depth of the mixed layer with time.

A computation of the enthalpy added to the mixed layer above 800 mb from 0920 to 1021 yields \(1.1 \times 10^8\) J m\(^{-2}\), corresponding to an average sensible heat flux of 300 W m\(^{-2}\). This is more than twice the sensible heat flux computed for the 6 August case,

shows, the 0817 Queen Air sounding was considerably warmer than the 0500 sounding at Albuquerque. The winds were also increasing with time from the north. As a result, it is difficult to separate the effects of large scale advection and subsidence from local effects.

Fig. 15 shows the evolution of divergence as well as box-mean potential temperatures at the two box levels. It is clear that the level of non-divergence was considerably lower than on 6 August 1980. This can be attributed to the greater static stability below mountain top level, which tends to suppress the development of convection. Nevertheless, the convergent flux in the inflow region reached \(-58\) km m s\(^{-1}\), which is almost as large as that occurring on 6 August.

The altitude changes between boxes were made \(-10\) km north of the northern end of the Magdalena Mountains during this flight. The potential temperature and mixing ratio soundings from these maneuvers are plotted in Fig. 16, along with soundings centered at 0817 and 1107. The early sounding was taken over the plains north and east of the mountains, while the late one was taken while descending over the Rio Grande valley. The apparent drying between these two soundings probably reflects spatial variations,
but still less than the sum of sensible plus latent flux for that day. Little rain occurred previous to the 26 July case, so the ground was very dry. The heat flux into the atmosphere was therefore probably mostly sensible in form, explaining its greater observed value in this case. Subsidence could have led to a slight overestimate of the heat flux.

Table 2 lists the results of all box maneuvers for the sample mentioned in Section 2, i.e., those cases with weak winds, strong insolation and appropriate measurements. These fluxes are plotted versus altitude in Fig. 17, with distinction being made between degrees of cloud development.

The overall pattern is clear: with minor exceptions, convergence occurs only below mountain top level, while divergence is dominant aloft. When significant clouds are present, the level of non-divergence appears to be somewhat higher than in the clear case—near mountain top level rather than roughly mid-level on the mountain. Also the level of maximum divergence is somewhat higher.

In the cloudy case, a secondary divergence maximum occurs near 500 mb. This outflow is invariably associated with a detraining stratus deck. When
clouds are present, weak divergence, possibly consistent with zero, exists between cloud base and the 500 mb outflow.

The above picture is certainly not consistent with the entraining plume model of a cumulus cloud. We now ask whether vertical mixing is present in the clouds.

c. Paluch plots

Paluch (1979) demonstrated by the use of conservative, linearly mixing tracers that vertical mixing was occurring in Colorado cumulus congestus clouds. Her technique is used in Figs. 18 and 19 on two case studies. The tracers used are equivalent potential temperature \( \theta_e \) and total water (liquid + vapor) mixing ratio \( Q \). For non-precipitating, non-radiating clouds, these quantities are (almost) conservative and mix (almost) linearly. Therefore, any mixture of air from a set of source regions must lie in the \( \theta_e - Q \) plane, inside the polygon defined by the \( \theta_e - Q \) values of the source regions. Paluch actually uses a quantity that more accurately accounts for the contribution of water vapor to the specific heat than \( \theta_e \). However, we believe that \( \theta_e \) is sufficiently accurate for our purposes.

For Fig. 18, the cloud had tops lower than 500 mb when the traverses of interest were made and was definitely not precipitating. Since environmental values below 500 mb lie along a nearly straight line in the \( \theta_e - Q \) plane, the polygon of allowable mixtures is very narrow. To good accuracy, all points measured in the cloud fell inside the polygon. This example yields no information about vertical mixing, but tends to confirm the accuracy of the linear mixing hypothesis.

![Paluch plot](image)

**Fig. 18.** Paluch plot for case of 29 July 1980 (flight 4).
With confidence thus established in the method, we find strong evidence of vertical mixing in traverse 2 of flight 1. This traverse was through a field of non-precipitating, towering cumulus clouds at a level of 544 mb. The traverse occurred at 0950 MST, or nearly an hour before the onset of visible precipitation. Tops at the time were ~400 mb, and the towers traversed were the first towers to reach this level. As Fig. 19 shows, cloud material in this traverse must include environmental air up to ~400 mb if the linear mixing hypothesis holds, indicating that some air has been transported down from upper levels.

The two most likely sources of error in this measurement are the calibration of the Johnson–Williams probe and possible wetting of the reverse flow thermometer. The temperature was above freezing, avoiding icing problems.

Subsequent to the observational program, the Johnson–Williams instrument was subjected to extensive calibration checks which showed that the instrument was reading 0–20% low, depending on the airspeed. (Keith Griffith, personal communication.) The maximum liquid water content measured was 0.6 g kg⁻¹ in the traverse, and 20% of this yields negligible error.

The reverse flow thermometer probably did not get wet in the short traverse through the cloud. However, if it did it would have produced artificially low temperatures due to the wet bulb effect. Correction for this would have the result of pushing the data points in Fig. 19 farther to the right, strengthening the case for vertical mixing.

The triangles in Fig. 19 actually correspond to data along the track of box 4 in clear air. As Fig. 20 shows, the low values of $\theta_e$ here are associated with very low values of the mixing ratio encountered in segments of this box. The values are in fact impossible to reconcile with the supposedly conservative nature of $\theta_e$.

One suspects instrument malfunction in a situation like this. However, both dewpointers and the Lyman-alpha hygrometer recorded the humidity anomaly, ruling out this possibility.

Examination of the measured flow in Fig. 20 reveals that the regions of low mixing ratio occurred downstream of the edges of the cloudy region (centered in, and somewhat smaller than the box), but not in the middle. Time-lapse pictures of the cloud at this stage revealed turrets rising well above the 490 mb box level, rapidly mixing with the environment, and then collapsing into a detraining stratus layer. This layer was ~300 m above the box level. As in the 6 August case, the region near the stratus was quite divergent with a flux measured in the box of 101 km m s⁻¹. No precipitation was visible at this time, but rain appeared ~25 min after the observation of the mixing ratio anomaly.

A possible explanation for the observed low values of $\theta_e$ is that extreme radiative cooling was occurring in a thin layer at cloud top, resulting in descent of cloudy air to the measurement level via buoyancy forces. The visual behavior of the turrets as they penetrated above 500 mb indicates very dry air aloft. The Albuquerque sounding does not match well with that obtained by the Queen Air, and the air above 450
mb is probably drier than indicated by Albuquerque at 1200 GMT. Taking the extreme case of a cloud top, radiating to a vacuum as a black body, one can relate the thickness \( d \) of the radiating layer to the time \( \Delta t \) available for radiative cooling and the radiative temperature drop \( \Delta T \) by

\[
\rho C_p \Delta T d = \sigma T^4 \Delta t,
\]

(5)

where \( \rho \) and \( C_p \) are the density and specific heat of air at constant pressure, \( \sigma \) is the Stefan-Boltzmann constant and \( T \) is the temperature.

Saturation over water of air with the observed mixing ratio of 0.7 g kg\(^{-1}\) occurs near 300 mb, assuming environmental temperatures. The cloud top was slightly higher than this level at this time. Assuming the dry air came from 300 mb, we take \( T = -21^\circ C \), which is the average of the 300 and 490 mb temperatures, and \( \Delta T = 4^\circ C \). Arbitrarily taking \( \Delta t = 10^3 \) s, we find \( d = 100 \) m. How a layer this thin, atop a cumulus cloud, can remain unmixed for 15 min during a descent of several kilometers is unknown. However, while this explanation is improbable, it is difficult to think of a better one.

5. Discussion

Perhaps the most striking result of the present work is simply the weakness of the flow through the observed systems. The low level convergence with cloudy conditions is only marginally stronger than that occurring when it is dry. Even this mild enhancement may be a selection effect, associated with conditions that produce storms, rather than a result of latent heat release.

The other striking result is that the clear air pattern of convergence below mountain top level and divergence above persists into the early thunderstorm stage. Only during the last box track of flight 10 did we find evidence that a storm was developing its own low level flow pattern. This pattern was basically non-divergent, with inflow and updrafts on the south and east sides balancing forced downdrafts and outflow on the west.

Fig. 21 is an educated guess of the mass budget of

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Fig. 20. As in Fig. 6, except for box 4, 24 July 1980 (flight 1).

Fig. 21. Schematic drawing showing estimated mass budget of a mountain-induced thunderstorm in its early stages.
a typical mountain-induced thunderstorm in its early stages. It is basically a guess because the flow through cloud base is the difference between the inflow below 700 mb and the outflow above. Since the numbers are comparable and not known to better than ±50%, we have arbitrarily assumed that half the estimated inflow 100 mb × 60 km m s⁻¹ = 6 mb km² s⁻¹ goes into outflow below cloud base. (Multiply by 10⁷ to get kg s⁻¹.)

A further uncertainty is the thickness of the detrainment layer near 500 mb. If we assume this layer to be 30 mb thick with a mean outward flux of 70 km m s⁻¹, it carries away 2 mb km² s⁻¹, which is a substantial fraction of the mass current through cloud base. If an additional 1 mb km² s⁻¹ is lost between cloud base and 500 mb, the upward mass current through 470 mb is zero! The uncertainties on this number are very large, so it is clear that we cannot even determine its sign. In any case, it is substantially less than the upward current at 700 mb, due to the strong divergence aloft.

It should not be construed from this that the vertical fluxes of moisture, heat, etc. go to zero at 470 mb. Eddy transports that do not affect the mass balance probably dominate here.

The profile of mass current with height that we see, greatly differs from that observed in thunderstorms over the plains and the tropical Pacific, where monotonically increasing values typically exist up to mid-levels. However, budget studies over the eastern tropical Atlantic show a profile quite similar to ours (Thompson et al., 1979). The simulation of a tropical Atlantic cumulus congestus cloud by Simpson et al. (1982) shows detrainment at 3.6 km MSL, which is consistent with the above budget study. It is important to discover what factors determine this profile of mass current.

If the vertical mixing exhibited in flight 1 (24 July 1981) is typical of the growing phase of our storms, this might provide an explanation of the mass current profile. Raymond (1981) showed that vertical mixing tends to inhibit the overall growth of a storm, while at the same time enhancing eddy development. This is because the resulting evaporation causes net cooling in the upper part of the storm, which steepens the lapse rate for convective eddy production, while generating overall negative bouyancy. Raymond (1981) also showed that this effect disappeared when the storm top rose well above mid-levels due to the higher values of z₁ in the upper troposphere.

An additional point of interest is the effect of subcloud mixed layer processes on storm development. The air in the convective core typically tracks with the mixed layer potential temperature, leading it by ~0.6°C. Thus, as the boundary layer warms up, the air in the core becomes more unstable to moist convection. Small changes in mixed layer moisture content also affect moist instability. These effects are clearly seen in the 6 August case, in which the mixed layer warming and moistening over a period of an hour increased the available convective energy by a factor of four.

Finally, we note the peculiarly low mixing ratios found in the 490 mb box on 24 July. These low mixing ratios and resulting low equivalent potential temperatures are not an invariable feature of the mid-level detrainment layer—the 6 August storm exhibited the same detraining status as well as comparable outward flux at mid-levels, and yet lacked the very dry outflow. At present, we have no explanation for this occurrence, but the radiative cooling hypothesis deserves further investigation.

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APPENDIX

Instrumental Errors

This Appendix is an analysis of instrumental errors involved in measuring the mean cross track wind $\bar{v}_n$ in an aircraft convergence box. We assume that the box track closes perfectly on itself and are not concerned with the effects of temporal change in the system being measured.

The wind velocity $v$ is the difference between the aircraft-relative wind $v_r$ and the velocity of the aircraft $v_a$:

$$ v = v_r - v_a. $$

The inertial platform is used to measure $v_a$, while the gust probe and pitot–static system measure $v_r$. The mean cross track wind is thus given by a time integral (assuming constant aircraft speed)

$$ \bar{v}_n = T^{-1} \int (v_r - v_a) \cdot \hat{n} \, dt, $$

where $\hat{n}$ is the outward normal unit vector to the box track and $T$ is the time required to complete the box.

The primary error in $v_a$ is an 84 min oscillation with amplitude up to ~1 m s⁻¹. Let us model this
error $\Delta v_a$ by a sinusoidal oscillation

$$\Delta v_a = v_0 [\hat{i} \sin(\omega_0 t + \delta_x) + \hat{j} \sin(\omega_0 t + \delta_y)],$$

where $v_0 = 1 \text{ m s}^{-1}$, $\omega_0$ is the angular frequency of an 84 min period ($T_0$), and $\delta_x$ and $\delta_y$ are phase factors. If the convergence box track is idealized by a circle, then the error in the instantaneous cross track wind due to this effect is

$$\Delta v_n = \Delta v_a \cdot \hat{n} = v_0 [\sin(\omega_0 t + \delta_x) \cos \omega t + \sin(\omega_0 t + \delta_y) \sin \omega t],$$

where $\omega = 2\pi / T$.

We assume that boxes are completed in a time $\ll T_0$, an assumption well justified for our flights. Thus, $\omega \gg \omega_0$. An expansion in powers of $t$ is therefore justified, with the result that the quadratic term is the first non-vanishing term in the error in the mean cross track wind $\Delta \bar{v}_n$. The result is an upper bound on $\Delta \bar{v}_n$:

$$\Delta \bar{v}_n < 2v_0 (T/T_0)^2.$$  

For $T = 20 \text{ min}$, $v_0 = 1 \text{ m s}^{-1}$ and $T_0 = 84 \text{ min}$, this yields a maximum error in $\bar{v}_n$ due to the inertial platform of 0.11 m s$^{-1}$.

For level flight the error in $v_\gamma \cdot \hat{n}$ reduces to the accuracy with which the vertical gust probe vanes measure the yaw angle of the aircraft. NCAR quotes a maximum error, exclusive of a constant offset, of 10% in the yaw angle measurement. Our examination of the computed wind, during extreme yaw maneuvers, supports this figure. For a typical true airspeed of $80 \text{ m s}^{-1}$, this means that $1^\circ$ of yaw yields $0.14 \text{ m s}^{-1}$ error in the cross track wind. In turbulence, the Queen Air can exhibit yaws up to $\pm 3^\circ$. However, this is an oscillatory phenomenon, and the average over a few tens of seconds is invariably much less. Examination of all boxes and reverse track maneuvers in flights 3 and 10 indicates that actual smoothed yaws stay within $\pm 0.6^\circ$ of the mean yaw, resulting in a maximum possible error due to this cause of $\pm 0.08 \text{ m s}^{-1}$.

The previously mentioned constant offset in yaw is removed by forcing both legs of a reverse track maneuver to yield the same crosswind. The accuracy of this procedure depends on how smooth the air is. It is difficult to give an estimate of the contribution of this procedure to the total error.

Banked turns in the box maneuver involve the pitch angle vanes in the cross track wind calculation. For shallow banks occupying a small fraction of the box track, the effect is not large, but is difficult to evaluate.

Excluding the last two effects, a conservative estimate for the accuracy of the mean cross track wind measurement is $0.2 \text{ m s}^{-1}$. Two actual boxes yield further insight. Box 2 of flight 2 was done at a level well above any convective activity in quiescent air. The resulting mean cross track wind was $0.09 \text{ m s}^{-1}$. Box 1 of flight 11 was a very narrow box, with the expectation of almost no flux as a result. The mean cross track wind in this case was $0.08 \text{ m s}^{-1}$. We therefore conclude that $0.2 \text{ m s}^{-1}$ is a conservative estimate of the potential error in the measurement of $\bar{v}_n$.

REFERENCES


