Small-Scale Motions Observed by Aircraft in the Tropical Lower Stratosphere: Evidence for Mixing and its Relationship to Large-Scale Flows

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ABSTRACT

Measurements of temperature and ozone from instrumented aircraft in the tropical lower stratosphere show the presence of small-scale disturbances generated by 1) underlying cumulus convection and 2) Kelvin–Helmholtz instability. The disturbances associated with underlying convection have peak-to-peak vertical parcel excursions of ~300 m. Flying conditions were smooth, suggesting an ensemble of gravity waves and little or no turbulent mixing. It is nevertheless possible that these waves break at other altitudes, leading to turbulent mixing and net fluxes of vertically stratified tracers.

Disturbances attributed to KH instability implied vertical parcel excursions of 300–400 m. The disturbances coincided with rough flying conditions, suggesting turbulent mixing. A linear stability analysis of the atmospheric basic state defined by high-resolution radiosondes shows fastest growing waves with horizontal wavelengths of 1.4–1.8 km, consistent with the aircraft observations. The strong shears responsible for the KH instability are due to large-scale waves propagating into a region of small intrinsic frequency. Radiosonde observations show that the zonal length scale of these waves is ~1000 km.

1. Introduction

An important component in evaluating man's impact on stratospheric chemistry is the rate at which photochemically inactive pollutants (e.g., fluorocarbons) are transported from their sources in the troposphere to photochemically active regions in the upper stratosphere. As noted by Holton (1984), the bottle-necks appear to be the tropopause and lower stratosphere. This has also been recognized by one-dimensional photochemical modelers (e.g., McConnell and McElroy, 1973), who have used eddy diffusion coefficients with sharp minima in the lower stratosphere.

Based on work by Dunkerton (1978), Mahlman et al., (1983) and many others, we have a conceptual model of tracer transport in the lower stratosphere, which is illustrated by an adaptation of Holton's (1984) schematic (Fig. 1). In this model, tropospheric air is transferred into the lower tropical stratosphere, either by cumulonimbus towers (Danielsen, 1982) or larger scale motions (Brewer, 1949). It is then transported upward by the "diabatic circulation" (Dunkerton, 1978) and exchanged with air in midlatitudes by large-scale eddies (Danielsen, 1980).

In this picture, small-scale mixing has two important roles in tropical lower stratospheric transport, illustrated by the dotted arrows in Fig. 1. First, it mixes air transferred by large-scale eddies from midlatitudes with the tropical environment. For quasi-linear wavelike eddies, rapid small-scale mixing is crucial to their effectiveness in latitudinal tracer transport, since advected tracers must be mixed with their new environment before the air mass returns to its original position (Danielsen, private communication, 1986). For large amplitude nonlinear eddies, the rapidity of small-scale mixing is less important. These eddies transfer air irreversibly, and the air can then be mixed with its new environment over a long period of time.

The second role of small-scale mixing is to transport tracers vertically down the gradient of the mean tracer distribution. In the tropics, this reinforces the diabatic circulation for tropospheric tracers and counteracts it for stratospheric tracers. The ratio of the magnitudes of transport due to vertical mixing and diabatic circulation can be expressed as

\[ r \sim \frac{w'X'}{wX} \sim \frac{K_{zz}\bar{X}}{K_{zz}} \sim \frac{K_{zz}}{Dw}. \]

Here \( D \) is the depth over which the zonally averaged tracer mixing ratio \( \bar{X} \) varies by order unity, \( w \) is the vertical velocity, and \( K_{zz} \) is the vertical eddy diffusivity due to small-scale motions. Overbars denote zonal averages and primes the small-scale deviations from those averages. For fluorocarbons in the tropics, \( D \) can be as
small as 5 km (Poppoff et al., 1979). Estimates of \( K_{oz} \)
due to stratospheric turbulence from radar, balloon and
aircraft data vary from 0.012–0.3 m\(^2\) s\(^{-1}\) (Woodman,
1981). These yield \( r \) values from 0.01 to 0.3, assuming
upward mean meridional motion (induced by radiative
heating) of 0.0002 m\(^{-1}\) in the tropical lower stratosphere.
Thus, as Woodman points out, present estimates of small scale mixing are inadequate even for
establishing whether such mixing is important for direct
vertical transfer of tracers in the tropical lower stratosphere. There is thus a clear need to understand
the mechanisms that generate small scale mixing in the
tropical lower stratosphere, and evaluate their effects
quantitatively.

This paper addresses two possible mechanisms: large amplitude waves generated by underlying cumulus
towers, and shear instability associated with upward
propagating large scale waves. There is evidence for
both of these mechanisms from the NASA ITCZ field
experiment, performed in Panama during July, 1977
(Poppoff et al., 1979). Section 2 discusses the available
data from this field experiment. Section 3 reviews the
aircraft measurements and discusses the cumulus mechanism. The shear instability mechanism is
discussed in sections 4 and 5: in section 4 we perform a
linear stability analysis of the observed winds to show
that KH shear instability can explain the aircraft ob-
servations; in section 5 we examine large scale motions
to establish how the unstable shears are generated.

2. Data

The NASA ITCZ experiment from 16 to 31 July
1977 was designed as a survey mission for tropical
stratosphere–troposphere exchange. Each day during
the two-week experimental period, a U-2 aircraft was
flown in a standard north–south flight pattern at six
standard pressures, whose nominal altitudes ranged
from 14.10 to 21.30 km, at an approximately constant
interval of 1.45 km. Each north–south flight leg was
\( \approx 25 \) minutes long, covering \( \approx 2.5 \) degrees of latitude
at the aircraft speed of \( \approx 200 \) m s\(^{-1}\). On four of these
flights, measurements of temperature and ozone at 1
sample per second were available. Unfortunately, air-
craft pressure data for these four flights was lost due
to a recording malfunction. However, since the nomi-
nal flight pressures on each day were identical, pressure
data from other flights could be used to estimate the
mean altitude for each leg, and the altitude variance.

High-vertical-resolution radiosondes taken at 6-h
intervals (Poppoff et al., 1979) were also available. The
radiosondes were tracked by radar to improve accuracy
and to eliminate the uncertainty in altitude produced
by errors in the radiosonde pressure measurements.
The measured variables (temperature, winds and rela-
tive humidity) were then filtered to remove vertical
scales less than 50 m (Danielsen et al., 1980).

3. Aircraft ozone and temperature variations

Figures 2 and 3 show traces for the temperature and
ozone mixing ratio at 17.00 km, 18.45 km, 19.85 km
and 21.30 km on July 27 (Starr et al., 1979). Here,
time has been converted to latitude using the position of
the aircraft in the flight pattern. The altitude for
each leg is arrived at as follows. First, the average
pressure for that leg was calculated from flights when
pressure measurements were available. Then, those
pressures were compared to pressures measured by the
radiosonde on 27 July.

Three basic types of short-period variations are
noted: 1) irregular variations that are highly correlated
between ozone and temperature, with a broad range
of horizontal scales (5–25 km (17.00 km leg); 2) vari-
ations having little or no correlation between ozone
and temperature (18.45 km leg); 3) short-period, regular
oscillations having periods with respect to the aircraft
of \( \approx 23 \) seconds (south portion of the 19.85 km leg).
These short-period oscillations also have high corre-
lation between ozone and temperature. There are also
larger scale variations in ozone and temperature, as
evidenced by the steplike shift in the middle of the
19.85 km flight leg. A general feature of the traces is
the very different character of variations at each of the
levels, in contrast to the good coherence between levels
apparent in HICAT data (Lilly and Lester, 1974). This
lack of coherence is consistent with the fine layering
in large scale tropical stratospheric wind fields (see
discussion in section 5).
tures, shown in Fig. 4a. These structures are from the radiosonde soundings made at 1000 and 1600 UTC 27 July (universal time, hereafter referred to as “Z” time—e.g., “10Z”). They bracket the 27 July flight in time.

At 17.00 km, the peak-to-peak temperature variations expected from either the standard altitude variations δz estád or the extreme aircraft altitude variations δz exa are much smaller than the observed variations of 5.8°K. Even with a higher average altitude of 17.15 km, (which places the aircraft in the middle of the strong vertical temperature gradient evident in Fig. 4a), the observed variations substantially exceed those expected even from the extreme altitude variations.

At 19.85 km for the 10Z sounding, the peak-to-peak temperature variations expected from altitude variations are again much smaller than the observed variations. For the 16Z sounding, though, the disparity between the amplitude of the short-period oscillations and altitude-induced temperature changes is much smaller. However, U-2 temperature and pressure data from the 1980 water vapor exchange experiment in

Fig. 2. U-2 temperature on 27 July 1977 at four altitudes as a function of time and latitude. Times are universal time. Arrows indicate direction of flight.

a. Variations due to altitude changes

Interpretation of these observed variations is hampered by the loss of pressure data, making it difficult to verify that the flight legs are indeed at constant pressure. This is important in establishing whether the variations are due to aircraft motions in regions with vertical temperature and ozone gradients, or to real horizontal gradients. Fortunately, pressure data is available from flights on several other days during the experiment on which the identical flight pattern and set of altitudes were flown. Though the pressure variations would certainly not be identical, the pressure variance on a “constant pressure altitude” flight leg would be similar.

Table 1 shows the peak-to-peak temperature variations due to characteristic aircraft altitude variations alone in a horizontally homogeneous atmosphere at 17.00 and 19.85 km. Observed peak-to-peak temperature variations are shown for comparison. We have assumed two alternative vertical temperature struc-

Fig. 3. As in Fig. 2 except for ozone in ppmv.
Table 1. Expected temperature deviations, $\delta T_{at}$ and $\delta T_{ext}$, due to standard aircraft altitude deviations $\delta z_{at}$ and extreme altitude deviations $\delta z_{ext}$, respectively. Observed aircraft temperature deviations $\delta T_{ob}$ are shown for comparison: for 17.00 km, these are the extreme peak-to-peak temperature variations during the flight leg; for 19.85 km, they are the peak-to-peak amplitude of the short-period oscillations (Fig. 2). The $\delta z$ are in meters and $\delta T$ in $^\circ$K. The vertical temperature structure in which the aircraft flies is defined by the indicated sounding on 27 July 1977.

<table>
<thead>
<tr>
<th>Altitude (km)</th>
<th>Sounding (UTC)</th>
<th>$\delta z_{at}$</th>
<th>$\delta T_{at}$</th>
<th>$\delta z_{ext}$</th>
<th>$\delta T_{ext}$</th>
<th>$\delta T_{ob}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>17.00</td>
<td>10</td>
<td>68</td>
<td>0.10</td>
<td>150</td>
<td>0.18</td>
<td>5.8</td>
</tr>
<tr>
<td>17.00</td>
<td>16</td>
<td>68</td>
<td>0.46</td>
<td>150</td>
<td>1.10</td>
<td>5.8</td>
</tr>
<tr>
<td>17.15</td>
<td>10</td>
<td>68</td>
<td>2.06</td>
<td>150</td>
<td>3.16</td>
<td>5.8</td>
</tr>
<tr>
<td>17.18</td>
<td>10</td>
<td>68</td>
<td>1.14</td>
<td>150</td>
<td>2.38</td>
<td>5.8</td>
</tr>
<tr>
<td>19.85</td>
<td>10</td>
<td>81</td>
<td>0.10</td>
<td>140</td>
<td>0.26</td>
<td>2.4</td>
</tr>
<tr>
<td>19.85</td>
<td>16</td>
<td>91</td>
<td>0.92</td>
<td>150</td>
<td>1.62</td>
<td>2.4</td>
</tr>
</tbody>
</table>

Panama (Margozzi, 1983) indicates that the time scale for aircraft altitude changes is much larger than the $\sim$23-sec period of the oscillations. Instances of quasi-regular oscillations similar to those in Figs. 2 and 3 were never associated with altitude changes. In fact, there were very few cases of quasi-oscillatory altitude variations, and of these, none had periods less than several minutes.

Aircraft altitude variations in both the 1977 and 1980 experiments (other than intentional ascents and descents) tended to be irregular in time, with occasional steplike behavior similar in appearance to that near 9.2°N latitude in the 19.85 km leg (Figs. 2 and 3). The $\sim$2.8°K magnitude of this step, however, substantially exceeds the extreme variation expected from altitude variations. A strong horizontal temperature gradient is the more likely explanation. In fact, Fig. 4a shows good temperature correspondence between the south portion of the 19.85 km flight leg and the 10Z sounding, and the north portion and the 16Z sounding, suggesting a large scale, southward moving temperature feature. The radiosonde ozone measurements in Fig. 4b are consistent with this interpretation. The absolute agreement between radiosonde and aircraft ozone is poor, but the radiosonde ozone tendency at 19.85 km from 10Z to 16Z is in the same sense as the step in the aircraft ozone near 9.2°N in Fig. 3.

b. Calculation of vertical parcel excursions

The existence of large amplitude, small scale temperature and ozone fluctuations does not necessarily imply vertical transport in the stratosphere. If the fluctuations represent completely conservative wave phenomena, for example, they will have no significance for vertical mixing or transport at all. To establish vertical transport, vertical wind measurements are required, which are not available from this dataset. However, by augmenting the aircraft data with nearby radiosonde measurements, minimum vertical parcel excursion distances associated with these fluctuations can be calculated. Such vertical parcel excursions are perhaps a better measure of the significance of the temperature and ozone disturbances than the variances themselves, for the following reasons. First, the vertical excursions yield a minimum vertical extent for the phenomenon, which, when compared to the vertical scale of large scale waves, indicates the potential effect the small scale phenomenon has on these large scale waves. Second, if irreversible mixing does occur during the parcel excursions, the excursion can be interpreted as a mixing length (Reed and German, 1967). Local eddy diffusivities associated with the phenomenon can then be estimated.

To calculate the vertical excursions, we make three important assumptions: first, that vertical aircraft motion plays no part in producing the observed ozone and temperature variations; second, that air parcel potential temperature and ozone are conserved along trajectories; and third, that trajectory slopes are much larger than the slopes of the contours of the large scale ozone and potential temperature fields. The first assumption was addressed in the previous subsection. The second implies no mixing, and would seem to defeat the purpose of estimating potential significance for irreversible vertical mixing. In fact, conservative motion yields a minimum vertical excursion distance for tracers whose variation with altitude is monotonic (like ozone and potential temperature). This is because mixing air from distant altitudes (with a large tracer deviation) with air from nearby altitudes (with correspondingly smaller deviations) reduces the observed tracer deviation. To produce the same tracer deviation, parcels would have to originate from more distant al-

![Figure 4](http://journals.ametsoc.org/journals/jas/article-pdf/43/24/3210/3423621/1520-0469(1986)043_3210_ssmoba_2_0_co_2.pdf)
titudes to balance the influence of air mixed in at altitudes closer to the level of observation. The third assumption of weak mean horizontal gradients is a good one in the tropics, as confirmed by the aircraft observations. Temperature changes from one end of the flight legs to the other are generally minimal. The exception is at 19.85 km (and to a lesser extent at 18.45 km); this, however, is due to a narrow steplike feature that should not affect calculations for the oscillations at that altitude.

With these assumptions, we can derive the vertical parcel excursion distances \( \delta \) required to produce the aircraft variations. For the basic states in which the gravity waves are embedded, we use the two soundings bracketing the flight (Fig. 4a). For temperature, the vertical excursion distance \( \delta T_a \) is defined by

\[
\delta T_a = \left( \frac{p_a}{p_0} \right)^{\frac{\gamma}{\kappa}} \int_{z_1}^{z_2} d\theta,
\]

where,

- \( p_0 = 1000 \text{ mb} \)
- \( p_a \) average aircraft pressure
- \( \delta T_a \) aircraft temperature change
- \( \theta \) radiosonde potential temperature.

The substantial disagreement between the aircraft and radiosonde ozone data (Fig. 4b) makes derivation of \( \delta \) for ozone more difficult. Previous studies (Geraci and Luers, 1978) show minimal bias in the total ozone derived from the electrochemical sonde (compared to Dobson measurements), so the error is probably due to line losses in the aircraft measurements. Note that the bias between aircraft and radiosonde measurements is not constant; instead the difference increases with altitude and increasing ozone. This suggests that we can express the true ozone mixing ratio \( \mu \) in terms of the aircraft measurements \( \mu_a \) as

\[
\mu = \mu_a \alpha
\]

Table 2. Radiosonde and aircraft ozone (ppmv) and their ratios \((\alpha \text{ in text})\) as functions of altitude. The radiosonde values are taken from the indicated soundings on 27 July 1977. Aircraft values are the averages over the flight legs except for 17.00 km and for 19.85 km (see text).

<table>
<thead>
<tr>
<th>Altitude (km)</th>
<th>Sounding (UTC)</th>
<th>Ozone</th>
<th>Radiosonde</th>
<th>Aircraft</th>
<th>Ratio</th>
</tr>
</thead>
<tbody>
<tr>
<td>17.00</td>
<td>10</td>
<td>0.227</td>
<td>0.145</td>
<td>1.57</td>
<td></td>
</tr>
<tr>
<td>17.00</td>
<td>16</td>
<td>0.190</td>
<td>0.145</td>
<td>1.31</td>
<td></td>
</tr>
<tr>
<td>18.45</td>
<td>10</td>
<td>0.434</td>
<td>0.266</td>
<td>1.63</td>
<td></td>
</tr>
<tr>
<td>18.45</td>
<td>16</td>
<td>0.476</td>
<td>0.266</td>
<td>1.78</td>
<td></td>
</tr>
<tr>
<td>19.85</td>
<td>10</td>
<td>0.915</td>
<td>0.580</td>
<td>1.58</td>
<td></td>
</tr>
<tr>
<td>19.85</td>
<td>16</td>
<td>1.060</td>
<td>0.640</td>
<td>1.66</td>
<td></td>
</tr>
<tr>
<td>21.30</td>
<td>10</td>
<td>1.870</td>
<td>1.015</td>
<td>1.84</td>
<td></td>
</tr>
<tr>
<td>21.30</td>
<td>16</td>
<td>1.970</td>
<td>1.015</td>
<td>1.94</td>
<td></td>
</tr>
</tbody>
</table>

Table 3. Vertical parcel excursions \( \delta \) in meters implied by the temperature and ozone fluctuations observed by the aircraft at the indicated altitudes. Also included are \( \delta \) implied by temperatures corrected for finite instrument response time. The vertical temperature and ozone structures are defined by the indicated sounding on 27 July 1977.

<table>
<thead>
<tr>
<th>Altitude (km)</th>
<th>Sounding</th>
<th>Temperature</th>
<th>Ozone</th>
<th>Temperature (corrected)</th>
</tr>
</thead>
<tbody>
<tr>
<td>17.00</td>
<td>10Z</td>
<td>192</td>
<td>308</td>
<td></td>
</tr>
<tr>
<td>17.00</td>
<td>16Z</td>
<td>332</td>
<td>410</td>
<td></td>
</tr>
<tr>
<td>19.85</td>
<td>10Z</td>
<td>268</td>
<td>321</td>
<td>395</td>
</tr>
<tr>
<td>19.85</td>
<td>16Z</td>
<td>129</td>
<td>166</td>
<td>194</td>
</tr>
<tr>
<td>19.58</td>
<td>16Z</td>
<td>252</td>
<td>273</td>
<td>371</td>
</tr>
</tbody>
</table>

where \( \alpha \) varies slowly with altitude. Table 2 shows the calculated \( \alpha \) for the four flight levels shown in Fig. 2. Aircraft and radiosonde ozone measurements are also shown. For the 18.45 and 21.30 km flight legs, average aircraft ozone values were used. At 19.85 km, though, south portion aircraft ozone is compared with 10Z radiosonde ozone, and north portion aircraft ozone with 16Z radiosonde ozone. This is consistent with the aircraft-radiosonde temperature relationships evident in Fig. 4a discussed in the previous subsection. At 17.00 km, the aircraft data suggests “background” values of temperature and ozone significantly smaller than the strict quantitative averages, with positive deviations, presumably associated with downward air motion. The aircraft ozone used for comparison with the radiosonde is taken as this background, rather than the actual average. The table shows that \( \alpha \) generally increases with altitude, but much more slowly than the ozone mixing ratio. The extreme variation is about 50% (from 1.31 to 1.94), compared to a factor of seven for the aircraft ozone.

Given the slow variation of \( \alpha \) with altitude, the deviation of true ozone mixing ratio associated with the vertical parcel excursions \( \delta \mu_a \) is

\[
\delta \mu = \alpha \delta \mu_a.
\]

The vertical parcel excursion \( \delta \) is then defined by

\[
\delta \mu = \left( \frac{p_a}{p_0} \right)^{\frac{\gamma}{\kappa}} \int_{z_1}^{z_2} d\theta,
\]

where \( \mu_r \) is the radiosonde ozone mixing ratio.

The calculated \( \delta \) is implied by the ozone and temperature fluctuations, assuming constant aircraft altitude, conservative, adiabatic motion and horizontal homogeneity in the basic state are shown in Table 3. We have included separate calculations for ozone and temperature, and for the two radiosondes that bracket the aircraft flight. At 17.00 km, the ozone excursion
values are probably too large, due to failure of the radiosonde data to properly resolve the sharp increase in ozone that most likely occurs at 17.15 km in conjunction with the sharp temperature increase apparent in Fig. 4a. (Radiosonde ozone data was available at 300 m intervals, temperature data at 100 m intervals.) Taking the values derived from the temperature, the vertical excursions at 17.00 km are then 200–300 m, depending on whether the 10Z or the 16Z sounding best represents the environment at the time of the flight. Since the flight time of the 17.00 km leg was 14Z, closer to 16Z than 10Z, the 300 m value is probably more appropriate.

At 19.85 km, the results should be adjusted for the response time of the Rosemount temperature probe on the aircraft, which is ~4.5 seconds at this altitude. Assuming that the measured temperature relaxes to the actual temperature with a coefficient of (4.5 sec)^{-1}, the measured temperature amplitude for a 23-sec oscillation would be 63% of the actual temperature amplitude. Correction for this factor, also shown in the table, does not improve the overall agreement between the ozone and temperature calculations. However, the fact that the corrected and uncorrected results for theta bracket those for ozone reinforces the basic conclusion—namely, that there is strong evidence for peak-to-peak vertical parcel excursions at this altitude of 150–200 m for 16Z, and 300–400 m for 10Z.

Because of the large discrepancy between calculated excursion distances for the two soundings, it is important to establish which one best represents the environment for the oscillations. As indicated above, comparison of the mean values of the north and south portions of the flight leg with the two radiosondes indicates that the portion of the flight leg containing the oscillations (“south portion” in Fig. 4) corresponds to the 10Z sounding. The 300–400 m value for the vertical excursion distance thus appears to be more appropriate (assuming no errors in aircraft or radiosonde temperatures). On the other hand, if the 16Z sounding does represent the environment for the oscillations and the aircraft and radiosonde temperatures are correct, the discrepancy between the radiosonde temperatures and the “south portion” aircraft temperatures in Fig. 4 would be due to errors in the aircraft altitude. Table 3 shows vertical excursion distances calculated for the 16Z sounding and a 19.58 km aircraft altitude, which is where 16Z radiosonde temperatures and “south portion” aircraft temperatures match. These vertical excursions are comparable to those for the 10Z case. Of course, matching the 16Z radiosonde temperatures to the aircraft temperatures implies a 280 m altitude deviation from the experimental average for this flight leg, much larger than the expected deviation of 150 m (Table 1). However, the comparable vertical parcel excursions calculated for the 10Z and 16Z environments reinforces the conclusion that the short-period oscillations in the southern portion of the 19.85 km altitude level are associated with 300–400 m vertical parcel excursions.  

### c. Interpretation—17.00 km fluctuations

The calculated \( \delta_{ex} \) for both altitudes are quite large. For example, they are within a factor of \(~2\) of the vertical scales of some of the large scale propagating waves apparent in Fig. 12 (e.g., the case outlined by the solid and dotted lines). It is important, then, to establish: 1) the mechanism generating the fluctuations; and 2) the net tracer transport associated with them, if any. Understanding their origins is also important for establishing their temporal and spatial frequency, and thus their importance to overall transport.

The mechanism generating these fluctuations appears to be underlying cumulus convection (Starr et al., 1979). Figure 5 shows the temperature trace from the 17.00 km flight leg for 31 July. As on 27 July, the same large amplitude irregular fluctuations conditions are apparent. Easterly wave passages over Panama occurred on both 27 and 31 July (Poppoff et al., 1979), with substantial cumulus activity noted over the flight track by the pilots on both days. On 31 July, however, the pilot was able to note the location in the flight track where cumulus tops were approached. This location was also where the large amplitude fluctuations were observed. The horizontal scale of the fluctuations, 5–25 km, is also consistent with the range of sizes of tropical convective cells (Leary and Houze, 1979). In contrast, the temperature fluctuations observed by the aircraft on 26 and 30 July, when convection in the flight track area was minimal, had much smaller amplitude at all altitudes.

In the absence of aircraft wind measurements, the only available indication of the reversibility/irreversibility of the fluctuations at 17.00 km is in the pilot report, which indicated smooth flying. This suggests that the observed fluctuations are an ensemble of gravity waves that are not producing any mixing, at least not at this altitude. However, wave breaking and mixing may occur at altitudes above and/or below the aircraft flight level, so the fluctuations may have vertical transport associated with them.

### 4. Interpretation—19.85 km fluctuations

The fluctuations at 19.85 km are fundamentally different from those at the lower altitude in three ways.

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1 Examination of the radiosonde data also explains the third type of fluctuation referred to at the beginning of this section—fluctuations where ozone–temperature correlation is poor, such as at 18.45 km. At this altitude, the aircraft is flying very close to an ozone minimum; potential temperature, however, still increases with altitude. Vertical air parcel motions convert vertical gradients of conservative tracers to horizontal gradients, yielding the observed horizontal temperature variations. In the absence of vertical ozone gradients, however, those same vertical motions will not produce horizontal ozone variations.
First, they are unusually regular. Between 8.5° and 9°N latitude, for example, a fairly constant period of ~23 seconds is maintained for nine cycles, with the same period recurring for shorter stretches elsewhere on the south portion of the flight leg. Second, the pilot reported "rough flying," suggesting turbulence, a unique observation for this experiment at such a high altitude. Third, strong and persistent zonal wind shears are apparent in both the 10Z and 16Z soundings bracketing the flight (Figs. 6 and 7). These shears were among the strongest and most persistent observed during the experiment, as shown by the time-height section of inverse Richardson number (see Fig. 10).

The regularity of the oscillations, coupled with the report of rough flying and the strong wind shears suggests that Kelvin–Helmholtz instability may be occurring near the 19.85 km. In fact, the minimum Richardson number as calculated directly from the sounding data, which occurs at ~19.5 km in both soundings, is 0.28 at 10Z, and 0.20 at 16Z. Given that the radiosonde data has been filtered, both soundings may well have subcritical Richardson numbers.

### a. Stability analysis

An approach for verifying whether KH instability is indeed responsible for the observed oscillations is a linear stability analysis of the radiosonde data. Such a stability analysis would test two things: first, whether the wavelength, propagation speed and propagation direction of the unstable modes are roughly consistent with the wave period observed by the aircraft; and second, whether the vertical structure of the temperature perturbation of the unstable mode can extend from 19.5 km (where the shear is strongest and the unstable mode is expected to have its largest amplitude) to the aircraft altitude of 19.85 km.

There are two major problems in interpreting stability analyses of data. First, the vertical filtering may mask a stronger vertical shear and lower Richardson number than the data shows. Second, even filtered data may include a significant contribution from the disturbance field. Both of these will result in errors in the basic state field, and the associated unstable wave properties. These problems are made more severe by the fact that $\text{Ri} < 0.25$ is required for instability, which means that fairly small vertical scales must be retained.

Our approach was to fit the data to quasi-analytic functions (shown in Figs. 6 and 7 with the data) as
faithfully as possible. The major modifications to the data were in the 10Z meridional wind field, where the short vertical scale oscillations were removed, and in the 10Z zonal wind field, where the vertical scale of the hyperbolic tangent was reduced by ~10% to reduce the minimum Ri from 0.28 to 0.22. We justify the modification to the meridional wind by noting that the common feature of the 10Z and 16Z meridional winds is an overall increase in the northward component in the 19–20 km region, with no evidence of the 10Z vertical oscillations in the 16Z data. On this basis, we assume that these oscillations are part of the disturbance field and not the basic state.

Implicit in taking the fields in Figs. 6 and 7 as appropriate basic states, rather than fields with even smaller vertical scales, smaller Ri and shorter unstable wavelengths (Lalas and Einaudi, 1976), is the hypothesis that the instability will tend to maintain the basic state shear at or near marginal stability, i.e., Ri ~ 0.25. Results from the HICAT experiment (Heck, 1977) indicate that this assumption is justified for frontal zones in the midlatitude upper troposphere. However, strong shears in the tropical stratosphere such as those in Figs. 6 and 7 are probably generated by upward propagating inertia–gravity waves (Cadet and Teitelbaum, 1979; Mackawa et al., 1984). As the waves approach critical levels, the vertical scales decrease (Booker and Bretherton, 1967) with corresponding increases in the vertical wind shears associated with the waves. If the time scale for the growth of these shears is short compared to the KH growth times, KH instability would not grow rapidly enough to dissipate the increasing vertical temperature gradients associated with the upward propagating wave, resulting in Ri < 0 and convective instability (Fritts, 1984).

However, as Fritts points out, this is only true if the strong observed shears are due to short period (i.e., doppler shifted frequency much greater than the Coriolis parameter) gravity waves. In this case, the persistence of the shears suggests a time scale of about a day (see section 5), implying a characteristic frequency of $1-2 \times 10^{-5} \text{s}^{-1}$, comparable to the Coriolis parameter of $2.3 \times 10^{-5} \text{s}^{-1}$ at 9°N. As Fritts indicates, the influence of rotation will increase the wind perturbations relative to the temperature perturbations of upward propagating waves. Thus, KH instabilities may well grow rapidly enough to govern the mean wind shears in which they are embedded. In fact, Barat's (1982) observational study has confirmed the presence of persistent turbulent layers with minimum Richardson numbers of 0.25.

Assuming horizontal homogeneity in the basic state wind and temperature fields, the linearized perturbation equations for a nonhydrostatic fluid reduce to

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2 For the winds, these were essentially hyperbolic tangents, with polynomials at the top and/or bottom. For the Brunt–Väisälä frequency, polynomials were used.

3 Wave-induced vertical shears will also amplify with altitude due to decreases in density. Generally, only in the upper stratosphere and mesosphere are the vertical shears associated with this mechanism large enough to generate instability.
\[ w_{zz} + w \left[ \left( \frac{k u_{zz} + l \bar{v}_{zz}}{\omega} \right) + \left( \frac{k^2 N^2}{\omega^2} \right) - k^2 \right] = 0 \] (5)

with

- \( w' \) perturbation vertical velocity
- \( k \) zonal wavenumber
- \( l \) meridional wavenumber
- \( \kappa^2 \) \((= k^2 + l^2)\)
- \( \omega \) wave frequency at the ground
- \( \bar{\omega} \) \((= \omega - k \bar{u} - l \bar{v})\)
- \( N^2 \) Brunt–Väisälä frequency squared
- \( \bar{u} \) basic state zonal wind
- \( \bar{v} \) basic state meridional wind

Equation (5) is identical to the equation used by Lalas and Einaudi (1976), except that 1) we have included the effects of shear in \( \bar{\omega} \) and 2) we have ignored the effect of a finite scale height. This approximation is certainly justified since the shear zone is \( \sim 0.5 \) km thick. We require that the solution decay exponentially at the boundaries, and that energy flux be away from the zone of instability. The boundaries are at 18 and 22 km.

We use a method similar to that adopted by previous investigators. With (5) expressed in finite difference form, we guess the eigenvalue \( \omega \), and evaluate the determinant of the appropriate matrix. A variant of Newton's method is used in an iteration process to find the value of \( \omega \) for which the determinant vanishes. The minimum values of Ri are quite close to the marginally unstable value of 0.25, so grid sizes of 1–10 m were required. All results were tested for sensitivity to grid size. In all cases, halving the grid size produced growth rate changes of less than 0.1%.

Previous theoretical work on hyperbolic tangent shear instability with constant \( N^2 \) (Lalas and Einaudi, 1976) has shown that only one growing mode (called the "KH mode") exists in the absence of rigid boundaries. With rigid boundaries, secondary unstable modes in addition to the KH mode are excited by reflection and overreflection between the boundaries and the steering level. There are no rigid boundaries in our formulation. Nevertheless, it is possible that the strong \( N^2 \) variations might play the role of rigid boundaries and cause sufficient partial reflection to excite secondary modes. We examined a large range of horizontal wavenumbers and phase speeds for each basic state but could find no evidence of such modes.

The properties of the fastest growing waves for the 10Z and 16Z calculations are shown in Table 4. Also included in Table 4 are the characteristics of the fastest growing waves for a number of other cases, "Constant \( N^2 \)" is the same as the respective cases, with \( N^2 \) everywhere equal to its value at the level of minimum Ri. "Pure Tanh" indicates constant \( N^2 \), with zonal and meridional winds represented by the dotted lines in Fig. 6—that is, without the polynomial fit at the top and/or bottom of the shear zone. These cases are included to examine the importance of the detailed structure in \( N^2 \) and the far-field structure in the winds.

The propagation angle, the wavelength of fastest growth, and the phase speed are not very sensitive to either \( N^2 \) or the far-field wind structure. The propagation angle \( \theta_p \) is close to

\[ \theta_p = \arctan(\bar{v}_z/\bar{u}_z) \] (6)

with \( \bar{v}_z \) and \( \bar{u}_z \) evaluated at the level of minimum Ri. The difference in \( \theta_p \) for the "constant \( N^2 \)" case is due to lower altitude of the point of minimum Ri, and the lower meridional shear there. The most sensitive parameter is the growth rate, which increases markedly for the constant \( N^2 \) case. This is due to the reduction in altitude-averaged \( N^2 \) (and consequent reduction in altitude-averaged Ri) from the standard case.

### b. Comparison with observations

Figure 8 shows a schematic of the unstable waves related to the aircraft flight path, with the parameters for the 16Z case. With this geometry, the expression for the wave period \( \tau \) as observed by the aircraft is

\[ \tau = \frac{\lambda}{c_a \sin \theta + c_w \cos \theta} \] (7)

<table>
<thead>
<tr>
<th>Case</th>
<th>Wavelength (m)</th>
<th>Propagation angle (deg)</th>
<th>Westward phase speed (m s(^{-1}))</th>
<th>Growth rate (per hour)</th>
<th>Aircraft period</th>
</tr>
</thead>
<tbody>
<tr>
<td>10Z</td>
<td>1775</td>
<td>276.0</td>
<td>16.44</td>
<td>2.53</td>
<td>47.6</td>
</tr>
<tr>
<td>16Z</td>
<td>1475</td>
<td>287.8</td>
<td>13.42</td>
<td>2.29</td>
<td>20.2</td>
</tr>
<tr>
<td>10Z</td>
<td>1690</td>
<td>275.5</td>
<td>15.51</td>
<td>5.03</td>
<td>48.8</td>
</tr>
<tr>
<td>(Constant ( N^2 ))</td>
<td>1690</td>
<td>275.5</td>
<td>15.51</td>
<td>5.03</td>
<td>48.8</td>
</tr>
<tr>
<td>10Z (Pure Tanh)</td>
<td>1680</td>
<td>275.2</td>
<td>15.42</td>
<td>4.95</td>
<td>50.2</td>
</tr>
<tr>
<td>Composite</td>
<td>1775</td>
<td>279.0</td>
<td>16.27</td>
<td>2.89</td>
<td>37.5</td>
</tr>
<tr>
<td>Composite</td>
<td>1775</td>
<td>287.6</td>
<td>16.07</td>
<td>2.19</td>
<td>23.4</td>
</tr>
</tbody>
</table>
where
\[ \begin{align*} 
\lambda & \quad \text{wavelength} \\
\alpha & \quad \text{southward aircraft speed} \\
\nu & \quad \text{westward wave speed} \\
\theta & \quad \theta_p - 270^\circ 
\end{align*} \]

The periods derived for the two cases are shown in Table 4. The agreement with the observed 23-sec period is marginal for 10Z and excellent for 16Z. This better agreement for 16Z seems to conflict with the conclusions in section 3.1, which indicated a correspondence between the 10Z temperatures and the portion of the 19.85 km flight leg with the wavlike oscillations.

The discrepancy can be resolved by considering two points. First, as shown by Fig. 8 and (7), the observed period is very sensitive to small changes in the propagation direction, which, as (6) indicates, is proportional to the meridional shear \( \tilde{\sigma} \), at the altitude of minimum Ri. In fact, the better agreement for 16Z is due in large part to the stronger meridional shear for that case. The time of the 19.85 km flight leg (15Z) is actually much closer to 16Z than 10Z. Hence, the shears at the flight time are probably stronger than those used in the stability analysis for 10Z. The fastest growing wave for a wind field consisting of the 10Z zonal wind and \( N^2 \), and the 16Z meridional winds, ("composite" in Table 4) has a wavenumber vector with a larger northward component, and somewhat better agreement with the aircraft observations.

The second point is that the growth rate is not highly sensitive to small changes in the direction of propagation. This is illustrated by the last line in Table 4, containing the parameter values for the fastest growing wave of the composite case, assuming a propagation angle large enough to yield agreement with the observations. The growth rate is over 75% of that for the fastest growing wave. In view of these two points, and the discrepancy between the observations and the "appropriate" basic state (due to wave feedbacks on that basic state), the stability analysis shows that the observed wave period is consistent with KH instability of the radiosonde profiles.

The amplitude structures of the perturbation vertical velocity and temperature fields for the 10Z and 16Z cases are shown in Fig. 9. The spread in the vertical velocity field is clearly sufficient to expect significant amplitude at the aircraft altitude of 19.85 km. The vertical decay in the perturbation temperature field is more rapid with a very narrow region of large relative amplitude. Nevertheless, excluding this large spike, the amplitude at the aircraft flight level is about 0.3 of the maximum temperature amplitude. Clearly, the calculated wave spreads sufficiently that one can expect it to be observable by the aircraft.

Establishing the character of possible turbulence associated with the unstable waves is more difficult. The only indication was the pilot's report of "rough flying" throughout the 19.85 km flight leg. He observed this in both halves of the leg, including the portion with no regular oscillations. Given the nominal constant altitude of the flight leg (as well as the absence of pressure

![Fig. 8. Pictorial depiction of wave fronts for the 16Z case relative to the aircraft flight path.](image)

![Fig. 9. Relative amplitudes of the perturbation vertical velocity (a) and temperature (b) for the calculated 10Z (solid) and 16Z (dashed) unstable waves.](image)
data), we cannot make an observational estimate of the vertical extent of the turbulence. However, it is unlikely that the unstable wave would generate turbulence everywhere that it has significant amplitude, which in this case is about 0.5 km. Such a thickness is far in excess of radar estimates of <200 m (Sato and Woodman, 1982). A 0.5 km turbulent layer thickness would also imply strong mixing over that depth and thus a quasi-adiabatic layer of comparable thickness centered near 19.5 km—not observed on either sounding.

It is more likely that the finite amplitude KH wave produces thin superadiabatic layers, which in turn generate the turbulence (Davis and Peltier, 1979). Their calculations of a finite amplitude KH wave in a hyperbolic tangent shear flow show superadiabatic layers (in the wave field) forming about one to two shear scale depths away from the center of the shear. In our case, the shear scale depths are $\sim 180$ m, suggesting that superadiabatic layers might form between 180 and 360 meters above the shear center at $\sim 19.5$ km. Though there are significant differences between their basic state and ours (their basic state, for example, has $\text{Ri} \sim 0.07$, vs $\text{Ri} \sim 0.22$ in our case), the nonlinear calculations do suggest that turbulence could be generated as a result of KH instability at the aircraft altitude.

5. Interpretation in terms of large-scale flows

Previous observations of stratospheric turbulence (e.g., Sato and Woodman, 1982) have shown that it frequently occurs in vertically thin (<200 m) layers, with horizontal scales of hundreds of kilometers. Moreover, these layers move downward with time, indicating that they are associated with large scale waves with downward phase propagation and upward energy propagation. In fact, Sato and Woodman have shown that enhanced turbulence, as measured by radar reflectivity, is directly related to the enhanced shear associated with upward propagating waves (see also, Maekawa et al., 1984). It is important to establish whether the strong shears generating the KH instability at 19.85 km are produced by a similar mechanism.

The wind profiles in Figs. 6 and 7 represent unusually large shears and low Ri for the two-week experimental period. That this is so can be seen from Fig. 10, which shows a time–height section of Ri for the entire two-week experimental period. The region from 19–20 km from 26 July through early on 28 July is clearly the most significant area of low Richardson numbers below 23 km. On 27 July the low Ri is due to the superposition of a strong negative shear in the time averaged zonal wind (Fig. 11), and a strong positive deviation in the

![Figure 10](https://example.com/fig10.png)

**Fig. 10.** Time–height section of the inverse of the Richardson number (Ri) at Panama during the NASA 1977 ITCZ experiment. The radiosonde data was filtered in time, removing periods less than 0.5 days prior to calculating Ri. The contour interval is 0.5, and values greater than 2.0 are stippled. The solid lines denote low Ri values associated with an upward propagating wave (see text).
zonal wind between 18.5 and 19.4 km (Fig. 12). This strong deviation is evident not only in the 10Z and 16Z soundings discussed earlier, but also in the 4Z and 22Z soundings of the same day. It is not a short period phenomenon. On the previous day, strong shears in the meridional wind (not shown) are responsible for the low Ri values.

We can obtain some idea of the mechanism for generating the positive bulge in the zonal wind deviation near 19 km by closer examination of Fig. 12. The positive and negative deviations in the zonal wind display an oscillatory character in both height and time, with predominant downward propagation throughout. This downward propagation has been noted by other investigators (Cadet and Teitelbaum, 1978; Maekawa et al., 1984). If the oscillations are assumed to be due to quasi-linear waves, this downward propagation is consistent with upward energy flux (Holton, 1975). Near 20 km, the character of the waves changes from generally short vertical wavelengths below that level, to longer vertical wavelengths above, up to about 24 km. Generally similar behavior can be seen in the meridional wind.

This picture can be explained by postulating that convective and/or synoptic activity in the troposphere excites a broad spectrum of waves, and that these waves

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**Fig. 11.** Time averaged zonal wind for the 1977 NASA ITCZ experiment 16–31 July 1977.

**Fig. 12.** Time–height section of the deviation of the zonal wind from the time average shown in Fig. 11. Contours are 2 m s⁻¹ and negative (easterly) deviations are stippled. See text for key to lines and symbols.
propagate upward into the stratosphere. As these waves propagate into the region of easterly shear between \(\sim 13.5\) and \(\sim 23.5\) km (Fig. 11), waves with westward phase speeds smaller than 32 m s\(^{-1}\) will reach critical levels at or below \(\sim 23.5\) km. For inertial–gravity waves that are smaller than planetary scale, this level will be where the absolute value of the doppler shifted frequency \(\hat{\omega} = \omega - k\bar{u}\) is equal to the local Coriolis parameter (Booker and Bretherton, 1967). For equatorial, planetary scale modes, this will be where \(\hat{\omega} = 0\). As the waves approach their critical levels, their vertical wavelength, \(\lambda_v\), decreases. For inertial–gravity waves on an \(f\)-plane, the expression is

\[
\lambda_v^2 = \frac{4\pi^2(\hat{\omega}^2 - f^2)}{N^2\kappa^2} \tag{8}
\]

with:
- \(k\) longitudinal wavenumber
- \(l\) latitudinal wavenumber
- \(f\) Coriolis parameter
- \(\kappa^2 = (k^2 + l^2)\)

At the critical level itself, they are largely absorbed or reflected (Booker and Bretherton, 1967). Hence, below the jet core, one expects strong variance in short vertical wavelengths, which is clearly evident in Fig. 12.

Near the jet core, the only upward propagating waves remaining have either high westward phase speeds, or eastward phase speeds. Above 20 km, for example, upward propagation requires that westward phase speeds must be greater than \(\sim 20\) m s\(^{-1}\) (for the equatorial wave modes—even larger for inertial gravity waves on an \(f\)-plane). Given that the amount of tropospheric wave energy at phase speeds greater than 20 m s\(^{-1}\) is relatively small, wave activity near the jet core should be dominated by eastward propagating waves. These have large doppler shifted phase speeds relative to the westward propagating waves absorbed at lower levels. As (8) indicates, their wavelengths would also be relatively large. This is consistent with the observations in Fig. 12.

One example of an upward propagating wave group with westward phase speed is highlighted by the solid and dotted lines in Fig. 12 near 19 km from 23 to 26 July. The solid and dotted lines label the extrema of the zonal wind deviation and the meridional wind, respectively. The behavior of these zonal wind deviations is well correlated with the meridional wind, which consistently leads the zonal wind by about \(\sim 90\) degrees. The period is about 2 days. (The relative phasing of the zonal and meridional wind does not indicate the sign of the phase speed; this phasing is the same for Rossby–gravity waves, and for inertial–gravity waves propagating in both directions.) The feature of this wave group that indicates the sign of the phase speed is the decrease of the vertical wavelength with altitude and the absence of the wave group above 20 km. Similar behavior for shorter period waves has been observed by Maekawa et al., (1984). The infrequency of the measurements (four times per day), as compared to the wave period (about 2 days), and the short duration of the wave makes it impossible to perform the spectral analyses required to establish firm numerical values. However, the qualitative behavior of the wave, and good correlation between the zonal and meridional wind, clearly shows it to be regular wave motion with a decreasing vertical wavelength with altitude.

An example of an eastward propagating wave in the jet core region above 20 km is highlighted by the dashed lines and large dots in Fig. 12 from 18–23 km from 18 to 22 July. The dashed lines and large dots label the extrema of the zonal wind deviation and temperature deviation, respectively. The lines of constant temperature phase are consistently parallel to those for the zonal wind deviation, with the temperature leading by \(\sim 135\) degrees. The period is somewhat less than two days. For inertia–gravity waves on an \(f\)-plane, the angle \(\phi\) by which the temperature perturbation (as observed at a fixed point on the ground) leads the zonal wind perturbation is

\[
\phi = -\arctan \left( \frac{k\hat{\omega}}{f} \right). \tag{9}
\]

(Here, we have assumed that the frequency \(\omega\) and the Doppler shifted frequency \(\hat{\omega}\) have the same sign.) For waves with purely zonal propagation, the temperature leads the zonal wind by 90 degrees for eastward Doppler shifted phase speed. With some positive meridional wavenumber, this phase shift is somewhat larger. Nevertheless, there must be some eastward component in the phase propagation for the temperature to lead the zonal wind (by less than 180 degrees). Temperature lagging the zonal wind indicates westward Doppler shifted phase speed.

As a wave approaches its critical level, and its vertical wavelength decreases, shear will increase, with particularly low Richardson numbers where the wave shear and the time-mean zonal wind shear have the same sign. This is apparent in Fig. 10, where the upward propagating wave group near 19 km on 23–26 July is in evidence. The region of low Ri propagating downward with time from late on 24 July at 20 km to early on 26 July at 19.8 km is due to this wave group; a region of larger, but still small Ri, is associated with another phase line of the same wave at 19.3 km.

It is more difficult to interpret the positive zonal wind perturbation near 19 km on 27 July (Fig. 12), and the associated low Ri near 19.5 km (Fig. 10), as a single, upward propagating wave. Connecting the positive zonal wind bulge at 19 km on 27 July with the positive zonal wind perturbation of the upward propagating wave group at \(\sim 20\) km on 25 July results in an inconsistency in the wave period, from \(\sim 2\) days
between 19 and 20 km on 23–26 July to ~3 days at ~19 km. It is more likely that the phenomenon results from the superposition of several upward propagating waves, one or more of which are approaching critical levels. Still, there is clearly wave motion involved, as evidenced by the strong peak in the meridional wind on 28 July which leads the positive zonal wind deviation on the 27th. The relevant positive and negative extrema in the meridional wind are denoted by a plus and minus, respectively, in Fig. 12.

The strong shear at 19.5 km on 27 July, and the corresponding strong positive deviation in the zonal wind, is also apparent in measurements made at other radiosonde stations in the region. Figure 13a and 13b show, respectively, time plots of zonal wind at 70 mb (18.7 km) and the difference between the 70 and 50 mb (20.7 km) zonal winds at three stations at latitudes comparable to Panama, including Panama itself. In all cases, there is a pronounced peak in the zonal wind at 12Z on the 27th. The wind differences between 70 and 50 mb are not as striking, but again, all three stations show significant peaks at either 12Z on the 27th or 00Z on the 28th. The fact that the phenomenon is nearly simultaneous at all three stations supports the hypothesis that the phenomenon is due to a superposition of waves rather than a single wave.

Data from stations at lower latitudes had insufficient time continuity for this type of analysis. Time plots in the zonal winds at higher latitudes (15–20 degrees) showed no similar features on the 27th, so the phenomenon is narrowly confined to low latitudes. In contrast, the zonal scale is quite large; (11°N, 299°E), for example, is nearly 2000 km from Panama. This disparity of zonal and meridional scales is consistent with the character of equatorially trapped waves (Holton, 1975).

6. Summary and conclusions

The goal of this observational study was the examination of two possible mechanisms for inducing small scale tracer fluxes in the tropical lower stratosphere: underlying cumulus convection and shear instability. Though data was inadequate to verify actual net tracer fluxes, both mechanisms are clearly in evidence during the 1977 NASA ITCZ experiment. Regarding the cumulus convection mechanism, we have established the presence of large amplitude disturbances with horizontal scales of 5–25 km on days with intense cumulus convection. On one flight, the disturbances were directly related to passage over convective cells. This, and the fact that disturbance amplitudes are substantially weaker on days with no convection, indicates a clear causal relationship.

Vertical parcel excursions due to cumulus-generated disturbances were calculated for the 17.00 km flight altitude on 27 July. The vertical excursions are 200–

![Figure 13](https://example.com/figure13.png)

**FIG. 13.** Time plots of zonal wind at 70 mb (a) and difference between 70 and 50 mb zonal winds (b) for the experimental period at three stations in the tropical, Northern Hemisphere Americas. Panama, denoted by the dotted lines, is at 9°N, 280°E.
300 m, not much smaller than the < 1 km vertical scales
of some large scale inertia–gravity waves. This indicates
that the cumulus-generated disturbances could have a
dynamical effect, presumably dissipative, on the larger
scale waves. However, the pilot’s report of smooth
flying at the altitude where these disturbances are ob-
served is also consistent with an ensemble of gravity
waves that do not induce mixing. Still, the waves may
break at other altitudes. If turbulent mixing at the ends
of the parcel excursions results from such wave break-
ing, the vertical excursions can be interpreted as mixing
lengths, and one could calculate the local eddy diffu-
sivity due to this phenomenon.

Strong evidence for shear instability was found at
the 19.85 km flight altitude on 27 July, in the form of
regular oscillations, whose amplitude indicate vertical
parcel excursions of 300–400 m. Rough flying condi-
tions were also observed, indicating possible turbulence.
Radiosonde wind fields nearest the flight time show
strong shears and low Richardson numbers at this alti-
itude. Stability analyses of these flows show that the
fastest growing Kelvin–Helmholtz modes have struc-
tures and periods consistent with the aircraft obser-
vations.

The strong shear responsible for the instability is
generated in part by westward propagating equatorial
waves, propagating upward into a region of small
doppler-shifted frequency associated with a strong
easterly jet. Data from other radiosonde stations in the
region indicate that the phenomenon is narrow in lati-
dude, but very extensive in longitude; there is evidence
for strong shears at the same altitude and time as much
as 2000 km east of Panama. The small vertical extent
and large horizontal extent of the shear region is con-
sistent with previous observations of the character
of stratospheric turbulence.

The clear next step in evaluating the importance of
small scale mixing generated either by underlying cu-
mulus or shear instability in this part of the atmosphere
are in situ measurements of tracer fluxes by aircraft.
For thin stratospheric turbulent layers generated by
shear instability, supplementary radar observations
would be useful in locating these layers for the aircraft.
With their good time and height continuity, the radar
observations would also allow the extrapolation of
global eddy diffusivities from aircraft case studies.

Of particular interest are the fluctuations associated
with underlying cumulus convection. Significant tracer
fluxes associated with these fluctuations would suggest
correlation of the strongest small scale mixing with
areas of the most intense convection, at least in the
lower stratosphere. Data from the 1980 Panama
Stratosphere–Troposphere exchange experiment indi-
cates strong cumulus-related fluctuations up to 1.5
km above the anvil tops (Margozzi, 1983). Globally,
the most intense convection is found in January over
intense fluctuations would be expected in this area,
with greater potential impact on diffusive stratospheric
tracer transport.

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