Maintenance of the Free-Tropospheric Tropical Water Vapor Distribution.  
Part I: Clear Regime Budget  

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ABSTRACT  

The water vapor budget of the free troposphere of the maritime Tropics is investigated using radiosonde observations, analyzed fields, and satellite observations, with particular attention paid to regions free of organized convection.  In these arid regions, time-average drying by subsidence must be balanced by moistening via horizontal advection from convective areas and via vertical turbulent transport from below.  It is found that for at least 25% of the maritime Tropics, 80% ± 10% of this source above 700 mb is by horizontal advection.  The remainder comes from vertical convective transport (scales < 250 km), with a pronounced local maximum at 500 mb.  The regions for which this is true are characterized by pentad outgoing longwave radiation > 270 W m⁻² and may be said to exist out of equilibrium with the surface as regards moisture.  Transport from below makes a significant contribution between 700 and 800 mb, despite the usual presence of an inversion below these levels, but is difficult to quantify accurately.  The convective transport convergence is estimated as a residual from large-scale budgets and directly from sounding time series by an independent method, which shows a narrow maximum at 500 mb.

Half of the paper addresses the question of data accuracy, including sounding and analyzed data, as it pertains to the question at hand.  It is concluded that the moisture budgets from the European Centre for Medium-Range Weather Forecasts (ECMWF) analyses are of useful accuracy despite some significant mean discrepancies between the analyses and sounding observations in convective areas.  The budget is found to be similar to that of a general circulation model based on the ECMWF forecasting model.  Humidity measurements from operational soundings appear responsive below 300 mb, but then abruptly become unresponsive.

1. Introduction  

This paper tries to answer the question, how does water get into subsiding portions of the free troposphere in the Tropics and subtropics?  These arid regions are subjected to the relentless loss of water vapor through subsidence; air that sinks into the boundary layer is inevitably moister than the air that converges in the upper troposphere, thus leading to a drying of the free troposphere.  In a steady state, this drying would have to be compensated by either upward mixing of vapor from the boundary layer or by horizontal advection from moister regions.  “Upward mixing” could be accomplished by penetrative, but isolated, convective clouds; these are distinguished from organized convection, which stands out in satellite photographs by producing cloud shields that trap large amounts of outgoing longwave radiation (OLR) and accounts for most of the atmospheric sensible heating (rainfall).  In this paper, the two sources will be compared with the subsidence drying in order to estimate their relative roles.

Most, but not all, of the subsiding regions under consideration here coincide with the traditional maritime “trades” regions: those between approximately 10° and 25° from the equator in either hemisphere (“warm pools” excluded), where surface trade winds blow easterly and toward the equator, and deep convection does not usually occur.  The trade wind regime over tropical oceans has long been known to play an important role in the global energy and water cycles (Riehl et al. 1951).  Study of these regions has been limited mainly to the boundary layer itself, which is viewed as a source of water vapor for convective regions (Riehl et al. 1951; Riehl and Malkus 1958).  Vertical transports in deep convective regions, which play an important role in the circulation, have received much attention, but such transports above the trade wind boundary layer have not.  Closures of the heat and moisture budgets within the trades boundary layer have, however, been attempted using Barbados Oceanographic and Meteorological Experiment data (Holland and Rasmusson 1973; Nitta and Esbensen 1974) and Atlantic Trade Wind Experiment data (Augstein et al. 1973).  Poor data quality in dry conditions, and perhaps a perceived lack of importance to the circulation, have presumably protected the trade wind free troposphere budgets themselves from closer scrutiny.
Trade inversions usually occur between 750 and 900 mb (e.g., Kloesel and Albrecht 1989). An example of a strong inversion is shown in Fig. 1. (This sounding will be revisited in the appendix, where sounding calibration errors are discussed.) It is often supposed that, in trades-type regions, inversions above the boundary layer such as this one prevent any vertical transport of water, heat, or momentum out of the boundary layer (see Augstein et al. 1973). This would leave horizontal advection as the only source for the trade wind free troposphere. However, Holland and Rasmusson (1973), based on a similar study in the same part of the Atlantic, found that eddy transport convergence of water vapor was a dominant term in the budget as high as 650 mb. The gradient between water vapor concentrations near the surface and amounts above this inversion is quite large, so that merely occasional erosions of the inversion might be adequate to admit enough water into the free troposphere to balance subsidence drying in the time-averaged sense. Such boundary layer disappearances might occur at the crests of passing gravity waves—which could lift or destroy the inversion—or as a result of intense surface fluxes of moist energy resulting from an ocean hot spot or burst of surface wind.

While previous studies have been based on intensive local observation programs, this work employs sounding, satellite, and European Centre for Medium-Range Weather Forecasts (ECMWF) analyzed field data over the entire Tropics to try to quantify the two sources in a climatic sense. Outgoing longwave radiation measurements from the National Centers for Environmental Prediction (NCEP, formerly the National Meteorological Center) will be exploited as a coordinate to establish different climate regimes. A general circulation model and alternate analysis product will each be examined briefly to aid in interpreting the results. Water vapor measurements and global atmospheric analyses each suffer from a number of inadequacies, so these will be dealt with extensively. Section 2 gives an overview of the relevant information, and section 3 gives budget results from the ECMWF analyses. Sounding data are introduced in section 4 and are used to estimate biases in the ECMWF analyses. The ECHAM3 GCM and Navy Operational Global Atmospheric Prediction System (NOGAPS) analyses are also examined briefly in this section to gain insight into the behavior of the ECMWF forecast model. Finally, local physical mechanisms for explaining the budget residuals in the analyses are discussed and quantified in section 5, and con-
clusions are discussed in section 6. Further information on sounding difficulties is given in an appendix.

2. Overview of data and methodology

We begin with a brief look at the available datasets during the period from 7 March to 7 April 1993. This period coincides with the Central Equatorial Pacific Experiment (CEPEX) (Williams 1993) and is an advantageous period of study because some additional measuring platforms were available at this time. It will be referred to throughout this paper as the “CEPEX period.” The network of sounding stations available in the tropical Pacific (20°S to 25°N) appears in Fig. 2a. At each station, an arrow indicates the direction of the 32-day mean “mass corrected” water vapor flux $u(q - q)$ at 500 mb (explained in section 2a), where $q$ is averaged over all the stations. Its convergence is an estimate, unaffected by mass budget errors, of the net convergence of water vapor due only to horizontal motions. The observed OLR is also shown to indicate cloud cover during the month (see section 2b). The observations show clearly that water is leaving the convective cloudy area over the warm oceans of the western Pacific and heading toward drier areas.

However, the paucity of stations and lack of coherence between relatively nearby flux vectors portend difficulty in obtaining the desired budget terms directly from observations. Furthermore, the stations tend to be clustered in the convective areas rather than the subsiding areas, which are under investigation here. Previous attempts to estimate advective moistening of atmospheric columns directly from radiosonde observations have met with horizontal sampling problems (Mcbride et al. 1989; Connolley and King 1993). These authors used denser sounding arrays than those available here.

For this reason, this study will rely principally upon upper-air analyses prepared by the ECMWF. Their method, and possible shortcomings of the result, will be discussed in section 4a. A special analysis set has been obtained at 1.1° and four-times-daily resolution for a limited part of the Tropics during the CEPEX period, which is used in Fig. 2. Standard 2.5° twice-daily resolution data will be used thereafter, however, due to their full coverage of the Tropics and their lesser computational burden. Both analyses are uninitialized.

Figures 2b,c show maps of moistening by horizontal advection $AH$ (defined below) and of vertical velocity $\omega$ from these analyses averaged over the entire 32-day period at 500 mb. Patches of strong analyzed rising motion appear concentrated within the areas of low OLR, with sinking motion in most of the trades areas. The low OLR regions are also high in humidity (not shown), while the clear portions are dry. As indicated by the sounding stations, water vapor is diverging horizontally from the convective regions, converging within the equatorial cold tongue east of the date line and the belt north of the ITCZ. We may conclude from this brief look that the analyzed water transport is consistent with that from soundings and that the analyses generally support the characterization of the convective and arid regimes that has been given, with the added characteristic that arid regions receive water vapor horizontally from convective ones (at least at 500 mb).

a. Water vapor budget decomposition

We wish to distinguish specifically between vertical and horizontal, or large- and small-scale, processes. To make this precise, we begin with

$$\frac{\partial q}{\partial t} = -\nabla \cdot q \mathbf{u} + Q,$$

where $Q$ is the net evaporation of liquid or ice (if any). If this equation is averaged over a finite volume having isobaric top and bottom surfaces of area $A$ and pressure thickness $h$, the divergence term can be expanded as

$$\nabla \cdot q \mathbf{u} = \frac{1}{A} \int \left( q \mathbf{u}_2 \cdot \mathbf{n}_2 \right) dl + \bar{q} \frac{\partial \omega}{\partial p} + \bar{\omega} \frac{\partial q}{\partial p} + \frac{1}{Ah} \int q_1' \omega_1' - q_2' \omega_2' dA,$$

where $\mathbf{n}_2$ is the outward normal unit vector from the sides of the volume, and $q$, $q_i$, and $\omega_i$ are the humidity and pressure velocity on the top ($i = 1$) and bottom ($i = 2$) surfaces. These are decomposed into averages over the volume, $\bar{\omega}$ and $\bar{q}$, a difference between the top and bottom average $h \partial \omega / \partial p$, and deviations from these averages, $\omega_i'$ and $q_i'$. The continuity equation

$$\frac{\partial \omega}{\partial p} = \int \mathbf{u}_2 \cdot \mathbf{n}_2 dl$$

can then be used to reduce this to

$$\nabla \cdot q \mathbf{u} = \frac{1}{A} \int \left( q - \bar{q} \right) \mathbf{u}_2 \cdot \mathbf{n}_2 dl$$

$$- \bar{\omega} \frac{\partial q}{\partial p} + \frac{1}{Ah} \int q_1' \omega_1' - q_2' \omega_2' dA \quad (1)$$

$$= -AH - AV - DV, \quad (2)$$

where $AH$ is the moistening by horizontal advection (as shown in Fig. 2), $AV$ the moistening by vertical advection, and $DV$ the convergence of transport across isobars by unresolved vertical eddies or plumes. Each of these quantities is a four-dimensional function of space and time. We will compute $AH$ and $AV$ directly from the ECMWF analyses and infer $DV + Q$ as a residual for two months: the CEPEX period (7 March to 7 April) and January 1993 during Tropical Oceans Global Atmosphere Coupled Ocean–Atmosphere Response Experiment (TOGA)
Fig. 2. West and central Pacific at 500 mb, mean over CEPEX period. (a) Map showing the locations of upper-air sounding stations in the region of interest. Arrows indicate the mass corrected moisture flux $u(q - \bar{q})$ at 500 mb for stations reporting winds. Arrows will tend to point upwind in dry areas and downwind in moist areas. OLR contours are also shown, contour interval is 20 W m$^{-2}$, and areas below 240 W m$^{-2}$ are shaded. (b) ECMWF analyzed vertical pressure velocity at 500 mb. Rising areas are shaded and sinking areas are clear, with dense shading for rising of more than $-50$ mb day$^{-1}$. The maximum rising is $-300$ mb day$^{-1}$. (c) Shows 500-mb AH as defined in the text from 1.1° ECMWF analyses. Areas of flux divergence are shaded, with dense shading below $AH = -0.4$ g kg$^{-1}$ day$^{-1}$. Minimum AH is $-1.0$ g kg$^{-1}$ day$^{-1}$. Data in all panels are means over the CEPEX period.

COARE). Evaporation and condensation processes may make some contribution to the source term $Q$, particularly in the upper troposphere, since even the most arid regions of the Tropics are frequented not only by boundary layer clouds but many other types of scattered or thin clouds (Warren et al. 1988). No effort will be made in this study to estimate $Q$ directly. In a later section, however, a method is intro-
duced whereby DV can be estimated directly from sounding time series.

Here is a brief description of the numerical details of the budget term computation. According to (1), AH is computed from the gridded analyses within boxes formed by four grid points. This calculation is conservative over large areas (interpolation errors will cancel in adjacent cells) and has the virtue of being automatically corrected for mass flux imbalances that may be present in the analyzed winds due to interpolation effects (Trenberth 1991). By centered differencing of $\log q$ in $\log p$, $\partial q/\partial p$, necessary for computing the sub-sidence drying, has been calculated at each grid point and then averaged over each cell by neighborhood averaging. Unlike the horizontal scheme, this point formulation is not conservative, but was preferred over linear methods in light of the exponential dependence of vapor on height. The analyses themselves are interpolated from spectral coordinates in a pointwise representative, rather than conservative, manner (Trenberth 1991). Results have been computed using $1.1^\circ$, four-times-daily and $2.5^\circ$, twice-daily analyses as a numerical check, with virtually identical results when averaged as maps or versus OLR (as described in the following section). This reassures us that the resolution problems found by Chen et al. (1996) do not affect the results here.

b. Use of pentad outgoing longwave radiation as a tropical climate coordinate

In this study, observed OLR will be used to establish climate regimes. That is, OLR—rather than longitude or latitude—will serve as the independent variable for exploring structure in other fields. Investigators of the tropical atmosphere commonly use this observable as an indicator of cloud cover, convection, and even rainfall (Arkin and Ardanuy 1989). The NCEP product used here, designated OLR_{pent}, is available on a $2.5^\circ$ grid averaged over 5-day periods. NCEP's 5-day averaging period is conveniently close to the water vapor recycling time of 4–7 days, which is obtained below. For further information on this OLR product see Gruber et al. (1994).

3. Results from ECMWF analyses

The budget terms outlined above have been calculated at each ECMWF grid location, level, and time. We have matched each location and time with the collocated OLR_{pent} observation and then found the mean of each budget component over points within each 5 W m$^{-2}$ OLR_{pent} bin. The behavior of these budget terms with changing OLR during the CEPEX period is indicated in Fig. 3 for the entire Tropics (25°N to 25°S), ocean regions only, using the $2.5^\circ$ analyses. As indicated in section 2, heavily convective regimes (low OLR) are associated with horizontal divergence of wa-

ter vapor (AH negative) but much vertical moistening, while inactive regions (high OLR) experience convergence and vertical drying. The discrepancy between them—accounting for temporal changes—is shown as an "apparent sink" in the figure. The net loss at low OLR is the effect of condensation and precipitation, while the small net gain (negative apparent sink) at high OLR is evidence for a positive contribution from other sources.

Since the budget terms level off above 270 W m$^{-2}$, OLR_{pent} values at or above 270 W m$^{-2}$ will be used from now on to define the "clear" regime whose budget will be examined further. Its spatial distribution during the CEPEX period is pictured approximately in Fig. 4. It tends to lie in the trades regions, covering about 25% of the maritime Tropics, as also seen from the position of 270 along the x axis in Fig. 3. The equatorial cold tongue of the central Pacific belongs to the regime during the first two pentads of the period, but mostly it becomes cloudy thereafter and does not appear clear in the figure. The size of the clear regime may seem surprisingly small in light of the common assumption that the tropical atmosphere is sinking everywhere except within a tiny, actively convecting fraction. However, the fraction of the Tropics that is sinking is not a single number but an increasing function of the resolution with which the field is measured. At 5-day and $2.5^\circ$ resolution, the fraction is less than half, according to ECMWF (Fig. 3). Furthermore, many sinking areas are excluded from the clear regime here by the OLR threshold.

![Fig. 3. ECMWF 500-mb moisture budget vs OLR_{pent}, CEPEX period. Variation of AH (solid line) and AV (dashed line) with OLR_{pent} during the CEPEX period. Horizontal moistening is AH; AV is vertical moistening, with negative values indicating drying by subsidence. Dot-dashed line is the apparent sink AV + AH - $\partial q/\partial t$ or $-\Delta V - Q$. Areas over ocean from 25°S to 25°N are included. The horizontal axis is scaled so that distance is proportional to frequency of occurrence. Variables are computed from ECMWF 2.5° analyses, as described in text.](image-url)
The budget terms have been averaged over this regime at each available height, shown in Fig. 5 divided by the mean $q(p)$ itself to give a fraction per day. The apparent source decreases with altitude from just over 50% of the subsidence drying at 700 mb to nearly zero at 300 and 400 mb during the CEPEX period. The computation has been cut off at 300 mb because this is the highest level at which the analyses are reliable (see appendix). The “corrected” budget terms appearing in the figure are based on a crude method of adjusting the results to agree with sounding data. This method will be explained in section 4a, where the analyses are tested against soundings. There are two important results here: first, the subsidence rate is strong enough so that the e-folding drying time due to subsidence, in the absence of a balancing source, is between 4 and 7 days; second, the apparent source is small in the clear regime compared to horizontal advection, except at the lowest layer (700 mb). This “apparent source” could be due to local vertical transports of vapor, due to phase change processes, or due to analysis errors. This calculation was repeated using only the Indo–Pacific region (45°E–140°W), with practically the same result despite the existence of substantially more convection within this subset of the Tropics. This Indo–Pacific computation was also repeated with the 1.1° analyses, with no significant change.

Since the OLR threshold is somewhat arbitrary, the result has also been computed with other OLR thresholds and is shown in Fig. 6. Consistent with Fig. 3, the
importance of the apparent source at 500 and 700 mb shrinks moderately as the threshold is made more stringent (higher OLR). Also, to test the generality of the results, January 1993 has been examined, with the result given in Fig. 7. The slightly bigger apparent source here is a result of stronger subsidence in the clear regime, as shown later in section 4b.

4. Observational concerns

The analyses that led to the above results are a blend of model forecasts and observations (primarily from soundings), each of which may suffer from biases. In this section the soundness of the above results will be addressed. Island station sounding data will be used to evaluate systematic errors in the analyses. (They will also be used in the next section to estimate the convective moisture source.) The stations spread across the western and central equatorial Pacific, as indicated in Fig. 2. In the appendix, these data and their shortcomings are discussed in some detail. Two new results that stem from the evaluation of the sounding data are 1) that the two instrument types used (VIZ and Vaisala) both show responsive behavior up to 300 mb in the Tropics, but no higher, and 2) that 10% RH appears to be an unbiased estimate of the true humidity of a "dry" (less than 20% RH) sounding observation, as reported for VIZ soundings prior to October 1993.

a. Evaluation of ECMWF moisture analyses

Moisture fields and vertical velocities are the most poorly known analysis products, differing widely between different forecasting centers (e.g., ECMWF and NCEP) and relying heavily on model forecasts (Trenberth and Olson 1988; Tiedtke et al. 1988; Kasahara and Mizzi 1992). Unfortunately, the poor observational sampling of trade wind areas renders the analyses particularly suspect there. However, even large daily errors in the analyses are acceptable for the present purpose as long as they are not systematic.

A simple indication of systematic moisture errors in the forecast model is the difference between the mean analyzed and observed humidities at the station locations. This difference has been calculated for two OLR regimes and appears in Fig. 8. The first OLR regime is
the clear regime again (OLR > 270 W m\(^{-2}\)), and the other is a convective regime chosen to include a similar number of soundings (OLR < 210 W m\(^{-2}\)). The analyses are fairly accurate in the clear regime, but show a marked departure from the observations in cloudy areas, which is unquestionably larger than the nominal 5% observational error. This magnitude of error is quite surprising considering that the analyzed values are near stations, where they should be accurate even if the forecasts are poor. Based on comparisons between analyses and forecasts (not shown), this would appear to be a result of deliberate data quality control procedures at ECMWF and low confidence generally in the sounding humidities at 500 mb and above. Previous efforts have also concluded that the moisture analyses are mostly forecast based (Tiedtke et al. 1988). As to the forecasts themselves, Mohanty et al. (1995) show that the forecasting model dries out the tropical midtroposphere excessively during long forecasts, attributing the problem to an overzealous convective scheme. This would be consistent with the present results. Sherwood (1995) combined satellite, sounding, and ECMWF data, finding that the discrepancy in Fig. 8 appears to be due to analysis errors rather than sounding biases. This information will appear in a future article.

Since the bias is present only in the convective areas, the horizontal transports will have been underestimated if the winds are accurate. Vertical drying may also be underestimated since the analysis vertical gradients in the clear regime are somewhat low. If we "correct" the analyzed budgets by rescaling \(AH\) proportional to the humidity difference between these regimes and \(\nabla q/\nabla p\), then the budget terms change as indicated in Fig. 5. This is not to imply that the corrected values are exactly right, but merely to gauge the uncertainty in the budget terms suggested by the level of error in the analyses. The cancellation of the terms in the residual is fortuitous, and the residual should still be considered uncertain to at least 0.02 day\(^{-1}\).

\textbf{b. ECMWF clear-regime vertical velocity test}

It has long been recognized that sinking of air into the atmospheric boundary layer must accompany the radiative cooling of the troposphere in undisturbed trade regions for energy balance, while rising motion occurs in the convective zones (Riehl et al. 1951; Riehl and Malkus 1958). This clear-sky energy budget constraint may be used to test the ECMWF \(\omega\) estimates there. Restricting our attention to the clear regime, the budget is

\[
\frac{\partial T}{\partial t} = -\mathbf{u} \cdot \nabla T - \omega \frac{T \partial \theta}{\theta \partial p} + R. \tag{3}
\]

The horizontal advection, \(\mathbf{u} \cdot \nabla T\), has been calculated directly from the analyzed fields and makes a small contribution. The vertical temperature structure has been obtained from a subset of the tropical island pentad-mean soundings, whose collocated OLR is greater than 265 W m\(^{-2}\) (the threshold was lowered slightly from 270 in order to include more soundings). The radiative heating \(R\) has been estimated using the LOWTRAN and CCM2 radiation codes for longwave and shortwave heating, respectively. The uncertainty in the radiative transfer calculated by this method is 0.1 K day\(^{-1}\); the column average longwave error should be no more than 0.05 K day\(^{-1}\) (Ellingson et al. 1991). Water vapor content above 300 mb was taken from the average of a set of airborne frost-point hygrometer soundings made during CEPEX (Kley et al. 1995). The overall uncertainty due to sounding errors is similar to the algorithmic uncertainty.

Figure 9 shows the full \(\omega\) profile, averaged over all clear-regime cases, for each of 4 months: the CEPEX period, January 1993, October 1992, and July 1992. Also shown is the sinking rate estimated by solving (3) for \(\omega\). The figure indicates that the analyzed subsidence is weaker than that expected from energy balance by 10%–30% depending on the month and level, except at 200 mb where the shortfall is more than 50%. The CEPEX month shortfall, indicated by the thick contour, is relatively small. The weakest subsidence shown is for July, and the strongest is for January. It is possible that some of this seasonality might be real, but it is more plausible that the sampling of subsident regions by the network varies seasonally due to shifting convective patterns. The equilibrium \(\omega\) estimate shown is from soundings during the CEPEX month, but it should not vary appreciably for the other months.

Another analysis has been examined as an additional check on ECMWF, the NOGAPS analysis (Hogan and

\begin{figure}[h]
\centering
\includegraphics[width=\textwidth]{figure9.png}
\caption{Clear-regime \(\omega\) analyses vs radiative equilibrium. Vertical velocity profiles from ECMWF CEPEX period (thick line); ECMWF July 1992, October 1992, and January 1993 (thin lines, left to right); and NOGAPS CEPEX period (thick dashed line) averaged over all ocean locations and pentads having OLR > 270 W m\(^{-2}\) from 25°S–25°N. The dot–dashed line indicates the possible impact of thin cirrus having an impact of 10 W m\(^{-2}\) on OLR.}
\end{figure}
Brody 1993). The vertical velocity from NOGAPS during the CEPEX period is shown by a thick dashed line and does not differ much from ECMWF, except perhaps at 300 mb. The similarity between the results indicates that the result is not highly analysis dependent.

Two additional sources of energy may account for the modest shortfall. One is radiative heating due to small cloud amounts, which are ubiquitous in the Tropics (Warren et al. 1988), contributing on the order of 10 W m$^{-2}$ to longwave forcing (Ramanathan et al. 1989). Another is additional clear-sky solar absorption; Ramanathan et al. (1995) found that a shortwave transfer model having on the order of 10 W m$^{-2}$ more absorption than CCM2 was necessary in order to match surface observations (Waliser et al. 1996). The dot-dashed line in Fig. 9 shows the effect of distributing either of these additional sources between 200 and 400 mb, an arbitrary distribution that might show the effect of heating by thin cirrus. Clearly, some subsidence shortfall remains for at least 3 out of the 4 months, but does not appear large, at least in the middle troposphere.

The levels where disagreement seems significant compared with the uncertainties in true radiative cooling, even in ECMWF's "good" months, are from about 150 to 200 mb and from 700 to 850 mb. The discrepancy near the tropopause may result either from dynamic top-of-atmosphere boundary conditions used in the analysis procedure, or from radiative heating by high-altitude thin cirrus. A reasonable hypothesis on the discrepancy at 700 mb is that occasional weakening and elevation of the trade inversion due to transient dynamic events allows small amounts of moist boundary-layer air to reach above the average inversion height, and that these then make a significant contribution to the energy budget as high as 600 mb through condensation. This would account for both the energy and moisture deficits that appear at this level in Figs. 5, 7, and 9. It is also possible that both analyses are seriously underestimating the sinking rate here for some reason.

c. Comparison with ECHAM3 general circulation model budget

Since the analyses seem to rely to a significant degree on forecasts, it makes some sense to have a look at what the forecasting model would do on its own. For this purpose, the ECHAM3 general circulation model has been examined. This model is based on the forecast model used at ECMWF, but with improvements to certain physical packages [see Roeckner et al. (1992) for further details on the model]. The numerical framework and cumulus schemes (presumed to be of greatest importance here) are unchanged.

Figure 5 has been duplicated with output from ECHAM3, resulting in Fig. 10. The budget terms are fairly similar from 300 to 500 mb, which is reassuring as to the model's fidelity, but may just reflect a lack of observation impact in the analyses. The apparent moisture source at 700 mb has grown, suggesting that perhaps it is exaggerated in the forecasts and (to some unknown degree) in the analyses. There is also a second maximum in the upper troposphere, above the level where observations are reliable. This is probably due to cloud evaporation processes in the model, but the source has not been verified.

An examination of the tendencies in arid regions, supplied from a different run of the model by U. Lohmann (1995, personal communication), indicates that the cumulus scheme is responsible for most of the source at 700 mb. This does not necessarily mean that the source is realistic. Navarra et al. (1994) have suggested that truncation errors in spectral GCMs will cause spurious convective moisture sources to appear in dry regions and also exaggerate local convective activity in stable oceanic areas, due to Gibbs phenomena in the orography. However, it is at least encouraging that the residual is mostly a physical effect rather than numerical diffusion (although the latter probably makes some contribution). The budget residuals in both ECHAM3 and the analyses are not confined to particular portions of the clear regime, but occur widely. They are particularly pronounced over the North Atlantic and are weakest over the Indian Ocean.

A final concern would be the fixer, a part of the advection calculation that redistributes water vapor after it has been advected by the standard dynamics in order to prevent negative humidity and enforce global conservation. This is an additional, unphysical source. In 1992, the ECMWF operational forecasting model was equipped with semi-Lagrangian transport (Ritchie et al. 1995), in which the fixer source is only a few percent of the total source or advective tendency (Williamson

![ECHAM tropical q budget, OLR > 275](image)
and Rasch 1994). The ECHAM3 model uses the older, spectral transport scheme, but does not fill negative values (which occur with significant frequency in the middle and upper troposphere). Thus, the fixer does not contribute significantly to the apparent source in either the analyses or the GCM.

5. Mechanisms for local vapor sources

Now that this digression into the behavior of analyzed fields and global models is over, we pause and assess where we stand. The examination of the analyzed budgets indicated that most clear-regime vapor in the maritime free troposphere arrives horizontally—but it also found a significant “apparent source” that must be explained some other way, most of it at 700 mb. Four possibilities have been identified that may account for it:

1) convergence of vertical correlation flux of water DV in isolated cumulus elements rising from the boundary layer (see section 2a);
2) evaporation of horizontally convergent advection of condensed water;
3) biases in the analyses due to model physics (either algorithmic or spectral truncation problems are possible); or
4) errors in the analyses due to numerical diffusion in the forecast model, or numerical problems in recreating the budget terms from the gridded data.

Since the apparent source in ECHAM3 can be traced to model physics (shallow convection) and was similar to that of ECMWF, we may conclude that numerical diffusion (item 4) is not a large contributor to the apparent source in either budget. Errors in model convection would be included in item 3, but if model cumulus effects are realistic they would correspond to item 1.

A convective dry bias in the analyzed moisture has been identified and traced to the forecast model, but a crude method of accounting for the moisture biases did not explain away the apparent source, except possibly at 300 and 400 mb. Biases in θ probably also exist, but are uncertain, due to uncertainties in cloud forcing and solar absorption. At worst they are 10%–20% during the 2 months for which moisture budgets have been calculated and would tend to make the apparent source somewhat larger if factored in. Thus, we conclude that item 3 above is not an adequate explanation of the apparent source, although it has the potential to make a significant contribution either way.

Estimates of turbulent transport from sounding time series

This reasoning implicates physical processes (items 1 and 2) as modest sources of free tropospheric water in subsident regions. In this section, soundings are used independently in an attempt to estimate DV directly by exploiting the time-lagged covariance between moisture levels and stability in time series from individual stations within the clear regime.

Previously, it was asserted that occasional disappearance of the capping inversion, or lifting of the inversion to a greater height, would be a possible mechanism for water transport from the boundary layer to the free troposphere. Although the inversion is usually around 800–850 mb, it often lies at other altitudes (Fig. 11). In this figure, boundary layer heights are shown from three different sources: six islands at which clear conditions persist for long periods (Willis, Wake, Hilo, Lihue, Guam, and Canton), ship-launched upsonde and dropsonde data from the CEPEX experiment, and dropsonde results published previously for the east Pacific by Kloesel and Albrecht (1989). In the histogram, only “inversion” soundings are included, according to the identification scheme of Kloesel and Albrecht (1989). Soundings where surface air never becomes buoyant (“stable” soundings, 40% of the total) are also not included. Boundary layer heights have been identified by finding the absolute minimum of the function p – p*, as described by Betts and Albrecht (1987), which was found to be the most reliable among the three methods they describe. The histogram shows that

![Fig. 11](image_url)
inversions form as high as 700 mb with significant frequency. In addition, 19% of the soundings were “unstable,” such that transport could conceivably occur to at least 600 mb. Some differences between the histograms from the different platforms will be discussed later.

Suppose we want to estimate the moistening at a level \( p \) from below due to stability-dependent turbulence. During episodes where the inversion disappears or rises above \( p \), transport of water to \( p \) from below will become possible. The episodes will be called “turbulent episodes,” although turbulence has not actually been measured—its possibility has only been inferred from stability. Elevated moisture levels during and after such an episode will indicate net moistening. We have used soundings from the six stations listed above, keeping only those colocated with pentad OLR values greater than 270 W m\(^{-2}\), and have assigned a boundary layer top height to each. As mentioned previously, the soundings may be classified as stable, inversion, or unstable. In order to assign an “inversion height” to unstable soundings, parcels from 960 mb were lifted reverse adiabatically (including condensed water loading) until reaching their level of neutral buoyancy, which was identified as the inversion height for that sounding.

A sample time series of this is given in Fig. 12 for events at 700 mb. To obtain these data, turbulent onset events were found in which only the third sounding of a three-sounding OLR\(_{pent} > 270\) W m\(^{-2}\) series from a given station lacked an underlying inversion. Likewise, turbulent cessation events must be an inversion-free sounding followed by at least two inversion soundings, all with OLR\(_{pent} > 270\) W m\(^{-2}\). The turbulent (inversion free) episodes lasted from one to four soundings. Thus, each point at −0.5 in Fig. 12 is the last sounding in a time series before a turbulent episode began, and each one at +0.5 is the first one after an episode ended, regardless of the length of the episode. These stringent requirements result in a small number of documented episodes, despite the large number of soundings to start with.

The composite time series shows significantly elevated moisture levels during the turbulent episodes and perhaps slightly elevated moisture levels afterward. A schematic picture of a composite time series at a given altitude is offered in Fig. 13. The moisture is low before the turbulent episode \( q_1 \) then rises to a higher value, which averages \( \bar{q} \) during the episode. The episode ends after \( T_c \) days, after which the humidity falls to \( q_2 \), then averages \( \bar{q} \) until another episode begins \( T_s \) days later. Of course, there is no periodicity to this process, and the times are averages over many events. We find \( T_c \) and \( T_s \) by

\[
T_c + T_s = \frac{\text{Total time}}{\text{Number of episodes}}
\]

\[
\frac{T_c}{T_s} = \frac{\# \text{ Turbulent soundings}}{\# \text{ Non-turbulent soundings}}
\]

At 700 mb, \( T_c = 1 \) day and \( T_s = 10 \) days. The values for \( T_c \), \( T_s \), \( \bar{q} \), and \( \bar{q} \) are estimated by considering all clear-regime turbulent episodes and not just those meeting the other criteria for inclusion in Fig. 12; values for \( q_1 \) and \( q_2 \), however, are determined only using series fragments that meet the additional criteria, about half the total.

How do we determine the net moistening associated with this process? One possibility is to assume that all moisture appearing during turbulence is advected away horizontally by the time the stabilizing inversion has
returned. In this case, the time-averaged net moistening can be estimated roughly as
\[ \bar{C} = \frac{\bar{q}_c - \bar{q}_s}{T_s}, \]
(4)
where all variables are functions of pressure. This will be designated method 1. The horizontal advection assumption behind this estimate is clearly incorrect, however, in cases where an inversion is temporarily lifted above \( p \) at the crest of a passing wave and then lowered again. In this case, all of the extra humidity present during the event is returned to the air below, and the net moistening is zero. A second method, therefore, is to assume local behavior thusly:
\[ \bar{q} = A + C, \]
(5)
where \( q \) is the humidity, \( A \) is net drying by all other processes, and \( C \) is equal to \( c_0 \) during turbulent episodes and zero in between. We model \( A \) as a relaxation process
\[ A = a_0 - a_1 q \]
(6)
and will set \( a_0 = 0 \). We average (5) over turbulent episodes, and then between them, to produce a two-equation system. This can be solved for \( c_0 \), with a result for \( \bar{C} \) of
\[
\bar{C} = \frac{c_0 T_c}{T_s + T_c} \left[ \frac{\bar{q}_c + \bar{q}_s T_s}{T_s(T_s + T_c)} + \frac{a_0 T_c}{T_s + T_c} \left( \frac{\bar{q}_c}{\bar{q}_s} - 1 \right) \right].
\]
(7)
This will be designated method 2 for \( a_0 = 0 \). The contribution to the moisture budget from local sources has been computed by both methods and is shown in Fig. 14, along with the residual estimates from the ECMWF and ECHAM3 budgets. The two sounding methods each agree, well within uncertainties, with the budget estimates at 500 and 400 mb. The sounding methods indicate a local maximum in the moistening at 500 mb, with little contribution between this level and the 700-mb level. At the 700-mb level itself, the sounding methods become wildly different, indicating that large excursions in the humidity occur during turbulent episodes (\( \bar{q}_c > \bar{q}_s \)) but that humidity returns to about the same level afterward (\( q_2 \not= q_1 \)). This probably reflects vertical displacement of the inversion by waves, so that method 1 is overestimating the transport. There is clearly some overestimate since the method 1 fraction goes above 100% at 800 mb. Nevertheless, the overall agreement of the various estimates is quite encouraging, and the interesting vertical structure of the moistening was somewhat of a surprise. This structure, in particular the maximum at 500 mb, reflects fluctuations in the amount of RH variance at different levels, which is correlated with stability.

The estimated DV profiles have been obtained using simplistic theories and should be regarded with some caution. The uncertainty in the method 2 estimate due to sample size is shown and is comparable to or larger than the uncertainty in method 1. If additional months of data were used, the uncertainties could be reduced, but this did not seem worthwhile since they are already smaller than the difference between the two methods. Nonetheless, the general picture emerges that DV is a likely explanation of the remaining apparent source, which was obtained as a residual from the analyzed fields.

A source of concern is the ability of island stations to represent the marine boundary layer. One very large island station (Hilo) has been used, but this station is on the windward coast of Hawaii within a fairly steady trade wind flow. Another island (Guam) is about 200 square miles, and the remaining four are reefs. The comparison of Fig. 11 shows that the island boundary layers appear somewhat thicker than those from genuine ocean measurements, showing particularly a lack of low inversions. Thus, this sounding dataset would tend, if anything, to overestimate DV. Figure 11b shows that if the ship and dropsonde measurements are restricted to clear-regime conditions, their height distribution looks more like that from the islands. The data from Kloesel and Albrecht (1989) were not restricted by cloud cover, but were restricted to exclude a small number of soundings having multiple inversions, which was not attempted here. The omitted soundings reportedly tended to include radiation inversions near the sur-
face, together with inversions at the higher end of the distribution.

To summarize, the estimates of DV obtained directly from soundings may be questioned, but seem to be consistent with the apparent source in the large-scale budgets. They also indicate that this source is relatively smaller between the analysis levels than at those levels.

6. Summary

Several conclusions can be drawn about the vapor budget in arid regions above the boundary layer. First, the sources of moisture to balance subsidence in the clear regime, defined by $OLR_{pent} > 270 \text{ W m}^{-2}$, are mostly nonlocal within the measurable free troposphere, averaging about 20% local above 700 mb with an absolute uncertainty of order 10%. This fraction increases as one moves to regimes of increasing cloudiness, until the dominant balance shifts to one between removal of water by precipitation and supply by low-level convergence. The quasi-equilibrium timescale of water vapor in the clear regime was found to be no more than a week; that is, the troposphere would dry out due to subsidence with an $e$-folding time of a week or less if moisture sources were cut off. This is approximately equal to the global residence time of water vapor in the atmosphere.

Direct estimates of turbulent transport of vapor from the boundary layer, obtained by looking at sounding variability, were based on crude reasoning, but proved consistent with the budget residual obtained from the analyses. It should be stressed that, although the analyses have been produced using soundings as input, this estimate of turbulent transport (whatever its other shortcomings) is thoroughly independent from the budget residuals. At 300, 400, and 500 mb, there were noticeable biases in the mean analyzed humidities, but they did not seem to have a large impact on the residual estimate of the local moistening.

At the 700-mb level itself, considerable uncertainty exists in the amount of local moistening. The ECHAM3 shallow convective scheme was found to insert significant amounts of moisture at this level, producing a large residual in the budget. The ECMWF 700-mb residual was somewhat smaller, but still large, about 50%–60% of the subsidence drying. Model forecasts at this level tended to be too moist in clear areas. Thus, model errors may have slightly exaggerated the local source here, but much is probably real. The sounding estimates of local turbulent moistening were not reliable here either, since the answer is very sensitive to the physical assumptions used to interpret the data. At any rate, the sounding methods indicate that the situation is changing so rapidly with depth that not too much should be made of the result at this particular analysis level.

Some conclusions have also been reached concerning the accuracy of sounding data and the analyses. An examination of apparent mixed layers showed that standard radiosonde data and analyses should not be trusted above 300 mb, but appear responsive below this. A recently improved sonde by Vaisala has been reported to give useful results above this level, however. It was also determined that NWS dry reports ($RH < 20\%$) tend to average to about 10% RH, according to both hygrostat types, if calibrated correctly. This is a lower value than has been used previously.

Overall, the performance of the models and analyses with respect to arid region water vapor seems to be reasonable, especially considering the numerical difficulties, poor knowledge of shallow convective processes, and lack (or mistrust) of observations. humidities in convective areas are underestimated, but when sufficiently averaged, errors in vertical velocity and moisture in arid regions are not as large as might be feared. The main shortcoming of the standard analyses for this study was the vertical sampling. Full spectral analyses have recently become available, with some additional vertical sampling, but the vertical structure in Fig. 14 was not anticipated, and they were not used.

An interesting local maximum in DV at 500 mb was found. It indicates preferential detrainment near the freezing level. Further investigation of this feature could provide some much-needed insight into the factors that control vertical profiles of the effects of cumulus on the large-scale variables.

The most interesting interpretation of the results in this paper is that, during any given week or month, the free troposphere is not in equilibrium with the surface in many parts of the Tropics, at least as regards moisture. The equilibrium reached in these areas depends mainly on conditions horizontally distant from the areas themselves. In a companion paper (Sherwood 1996), it is shown by explicit calculation that observed water vapor levels away from organized convection may be predicted accurately on the basis of large-scale processes alone.

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APPENDIX

Sounding Data

The soundings used in this study are listed in Table A1. Most are routine upper-air sounding sites operated by various national agencies, but some are supplemental soundings associated with the CEPEX and TOGA-COARE experiments. The data from some of these stations (indicated by ‘H,’ or high resolution) were obtained in their original form on magnetic media from the sounding station. The resolution of these soundings is at least 50 meters (for an example, see Fig. 1). The period considered is from March and April 1993. Some of these stations (Canton and ISS stations) only operated during part of the 2-month period. The remainder of the stations, indicated by ‘L’ in the table, were obtained through the World Meteorological Organization (WMO) archive at NCAR from Global Telecommunication System (GTS) broadcasts. Also listed in Table A1 are the system and thermodynamic sensor types used at each station. The mixing ratio is calculated from the observed temperature and relative humidity using the Goff–Gratch formula (Goff and Gratch 1946; Elliot and Gaffen 1991).

Previous investigations have found that the Vaisala sonde obtains humidities accurate to 5% through nearly the entire troposphere (Kley et al. 1995). The VIZ hygrometer suffers from a number of problems, however (Brousaides and Morrissey 1971; Brousaiides 1975; Elliot and Gaffen 1991; Wade and Schwartz 1993; Wade 1994), but appears to achieve similar accuracy through most of the troposphere, except under very dry or wet conditions. The most important problem is one of calibration at low relative humidity, which has been corrected according to Wade (1994), bringing the VIZ humidity distributions into line with those of the Vaisala sensor (see below). Figure 1 shows an example of this correction. A MicroART software error has also been removed (Wade and Schwartz 1993).

All soundings with any suspiciously large deviations from the average temperature, lapse rate, wind speed, or wind shear have been inspected by eye to ensure that they have no apparent flaws. A small number of soundings were removed from the database, and a larger number required truncation or partial removal. An occasional problem was found with the VIZ temperature sensor. Although the precision of the thermistor has been established at 0.3°C by comparison flights (Elliot and Gaffen 1991), a few soundings in the present study showed too shallow a lapse rate above the freezing level, reaching temperatures near the tropopause that are as much as 15°C too warm. This happened in about 1%–2% of soundings, and the cause is unknown. Furthermore, the temperatures reported by the VIZ thermistor generally show more variability than those of the Vaisala Thermap, with occasional anomalies throughout the upper troposphere of several degrees relative to adjacent soundings. The quality control pro-

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* Indicates no wind data.
cEDURE adopted here was to truncate a sounding at 450 mb if its average temperature between 350 and 200 mb deviates by more than 2.5σ both from the median at that station and from the mean of its two nearest neighbors (a total of 13 MicroART and 0 other soundings.)

a. Useful sensor range

The VIZ sensors are rated by the manufacturer only to −40°C (below approximately 250 mb in the Tropics) and the Vaisala sensors to −80°C (the entire troposphere). Data are typically used to 300 mb. However, the usefulness of the humidity sensors above 500 mb (particularly the VIZ hygrometer) is often questioned, and the author is not aware of any previous quantitative determination of the useful range of actual operational data.

To address this issue, the high-resolution data have been searched for apparently mixed layers such as the one near 550 mb in Fig. 1. Layers like this one, mixed by turbulence, should be uniform in conserved properties such as potential temperature and mixing ratio. The criterion for identifying one of these in the sounding data is that the potential temperature vary by no more than 0.5°C between two points 20 mb apart. Across each identified layer, a functional

\[ F = \frac{\delta \log q}{\delta \log q_s} \]  

(A1)

has been calculated, where \( q_s \) is the saturation mixing ratio, to see if the mixing ratio is also well mixed. Layers of constant mixing ratio \( q \) yield \( F = 0 \), while layers of constant \( RH = q/q_s \) yield \( F = 1 \). Not all layers identified by this technique are truly mixed, so only values \(-1 < F < 2\) are retained and averaged at each pressure level. The results are plotted in Fig. A1. The Vaisala sensor boasts a faster response time, which is apparent through most of the troposphere. Both sensors appear to lose sensitivity abruptly at about 250–300 mb though, registering constant RH in apparently mixed layers above this height. Note that \( F \) is an indicator of the sensor’s ability to measure \( \delta q/\delta z \), not necessarily \( q \) itself. Thus, these results are not inconsistent with the finding of Kley et al. (1995) that the Vaisala sensor achieves accurate mean values up to 100 mb. As a result of this test, however, the use of all island sounding data and analyses will be restricted to 300 mb and below.

b. Archived data and low RH

Some of the island data could not be obtained directly and came instead from National Weather Service (NWS) archives of data transmitted by GTS. These stations are indicated by \( \Delta \) resolution in Table 1. Data obtained in this fashion have additional problems not associated with the instruments themselves, including erratic spikes and inconsistent vertical resolution (Gaffen 1992).

Additionally, prior to October 1993, the NWS policy regarding the VIZ hygrometer was to set the dewpoint depression to 30°C whenever the RH dips below 20%. This makes it impossible to correct these soundings since the sensor information has been discarded, so the dry values must be set arbitrarily. Previous investigators have set these to various constant RH values such as 15% (e.g., Gutzler 1993), while the NWS reports them as 19%. Figure A2 shows the actual mean RH among cases where RH is below 20% from the Vaisala soundings (indicated by triangles) and also from the corrected high-resolution VIZ soundings (crosses),

![Fig. A2. Average RH when RH < 20%. Vaisala stations are indicated by Δ and corrected VIZ stations by ×. The horizontal axis is the fraction of reported humidities (pressure weighted) that fall below 20%. Only Vaisala and high-resolution VIZ stations are included, using all soundings available.](image-url)
plotted against the frequency of occurrence. While stations that rarely observe humidity below 20% see means of up to 15% over these cases, the means quickly drop to approximately 10% for the islands that experience such dryness with significant frequency. Thus, all NWS dry data are set to 10% RH in this study. Figure A2 also shows that, with the correction of Wade (1994), the distribution of dry humidities measured by the VIZ hygrometer is not too far from that obtained by the Vaisala Humicap. The stations at Wake and Guam observe RH of less than 20% over half the time at some altitudes, so mixing ratio averages there could be in error by as much as 25% at the midtroposphere, according to the scatter in Fig. A2.

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