On the control of stratospheric humidity
Steven C. Sherwood
Universities Space Research Association, Seabrook, MD
Andrew E. Dessler
Earth System Science Interdisciplinary Center, University of Maryland, College Park, MD

Abstract. We present a hypothesis on the dehydration and transfer of air from the tropical troposphere into the stratosphere. The hypothesis is based on the existence of a thick “tropopause layer,” in which vertical and horizontal mixing are both significant. Air is rapidly dehydrated upon entering this layer in vigorous convective overshoots, then slowly ascends through the layer before fully entering the stratosphere. Dehydration and genuine entry into the stratosphere are separate processes that happen on much different time scales.

Introduction

The humidity of the stratosphere today corresponds within uncertainties to the saturation mixing ratio of the tropical tropopause [e.g. Dessler, 1998]. However, the exact mechanism by which air is dehydrated as it passes through the tropical tropopause and into the stratosphere is unknown. The issue is important since stratospheric moisture is believed to play an important role in ozone depletion [Kirk-Davidoff et al., 1999] and the stratospheric radiative balance [Forster and Shine, 1999], and seems to be increasing with time [Oltmans and Hofmann, 1995]. A variety of mechanisms have been proposed to explain the observed dehydration. Here we review them and present a new view incorporating overshooting convection and slow ascent that best fits the available data.

1. “Cold trap” Brewer [1949] envisioned the flux into the stratosphere as slow, large-scale upwelling through the tropical tropopause (located at ~100 hPa). In this model, it is assumed that the tropopause region acts like a “cold trap,” where air passing through it is dehydrated to the region’s local minimum saturation mixing ratio. The recent observation of thin cirrus near the tropopause [e.g., Prabhakara et al., 1993; Winker and Trepte, 1998] suggests the existence of frequent uplift near the tropopause, but the absence of a thick cirrus deck throughout the tropopause region argues against uniform, steady uplift [Robinson, 1980].

2. Quasi-stationary disturbances One mechanism that overcomes those problems is in situ dehydration over meso- or synoptic-scale convective systems. These systems can lift the lower stratosphere, producing substantial cooling in some cases [e.g. Fritsch and Brown, 1982]. They can persist for up to several days, giving them time to condense and remove significant amounts of vapor.

3. Propagating waves Cooling also occurs in the crests of buoyancy waves near the tropopause [Potter and Holton, 1995]. Though it is not clear that wave crests will pass slowly enough—or ice removal proceeds rapidly enough—to remove significant amounts of moisture before a wave crest passes [Jensen et al., 1996], the mechanism could still work given repeated lifting of the same airmass, or with slow enough waves. In the limit of very slow waves, this mechanism becomes essentially identical to the previous one.

4. Convective overshooting A rather different theory is that energetic convection originating near the surface overshoots the tropopause and mixes with stratospheric air, thereby transferring mass from the troposphere directly into the stratosphere [Danielsen, 1982, 1993]. Overshooting turrets ascend adiabatically, becoming much colder than the surrounding environment and producing the possibility of great drying. There is evidence to support this mechanism [e.g., Danielson, 1993], but not to establish that it is quantitatively significant. Mote et al. [1996] added that the overshooting upward mass flux might exceed that of the Brewer-Dobson circulation, “hyperventilating” or driving sinking through the tropopause elsewhere—though the latter would be inconsistent with modeled and observed net heating and ascent near the tropopause (see below).

The Mixing Layer Hypothesis

Our proposal focuses on a region that we will hereafter refer to as the tropical tropopause layer (TTL), that region of the tropical atmosphere extending from the zero net radiative heating level (355 K, 150 hPa, 14 km) to the highest level that convection reaches (~420-450K, 70 hPa, 18-20 km). The TTL can be thought of as a transition layer between the troposphere and stratosphere. There is strong evidence that air in the TTL is rising at a few tenths of a Kelvin (in potential temperature, θ) per day, in equilibrium with net radiative heating [Folkins et al., 1999].

Figure 1 shows a schematic diagram comparing our hypothesis with hypotheses 1-3 above, collectively identified as “slow-dehydration” mechanisms. In our hypothesis, air is dehydrated rapidly as suggested by Danielsen [1982] but detrains at variable levels throughout the depth of the TTL. The bottom of the TTL is close to the typical level of neutral buoyancy (LNB) of tropical convection [Selkirk, 1993]. Air detraining near the bottom of the TTL need not have over-
slowly risen to this altitude. Air on any given altitudes, however, must have originated in more energetic shot and will typically be relatively moist (compared to the moister air that detrained earlier at lower altitudes and has slowly and ubiquitously above 150 hPa, undergoing considerable horizontal movement before passing the hygropause. As mentioned above, this air can be dehydrated to moisture levels far below average for the TTL. This “young” dehydrated air spreads horizontally from the convective region, where it intermingles with “older,” moister air that detrained earlier at lower altitudes and has slowly risen to this altitude. Air on any given \( \theta \) surface will therefore have a spectrum of “ages” ranging from air that has just detrained, to air that has slowly risen all the way from the bottom of the TTL. At locations farther from intense convection, this spectrum will be weighted more toward older air, causing mean vapor levels to be greater.

Mixing and stirring processes within the TTL are a critical issue. Horizontal stirring is effective within the TTL regardless of how dehydration occurs, since the vertical transit time is long enough for horizontal winds to transport air far from its entry location. Our proposal amounts to an additional, convectively-driven, vertical mixing component within the TTL. This is because (as noted by Danielsen [1993]) air masses that detrain significantly higher than the LNB must have mixed irreversibly with high-\( \theta \), stratospheric air. The final composition of such air parcels is therefore a mixture of air from the planetary boundary layer (PBL) and from somewhere higher in the TTL, with conserved properties equal to the weighted mean from those two locations. Thus, newly dehydrated air in the middle and upper TTL would tend to follow mixing-line behavior, while “old” air could deviate from the mixing line. Interestingly, Folkins et al. [1999] found approximately linear variation of \( \theta \) and ozone within the TTL, consistent with vertical mixing.

Supporting Data

Our hypothesis is supported by data collected from the ER-2 during the STRAT mission, based in Hawaii during 1995-96. Data shown here were collected from 5S-25N at \( \theta \)-levels from 330 to 460 K. Points with \( \mathrm{N}_2\mathrm{O} < 308 - [0.025(\theta - 300)]^2 \) ppbv are omitted, to exclude air that has recycled through the stratospheric “overworld” where \( \mathrm{N}_2\mathrm{O} \) destruction occurs. This curve is our empirical fit to the upper envelope of \( \mathrm{N}_2\mathrm{O} \) values, with a ~8 ppbv offset. Results below do not change for reasonable changes in threshold, nor if \( \text{NO}_x \) is used to screen data instead. See Boering et al. [1994] for further information on the measurement and behavior of \( \mathrm{N}_2\mathrm{O} \).

First we examine horizontal water vapor [Weinstock et al., 1994] and ozone [Proffitt and McLaughlin, 1983] variations. Ozone indicates how long air has been in the TTL, since air transported into the TTL by convection contains low values (O(10) ppbv), which then increase steadily due to production. Figure 2 shows data from near the 380 K surface (which is close to the climatological tropopause and hygropause, or vapor minimum located just above the tropopause). These scatter plots clearly show that lower ozone (i.e. more recent arrival at this \( \theta \)-level) is associated with increasing dryness. This result, shown for all four STRAT observation periods, also holds at other altitudes within the TTL (not shown). Variations in \( \theta \) do not contribute significantly to the signal in Figure 2; in fact, the overall trends in vapor and ozone with height (not shown) would produce the opposite correlation. The \( \mathrm{H}_2\mathrm{O} \) variations are also uncorrelated with \( \mathrm{N}_2\mathrm{O} \) and therefore not caused by inmixing of overworld air, whose \( \text{O}_3-\text{H}_2\text{O} \) relation is in any case quantitatively much different (dotted lines in Figure 2).

Note that in the “slow dehydration” hypotheses, dehydration occurs during (or at the end of) the many weeks required for air to ascend from the bottom of the TTL to the hygropause (see Figure 1). By this time, due to the isentropic scrambling of the air, there is no reason to expect any correlation between water vapor and ozone abundances at a fixed \( \theta \). Thus, we believe that the observed correlations stand in opposition to any theory whereby the air is dehydrated primarily during slow ascent (though we cannot rule out some rapid, large-scale process, nor can we rule out a secondary contribution from slow ascent).

Further investigation of the data bolsters the vertical-mixing aspect of our hypothesis. Ozone profiles (Figure 3)
Figure 2. H$_2$O vs. ozone from four phases of the STRAT experiment, for data collected from 380±5 K. Dotted lines show best linear fit to 380 K data that were excluded based on N$_2$O.

Figure 3. O$_3$ vs. θ from two STRAT phases. Dryest 10% of points between 360 K and 420 K in each panel are surrounded by a gray diamond.

Figure 4. As in Figure 3, except CO$_2$ vs. ozone.

show that the most dehydrated air falls very close to a mixing line between tropospheric properties (low ozone and θ ~ 355 K, for air having the greatest likelihood of reaching these heights) and values above the tropopause. Other air often falls well to the right of this. This behavior also occurs in the two other STRAT periods (not shown), and suggests a connection between dehydration and vertical mixing.

If we examine carbon dioxide [Boering et al., 1994] as a function of apparent age of air (Figure 4), the dehydrated air again sometimes looks anomalous. Though CO$_2$ behavior is dominated by the clear upward propagation of the seasonal variation in tropospheric CO$_2$, some points depart substan-

tially from the majority. The dehydrated air again appears to fall on a mixing line in late 1995 (this is less obvious in July-August 1996 but recurs in Dec. 1996, not shown). Also, in all four periods, CO$_2$ variability is greatest between 370 and 400 K and CO$_2$ is generally higher than average in dehydrated air. Speculation is possible about a convective transport role in establishing this behavior, but more work would clearly be needed to draw conclusions.

Final thoughts

In agreement with Danielsen [1982, 1993], we argue that dehydration of air bound for the stratosphere occurs in convective systems. However, convection can introduce mass only into a layer which shares tropospheric and stratospheric characteristics. In particular, since this layer is vertically mixed, it lacks a key characteristic of the stratosphere. To completely escape tropospheric mixing, all detrained air must also undergo a period of slow, (roughly) zonally-uniform ascent to the layer top. Thus, stratospheric entry is fundamentally a two-step process.

Our hypothesis explains how this ascent at tropopause heights fails to produce an overcast layer. Air entering the TTL after overshooting is sufficiently dry—and the time taken for air to rise through the layer sufficiently long compared with horizontal mixing times—that the tropopause throughout the tropics can remain mostly near or below saturation even without significant precipitation of ice crystals. The idea also accounts for the observed increase in water vapor as one moves horizontally away from deep convection, as observed near the tropopause by Vomel et al. [1995].

The observational evidence for this process is interesting but far from conclusive. More observations are needed, particularly closer to deep convection. If the theory were true it would require:

- TTL air parcels, traceable to deep convection, with
properties representing mixtures with PBL air beneath the convection—including CO$_2$, radon, O$_3$, and others [e.g. Kritz et al., 1993];

- rapid removal of ice from overshooting towers;
- a modest climatological horizontal gradient of water vapor near and above the tropopause, with lowest values over deepest convection; and
- net downward mass flux near the tropopause over the deepest convection.

The last point—contrary to the usual interpretation of Danielsen’s mechanism—is required since individual mixtures, having lower $\theta$, would detrain below the mixing location. This would produce a net mass flux that is upward below the detrainment level, but downward between this level and the mixing level. Net downward motion was estimated over Indonesia by Sherwood [2000], though the estimate is prone to possible errors and it is not known whether the sinking is actually produced by overshooting.

An important implication of this hypothesis is that stratospheric moisture may depend on cloud physics. This warrants further investigation of such physics at the top of the convective layer.

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References


S. Sherwood, NASA/Goddard Space Flight Center, Mail Code 916, Greenbelt, MD (e-mail: ssherwood@lanl.gov)

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