Can overshooting convection dehydrate the tropical tropopause layer?

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[1] A numerical model is used to investigate the potential for irreversible dehydration near the tropical tropopause caused by overshooting deep convection. We show that convective updrafts overshooting the cold point tropopause can generate extremely cold, dry air within the updrafts. However, the updrafts contain sufficient mass in small ($\leq$20 μm radius) ice crystals that do not sediment out of the short-lived overshoots, such that when the overshoots collapse and warm back to the ambient temperature, these small crystals sublimate and rehydrate the air, resulting in no irreversible dehydration. Despite maximizing crystal size (and fall speed) by assuming low aerosol concentrations and using large ice-ice collection efficiencies, we find no evidence to support the hypothesis that overshooting convection can dehydrate the tropical tropopause layer (TTL) when it is initially ice subsaturated. Only when the TTL is initially supersaturated with respect to ice do we find that deep convection can draw down the humidity, as vapor in excess of saturation condenses on the ice crystals. The overall impact of deep convection on the TTL water vapor budget depends on the climatology of TTL relative humidity in convective regions.

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1. Introduction

[2] The aridity of the stratosphere is caused by freeze drying of air ascending across the cold tropical tropopause. Given the importance of stratospheric humidity for the ozone budget and infrared radiative fluxes, considerable effort has been devoted to understanding in detail how air entering the stratosphere is dehydrated and, in particular, how the stratospheric humidity might change in response to climate change. Several recent studies have shown that slow ascent through the tropical tropopause layer (TTL) coupled with large-scale horizontal transport through cold regions in the tropics and in situ formation of thin cirrus can dehydrate air to the observed lower stratospheric water vapor concentrations [*Gettelman et al.*, 2002; *Jensen and Pfister*, 2004; *Fueglistaler et al.*, 2005]. However, an alternative mechanism involving rapid cooling and dehydration in overshooting convection has also been suggested [*Sherwood and Dessler*, 2000, 2002].

[3] When strong convective updrafts overshoot beyond the level of neutral buoyancy into the relatively stable TTL, air in the undilute updrafts will continue to follow a moist adiabat and extremely cold temperatures can be achieved. Deposition of water vapor on the ice crystals in the cold updraft could generate extremely dry air. This dry air will then warm up, either as the updraft collapses or as overshoot air detrains into the warm environment (possibly resulting from breaking gravity waves [*Wang*, 2003]). The convective dehydration mechanism requires removal of most of the ice by sedimentation, such that when the dry updraft air warms and mixes with the environment, ice sublimation does not rehydrate the air. Hence the key requirement is that ice crystals are large enough to rapidly sediment out of the short-lived convective overshoots. Further, since convective overshoots penetrating to near the tropopause are relatively rare [*Dessler*, 2002; *Liu and Zipser*, 2005; *Dessler et al.*, 2006] and the vast majority of convective detrainment occurs at lower, warmer levels, the cold overshoots must detrain air that is much drier than the mean tropopause humidity if they are to contribute significantly to the water vapor budget.

[4] We consider the potential impact of deep convection on the tropopause temperature as a separate issue. Several studies have suggested that convection penetrating to the tropopause can cool and sharpen the temperature minimum [*Johnson and Kriete*, 1982; *Selkirk*, 1993; *Danielsen*, 1993; *Sherwood et al.*, 2003; *Kuang and Bretherton*, 2004]. Inasmuch as temperature controls freeze drying of air via the slow ascent mechanism, convective influences on TTL temperature could be an important factor affecting the stratospheric water vapor concentration. This issue is not addressed in this study.

[5] The impracticality of sending high-altitude aircraft into turbulent convective overshoots prevents direct measurements of the crystal sizes within the updrafts. Here, we

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use a numerical model to investigate overshooting convection physical processes and impact on TTL humidity. We show that air within the simulated convective updrafts penetrating the tropical tropopause can get extremely cold and dry. However, the mass of ice in relatively small crystals that do not precipitate out is sufficient to rehydrate the updraft air as it detrains and warms. As a result, our simulations of overshooting convection produce no net dehydration (unless the TTL is initially supersaturated with respect to ice).

2. Model Description and Initialization of Simulations

[6] For our simulations, we use DHARMA (Distributed Hydrodynamic Aerosol-Radiation-Microphysics Application), a three-dimensional Eulerian cloud model that simulates the coupling between atmospheric dynamics, cloud microphysics, and radiation [*Ackerman et al.*, 2003]. The dynamical component of DHARMA is a massively parallel, large-eddy simulation code [*Stevens and Bretherton*, 1996; *Stevens et al.*, 2002]. Embedded within this dynamics code is CARMA (Community Aerosol and Radiation Model for Atmospheres), a detailed microphysical model [*Ackerman et al.*, 1995; *Jensen et al.*, 1998]. Aerosols, cloud liquid, cloud ice, and graupel are each tracked in 24 size bins, spanning 5 nm to 1 μm for aerosols and 1 μm to 1.2 cm equivalent-volume radius for cloud hydrometeors [*Fridlind et al.*, 2004]. A wide range of microphysical processes are treated, including condensational growth, evaporation, sedimentation, melting, riming, gravitational collection, drop breakup, homogeneous and heterogeneous freezing of aerosols and drops [*Koop et al.*, 2000], and Hallett-Mossop rime splintering. Graupel is formed when droplets collide with cloud ice and the droplet bin mass is greater than the ice bin mass.

[7] Hydrometeor sedimentation rates are calculated following *Pruppacher and Klett* [1997]. For cloud ice (which dominates in the convective overshoots), density versus radius is taken from anvil cirrus measurements reported by *Heymsfield et al.* [2004]. For small cloud ice crystals (radii $\leq$ 40 μm) that are critical to the overshooting dehydration issue, the equivalent-volume sphere densities range from about 0.25 to 0.9. (Very little graupel mass is present in the overshoot.)

[8] For this study, we are aiming to maximize the potential for convective dehydration, which means minimizing ice crystal concentration such that the ice crystals will be as large as possible and sediment from the overshoots as rapidly as possible. First, we have chosen an aerosol concentration of 100 cm^{-3} (throughout the model domain) corresponding to relatively clean, maritime conditions [*Clarke and Kapustin*, 2002; *Brock et al.*, 2004]. We use a lognormal size distribution with a mode radius of 20 nm and a geometric standard deviation of 2.1. Next, we assume that ice-ice collection efficiencies are the same as those used for droplet-droplet collisions [*Pruppacher and Klett*, 1997; *Hall*, 1980; *Jacobson et al.*, 1994]. Actual collection efficiencies for ice-ice collisions are certainly lower than those for droplet collisions [see *Field et al.*, 2006, and references therein]; hence we are likely overestimating ice crystal growth by aggregation.

[9] We assume here that homogeneous freezing of droplets at about -37°C dominates production of ice crystals in the updrafts. If updraft speeds are sufficiently slow and heterogeneous ice nuclei (IN) are relatively plentiful, then ice crystals generated from IN at temperatures warmer than -37°C can rob water from remaining droplets via the Bergeron-Findeisen process and effectively quench the homogeneous freezing process, with the net result potentially being fewer, larger crystals. However, *Heymsfield et al.* [2005] estimate that this quenching will only occur in relatively weak updrafts (speeds < 5 m s^{-1}). Since we are focusing here on strong, overshooting updrafts with speeds much larger than this threshold, it is very unlikely that heterogeneous IN would significantly affect the ice crystal size distributions in the overshoot regions.

[10] The initial thermodynamic state for the simulations is based on the 19 November 2001 08:15 UTC sounding launched at Darwin, Australia during DAWEX (Darwin Area Wave Experiment) [*Tsuda et al.*, 2004]. This sounding was chosen because of the unusually large convective available potential energy of 5.5 kJ kg^{-1}. The TTL relative humidity with respect to ice in this sounding is $\simeq$90%. If the TTL is initially supersaturated, ice injected by convection can draw down the humidity to ice saturation, as demonstrated below. We are focusing here on the possibility that cold overshoots can dry the air well below the ambient ice saturation humidity. Open boundary conditions are used at each of the horizontal faces, with nudging toward the initial sounding applied to the outer 10% of the grid with a time constant of 100 s. Convection is triggered by specifying sensible and latent surface heat fluxes of 500 W m^{-2} each over a 20 km radius area in the center of the 72 $\times$ 72 km horizontal domain. Results are not sensitive to the magnitudes of these fluxes, so long as they are large enough to trigger overshooting deep convection. The grid resolution is 500 m in the horizontal and 275 m in the vertical, with a vertical domain of 0 to 24 km.

3. Results and Discussion

[11] The surface heat fluxes imposed trigger deep, strong convection with updraft speeds as high as $\simeq$35 m s^{-1} in the uppermost troposphere about 2 hours into the simulation. The most energetic updraft core overshoots the level of neutral buoyancy (16.5 km) and reaches about 18.2 km. When the updraft reaches about 14 km, the concentration and sizes of ice crystals at the top of the cloud have reduced sufficiently to allow buildup of ice supersaturation and homogeneous freezing of sulfate aerosols entrained into the plume. (At 14 km, the temperature is well below -37°C; hence the aerosols homogeneously freeze before the relative humidity with respect to liquid water exceeds 100%.) This process produces large concentrations (>10 cm^{-3}) of small crystals (see *Jensen and Ackerman* [2006] for details). Figure 1 shows an *x-z* cross section of water vapor mixing ratio, ice water content, and temperature perturbation (defined as the difference between the local temperature and the initial, horizontally uniform value) through the overshooting convective core at about 2.3 hours. In the overshoot region, the temperature is as much as 21 K colder than the surrounding environment; the minimum temperature in the overshoot is 174.6 K. The air is highly

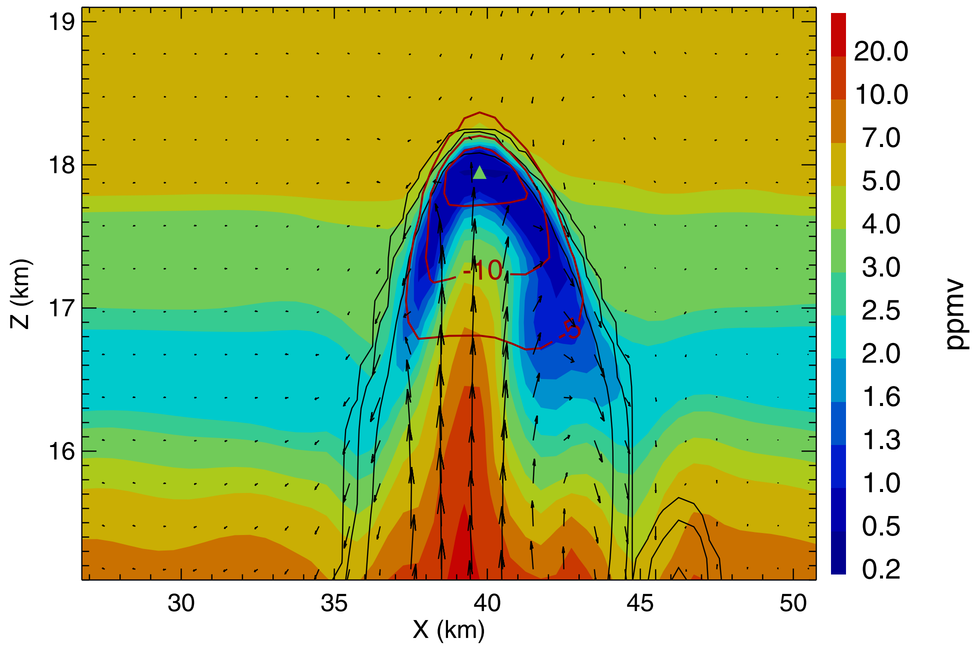

Figure 1. An *x-z* slice through the highest overshooting updraft in the simulation is shown. Fields include water vapor mixing ratio (color shading), ice water mixing ratio (black contours), and temperature perturbation (red contours). Ice water mixing ratio contours are 10, 100, and 1000 ppmv. The convective overshoot drives the temperature as much as 21 K colder than the environment and dries the air in the overshoot to as low as 0.4 ppmv. The green triangle shows the location of the ice mixing ratio distribution shown in Figure 2.

supersaturated at this location (relative humidity with respect to ice, $RHI \simeq 175\%$), but the water vapor mixing ratio is very low (0.4 ppmv). Hence the simulated overshoot does generate extremely cold, dry air, as required for overshooting dehydration.

[12] As discussed above, the potential for irreversible dehydration ultimately depends on the ice crystal size distributions in the overshoot. The crystals must be large enough to sediment out before the overshoot collapses and warms up. The cumulative distribution of ice mass mixing ratio in the overshoot as a function of crystal equivalent volume radius is shown in Figure 2. Also shown is the time to sediment 200 m versus crystal radius. Choosing 200 m as the length scale for sedimentation time calculations should underestimate the physically relevant sedimentation time since the depth of the overshoot is about 1.7 km. At the driest point in the overshoot, the water vapor mixing ratio is about 1.5 ppmv lower than that outside the updraft, but the mass of ice in crystals smaller than $\simeq$20 μm radius is >4 ppmv. These small crystals would take at least about 100 min to fall 200 m; this time is larger than the overshoot lifetime of $\leq$20 min, where the overshoot lifetime is the time required for the overshoot to collapse and warm up to within a few K of the initial temperature. This analysis indicates that the overshoot contains more than enough mass of small, slowly sedimenting ice crystals that will rehydrate the air when it warms back to the ambient temperature, thus preventing irreversible dehydration.

[13] Note that in this simulation the vertical shear of the horizontal wind near the tropopause is relatively weak, such that the overshoot simply collapses on itself. When significant shear exists near the tropopause, gravity wave breaking can rapidly entrain surrounding air into the overshooting plume [*Wang*, 2003]. However, such rapid mixing would only decrease the time allowed for sedimentation of ice crystals and thus decrease the potential for irreversible dehydration.

[14] The ultimate question here is whether air detrained from the convective overshoot in the TTL is drier than the ambient TTL air. For identification of the TTL air detrained from convection, we use a tracer with a specified initial mixing ratio of unity below 14 km and zero above. Hence

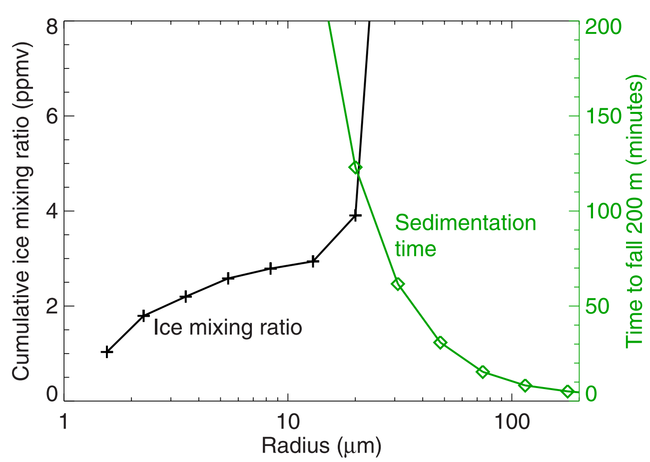

Figure 2. Cumulative distribution of ice mixing ratio in the overshoot (location indicated by the green triangle in Figure 1) is shown. Also shown (green curve) is the time required for crystal sedimentation of 200 m versus radius. At this location, the temperature is 21 K colder than the ambient, and the water vapor concentration is 0.4 ppmv.

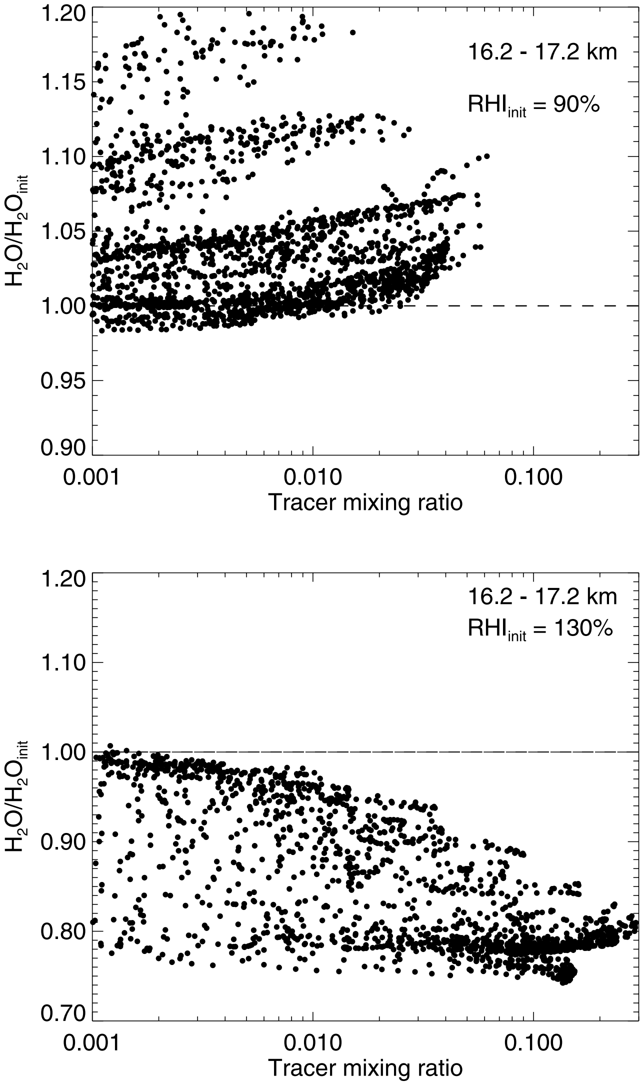

Figure 3. Ratio of final TTL (16.2–17.2 km) water vapor concentration to initial water concentration is plotted versus the convective tracer concentration. (We remind the reader that the level of neutral buoyancy is 16.5 km.) All ice has been removed from the TTL at this stage, by either sedimentation or sublimation. The tracer was initially restricted to altitudes below 14 km, implying that any grid cells containing the tracer at the end of the simulation were influenced by the overshoots. The top plot shows results from the baseline simulation with the initial TTL *RHI* of $\simeq$90%. The predominant effect of the overshooting convection is to hydrate the TTL. The bands apparent are a result of the distinct model vertical layers. The bottom plot shows results from a simulation with the initial TTL *RHI* specified as 130%. In this case, deposition of vapor in excess of saturation on convectively injected crystals, followed by sedimentation, reduces the humidity.

the presence of the tracer in the TTL after the convective event indicates outflow from the convection. In the region near the hygropause (16.2–17.2 km) where the initial water vapor concentration was 2–2.6 ppmv, the water vapor concentration in grid boxes containing appreciable amounts of the tracer is almost always larger than before the convection (see Figure 3, top plot). Further, within each model layer, the water enhancement tends to increase with the tracer mixing ratio. indicating that convective detrainment with the least dilution has the greatest degree of hydration.

[15] In summary, we find no evidence for irreversible dehydration of subsaturated TTL air in our simulation of overshooting convection. Despite maximizing the potential for overshooting dehydration by using relatively low aerosol concentrations and relatively large crystal aggregation efficiencies, the simulated overshooting plume still contains substantial mass of small crystals that do not sediment out. We note again that since convection overshooting to the hygropause level is very rare, this dehydration process would only be important for the water vapor budget if air much drier than the ambient were produced.

[16] Ultimately, once ice crystals injected by deep convection sublimate due to entrainment of dry air, the convection has left behind a volume of saturated air in the TTL. This saturated air is more humid than the surrounding air simply because the initial *RHI* was less than 100%. One would expect that if the TTL were initially supersaturated with respect to ice, then ice crystals injected by deep convection would grow and deplete the TTL humidity. As a demonstration of this dehydration process, we have run our simulation with the initial TTL *RHI* increased to 130% by increasing the TTL water vapor concentration. As shown in Figure 3 (bottom plot), after all ice crystals have sedimented out of the TTL, grid boxes with the convective tracer are drier than the initial conditions in this case.

[17] In agreement with past studies [*Smith et al.*, 2006], our simulations suggest that the impact of deep convection on the TTL water vapor concentration is to drive it toward saturation. If the preconvection TTL is initially supersaturated with respect to ice (as is often the case [*Jensen et al.*, 2001]), then vapor in excess of saturation will condense on ice injected by deep convection, potentially allowing irreversible removal of water from the TTL resulting from crystal sedimentation. If the TTL starts out substantially subsaturated, then sublimation of convectively injected ice will provide a source of water vapor. The net impact of deep convection on TTL humidity depends on the climatology of TTL relative humidity in regions with frequent convection penetrating to the tropopause region.

[18] We emphasize here that even though the overshooting dehydration mechanism appears to be ineffective, deep convection still necessarily plays key roles in the TTL water vapor budget for a variety of reasons. First, with the definition of the TTL as the layer where radiative heating drives slow ascent into the stratosphere, deep convection is the primary pathway for air entering the TTL. As discussed above, convective injection can be either a source or sink of water, depending on the preexisting relative humidity. Second, deep convection can affect the structure of the TTL temperature profile and the value of the cold point temperature. Third, deep convection is the dominant source of gravity wave activity in tropical regions remote from topography [*Fritts and Alexander*, 2003]. Temperature variability driven by gravity waves can drive in situ formation of TTL cirrus that can effectively dehydrate air slowly ascending into the stratosphere [*Jensen and Pfister*, 2004]. The gravity waves effectively decrease the minimum tem-

perature (and corresponding minimum H_2O saturation mixing ratio) experienced by air parcels ascending through the TTL, thus allowing dehydration to lower water vapor mixing ratios.

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