I. Microphysical and Dynamical Control of Tropospheric Water Vapor

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1. Introduction

Water vapor and condensed water are by far the most important greenhouse substances in the Earth's atmosphere, together accounting for the great majority of the net back-radiation from the atmosphere. In addition, clouds reflect incoming solar radiation, constituting about two-thirds of the planet's albedo. The condensed water in clouds is also strongly related to atmospheric water vapor content, occurring only where the actual vapor pressure locally approaches its saturation value. For these reasons, understanding the control of atmospheric water vapor is an essential part of understanding climate change.

Cycling of water through the troposphere takes on the order of 2 weeks, which is much shorter than the time scale of climate variability, so that we may regard the distribution and amount of water vapor and clouds as being in equilibrium with the rest of the climate system. (Residence time of water in the stratosphere is much longer, but stratospheric water has relatively little influence on radiative transfer, so we shall not concern ourselves with it here.) We argue here that the water content of the troposphere is controlled by both dynamical and microphysical processes, and that neither of these controls is well understood or represented in climate models. We demonstrate, using simple process models, that climate is sensitive to poorly known microphysical and dynamical processes and argue strongly for a program of research and field measurements designed to help resolve this problem.

2. Microphysical control

Consider a randomly chosen sample of air from the tropical subcloud layer, which contains the highest moisture content of any air in the atmosphere, and follow it (in a Lagrangian sense) as it moves around in the troposphere. It is advantageous to think of that movement in potential temperature (θ) coordinates, as shown in Figure 1.

Air crosses upward through θ surfaces principally through latent heat release in clouds; mostly in tall cumulonimbus in the tropics and over the oceans in general (even in winter), and in the moist slantwise ascent regions of baroclinic cyclones. Air descends through θ surfaces mostly in rainy downdrafts (at low levels) and through loss of heat by longwave radiation to space.

Now consider the water content of the test sample, averaged over a large ensemble of such samples. As air ascends through θ surfaces, it is saturated and generally contains cloud water. Air that ascends through the depth of the troposphere loses the vast majority of its water content by precipitation; still, it may exit the tops of clouds with condensed water content vastly in excess of the vapor content, owing to the very small saturation vapor pressure of the cold air near the tropopause. As it begins its descent, the sample's water content is the sum of its condensed water and water vapor content when it exits deep clouds.

It is important at this point to recognize that typical ascent velocities in clouds are at least an order of magnitude greater than the magnitudes of descent velocities in clear air, so most of an air sample's lifetime will be spent in its clear descent phase (or in passive stratiform clouds). Thus, much of the problem of determining the atmosphere's water content involves accounting for water outside of *deep* cloud systems.

The descending sample may take many weeks to return to the subcloud layer, during which time it may traverse considerable horizontal distances. As it descends, it is moistened (in a statistical sense) by detrainment from shallower clouds and by the partial evaporation of precipitation falling into it from above. (If these processes were absent, the sample would arrive at the surface with a relative humidity of only a few per cent.)

From this point of view, it is easy to see that the average water vapor content of the atmosphere is strongly influenced by those cloud microphysical processes, which determines how much condensed water air has on detraining from clouds, and the microphysics of the evaporation of precipitation as it falls through subsaturated air. (It is less easy to see that lateral mixing can strongly redistribute water between samples; this will be the subject of the next section.) Because microphysical processes are poorly understood, and because water vapor content in the upper troposphere is not well- measured, there is no reason to believe that these microphysical processes are well represented in climate models. Moreover, a typical climate model has seven or eight levels in the troposphere, vastly inadequate to resolve water vapor variations of four or five orders of magnitude.

Much of the vertical water transport in the tropics is accomplished by deep convection. But the convection schemes in use make extremely crude assumptions about moistening effects of convection, by-passing any attempt at a detailed accounting of microphysical processes, for computational and, perhaps, philosophical reasons. One example is the Kuo scheme (Kuo, 1965, 1974), which simply moistens in proportion to the precipitation rate and the departure of the air from saturation. Another is the Manabe scheme (Manabe et al., 1965) which moistens entire grid boxes to saturation (or near saturation) and precipitates any condensed water without allowing for re-evaporation. The Arakawa-Schubert scheme (Arakawa and Schubert, 1974) in its original form does not allow for re-evaporation of precipitation either. Perhaps the best scheme with some use in climate models is the Betts-Miller scheme (Betts, 1986; Betts and Miller, 1986), which relaxes moisture toward an empirical profile. But there is no physical reason to attach any universality to such a profile.

The sensitivity of one-dimensional, radiative-convective models to the type of convection scheme used, its parameters, and the model numerics has been documented by Rennó et al. (1994). An example of this sensitivity is demonstrated in Table 1, which shows the equilibrium surface temperatures in a one-dimensional radiative-convective equilibrium model in which the surface is a “swamp” (water with no heat capacity). The model uses the radiative transfer scheme of Chou et al. (1991). The scheme uses several broad-band representations of longwave and shortwave radiation, accounting for longwave absorption by water vapor, carbon dioxide, ozone, and clouds and shortwave absorption by ozone and water vapor, also accounting for shortwave reflection by the surface and by clouds. (Cloud-radiation interaction is turned off in these runs.) The convection scheme is that of Emanuel (1991) and is based on an episodic mixing model in which unstable air ascends adiabatically and then mixes with its environment at each model level, forming a spectrum of mixtures each of which then ascends or descends to its level of neutral buoyancy, where it detrains. At each step, a certain fraction of the condensed water is converted to precipitation, which falls and partially re-evaporates, driving an unsaturated downdraft that transports heat and water.

Table 1 shows the sensitivity of the equilibrium surface temperature to NL, the number of model levels, k_v, the vertical diffusion coefficient, $C_D\mathbf{v}_a$, the coefficient of surface heat and moisture fluxes, and the model time step. Note that the run with only 8 levels yields a surface temperature 5°C higher than the control run.

Table 2 shows the sensitivity of equilibrium surface temperature to microphysical parameters in the model. The most important parameter set is that which governs the conversion of cloud water to precipitation. Provisionally, this parameter is set as a function of the depth of the updraft:

$$\epsilon_i = \begin{cases} 0 & p_{i\mathrm{cb}} - p_i < PB_{\mathrm{crit}} \\ \frac{p_{i\mathrm{cb}} - p_i - PB_{\mathrm{crit}}}{PT_{\mathrm{crit}} - PB_{\mathrm{crit}}} & PB_{\mathrm{crit}} < p_{i\mathrm{cb}} - p_i < PT_{\mathrm{crit}} \\ 1 & p_{i\mathrm{cb}} - p_i > PT_{\mathrm{crit}} \end{cases}, \tag{2.1}$$

where ϵ_i is the fraction of condensed water converted to precipitation at level i, p_i is the pressure at that level, $p_{i\mathrm{cb}}$ is the pressure at cloud base, and PB_{crit} and PT_{crit} represent the minimum pressure depth below which no water is converted to precipitation, and the maximum pressure depth above which all water is converted to precipitation, respectively.

The other parameters shown in Table 2 include the fractional area covered by precipitating downdrafts, σ_d, and the fraction of precipitation assumed to fall through unsaturated air, σ_s.

Clearly, there is large sensitivity to all of the microphysical parameters, as expected from the arguments presented earlier. Unfortunately, it is extremely difficult to determine the optimal values of such ad hoc parameters, and to evaluate how well the scheme performs in determining water vapor content. The only conceivable way of validating convective schemes for use in climate models is to force them with vertical velocities, surface fluxes and radiation from a sounding array operated for at least 3 or 4 weeks, and to compare the predicted relative humidities to those observed.

Although the above experiments are concerned with cumulus convection, some of the issues that arise clearly carry over to the representation of microphysical processes in extratropical cloud systems. Moistening by such systems must also depend on conversion of cloud water to precipitation and on re-evaporation of precipitation.

The main conclusion to be drawn from these theoretical considerations and one-dimensional experiments is that *there is no reason to believe that any of the extant cumulus parameterizations is performing adequately in its control of atmospheric water vapor.* We shall see that this problem is compounded by horizontal mixing in the atmosphere.

3. Dynamical control

The microphysical processes described above would pose a formidable problem for modeling even if they took place in a relatively smooth larger scale environment. In reality, they are embedded in a dynamical environment of extraordinary intricacy. Advection by spatially smooth but time-dependent velocity fields can easily lead to chaotic wandering of fluid parcels. This generates small-scale structures in the advected field exponentially rapidly and can bring dry and moist air into close proximity in the vertical or horizontal. Nearby air parcels can have radically different Lagrangian histories, which in turn affects their moisture content. In the following, we illustrate some of these processes as manifested in passive tracer mixing calculations carried out on surfaces of constant dry potential temperature θ. A more complete exposition of isentropic mixing properties in the atmosphere can be found in Pierrehumbert and Yang (1993), and some of the effects of condensation are taken up in Yang and Pierrehumbert (1994). Of course, much essential physics is lost when we ignore exchange of fluid amongst θ surfaces. A full understanding of the typical Lagrangian history of air parcels will be achieved only when the ideas presented below are melded with the primarily cross-isentropic mixing processes detailed in the preceding section.

Figure 2 shows the zonal mean disposition of a typical "middleworld" θ surface such as we shall be concerned with, characterized roughly by the range θ between 300K and 380K. Lower, or "underworld" θ surfaces outcrop at the ground instead of in the tropics, and do not generally intersect the tropopause; higher "overworld" surfaces are wholly stratospheric, apart from cross-isentropic processes. The θ surfaces are not as rigidly fixed in space as are isobaric surfaces, but midlatitude baroclinic transients tend to have parcel trajectories that are only somewhat shallower than the slope of the mean isentropes. Hence, it is generally fair to think of the higher latitude portions of the undulating middleworld surfaces as being of consistently higher altitudes than the lower latitude portions.

Formation of *filamentary structures* is the phenomenon of primary interest. The isentropic trajectory problem constitutes a two dimensional, though time- dependent, dynamical system, and is thus characterized by two Lyapunov exponents.

These exponents give the exponential rate of growth or decay of the principal axes of the ellipse into which a small disk of tracers is transformed under action by advection for some specified time. For non-divergent flow, the two exponents must sum to zero, since the area of the disk is preserved. Since the isentropic flow has non-zero divergence, it could in principle lead to situations where both exponents are positive or both negative. Evidently, however, the parcels do not remain in regions of strong divergence or convergence long enough for this to happen. Instead, there is typically one positive and one negative Lyapunov exponent as for the incompressible case, and air parcels are systematically elongated in one direction while they are being systematically contracted in the other. This naturally leads to production of filaments, and to exponential amplification of gradients. Over finite times, the elongation rate is a sensitive function of the initial position of the trajectory, as discussed in Pierrehumbert and Yang (1993).

In Figures 3 and 4 we show how ten days of mixing re-arranges the air on the 315K, 330K and 350K θ surfaces. The winds used in the calculation are taken from the 12-hourly ECMWF operational analyses for March 1-10, 1993 (the beginning of the CEPEX period). We show snapshots on March 10, with the air colored according to the *latitude* at which it was located on March 1. A great deal of fine-scale structure has been generated after only 10 days. In fact, a direct calculation of the Lyapunov exponents shows that even more fine structure is generated than can be fully resolved in this mixing simulation. The typical midlatitude exponent of

.7/day implies that a 500km wide smooth feature would be elongated into a filament 500,000km long (which would have to be folded many times to fit on the globe) and only 500m thick after 10 days. The explicitly resolved filaments in Figures 3 and 4 arise from trajectories with anomalously weak stretching.

The joint action of planetary waves, shear in the jet streams, and especially synoptic eddies causes rapid and efficient mixing withing the two extratropical "ergodic" or "mixing" zones. The meridional displacement of air parcels in conjunction with the upward slope of the isentropes systematically processes air through the high-latitude, cold portions of the isentropic surface (Kelly, Tuck and Davies 1991; Yang and Pierrehumbert 1994), and constitutes a potent drying mechanism. Dry air produced in this manner is distributed throughout the midlatitudes in the form of filaments. General circulation models which fail to resolve these filaments would mix the dry air into adjoining moist air, spuriously lowering the relative humidity of the moist filaments and thus inhibiting condensation.

The extratropical mixing zone is bounded by a subtropical barrier, across which mixing is much weaker. Nonetheless, vigorous recirculating eddies of the subtropical "surf zone" episodically inject dry subtropical air and dry air of high-altitude midlatitude provenance into the tropics. Such an event can be seen in the Central Pacific on the 330K level shown in Figure 3. The injection process is very intermittent in the vertical, and (in the Central Pacific) shows up neither on the 350K nor 315K surfaces. It should thus leave its mark on suitable Central Pacific soundings in the form of a distinct mid-tropospheric layer of anomalously dry air. Aside from their direct effect on tropical humidity, the dry air incursions affect the relative humidity of the air through which precipitation is falling, and hence influence the precipitation efficiency.

Our mixing calculations were performed using a fully Lagrangian method, in which particles tagged with their initial latitude were placed initially on a 1-degree longitude by 1/2 degree latitude grid, whereafter their subsequent positions were tracked as they are advected by the observed isentropic wind field. Since no new particles are added in this calculation, a sufficiently strong and sufficiently persistent divergence of the velocity field could in principle sweep an entire large-scale area clear of particles. This situation does not occur on the 315K or 330K surface, but it is very prominent in the tropics on the 350K surface, as shown in the lower panel of Figure 4. Persistent divergence implies a mass source supplied to the isentropic surface from other surfaces; in the present case it presumably arises from the outflow of the upward branch of the Walker circulation in the Western Pacific. The ouflow on 350K is so strong that it creates a novel type of mixing barrier; it leads to a tropical "forbidden zone" into which alien air cannot readily penetrate. Aircraft or sonde observations in this region should show a fairly uniform body of air of low-altitude provenance. Longer term integrations (not shown) indicate that the air mass remains quite isolated until March 20, whereafter appreciable amounts of air from elsewhere in the tropics begin to leak in.

Finally, there is the matter of mixing *within* the tropics due to large-scale tropical transients. One is accustomed to the simple Eulerian picture of the Hadley or Walker circulation in which air goes up, outwards and down, ultimately closing the loop in a low level return current. This would lead to extremely dry air in the subsiding branches, in the absence of compensating moistening processes. However, trajectories based on time-averaged winds gives a misleading picture of the true Lagrangian histories of air parcels. Tropical transients are weak compared to extratropical transients, but the mean flow in which they are embedded is also weak, and so considerable mixing is still possible, albeit on a slower time scale than prevailing in the extratropics. The signature of lateral mixing is clear in the tracer calculations shown in Figures 3 and 4. Lateral mixing of this type may export moist air from the warm-pool atmosphere, and thus provide an important moisture source for the regions of time-mean subsidence in the Hadley and Walker circulation. This mechanism provides an alternative to the proposal of Sun and Lindzen (1993), who imply that evaporation of hydrometeors is the primary moistening mechanism throughout the tropics. Their proposal seems problematic, because it is far from clear that there are enough deep precipitating systems in the vicinity of the large-scale subsiding regions to provide the necessary moisture. If lateral intratropical mixing indeed turns out to be a significant moisture source, an additional burden would be imposed on the climate modeling enterprise: climate models would need to accurately reproduce tropical transients as well as the time-mean Hadley and Walker circulations. The representation of such features is not wholly satisfactory even in operational forecast models, and is still less so at the lower resolutions currently typical of climate-oriented GCM's.

4. Summary

We argue here that the distribution of water vapor in the atmosphere is controlled by cloud microphysical processes and by detailed advective processes which tend over time to lead to fine-scale filamentary structures. There is little if any reason to believe that microphysical processes are handled in any adequate way within either parameterized or explicit clouds within general circulation models, or that the eddy activity in such models is correctly advecting water vapor, particularly in the tropics. Since water vapor and the cloudiness which is quintessentially related to it are the principal greenhouse substances in the atmosphere, it is clear that a far better understanding of observed water substance distributions will be necessary before GCM's can be tested in a meaningful way. We believe that it is critical that suitable field programs be designed to quantify the distribution and budget of atmospheric water substance, particularly in the upper tropopsphere.

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