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Thunderstorms as Possible Micrometeorological Sink for Stratospheric Water

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Abstract

The dryness of the stratosphere has been explained, in general terms, as water condensation from the rising branch of the Hadley cell at the tropical tropopause [Brewer, 1949], but Ellsaesser [1974] suggested that the mean tropical tropopause is not cold enough to account for the observed water-vapor mixing ratios. During intense thunderstorms that, in part, penetrate the tropopause, investigators have observed (a) an increase in local stratospheric water vapor and (b) the temporary presence of air parcels substantially colder than and higher than the tropopause. These cold parcels are calculated to have extremely low water-vapor mixing ratios, and their occurrence in the stratosphere suggests a mechanism whereby the effective condensation temperature could be systematically colder than the tropopause. Ice crystals from the cloud, evaporating in the warmer stratosphere, presumably cause the observed increase in water vapor, but mixing of cold dessicated air parcels with lower stratospheric air would tend to decrease its water content. Thus there are opposing factors concerning the role of severe cumulonimbus storms on stratospheric water, and it may require detailed, microphysical analysis to see which effect is larger.

Introduction

The stratosphere has been observed to have a relatively low mixing ratio of water [Mastenbrook, 1971; Harries, 1976]. In his classic paper, Brewer [1949] explained this phenomenon by a "dynamic process ... of a circulation in which air enters the stratosphere at the equator, where it is dried by condensation, travels in the stratosphere to temperate and polar regions, and sinks into the troposphere." The dryness of the stratosphere was interpreted in terms of what has come to be called a "cold trap" mechanism. The rising branch of the Hadley-cell model of the general circulation supplied stratospheric water vapor, and the descending branches provided the sink. In recent years additional sources and sinks have been proposed.

Stanford [1974] suggested an additional sink for stratospheric water as condensation and precipitation during the Antarctic winter night.

Ackerman et al. [1977] reviewed the various measurements of stratospheric methane and gave an average profile showing 1.5 ppmv (parts per million by volume) at 12 km and about 0.1 ppmv at 45 km. This decrease of mixing ratio with altitude demonstrates in a direct manner the destruction of methane in the upper atmosphere. This destruction is primarily initiated by hydroxyl free radicals, involves a complex set of reactions with formaldehyde and carbon monoxide as intermediates, and occurs according to the overall chemical equation

$$CH_4 + 2\ O_2 = 2\ H_2O + CO_2 \quad .$$

The global rate (kg yr^{-1}) of this reaction in the stratosphere has been estimated by several investigators, for example: > 1×10^{10} [Nicolet, 1971], 5 x 10^{10} [Nicolet and Peetermans, 1973], 5.8 x 10^{10} [Crutzen, 1973], 6 to 19 x 10^{10} [Ehhalt and Schmidt, 1978]. These rates of water production in the stratosphere greatly exceed the rate of exospheric escape of hydrogen from the upper atmosphere [Liu and Donahue, 1974], so that the transfer of water from the stratosphere to the troposphere must exceed the transfer of tropospheric water to the stratosphere.

Barrett et al. [1973] proposed an additional source of stratospheric water vapor in terms of thunderstorms that penetrate the tropopause. They measured the water-vapor overburden by an infrared radiometric technique from a U-2 aircraft that flew around and over two penetrating thunderstorms in the southwestern U.S.A. In clear air away from the storm cloud, they observed water vapor mixing ratios of 2.7 ppmm (parts per million by mass) at the tropopause height and 2.4 ppmm at the peak height reached by the storm. At the mature stage of the storm they observed an increase in water vapor downwind of the storm cloud; the values were 7.2 ppmm at the tropopause height and 18.6 ppmm at the maximum cloud height. The authors concluded "that a significant fraction of thunderstorms in the plains and southwest of the U.S. do penetrate the tropopause and deposit significant amounts of water vapor in the stratosphere near and downwind of their tops."

Ellsaesser [1974] reviewed the literature on the net mass flux of air from troposphere to stratosphere in the tropical Hadley call and on the mean tropical tropopause temperatures. He stated that there appeared to be a discrepancy in the water budget with respect to the global circulation model and asked: "How does the stratosphere maintain a mean mixing ratio of around 2.5 ppmm [Mastenbrook, 1971] when all air entering the stratosphere passes through a cold trap whose characteristic mixing ratio is not less than 3.4 ppmm" (saturation at mean tropical tropopause temperature) "and when there is presumptive evidence for direct stratospheric injections of additional H_2O via CH_4 oxidation, the subtropical tropopause gaps, and cumulonimbus penetration of the tropopause?"

In part responding to Ellsaesser, Harries [1976] reviewed the observational data for stratospheric water vapor as gathered over the past 25 years. He emphasized the great experimental difficulties involved in measuring stratospheric water vapor and cautioned that all measurements should be regarded as possibly subject to large systematic errors. There are large variations of measured water vapor mixing ratios with height, latitude, time, and season. Harries noted that a small change in the temperature of the tropical tropopause "cold trap" could markedly affect the magnitude of (or even the existence of) a conflict with Brewer's general model.

We believe that the discrepancy Ellsaesser perceived in the general circulation model proposed by Brewer [1949] may be removed if one examines the physical processes occurring in the upward branch of the Hadley cell. Bates [1972] stated: "The outstanding feature of the tropical circulation is the Hadley cell.... In the ascending branch, where the latent heat of evaporation from the tropical ocean is converted into sensible heat, the primary activity takes place in motions of cumulonimbus scale." In discussing an intense cumulonimbus storm Roach [1967] stated: "It seems likely that the rate of exchange of air between the stratosphere and troposphere is small compared to the total flux through the storm." These two statements, taken together with Brewer's model, suggest that a major net input of tropospheric air into the stratosphere may be the small difference between large rising and sinking motions in cumulonimbus storms.

We have surveyed the extensive meteorological literature on this problem, and by means of direct and indirect quotations we seek to frame a hypothesis about how the rising and falling topmost towers of strong cumulonimbus storm clouds could act as a mechanism for limiting the mixing ratio of water in the lower stratosphere and upper troposphere, both in tropical and mid-latitude regions. This hypothesis might be termed the "cold finger" modification of the "cold trap" mechanism.

Classical Parcel Theory and Penetrating Thunderstorms

There is a standard, meteorological, idealized theory that gives a first-order approximation to the physical processes that occur during cumulonimbus activity [Rogers, 1976]. Although many of the assumptions of the theory are not realized under actual conditions, classical parcel theory provides simple limiting predictions against which real phenomena can be compared and classified.

Malkus [1960] used classical parcel theory to indicate that the temperature of air parcels in severe storms could be expected to be around 12.5 K colder than the environmental temperature. Briefly, a diagram such as the inset in Figure 1 is constructed [Wurtele and Finke, 1961]. The observed environmental temperature profile is indicated by the line A B C D. In our example, the temperature profile is that for Panama, 9 degrees north, 21 February 1963 [Hering, 1964]. It was assumed that the temperature at the surface increased to 304 K and that the parcel adiabatically rose to 1.2 km where condensation set in at 292 K, 0.0158 grams water per gram of air, and equivalent potential temperature of 347.46 K. It was further assumed that water precipitated from the parcel as it cooled to 273 K and ice precipitated from the parcel at lower temperatures. The calculated parcel temperature profile is A E C F of Figure 1. The parcel cools less rapidly than the environment along A E C, is warmer than the surrounding air, and being buoyant rises upward. In cumulonimbus clouds, this region is sometimes referred to as the "hot tower." At the neutral point C, the parcel has the same temperature and pressure as the surrounding air; but it has upwardly directed kinetic energy, equal to the area I. The idealized parcel overshoots the neutral point C and rises to a maximum height F, where area II equals area I. In rising to F, which is the stratosphere, the parcel has become colder, denser than, and negatively buoyant with respect to the surrounding air. The region C F is sometimes referred to as the "cold dome" or "stratospheric turret." After it comes to rest at F, the parcel spontaneously descends. For this example, the numbers written in on the expanded diagram in Figure 1 represent the saturation mixing ratios of water vapor (ppmm) with respect to ice on both the environmental and parcel profiles. At the tropopause, the saturation mixing ratio of water vapor is 1.2 ppmm in the parcel but 3.5 ppmm in the environment. At its maximum altitude, the parcel has a temperature of 170 K and a saturation mixing ratio of water vapor over ice of 0.05 ppmm. However, for actual storms, the process of entrainment (mixing of environmental air with the parcel) acts to prevent the parcel from reaching the heights predicted by the theory, and a line intermediate between C F and C D is more nearly to be expected [Roach, 1967].

Newton [1966] studied the air flow in an intense supercell storm. The following quotations are taken from Newton's article: "The cores of the updrafts are considered to be essentially unmixed, while their outer sheaths undergo strong mixing with the environment... Although the air in the core at times rises to heights predicted by parcel theory, only a limited portion of the air in the updraft can do so... At the height of maximum penetration into the stratosphere, a draft parcel, being around 30° colder than the environment, is subjected to powerful downward acceleration. Its subsequent behavior depends on the degree of entrainment ... the downdraft would lose all its momentum at 10.5 km. Being buoyant at that level, this air would accelerate upwards again, and under the influence of further entrainment undergo a damped oscillation ... eventually remaining below the tropopause but in the upper troposphere."

Roach [1967] analyzed photometric and radiometric records of aircraft flying above and around a series of severe thunderstorms in Oklahoma in 1962. Storms were observed which were 10 K colder at the top of the cloud than in the surrounding air. Because of cloud particles, the radiometer could penetrate only 100 to 200 meters into the cloud, leaving open the possibility that the central core might be even colder. The observed maximum cloud heights were often close to those expected from idealized parcel theory. The cold dome was observed to contain a large amount of condensed water (presumably ice), which was lifted to this great height by the strong updraft.

In studying an intense cumulonimbus cloud at 16.4 km above Brownsville, Texas, Fujita's [1974] aircraft entered a relatively clear layer of air above the mean tropopause with a temperature of 189 K, which he characterized as a storm-produced "meso-high aloft." Schereschewsky [1977] observed cases where stratospheric turrets above intense cumulonimbus storms had temperatures about 10 K below the environmental temperature.

Suggested Mechanisms

Intense thunderstorms that penetrate the stratosphere are exceedingly more complicated than the representation of Figure 1, but the regions in Figure 1 still supply useful nomenclature. Roach [1967], Fujita [1974], and Schereschewsky [1977] have observed cases where the stratospheric towers of intense thunderstorms were 10 K or more colder than the environment. The temporary existence of large parcels of air above and colder than the tropopause suggests several mechanisms whereby the overshooting cumulonimbus clouds could provide a "cold trap" effectively colder than the mean tropopause.

Schereschewsky [1977] proposed that stratospheric water vapor might condense on the surface of the cold dome.

An extension of the direct contact mechanism is that stratospheric air lifted, say 2 ± 1 km, by the rising dome would cool 18 ± 9 K by adiabatic expansion and form a pileus cloud in the stratosphere, which in part might precipitate into the cold dome before it retreated back into the troposphere. Since the life-time of a given stratospheric turret is only a matter of a few minutes [Roach, 1967; Fujita, 1974], one cannot expect the ice crystals in such a stratospheric cloud to grow very large or to fall very fast [Stanford, 1974]. Even so, this mechanism may be worthy of a quantitative study.

The primary mechanism offered here for thunderstorms acting as a net removal mechanism for water from the stratosphere concerns the processes that occur in overshooting cold domes. Whether thunderstorms act as a source or sink for stratospheric water strongly depends on what happens to the ice particles in the cold dome. At the point of maximum height in a thunderstorm (compare F in Figure 1), it is colder than the surrounding air, the mixing ratio of water vapor is extremely low, but the parcel contains a large burden of ice particles. These particles formed over a relatively long period of time as the cloud rose through the upper troposphere, and presumably they are large. As the cold dome sinks from F toward C in Figure 1, it will heat by adiabatic compression, tend to become unsaturated, and start to evaporate its ice particles. However, the ice particles would fall relative to the air parcel and would not necessarily remain in equilibrium with respect to water vapor. Ice particles have been shown experimentally [Braham and Spyers-Durran, 1967] and theoretically [Hall and Pruppacher, 1976] to survive in unsaturated air over significant distances of fall (order of magnitude of kilometers, depending on initial particle size, degree of unsaturation, temperature). This precipitation of ice particles constitutes a non-equilibrium, physical separation of solid water from extremely dry air.

Schereschewsky [1977] investigated the outflow in the clouds that overshoot the neutral buoyancy point, C of Figure 1. He stated that there is relatively little mixing between the rising cold dome and the surrounding stratosphere. Such mixing as does occur with the rising ice-rich cold dome would be expected to add ice particles which would evaporate in the stratosphere. Barrett et al. [1974] observed high stratospheric water vapor mixing ratios during the mature stage of a storm.

Typically, it requires about 4 minutes for a cold turret to break through the buoyancy neutral point, rise to maximum height, and sink back through the neutral point [Roach, 1967; Fujita, 1974]. For a wide range of intermediate sizes, ice particles have a fall velocity of about 50 cm sec^{-1} [Rogers, 1976, p 127]. In the approximately 120 seconds it takes a penetrating turret to sink from its maximum height back to its neutral point, ice particles would fall about 60 meters relative to the air parcel, tending to dessicate the topmost layer of the retreating parcel. Only a small fraction of the air that circulates through the stratospheric dome remains in the stratosphere [Roach, 1967]. The region between the top of the penetrating turret and the stratosphere above has a steep temperature inversion (Figure 1), which inhibits mixing; but the inversion decreases to zero as the sinking parcel passes through the buoyancy neutral point, leading to relatively fast mixing between the stratosphere and the penetrating parcel as it exits the stratosphere [Schereschewsky, 1977]. Thus it is reasonable to postulate systematically favored mixing between the stratosphere and the top surface of the parcel, which over the micrometeorological scale of a few tens of meters had been dried by particle precipitation relative to its contained cold dry air. Any oscillations of the parcel, though damped [Newton, 1966], would tend to prolong the period of particle precipitation.

In general, lower stratospheric and upper tropospheric air would have its water content lowered as it mixed with the dessicated portion of the sinking parcel. As the cold dome leaves the stratosphere, its vertical motions are damped by entrainment with the environment, and eventually it comes to rest typically in the upper troposphere [Newton, 1966]. In tropical regions subsequent upward motions of the Hadley cell (whether generalized upwelling driven by large scale convergence or by cumulus activity on subsequent days or both) would tend to propagate into the stratosphere this drying action of the upper troposphere.

To the extent that severe thunderstorms occur at mid-latitudes, this mechanism of drying the lower stratosphere and upper troposphere should act to supplement the primary drying effect that occurs in the tropical zone. For example, the saturation mixing ratio of water vapor over ice would be about 1.6 ppmm in Fujita's [1974] "meso-high aloft" (189 K, 16.4 km, about 104 mb).

Order of Magnitude Calculations

We cannot offer calculations that confirm the speculations presented here, but order-of-magnitude calculations are appropriate and interesting.

If 3.4 ppmm is taken to be the saturation water vapor mixing ratio at the mean tropical tropopause of about 194 K, then the dynamical processes discussed here need to reduce the temperature of the air that enters the stratosphere by only 2 K below the mean tropopause temperature to provide a saturation mixing ratio of about 2.5 ppmm. This temperature difference is small compared to the difference between the tropopause and the minimum temperature reached by a penetrating cumulonimbus cold dome.

Similarly, one may ask what must be the difference in water mixing ratio between the input and output of the Hadley cell in order for it to remove the water vapor produced by methane oxidation, for example. Ellsaesser [1974] reviewed 12 estimates from the literature of the

Hadley cell air flux, the values quoted ranged from 0.76×10^{17} to 9.2×10^{17} kg yr^{-1}, and the average of the 12 values was 3.8×10^{17} kg yr^{-1}. We adopt 6×10^{10} kg yr^{-1} as an estimate of the mass of water produced by methane oxidation. If 3.8×10^{17} kg air yr^{-1} removes 6×10^{10} kg water yr^{-1} from the stratosphere, the mixing ratio of water entering the stratosphere must be 0.16 ppmm less than that which leaves the stratosphere, which is only 6 percent less than a typical stratospheric mixing ratio of 2.5 ppmm. At 194 K and 110 mb on the environmental curve of Figure 1, a change of 0.16 ppmm in water vapor mixing ratio over ice corresponds to less than 0.4 K. Thus a very small fraction of the temperature difference between the tropopause and F in Figure 1 when applied to the entire Hadley cell flux is sufficient to remove the water produced by methane oxidation. An additional temperature decrease for the air that stayed in the stratosphere would be required to balance the input of ice crystals to the stratosphere [compare, Barrett et al., 1974].

These considerations concern the field of cloud physics, and we hope that cloud physicists will take an increased interest in this problem.

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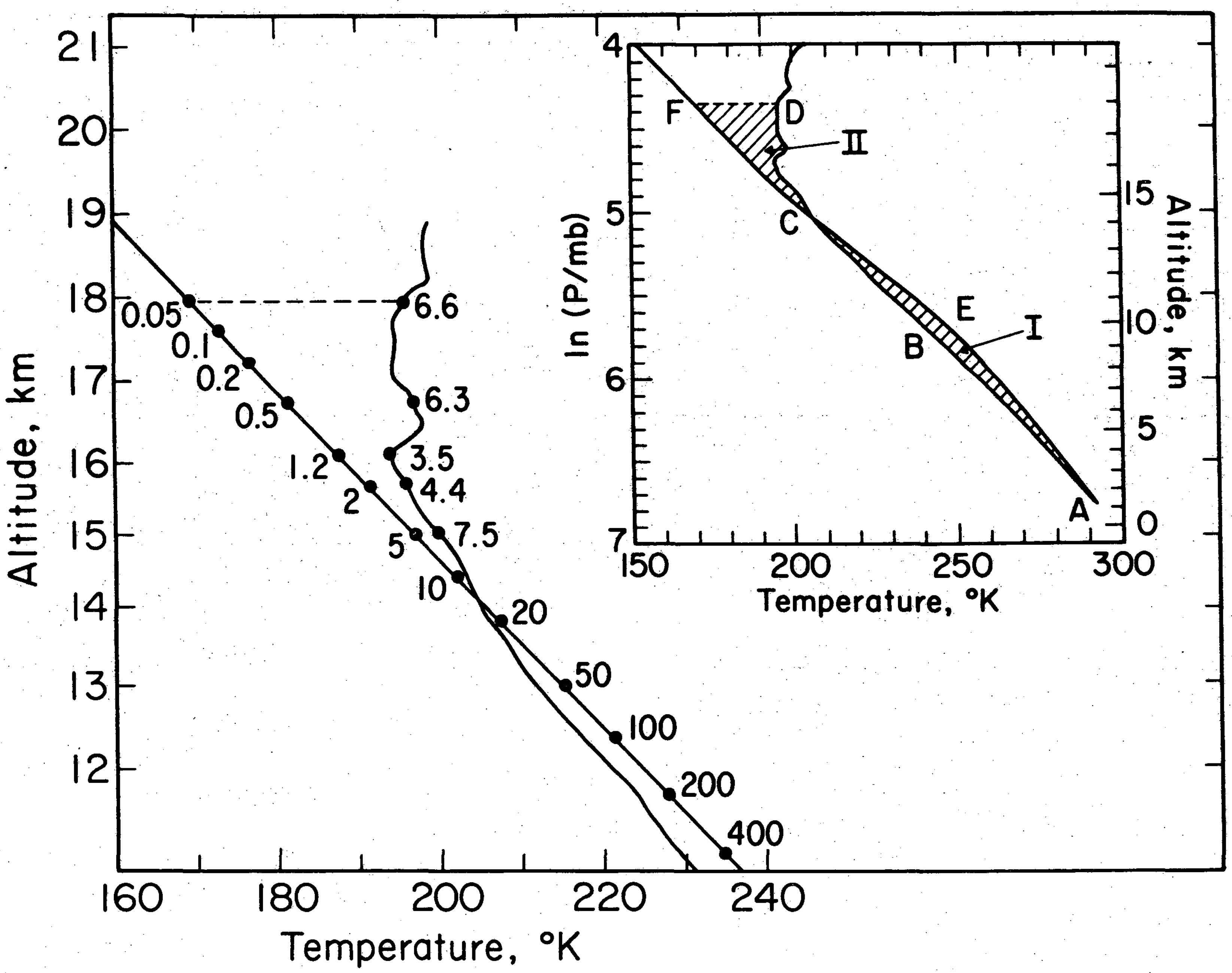
Figure Caption

Figure 1. Inset: Pseudo-adiabatic lapse rate and an environmental profile for Panama, 9°N, 21 February 1963 [Hering, 1964]. Surface temperature assumed to rise to 304 K. Absolute temperature T and natural logarithm of pressure in mb. Expanded scale: Portion of curve above plotted as absolute temperature and altitude in km. Numbers indicate saturation water vapor mixing ratios with respect to ice on the environmental and calculated "parcel-theory" profiles ppmm.



Altitude, km
21
20
19
18
17
16
15
14
13
12
160
180
200
220
240
Temperature, °K
0.05
0.1
0.2
0.5
1.2
2
5
10
20
50
100
200
400
6.6
6.3
3.5
4.4
7.5
ln (P/mb)
4
5
6
7
150
200
250
300
Temperature, °K
Altitude, km
15
10
5
0
F
D
II
C
E
I
B
A

Figure 1

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