

Radiative effects of changing atmospheric water vapour

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ABSTRACT

We have used the radiative computation scheme of Dopplnick to investigate the purely radiative effects resulting from variations in the total water vapour content in a clear atmospheric column. We consider a given water vapour profile uniformly scaled in the vertical by a factor over the range 0.2 to 2.4. For the case of varying water vapour but fixed temperature profile, we find that additional vapour produces greater cooling of the column as a whole while supplying energy to the surface and reducing the net energy flux out of the top of the column. For the same changes in vapour content, but atmospheric temperature varying so as to maintain constant relative humidity, we find even greater relative cooling of the atmospheric column as a whole, but less effective energy supply to the surface and an actual increase in the net flux upward at the top of the column.

Radiative heating rates were calculated for temperature and moisture profiles measured over the Arabian Sea during the 1979 summer MONEX and show radiative cooling of the lower layer and energy flux to the surface associated with the presence of water vapour in the lower layers of the atmosphere which may be important in the evolution of this vapour-rich surface layer preceding the monsoon.

1. Introduction

The importance of the radiative rôle of atmospheric water vapour in maintaining the thermal energy balance of the earth/atmosphere system is well established and has been discussed by many workers: Goody (1949) and Yamamoto (1953) pointed out that efficient radiative cooling by the water vapour bands tends to result in a high tropopause at latitudes in which the upper air is relatively humid. Manabe and Möller (1961) further emphasized this point and explored the dependence of radiative equilibrium temperatures upon the vertical distribution of tropospheric moisture. Subsequently Manabe and Wetherald (1967) investigated radiative-convective equilibrium in an atmosphere with specified relative humidity distributions, and studied the effect of doubling CO₂ concentrations in such a model. They found that higher tropospheric relative humidity resulted in a warmer tropospheric equilibrium temperature, and that stratospheric equilibrium temperature was quite insensitive to tropospheric relative humidity.

In this paper, we concern ourselves with the direct radiative effects arising from an increase or decrease of the total water vapour content of the atmospheric column. We present the results of computations of radiative heating rates and fluxes in the atmosphere, using the radiative computation scheme developed by Dopplnick (1970) after Rodgers (1964, 1967), which has been fully documented by Hoffman (1981). For clarity we have restricted the discussion to the case of a cloud-free column. We have chosen Dopplnick's (1974) zonally-averaged water vapour, temperature, and ozone profiles for January at the equator as our point of reference (Table 1). We have constructed vertical profiles containing more-or-less water vapour, simply by multiplying the mass mixing ratio at all levels below 150 mb, in this reference profile, by a specific scaling factor. At 100 mb and all levels above, the water vapour mass mixing ratio was held constant.

This representation of variable atmospheric water vapour content has been chosen for ease of analysis and interpretation. We do not mean to

Table 1. *Input parameters for the atmospheric profile at the equator for January*

Atmospheric data press (mb)	Lat = 0 Month = 1		
	Temp (K)	H ₂ O (g kg ⁻¹)	O ₃ (mg kg ⁻¹)
1000.0	298.9	15.400	0.020
850.0	290.4	10.600	0.020
700.0	282.5	6.100	0.030
600.0	275.6	3.902	0.034
500.0	267.5	2.300	0.040
400.0	257.1	1.000	0.040
300.0	241.6	0.300	0.050
200.0	219.5	0.020	0.060
150.0	206.2	0.003	0.090
100.0	192.8	0.002	0.140
70.0	196.8	0.002	0.800
50.0	205.2	0.002	2.600
30.0	214.5	0.002	6.900
20.0	222.0	0.002	11.500
15.0	227.6	0.002	13.000
10.0	235.5	0.002	14.000
5.0	249.0	0.002	14.000
2.0	266.0	0.002	6.300
0.5	280.0	0.002	3.100
0.1	233.0	0.002	1.400

imply that in the real atmosphere water vapour changes in this manner.

2. Cooling within spectral divisions

In our model, the thermal H₂O spectrum is divided into 20 subdivisions, two of which include intervals of overlapping with the O₃ 9.6 μ m and CO₂ 15 μ m bands. Water vapour dimer absorption is included within the region 720 to 1200 cm⁻¹ following Lee's (1973) formulation. For a detailed account of the frequency integration and data sources in the model, the reader is referred to Hoffman's (1981) report.

In Fig. 1 the cooling rates in the January equatorial profile due to thermal flux divergence within each of these intervals are presented. The contribution of the various spectral regions to cooling at different levels within the atmosphere is immediately apparent and can fairly readily be understood.

Rodgers and Walshaw (1966) have demonstrated that, in general, the radiative cooling rate at

any given level in the troposphere is dominated by the contribution due to direct radiative losses from that level to space (which is at ~ 0 K). This term is normally far in excess of those due to radiative exchanges with the ground or other atmospheric levels. One particularly relevant example, however, where these workers found the "cooling to space" approximation to be less applicable is the case of a strong low-level temperature inversion where the latter terms contribute more significantly to the total cooling rate.

Bearing this in mind, let us consider the "cooling to space" term. It is well known (Houghton and Taylor, 1973) that the thermal energy reaching space from any given level in the atmosphere is proportional to the local source function multiplied by the gradient of transmission upwards from that level to space. In the troposphere, the source function for all the specified spectral intervals decreases with temperature and therefore, height. The transmission gradient depends on the vertical distribution of water vapour and the H₂O band absorption coefficients, which in turn depend on line shape and strength. However, it can be shown for realistic line shapes and band models (Houghton and Taylor, 1973) that the transmission gradient reaches a maximum value, within a given spectral interval, at the level of unit optical depth in the atmosphere. For most intervals, this term is more rapidly varying than the source function, so that the level of most efficient radiation to space, and therefore local radiative cooling, occurs at or close to unit optical depth within any given spectral interval.

Note that optical depth is defined with reference to the top of the atmosphere: unit optical depth for any band is that level from which the transmission upwards to space is e^{-1} . For the purposes of the discussion which follows, we have identified the height corresponding to unit optical depth in the column by examining the calculated values of the transmission to space from each level for each of the individual bands shown in Fig. 1.

Referring to Fig. 1, we see that the rotation band of water vapour at wavelengths greater than 20 μ m, in which optical depth unity occurs between 200–400 mb, is largely responsible for cooling of the upper troposphere. The rotation-vibration band centred around 6.3 μ m also contributes to cooling in the upper levels. In both cases, the maximum in the transmission gradient at unit optical depth

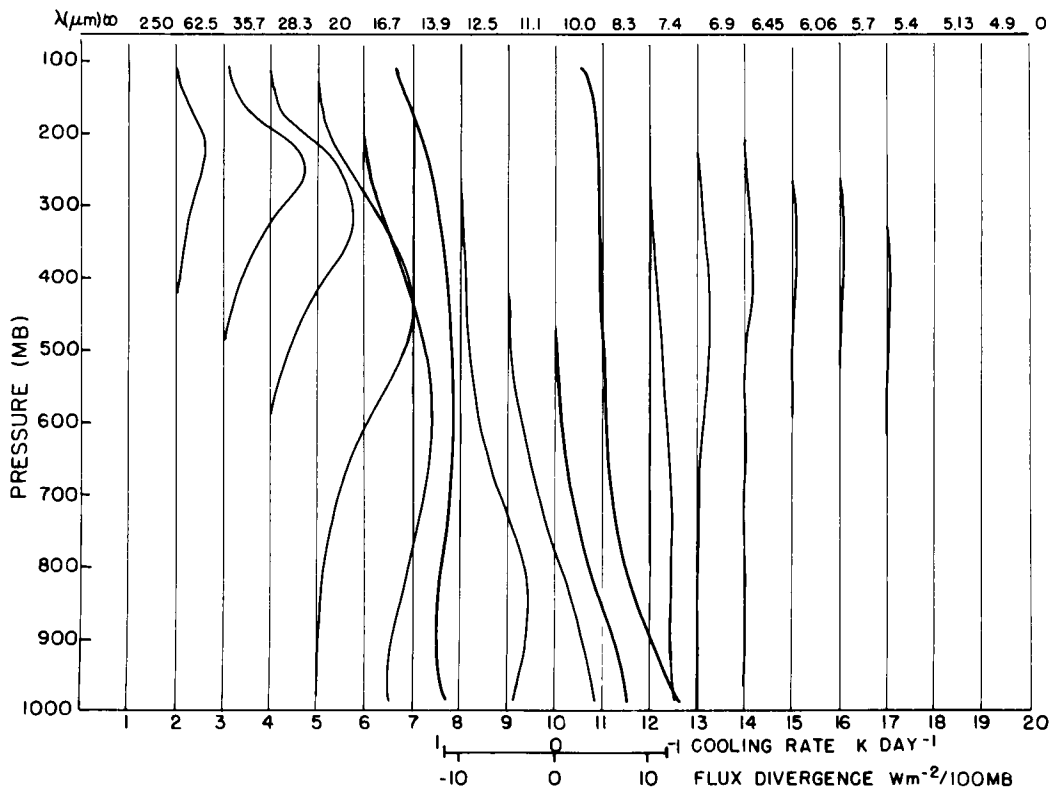


Fig. 1. Radiative cooling rates within each spectral interval for the January equatorial profile. Numbers 1–20 on the lower abscissa refer to spectral intervals defined by wavelengths on the top abscissa; for example, band 3 refers to wavelength interval from 62.5 to 35.7 μm .

occurs high in the troposphere because of the abundance of water vapour and the strength of these absorption bands. However, the former spectral band encompasses a large fraction of the spectrally integrated black-body source function and therefore contributes significantly more to cooling at these levels than does the 6.3 μm band.

In the so-called "atmospheric window" from 8–12 μm , absorption by the water molecule is weak, but there is an additional contribution due to dimer absorption. Thus, unit optical depth occurs very close to the ground within band 9 and is not in fact realized with bands 10 and 11. Thus these intervals, which encompass a spectral region close to the peak of the local black-body source function, contribute significantly to the radiative cooling of the lower troposphere. We note, however (Rodgers and Walshaw, 1966), that in the presence of a temperature inversion at the low levels, radiative

exchanges with the ground and other levels may contribute significantly to the net cooling rate. The net radiative cooling due to all bands, and including short-wave heating is shown in Fig. 2. The vertical distribution of the heating rates may be roughly categorized into four distinct regions: 1000–750 mb: very efficient cooling to space within the interval 7–17 μm incorporating the atmospheric window; 450–200 mb: very efficient cooling due to the water vapour rotational band; 750–450 mb: weaker, but significant cooling due to both these intervals resulting in a minimum in the cooling rate; 200–100 mb: water vapour concentration decreasing rapidly and cooling approaching zero.

The band (number 7) centred around 15 μm incorporates regions of overlap of the H_2O band and the CO_2 15 μm band as well as pure H_2O . In the absence of CO_2 , optical depth unity would occur near 850 mb, and maximum cooling would

be close to that level, whereas with CO_2 uniformly mixed at 320 p.p.m., optical depth unity occurs near 300 mb, radiative cooling is effective throughout the entire column up to ~ 250 mb (Fig. 1) and the analysis in terms only of cooling of H_2O to space is clearly no longer appropriate.

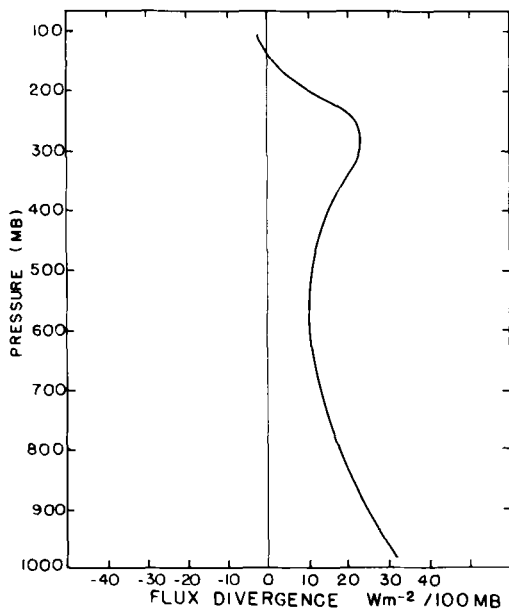


Fig. 2. Total radiative cooling rate (including short-wave heating) for January equatorial profile.

3. Effect of increasing water vapour

We are now in a position to examine the effect within each of these intervals, and therefore at different heights, when the total water vapour content of the troposphere is incremented.

In Fig. 3 we show, in a similar manner to Fig. 1, the incremental cooling rates resulting from an increase by 12.6% of the water vapour content at all levels below 150 mb while maintaining a constant temperature profile. As an example for discussion, consider the results in band 5 ($20\text{--}28.3\ \mu\text{m}$) separately. One sees that the increased atmospheric water vapour has produced more efficient cooling of the column between 450 and 200 mb, but has produced less efficient cooling (relative warming) between 450 and 900 mb. The radiative consequences of increasing the optical thickness of the atmosphere (for example, by increasing the CO_2 concentration) are commonly interpreted in terms of the effect of raising the height at which unit optical depth occurs, and thereby (in the troposphere) reducing the effective temperature at which the spectral band in question radiates to space (e.g. Hansen et al., 1981). This reduces the net radiative loss of energy from the atmosphere to space and ultimately is thought to produce a temperature increase in the column such that the overall radiative energy balance of the earth/atmosphere system is restored.

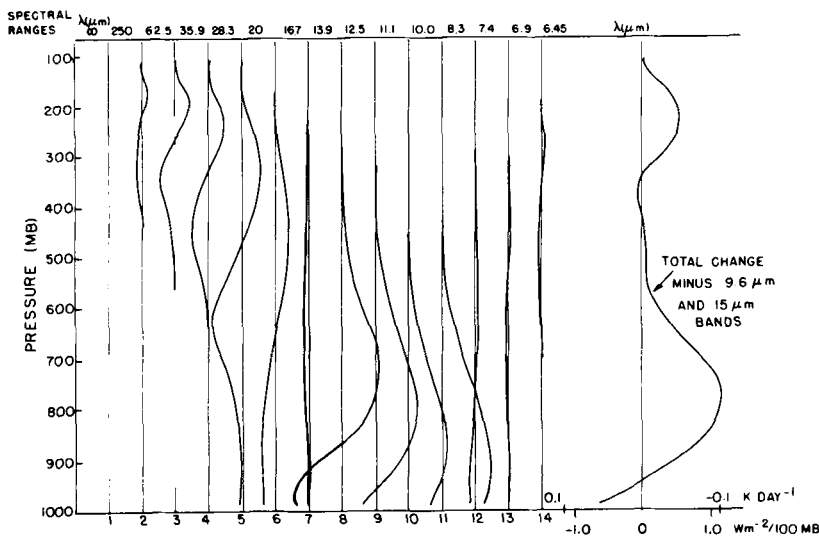


Fig. 3. Incremental radiative cooling rates resulting from 12.6% increase of water vapour concentration below 150 mb. Abscissa nomenclature as for Fig. 1. Scale on right-hand side applies to all bands.

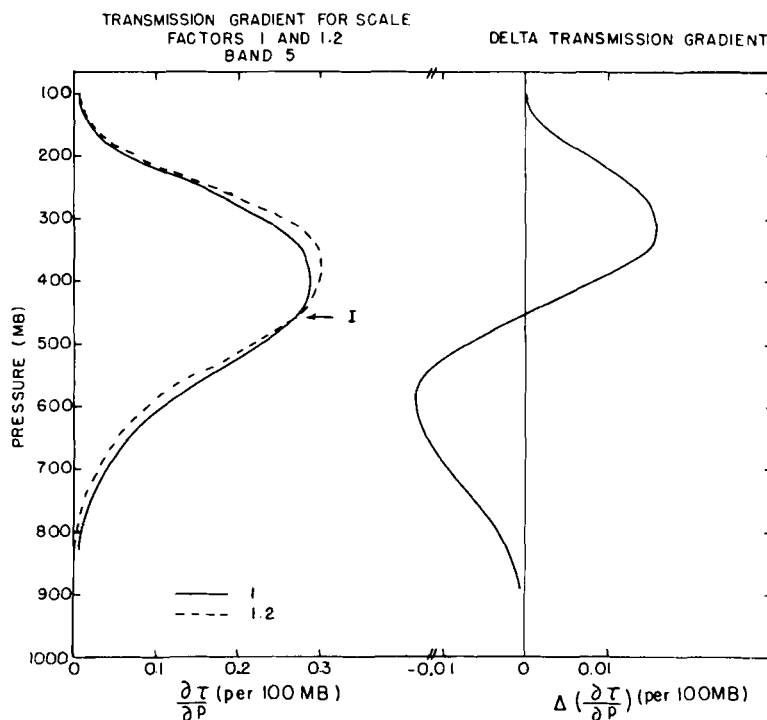


Fig. 4. Transmission gradients for scale factors 1 and 1.2 and Δ (transmission gradient) in band 5.

From the previous discussion, we would expect the local changes in radiative heating to be primarily determined by modifications to the transmission gradient, since in the present example we have left atmospheric temperature and therefore the source function unchanged. In Fig. 4 we show the transmission gradient $\partial\tau/\partial p$ for water vapour profiles corresponding to scaling factors of 1 and 1.2, as well as the difference between the transmission gradients in each profile ($\Delta(\partial\tau/\partial p)$) for band 5. Notice that these results correspond to a larger increase in water vapour content (20%) than those shown in Fig. 3 (12.6%). The transmission gradients ($\partial\tau/\partial p$) are estimates obtained by differencing the transmissions calculated between each specified level in the input profile and the top of the column.

We find, as expected, that in the more optically thick column, the maximum of the transmission gradient occurs slightly higher in the atmosphere. Consequently, the two graphs of transmission gradients intersect in a common point at level I (Fig. 4). When the water vapour content of the

troposphere is incremented by the specified amount, the transmission gradient above this level is increased and below it is decreased. Moreover, the graph of Δ (transmission gradient) exhibits a maximum at 300 mb and a minimum near 600 mb, as well as this zero point near 450 mb.

If we extend these results in order to estimate the form of the changes in the cooling to space term associated with the presence of additional water vapour in the column, we would anticipate relative heating (decreased transmission gradient) below 450 mb, with a maximum near 600 mb, passing through a null effect near 450 mb to radiative cooling of the upper troposphere with maximum effect near 300 mb.

This accounts well for what we see in band 5 and the other bands (compare Figs. 1, 3 and 4), when the water vapour concentration is increased by 12.6%. Notice that the maximum, zero and minimum of the relative cooling shown in Fig. 4, occur ~20 mb higher than the corresponding points for that band in Fig. 3. This is simply because the increase of water vapour in Fig. 4 is

20%; thus, unit optical depth rises to an even higher level in the column than it does when the water vapour is increased by 12.6%.

For all bands we expect to find increased cooling to space in the column above the level of unit optical depth and decreased cooling to space below that level; these effects should be significant over the height interval within which the band in question already contributes significantly to local radiative cooling. Thus it is no surprise to see (Fig. 3) that in the far infrared, the water vapour rotational band contributes relative cooling to the upper troposphere but relative heating to the middle troposphere. Progressing towards shorter wavelengths into the vicinity of the 8–12 μm atmospheric window, unit optical depth occurs lower and lower in the column and consequently, the relative cooling is effective throughout the middle and upper troposphere, while the lower troposphere experiences relative heating. Moving out of the window region into the 6.3 μm band, the cooling effect is again felt in the upper air, but because the source function is of small magnitude, the contribution of this band to relative cooling/heating is slight.

Within the window, in bands 8, 9, 10, 11, the effects are seen primarily below 500 mb with relative cooling of significant magnitude above 900 mb in all 4 bands and relative warming of the near surface air in bands 8 and 9.

The combined effect of the relative cooling/heating in all of these bands is shown at the right-hand side of Fig. 3. This includes the relative heating due to increased NIR absorption when atmospheric water vapour is increased. The NIR contribution is far out-weighted by the long-wave effects.

It is interesting to note that the *purely radiative* effect of a more humid atmosphere is to produce cooling of the upper troposphere 350–100 mb, and cooling for 550–950 mb. In the middle troposphere for 350–550 mb, the relative warming effect in the far infrared regions of the H_2O rotational band is almost exactly balanced by cooling in the shorter wavelengths of this band and within the 8–12 μm window region. Cooling of the lower troposphere is dominated by the interval 8–17 μm . There is a net warming of the near surface air below 950 mb which is mainly contributed to by the intervals 16.7–20, 12.5–13.9, 11.1–12.5 and 7.4–8.3 μm .

In summary, we find that the addition of more

water vapour to the January equatorial profile produces relative cooling of the upper and lower troposphere, relative heating of the surface air and has little effect on the mid-tropospheric cooling rates. These are direct radiative effects alone; we wish to distinguish them clearly from any convective heating which may subsequently result.

Our results suggest, in general, that for a given band, the impact of increasing the total absorbant amount depends upon the strengths of absorption within that band before the increase. Specifically, we expect relative heating below the level of unit optical depth and relative cooling above.

By extension, we then expect that the total effect of all spectral intervals will depend on the integrated optical depth of the atmosphere, specifically upon the absolute water vapour content of the column.

4. Total cooling rates

We continued this study by calculating the total heating rates in an atmosphere with fixed temperature profile but total tropospheric moisture content varying in the manner previously specified. Naturally this means that for most of the profiles corresponding to a scaling factor in excess of ~ 1.2 , much of the lower troposphere is supersaturated. However, our intention is to isolate radiative effects related only to the presence of additional absorber amount from those due to any concomitant temperature changes. The latter effect we will deal with later.

In Fig. 5 we show contours of the net tropospheric cooling rates for scaling factors in the range 0.2–2.2. These calculations were made with CO_2 content set at 300 p.p.m.

The water vapour mixing ratio at the surface in the January equatorial profile is 15.4 g kg^{-1} which corresponds to a fairly humid atmosphere (73% RH at 25°C). Average water vapour mass mixing ratio varies from about 15 g kg^{-1} at the equator to $\sim 2 \text{ g kg}^{-1}$ near 80° (Newell et al., 1972, Fig. 5.4); we therefore expect that as an approximation to most physically realistic situations, scaling factors within the range 0.2–1.0 would be appropriate. Nevertheless, higher levels of humidity do occur in the Earth's atmosphere and therefore are not excluded from this discussion.

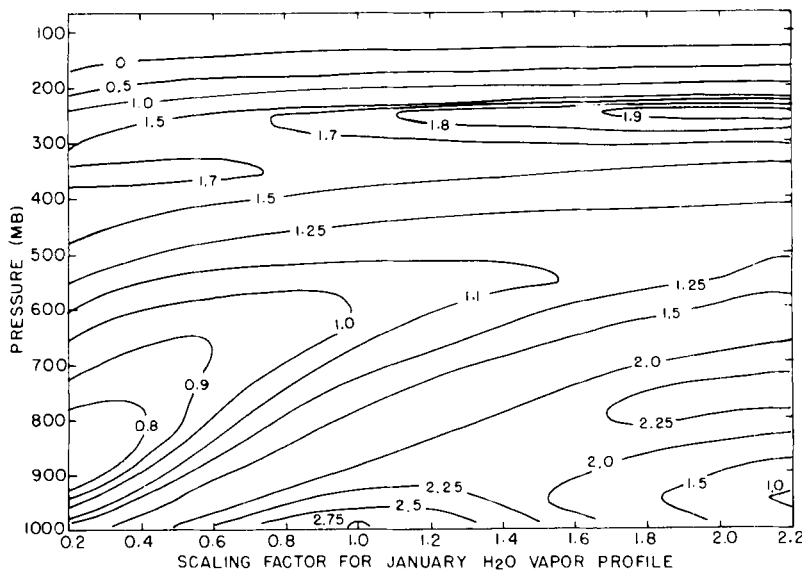


Fig. 5. Total tropospheric cooling rates as water vapour concentration is scaled with fixed temperature profile. Units: $^{\circ}\text{K day}^{-1}$.

The general structure of Fig. 5 appears reasonable on the basis of our previous results. For all scaling factors, we see a maximum in cooling from 450 mb to 120 mb centred roughly along 275 mb. This is associated with cooling to space within the far infrared H_2O rotational band. As the water vapour concentration increases, this band cools to space more efficiently and the level of maximum cooling rises from ~ 350 mb for a scaling factor of 0.2, to 250 mb for a scaling factor of 2.0. Concurrently, the level of transition from net cooling to warming just below the tropopause is extended slightly higher as water vapour is increased. This is evidenced in Fig. 5 by the very gradual sloping upwards of the zero heating rate contour. The height to which the level of most efficient cooling in these bands may rise is, however, limited by the rapid fall-off of water vapour concentration in the upper troposphere, implicit in our constraint that the stratospheric H_2O concentration remain fixed at $2 \mu\text{g g}^{-1}$. We see that the mid-tropospheric minimum of cooling from 450–750 mb which we identified in Fig. 2, for scale factor 1, actually extends from ~ 550 to 950 mb for scale factor 0.2 (Fig. 5), but narrows considerably in depth and magnitude as water vapour is increased. For a scale factor of 2.2, this minimum of cooling is localized to the layer ~ 400 –600 mb. The reason for this is

that as the total water vapour content increases and the atmospheric window becomes “closed”, the level of optical depth unity and thus the level of maximum cooling to space in that interval rises rapidly within the column. In contrast, the level of effective cooling of the upper troposphere by the long-wave rotational band rises more slowly; thus, the intermediate region, where cooling due to both bands is weak, becomes progressively less deep as water vapour increases in the column.

The importance of this additional cooling by the 7 – $17 \mu\text{m}$ region in the lower layers is most evident for scaling factors 0.2–1.0. At all levels below 800 mb, the local cooling increases as the water vapour content is increased. This accounts for a change in cooling rate from 1.24 K per day to 2.76 K per day at 1000 mb and from 0.71 K per day to 1.67 K per day at 800 mb over the range of scaling factors 0.2–1.0. Fig. 5 shows that in the prescribed set of water vapour profiles, most efficient cooling of the near surface air is attained at scaling factor 1. Given that this represents quite a humid atmosphere, we infer that in less humid and perhaps more typical atmospheric conditions, an increase of humidity will produce additional (purely) *radiative* cooling of the lower layers. Again we stress that this is distinct from any attendant convective heating.

It is interesting to note that as the scaling factor is increased beyond a value of 1, the near surface air is actually less effectively cooled by radiation: there is a relative warming. Thus for a scaling factor of 1.8, the cooling rate at 1000 mb is reduced to 1.75 K per day. The upper air, however, continues to be cooled and has reached 2.24 K per day at 800 mb for this scaling factor.

The reason for the relative heating of the lower layers is that, as water vapour increases and unit optical depth within the window is realized at higher and higher levels in the atmosphere, the magnitude of the transmission gradient within the near surface air becomes progressively smaller (see Fig. 4). Therefore, the contribution of the cooling to space term to the total radiative cooling rate is concomitantly reduced at the lower levels.

5. Experiments with fixed relative humidity

All the results presented so far were for a fixed temperature profile. Therefore, the column is represented as being supersaturated for most scaling factors greater than 1, and for small scaling factors, the relative humidity of the atmosphere is unrealistically low. It has been claimed (Möller, 1963) that the atmosphere tends to restore a certain climatological distribution of relative humidity in response to a change in temperature. Thus, we performed the same calculations but this time modifying the atmospheric temperature profiles such that a fixed profile of relative humidity was maintained when the water vapour content was scaled as specified. The resultant temperature profiles are shown in Fig. 6 for scaling factors 0.2–2.2. The temperature of the ground was in all cases set equal to that of the air in the lowest level. The normal temperature profile for the unscaled January equatorial profile is identified in Fig. 6 by the crosses. Notice that the temperature increments required to maintain constant relative humidity become progressively smaller as the atmospheric temperature is increased. This is because the water vapour saturation pressure rises exponentially with temperature. One consequence of this which we can anticipate is that the influence of changing temperatures upon the radiative heating rates and fluxes will be proportionally smaller in the case of a fixed increment of water

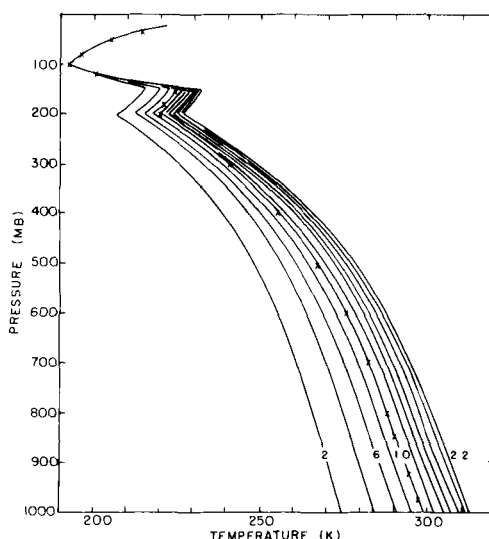


Fig. 6. Temperature profiles corresponding to various scaling factors with constant relative humidity profile. The crosses represent the unscaled January equatorial temperature profile.

vapour content in a humid rather than a relatively dry atmosphere.

The heating rates computed with varying water vapour and fixed relative humidity are shown in Fig. 7. The general form of these results is similar to those obtained for constant temperature (Fig. 5). Additional cooling due to more efficient radiation to space by the H_2O long-wave rotational band is seen above 400 mb and increased cooling of the lower layers by the shorter wavelengths is also evident. The major difference between these results and the results with a fixed temperature profile is that cooling rates here are generally greater for high scaling factors and smaller for low scaling factors. This is an obvious consequence of the variations in temperature profiles shown in Fig. 6. One significant aspect of this is that the additional flux divergence produced by temperature increases in the lower layers delays the occurrence of maximum cooling in the surface layers until a scaling factor of ~ 1.2 is reached. Moreover, the onset of the relative warming of these layers due to radiation occurs less rapidly, so that by a scaling factor of 2.0 the cooling rate of the lowest layer is 2.18 K per day, as opposed to 1.44 K per day in the constant temperature case.

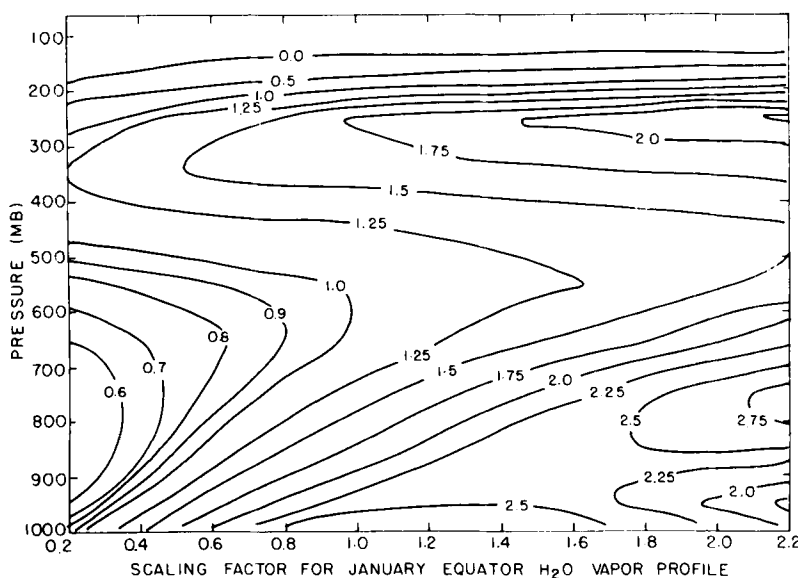


Fig. 7. Total tropospheric cooling rates as water vapour concentration is scaled while maintaining constant relative humidity profile. Units: $^{\circ}\text{K day}^{-1}$.

6. Radiative fluxes

As yet, we have not considered the impact of these changes upon the total fluxes of radiation both into the surface and out of the top of the atmosphere, nor have we explicitly considered the total radiative gain or loss of energy by the atmosphere as a whole. These factors are of fundamental importance to the maintenance of the total energy balance at the earth's surface, in the atmosphere itself, and for the combined earth-atmosphere system; therefore we consider them separately from the heating rates.

Fig. 8 shows the changes in total flux at various levels in the atmosphere when the water vapour content of the atmosphere is increased as specified. For graphical convenience, we have chosen the fluxes in the atmosphere represented by scaling factor 0.2 as a point of reference. Flux changes at a given level are represented as departures from the flux at that level calculated for the atmosphere corresponding to scale factor 0.2. The sign convention is chosen such that positive ΔF represents an increase in the net flux downwards relative to that obtained with this 0.2 scale factor profile. Flux changes are shown both for fixed atmospheric temperature profile and for fixed relative humidity. The absolute fluxes at each

level in the two reference profiles (scale factor 0.2) therefore differ by an amount determined only by the temperature differences of these two profiles (Fig. 6). The reader should note that since the models for fixed relative humidity and fixed temperature profile share a common water vapour and temperature profile at scale factor 1, the absolute fluxes at any selected level in the atmosphere are identical for the two models for scale factor 1. Consider firstly the case of fixed atmospheric temperature when the scale factor varies from 0.2 to 1.0, a five-fold variation in total vapour content which we suggest may encompass plausibly realistic atmospheric profiles. We see that at the surface (1000 mb), the tropopause (100 mb) and at the top of the atmosphere (0 mb), the net fluxes downwards increase by 74, 36 and 35 Wm^{-2} respectively. Thus the surface gains energy and the atmosphere alone actually loses energy to the ground because the net downward flux out of its lower boundary increases more rapidly than does the net downward flux at its upper boundary. These fluxes continue to increase over the range of scale factors 1.0–2.4, but we see that the rate of increase of the downward flux at 1000 mb begins to fall off at higher scale factors, while the flux at the top of the column increases at a rate only slightly less than over the scale factor range 0.2–1.0. Thus, in

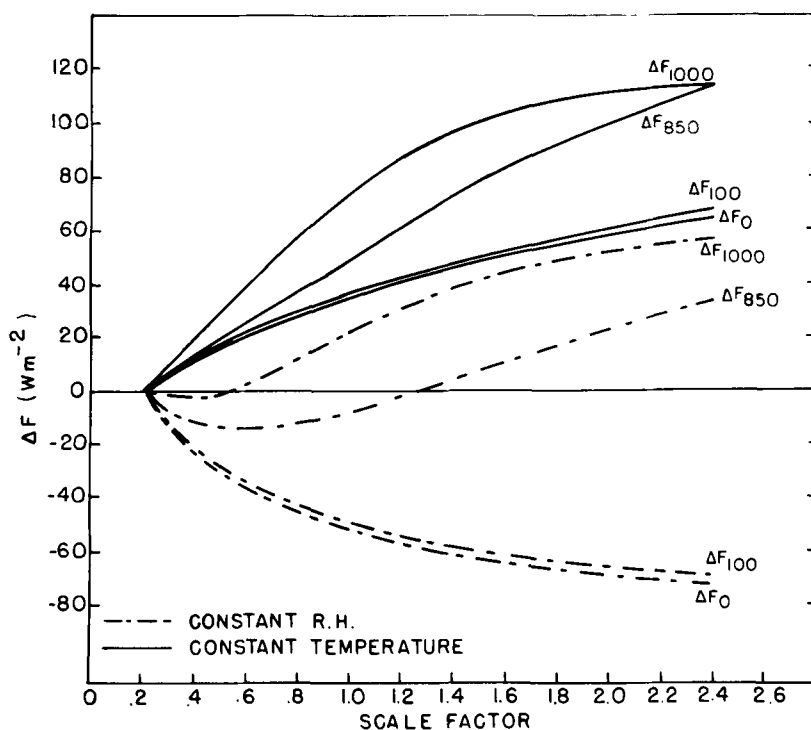


Fig. 8. Flux changes at various levels resulting from scaling water vapour profiles.

the very humid cases, the total radiative flux divergence of the column begins to decrease when the water vapour content is increased. Looking at the flux at 850 mb (Fig. 8) we see that most of this effect occurs within the layer 850–1000 mb. This is simply another representation of the previously discussed relative warming of the lower layers which occurs when the column becomes optically thicker. What we learn from Fig. 8 is that with the onset of this, the ability of the column as a whole to lose energy by radiation is diminished.

Consider now the flux changes with relative humidity held constant. From Fig. 6 we know that the most rapidly varying temperature in the system is that of the surface air, and the ground which is black. Thus we expect that at the lower boundary, the temperature increases will tend to produce a greater net upward flux of thermal radiation, corresponding to a negative ΔF in Fig. 8. Furthermore, we would expect the increasing temperatures throughout the column, as well as at the surface, to produce a greater thermal flux upwards at the top of the column. Both these effects are in the opposite

sense to those related purely to the presence of additional water vapour and the question to be answered is how they balance. In Fig. 8 we see that in the very dry atmospheres (scale factor 0.2–0.4) the temperature effects dominate those due to water vapour, and the net upward flux at the surface increases. However, as the atmosphere becomes more humid and the rate of change of temperature decreases (Fig. 6), water vapour effects become more important and there is a decrease of the net flux upwards at the surface (positive ΔF_{1000}); the surface gets heated. In contrast, the flux upwards at the top of the column increases over the entire range of scale factors (negative rate of change of ΔF). Fig. 8 indicates that the temperature increases associated with the constraint of maintaining relative humidity constant, far outweigh the tendency of the water vapour to blanket the earth/atmosphere system and reduce the amount of energy lost.

If we consider the total flux divergence of the column, we find, much as we might expect, that greater atmospheric temperatures result in greater

flux divergence within the entire column, therefore the rate of increase of cooling in the column as a whole is significantly higher when relative humidity is held constant than when the temperature profile is held constant. Moreover, the total flux divergence within the column continues to increase even at the highest scale factors considered here.

7. An example

Having so far considered only idealized variations in atmospheric temperature and water vapour profiles, we now present an example where major changes of atmospheric water content were observed in association with varying temperature profiles, for which we have calculated the radiative fluxes and cooling rates. These changing profiles were observed with dropsonde measurements made over the Arabian Sea to study the development and onset of the monsoon during the 1979 summer MONEX program.

A concise analysis of the summer circulation over the Arabian Sea has been given by Ramage (1966). The upper troposphere is characterized by large scale subsidence both before and during the monsoon. Convective activity and low cloud formation is therefore severely restricted over the region. A subsidence inversion develops between 500 m and 2000 m altitude and almost no precipitation is observed over the western Arabian Sea during this time. However, the southwest monsoon wind regime becomes established in the near surface layers (Fig. 9) and the combination of clear skies, persistent surface winds and high sea surface temperatures are particularly favourable for evaporative loss of energy from the sea surface to the air. This is the source for much of the water vapour and latent heat which is ultimately released into the troposphere when the monsoon rains break over the western coast of India. Ramage has pointed out that between 800 and 600 mb, the western Arabian Sea is actually overlain by desert dried air originating from the Arabian desert to the west. Thus the ability of the thin layer of surface air below the inversion to bleed off and hold water vapour can be seen as crucial to the development of the monsoon. In summer, the height of the inversion tends to increase from about 1 to 3 km from west to east across the Arabian Sea (Fig. 2 of Ramage, 1966) and thus is undoubtedly related to the

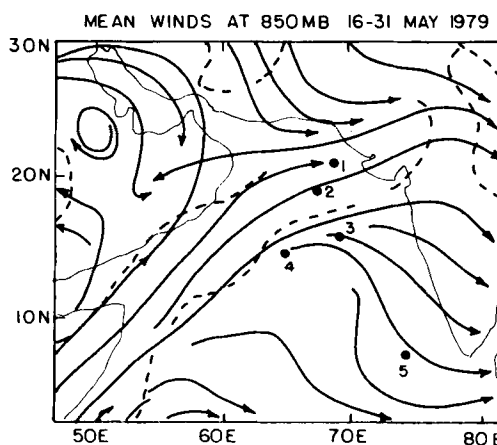


Fig. 9. Mean winds at 850 mb 16–31 May 1979 for MONEX area (taken from Krishnamurti et al., 1979).

strength of the subsidence in the upper air over these locations.

In view of our study of the general radiative effects arising from the presence of additional water vapour in the column, we felt that radiative cooling by water vapour in these lower layers might be significant in relation to the monsoon development.

We have chosen five of the dropsondes from the MONEX data set to represent the range of variation both of the height of the inversion and the water vapour content of the atmosphere over the Arabian Sea during the month preceding the monsoon. The locations of the sondes are indicated on Fig. 9 and the temperature and water vapour profiles are shown in Fig. 10 (interpolated at standard levels) for each sounding.

The sonde profiles suggest that under these conditions, the troposphere assimilates additional water vapour evaporated from the sea surface, not by a uniform increase of concentration throughout the column, but by increasing the depth of the bottom layer beneath the inversion. It is of interest to note that, as the thickness of this lower layer more than doubles, and the total atmospheric water vapour content increases concomitantly, the actual water vapour concentration at the lowest level remains within 2 g kg^{-1} of 20 g kg^{-1} and the near surface temperature remains within 2°C of 28°C .

Clearly, at this point in the development of the monsoon, the additional water vapour in the lower troposphere contributes little or no latent heat to

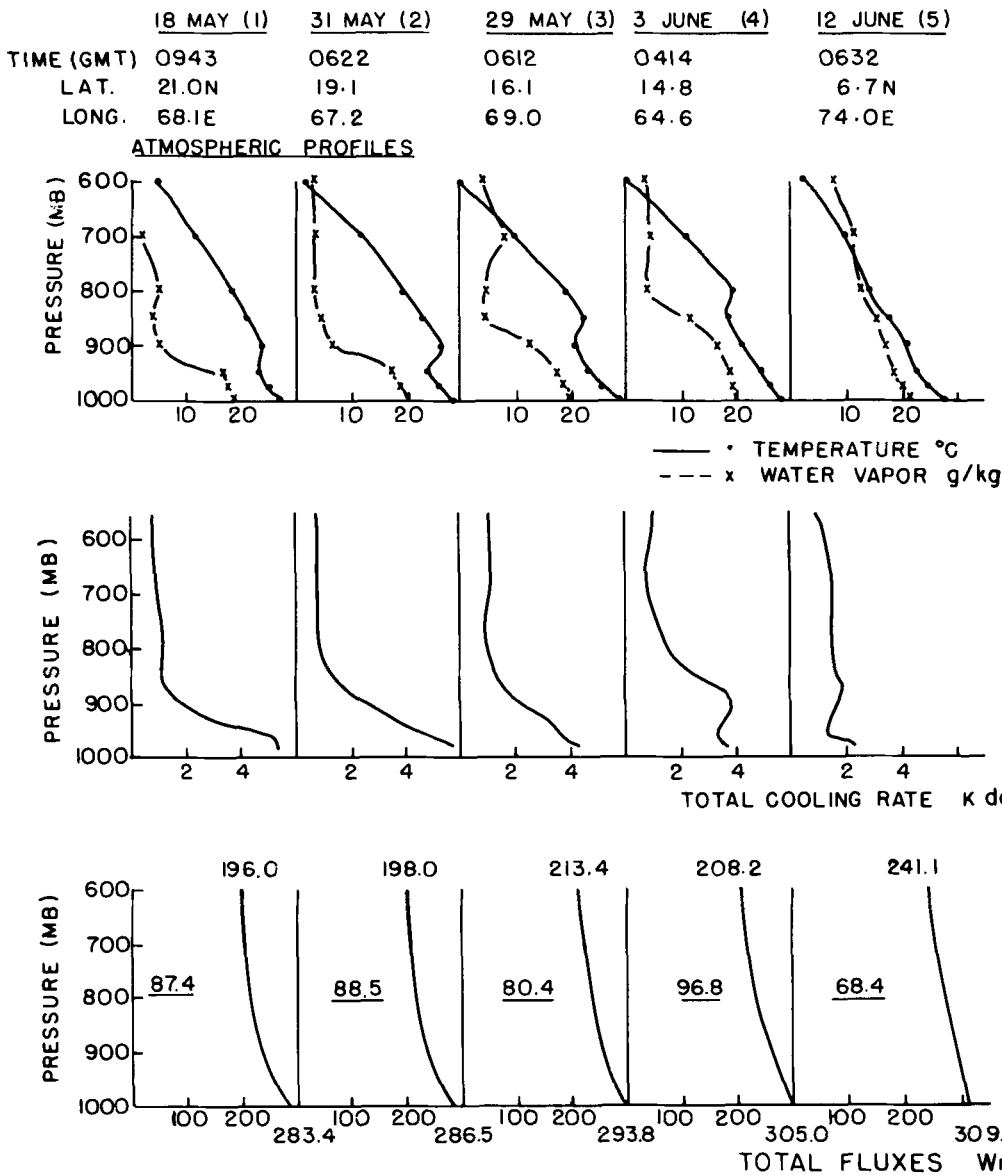


Fig. 10. Development of Arabian Sea inversion in the period 18 May–12 June, 1979 with calculated radiative heating rates. Bottom section represents the total solar and infrared flux (positive downwards). Values at 600 mb and 1000 mb are written on the curves, while the underlined values represent the divergence in the 600–1000 mb layer.

the column. Rather, the growth in thickness of the vapour-laden layer is marked by a reduction of temperatures near its upper boundary within the layers into which the additional vapour is deposited (Fig. 10). Thus the temperature inversion moves upwards by a relative cooling of the air near its base.

In the profile for June 12, however, which contains far more water vapour than the other cases, the temperature inversion is not evident. This profile in fact corresponds to conditions on the first day of the monsoon; clearly the temperature constraint restricting vertical convection is absent.

If we examine the calculated cooling rates for

each of these profiles (Fig. 10), we see a pronounced maximum of radiative cooling in the lowest layers, which on 31 May was close to 6°K per day. We also see that as the water vapour content of the column increases, the general trend is one of relative warming in the very lowest, near surface layer (1000–950 mb). Thus, cooling within the layer 1000–975 mb falls from 5.7°K per day on 31 May to only 2.4°K per day on 12 June.

At the levels immediately above this, however, the general trend is one of relative cooling as more water vapour bleeds into the column. Cooling between 850 and 900 mb is increased from 1.2°K per day to 3.8°K per day from the 18 May profile to the 3 June profile.

This behaviour is quite similar to that which we observed in the near surface air of humid atmospheres (scale factors greater than 1) when the water vapour content is increased (Fig. 5). Moreover, we see that the water vapour increase provides a radiative mechanism which is capable of producing the relative cooling at the base of the inversion which is required in order that the growing moist layer remain confined near the surface.

However, as the water vapour content of the column is thus increased, relative radiative warming near the surface begins to take effect and, as can be seen from the cooling rates shown for 12 June, the strong low-level maximum of radiative cooling disappears and cooling rates diminish from 4 to 2°K per day. Clearly, radiation no longer contributes to the maintenance of the inversion when the water vapour burden exceeds some limiting value. We point out that the tendency towards colder temperatures in the air below the inversion base would result in reduced flux divergence there, rather than increased cooling as calculated. Thus we are confident in ascribing these changes in cooling rate to the water vapour alone.

Another important point here is that the increased flux downward into the surface generated by the extra water vapour (see Fig. 8) will actually contribute to additional evaporation into the column, since the Bowen ratio over this area is of order 0.1 or less.

8. Conclusions

In summary, we have examined the radiative cooling effects associated with an increase of water

vapour in an atmospheric column. We have found that the result for the typical range of atmospheric humidity profiles, is to produce additional radiative cooling of the atmosphere as a whole, to increase the net flux into the lower surface and to reduce the net flux out of the top of the column. The latter two consequences will ultimately lead to convective heating of the column, and when this is crudely represented by maintenance of a constant relative humidity while water vapour is increased by the same amounts, we find that: (i) the net flux divergence of the atmosphere as a whole increases more rapidly; (ii) the net flux into the lower surface increases less rapidly; (iii) the total flux out of the top of the column actually increases, illustrating the fact that the increased energy loss of the atmosphere/surface is a consequence only of the convectively-induced temperature changes.

In conditions which are already very humid, we find that the additional water vapour, independently of any temperature changes, may actually reduce the radiative cooling rates within the lower layers.

We cite the example of the vapour-laden boundary layer which develops over the Arabian Sea before the monsoon as one instance where both the relative cooling due to increased water vapour and the subsequent relative warming of the lowest layers may be of meteorological significance. The enhanced cooling by water vapour below the inversion acts to maintain the inversion, and the relative warming occurs when the inversion is destroyed and the water vapour reaches higher levels. These radiative effects may therefore reinforce the effects of the large-scale dynamics.

We believe it would be of value to conduct a sequence of measurements of total radiative fluxes at selected levels as well as detailed temperature and humidity profiles over the warm tropical ocean in order to directly observe these effects and to establish their size relative to the size of the large scale dynamical effects in the evolution and breakdown of the moist boundary layer.

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