

## An Improved Estimate of the IR Cooling in the Atmospheric Window Region

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### ABSTRACT

A new treatment of absorption in the window region depending on water vapor pressure is incorporated into the IR cooling rate model of Rodgers [for review see Rodgers and Walshaw (1966)]. The appropriate mean diffuse transmission gradient is derived. The results are presented in the form of hemispheric cross sections and indicate that for tropical climates, the new absorption has a considerable effect on radiative transfer in the lower troposphere. The results obtained are very encouraging, showing a reduction in the differences between the measured cooling rates of Cox (1969) and the calculated cooling rates in the lower troposphere.

### 1. Introduction

The understanding of the radiative transfer of the atmospheric window between 8 and 13  $\mu\text{m}$  is very important. This is because the window region lies near the peak of the blackbody function at terrestrial temperatures and thus considerable amounts of energy can be exchanged there. In the past, observations of IR cooling by the window have not agreed with calculated cooling rates. In particular, observations exhibited larger cooling in the lower troposphere than those predicted by theory.

The continuum-like absorption in the window region was mainly attributed to the absorption of the wings of distant strong lines of water vapor and other minor constituents. Bignell (1970) reported a new type of absorption showing a linear dependence on water vapor and also found, contrary to previous work, that this absorption had a negative temperature dependence of 2% per degree Kelvin. While no firm explanation of this was given, Bignell and Burroughs *et al.* (1969) suggested the possibility that the water dimer molecule  $(\text{H}_2\text{O})_2$  was responsible for the observed characteristics. However, Grassl (1974) suggested that such a dimer would have a quadratic dependence on water vapor pressure. The absorption mechanism remains unsolved, although it appears that the absorption is probably due to a complex molecule of water vapor and another abundant gas in the atmosphere or a polymer molecule of water vapor.

Platt (1972) showed comparisons of observed and calculated upward radiance in the 830–1000  $\text{cm}^{-1}$  spectral region as a function of height. These results showed qualitative agreement between observations in the atmosphere and the laboratory-deduced water

vapor pressure dependent absorption coefficients of Bignell. Lee (1973) found excellent agreement of the measured radiation flux profiles and the calculated profiles using similar absorption data to those presented by Burch (1970).

In this article the new window absorption is introduced into the cooling rate model described by Rodgers and Walshaw (1966). The results are presented in the form of infrared cooling rate cross sections for the Southern Hemisphere. A comparison between the calculated cooling rate profile and the observed profile (Cox, 1969) is also included.

### 2. The mean diffuse transmission gradient

Extensive discussions concerning the IR cooling rate model used in this paper were made by Rodgers and Walshaw (1966). For the spectral interval denoted by  $\Delta\nu$ , the cooling rate equation becomes

$$\frac{\Delta\theta}{\Delta t} = \frac{1}{\rho C_p \Delta\nu_i} \left[ \int_0^z \frac{\partial \bar{T}_{f_i}}{\partial z}(z, z') \frac{\partial B_i(z')}{dz'} dz' + \frac{\partial \bar{T}_{f_i}}{\partial z}(z, 0) [B_i(0) - B_i(g)] + \frac{\partial \bar{T}_{f_i}}{\partial z}(z, Z) [B_i(\text{top}) - B_i(Z)] \right] \Delta\nu_i, \quad (1)$$

where the frequency integration has been replaced with a summation over a finite number of subintervals. The frequency interval  $\Delta\nu$  is divided into  $n$  subintervals and the function  $\bar{T}_{f_i}$  and  $B_i$  are the mean values over these subintervals.

The calculation of the cooling rate using (1) depends on the adequate value of the mean diffuse transmission gradient. In this section the mean diffuse transmission gradient is presented. The total absorption coefficient used is (Bignell, 1970)

$$k(\nu, \theta) = k_1(\nu, \theta)p + k_2(\nu, \theta)e, \quad (2)$$

where  $e$  is the water vapor pressure and  $p$  the total atmospheric pressure. The first term is expected for ordinary collision broadening while the second is due to the contribution by the new water vapor dependent absorption. For the 8–13  $\mu\text{m}$  window Bignell stated that  $k_1 \ll k_2$ , and the first term in (2) could be neglected.

The window region from 810–1210  $\text{cm}^{-1}$  was subdivided into four subintervals (i.e.,  $n=4$ ). For each subinterval, the absorption coefficient was assumed to be a linear function of wavenumber (Cox, 1973):

$$k_2(\nu, \theta) = a(\theta) + b(\theta)\nu. \quad (3)$$

An exact integration over frequency could then be performed to obtain the transmission gradient. The values of  $a$  and  $b$  for each transmission gradient were taken from Cox (1973) and were determined using Bignell's tabulation of  $k_2$  at a reference temperature of 20°C.

The temperature dependence of  $k_2$  used by Cox (1973) is adopted. The dependence is specified as proportional to the dimer concentration. The dimer mole fraction  $C$  is determined from

$$C(\theta) = \frac{1}{2} \left\{ \left[ 2 + \frac{p_0 \exp(C_1 - C_2/\theta)}{e} \right] - \left[ \left( 2 + \frac{p_0 \exp(C_1 - C_2/\theta)}{e} \right)^2 - 4 \right]^{1/2} \right\}, \quad (4)$$

where  $C_1 = 13.0096$ ,  $C_2 = 2878.225$ ,  $p_0$  is the reference pressure of one atmosphere, and  $\theta$  is the temperature (K). By using this temperature factor, the temperature dependence of  $a$  and  $b$  in (3) could be separated and the absorption coefficient is rewritten as

$$k(\nu, \theta) = k_2(\nu, \theta)e = (a + b\nu)\Phi(\theta)e, \quad (5)$$

where the temperature dependence factor  $\Phi(\theta)$  is taken as

$$\Phi(\theta) = C(\theta)/C(293 \text{ K}). \quad (6)$$

Using an absorption coefficient of the form  $k_2(\nu, \theta)e$ , and employing (5) and (6), the mean diffuse transmission gradient can be expressed as

$$\frac{dT_f}{dz}(z, z') = \frac{\pm 1}{c_2(z, z')\Delta\nu} \left\{ \left[ \frac{e_1(z, z') - e_2(z, z')}{c_2(z, z')} \right] \frac{dc_2}{dz}(z, z') - [ \nu_2 e_2(z, z') - \nu_1 e_1(z, z') ] \frac{dc_2}{dz}(z, z') - [ e_2(z, z') - e_1(z, z') ] \frac{dc_1}{dz}(z, z') \right\}, \quad (7)$$

where

$$\left. \begin{aligned} e_1(z, z') &= \exp[-c_1(z, z') - c_2(z, z')\nu_1] \\ e_2(z, z') &= \exp[-c_1(z, z') - c_2(z, z')\nu_2] \end{aligned} \right\}, \quad (8)$$

with

$$\left. \begin{aligned} c_1(z, z') &= \frac{\beta a}{0.622g} \sum_z^{z'} w(z)^2 p(z)^2 \Phi(z) \Delta z \\ c_2(z, z') &= \frac{b}{a} c_1(z, z') \end{aligned} \right\} \quad (9)$$

and

$$\left. \begin{aligned} \left[ \frac{dc_1}{dz}(z, z') \right]_{z=z''} &= [w(z'')p(z'')]^2 \Phi(z'') \frac{\beta a}{0.622g} \\ \left[ \frac{dc_2}{dz}(z, z') \right]_{z=z''} &= \frac{b}{a} \left[ \frac{dc_1}{dz}(z, z') \right]_{z=z''} \end{aligned} \right\}. \quad (10)$$

In (9)  $w(z)$  is the mixing ratio at level  $z$  and  $p(z)$  is the pressure of the atmosphere at the same level. Using a mean diffuse transmission gradient of the form expressed in (7), the IR cooling by the spectral interval  $\Delta\nu_i = \nu_2 - \nu_1$  can be calculated from (1).

### 3. Results and discussions

The general meridional distribution of IR cooling in the atmosphere is shown in Figs. 1a and 1b. The cooling rates have been calculated with and without the inclusion of  $e$  type absorption. The hemispheric data were taken from Climate of the Upper Air: Southern Hemisphere, Vol. 4, for January. Fig. 1a shows a maximum in cooling in the upper troposphere, diminishing toward the pole. Water vapor is the principal cooling agent in the model; the most significant cooling results from the strong absorption by the rotation band. The 6.3  $\mu\text{m}$  band, and to a smaller extent the 15  $\mu\text{m}$   $\text{CO}_2$  band, contribute moderately to the IR cooling in the troposphere.

In Fig. 1b, the effect of  $e$  type absorption on the IR cooling rates becomes more discernible. Generally, the cooling in the lower troposphere is enhanced for all latitudes, although only minimally in the polar regions. In tropical climates, the cooling by the window in the lower troposphere is very significant and contributes substantially to the total IR cooling in this region. This large cooling in the lower layers, exemplified by the large maximum between 0°S and 30°S, would tend to stabilize the atmosphere and help explain why the lower layers in the tropics are not convectively unstable. The variation with height shows a similar behavior with cooling rates in tropical regions being larger in all layers, especially up to 600 mb.

A comparison between the calculated and measured profiles of the IR cooling rate, for a tropical atmosphere under "clear" conditions, is presented in Fig. 2.

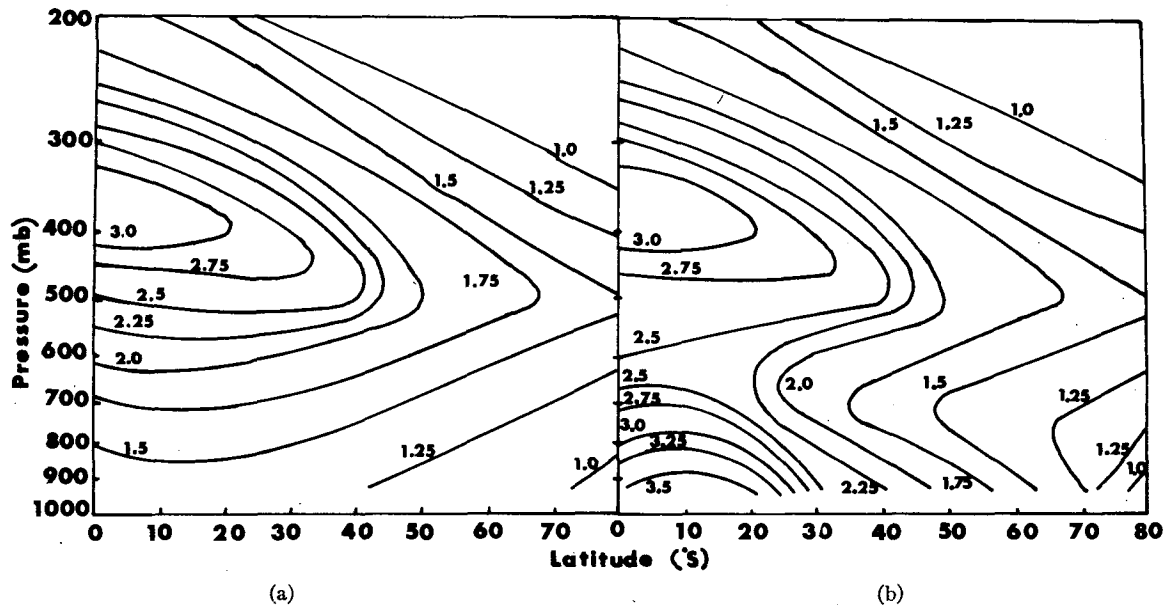


FIG. 1. Meridional distribution of the total IR cooling rates for the Southern Hemisphere without absorption by the window (a) and including the effect of  $e$  type absorption (b). Cooling rate contours are in  $^{\circ}\text{C day}^{-1}$ .

The observed cooling profile, temperature and moisture data used in the computation were taken from Cox (1969). These data are very similar to the  $0^{\circ}\text{N}$ – $10^{\circ}\text{N}$  March atmosphere presented by London (1952). Rodgers and Walshaw (1966) used these data in their computation of the IR cooling rate. The absorption by the atmospheric window was, however, crudely

incorporated in their model by assuming a constant absorption coefficient. Their cooling profile is reproduced in Fig. 2.

The agreement between the observed cooling rates and those calculated in the present study is extremely good. Above 350 mb, the IR cooling rates are overestimated by both computational models. Cirrus clouds could account for this discrepancy. The existence of cirrus, just below the tropical tropopause, would tend to cause IR heating in those levels. Unreported cirrus, as suggested by Cox (1969), may well have been present when the clear radiometer observations were made. London's temperature profile is systematically larger than the observed temperature profile and could explain why the cooling rates calculated by Rodgers and Walshaw are generally larger than the observed values and those calculated in this paper.

Below 700 mb, the computation of Rodgers and Walshaw seriously underestimate the IR cooling by as much as  $1.5$ – $2.0^{\circ}\text{C day}^{-1}$ . The comparison between the calculated profile and the observed profile is, however, greatly improved when the new window absorption is included in the theoretical model. The presence of this absorption accounts for the large discrepancies that previously existed between the observed and calculated cooling rates in the layer below 700 mb.

The existence of aerosols can also partly explain these earlier discrepancies. Grassl (1973) estimated that the contribution by aerosols in the lowest few hundred meters of the atmosphere ranged from  $0.1$  to  $0.7^{\circ}\text{C day}^{-1}$ . Kuhn *et al.* (1975), using observed haze properties, confirmed this estimate. This contribution is relatively small and may be balanced by solar

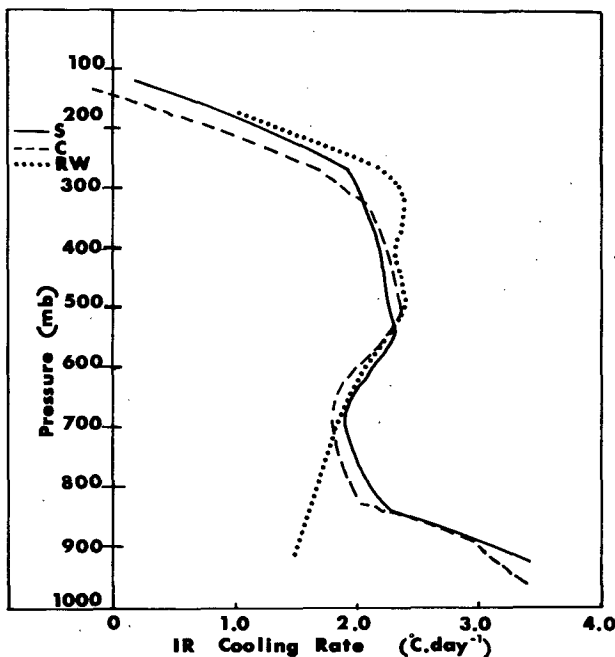


FIG. 2. Comparison of calculated and observed cooling rates in a clear atmosphere: C, observed profile of Cox (1969); RW, the calculated profile of Rodgers and Walshaw (1966) using the data of London (1952); S, present study.

warming in this layer. In polar climates, however, the contribution by water vapor to the IR cooling is much smaller and the cooling by aerosols can be significant, especially in the presence of a temperature inversion.

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