# 10.4 Applications to Meteorology, Climatology, and Remote Sensing

The previous section outlined the methods used to compute broadband upwelling and downwelling fluxes in a cloud-free atmosphere. Of course hardly anyone computes fluxes just for the fun of it; rather, they do it because it is an essential part of modeling the energy budget of the atmosphere.

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## 10.4.1 Fluxes and Radiative Heating/Cooling

### Radiative Heating Equations

In Section 2.7, we defined the net flux as

$$F^{\mathrm{net}}(z) \equiv F^{\uparrow}(z) - F^{\downarrow}(z) \,. \tag{10.53}$$

Depending on the context, we might be interested in the net flux computed (or measured) for all wavelengths, for just the shortwave or longwave band, or even for a narrow spectral interval. In any of these cases, $F^{\mathrm{net}}$ represents the corresponding net upward flow of radiative energy, measured in watts per meter squared, through a unit horizontal area at level $z$.

Now consider a thin layer of the atmosphere with its base at altitude $z$ and its upper boundary at $z + \Delta z$. The net flux $F^{\mathrm{net}}(z)$ represents the rate at which radiative energy enters the bottom of this layer; likewise, $F^{\mathrm{net}}(z + \Delta z)$ gives the rate at which energy leaves at the top of the layer. If the two fluxes are equal, then there is no net change over time in the internal energy of the layer. If they are not equal, then the layer must be experiencing a net gain or loss of energy.

It follows that the *radiative heating rate* at level $z$ is given by

$$\boxed{\mathcal{H} \equiv -\frac{1}{\rho(z) C_p} \frac{\partial F^{\mathrm{net}}}{\partial z}(z) \,,} \tag{10.54}$$

where $\rho(z)$ is the air density at level $z$ and $C_p = 1005\ \mathrm{J/(kg\ K)}$ is the specific heat capacity of air at constant pressure. The minus sign is needed because an *increase* in $F^{\text{net}}$ with height implies a net *loss* of energy from level $z$. Traditionally, $\mathcal{H}$ is expressed in units of °C/day. When the value of $\mathcal{H}$ is negative (as it is more often than not), one might prefer to speak instead of positive *radiative cooling rate*.

In order to utilize the band model machinery we developed earlier, let's confine our attention to the heating/cooling rate associated with a particular spectral interval $\Delta\tilde{\nu}_i$. The complete expressions for upwelling and downwelling flux, including boundary contributions, are then

$$F_i^{\uparrow}(z) = F_i^{\uparrow}(0)\mathcal{T}_i(0,z) + \Delta\tilde{\nu}_i \int_0^z \pi\overline{B}_i(z')\frac{\partial\mathcal{T}_i(z',z)}{\partial z'}\,dz' \,, \tag{10.55}$$

$$F_i^{\downarrow}(z) = F_i^{\downarrow}(\infty)\mathcal{T}_i(z,\infty) - \Delta\tilde{\nu}_i \int_z^\infty \pi\overline{B}_i(z')\frac{\partial\mathcal{T}_i(z,z')}{\partial z'}\,dz' \,, \tag{10.56}$$

where, as usual, $\mathcal{T}_i(z,z')$ is the band-averaged flux transmittance between levels $z$ and $z'$. Note that we are using $\overline{B}_i(z)$ as a shorthand notation for the average value of $B[T(z)]$ in the $i$th spectral interval.

Let's use the above expressions to evaluate $F^{\text{net}}$ and, from there, $\partial F^{\text{net}}/\partial z$, as required for the heating rate in (10.54):

$$\begin{aligned} F_i^{\text{net}}(z) = {} & F_i^{\uparrow}(0)\mathcal{T}_i(0,z) - F_i^{\downarrow}(\infty)\mathcal{T}_i(z,\infty) \\ & + \Delta\tilde{\nu}_i \int_0^z \pi\overline{B}_i(z')\frac{\partial\mathcal{T}_i(z',z)}{\partial z'}\,dz' \\ & + \Delta\tilde{\nu}_i \int_z^\infty \pi\overline{B}_i(z')\frac{\partial\mathcal{T}_i(z,z')}{\partial z'}\,dz' \,, \end{aligned} \tag{10.57}$$

$$\begin{aligned} \frac{\partial F_i^{\text{net}}(z)}{\partial z} = {} & F_i^{\uparrow}(0)\frac{\partial\mathcal{T}_i(0,z)}{\partial z} - F_i^{\downarrow}(\infty)\frac{\partial\mathcal{T}_i(z,\infty)}{\partial z} \\ & + \Delta\tilde{\nu}_i\frac{\partial}{\partial z}\left[\int_0^z \pi\overline{B}_i(z')\frac{\partial\mathcal{T}_i(z',z)}{\partial z'}\,dz' \right. \\ & \left. + \int_z^\infty \pi\overline{B}_i(z')\frac{\partial\mathcal{T}_i(z,z')}{\partial z'}\,dz'\right] . \end{aligned} \tag{10.58}$$

To evaluate the partial derivatives of the integral terms, in which $z$ appears both as one of the limits of integration *and* as an argument to the integrand itself, we invoke the following mathematical identity:

$$\frac{\partial}{\partial x}\int_{x_0}^x f(x,y)\,dy \equiv \int_{x_0}^x \frac{\partial f(x,y)}{\partial x}\,dy + f(x,x) \,, \tag{10.59}$$

so that the heating rate $\mathcal{H}$ at level $z$ is proportional to

$$\begin{aligned}\mathcal{H}(z)\rho(z)C_p &= -F_i^{\uparrow}(0)\frac{\partial \mathcal{T}_i(0,z)}{\partial z} + F_i^{\downarrow}(\infty)\frac{\partial \mathcal{T}_i(z,\infty)}{\partial z} \\ &+ \Delta\tilde{\nu}_i\left[-\int_0^z \pi\overline{B}_i(z')\frac{\partial^2 \mathcal{T}_i(z',z)}{\partial z'\partial z}\,dz' - \int_z^\infty \pi\overline{B}_i(z')\frac{\partial^2 \mathcal{T}_i(z,z')}{\partial z'\partial z}\,dz' \right.\\ &\left. - \pi\overline{B}_i(z)\left.\frac{\partial \mathcal{T}_i(z',z)}{\partial z'}\right|_{z'=z} + \pi\overline{B}_i(z)\left.\frac{\partial \mathcal{T}_i(z,z')}{\partial z'}\right|_{z'=z}\right].\end{aligned} \tag{10.60}$$

Before we make further adjustments, let's take a second to interpret the terms in the above equation.

The first term is the contribution of the upward flux at the surface to heating at level $z$. Note that $\partial\mathcal{T}(0,z)/\partial z$ (which is negative) plays exactly the same role as the "absorption weighting function" we derived in (7.51), except that it now applies to the band-integrated flux rather than the monochromatic intensity. The second term is of course the same as the first, except that it represents the absorption of radiation incident at the top of the atmosphere. Both the first and the second terms are always greater than zero and represent positive heating terms.

The two integrals in the second line of the equation represent the contributions of emission at all other levels $z'$ to heating at level $z$. These are also positive terms.

The final two terms,[4] respectively, represent *losses* of energy due to upward and downward emission of radiation by the atmosphere at level $z$. Note that both terms are equal and negative, because

$$\left.\frac{\partial \mathcal{T}_i(z',z)}{\partial z'}\right|_{z'=z} = -\left.\frac{\partial \mathcal{T}_i(z,z')}{\partial z'}\right|_{z'=z}. \tag{10.61}$$

Although we could combine the last two terms in (10.60) using the above identity, we will leave them separate in order to make the next steps clearer.

Although (10.60) is certainly an accurate and complete statement of how local heating depends on the profile of transmission and temperature (via $\overline{B}(z)$), we're less interested here in computational

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[4] These terms are omitted in Equation (11.29) of TS02.

validity than we are in physical insight. Watch carefully while we rearrange our equation into a new form that clarifies some aspects of the relationship between local heating and the surrounding environment.

We will start with an expression that has the form we are seeking and show that it is mathematically equivalent to, and can be substituted for, part of (10.60).

$$\begin{aligned}
&\int_0^z \pi\left[\overline{B}_i(z') - \overline{B}_i(z)\right] \frac{\partial^2 \mathcal{T}_i(z',z)}{\partial z' \partial z}\, dz' \\
&\quad = \int_0^z \pi \overline{B}_i(z') \frac{\partial^2 \mathcal{T}_i(z',z)}{\partial z' \partial z}\, dz' - \int_0^z \pi \overline{B}_i(z) \frac{\partial^2 \mathcal{T}_i(z',z)}{\partial z' \partial z}\, dz' \\
&\quad = \int_0^z \pi \overline{B}_i(z') \frac{\partial^2 \mathcal{T}_i(z',z)}{\partial z' \partial z}\, dz' - \pi \overline{B}_i(z) \int_0^z \frac{\partial^2 \mathcal{T}_i(z',z)}{\partial z' \partial z}\, dz' \\
&\quad = \int_0^z \pi \overline{B}_i(z') \frac{\partial^2 \mathcal{T}_i(z',z)}{\partial z' \partial z}\, dz' - \pi \overline{B}_i(z) \left[\frac{\partial \mathcal{T}_i(z',z)}{\partial z}\right]_{z'=0}^{z'=z} \\
&= \int_0^z \pi \overline{B}_i(z') \frac{\partial^2 \mathcal{T}_i(z',z)}{\partial z' \partial z}\, dz' - \pi \overline{B}_i(z) \left.\frac{\partial \mathcal{T}_i(z',z)}{\partial z}\right|_{z'=z} + \pi \overline{B}_i(z) \frac{\partial \mathcal{T}_i(0,z)}{\partial z} \\
&= \int_0^z \pi \overline{B}_i(z') \frac{\partial^2 \mathcal{T}_i(z',z)}{\partial z' \partial z}\, dz' + \pi \overline{B}_i(z) \left.\frac{\partial \mathcal{T}_i(z',z)}{\partial z'}\right|_{z'=z} + \pi \overline{B}_i(z) \frac{\partial \mathcal{T}_i(0,z)}{\partial z}
\end{aligned} \quad , \quad (10.62)$$

or, if we subtract the right side of the above equality from the left side,

$$\begin{aligned}
&\int_0^z \pi\left[\overline{B}_i(z') - \overline{B}_i(z)\right] \frac{\partial^2 \mathcal{T}_i(z',z)}{\partial z' \partial z}\, dz' - \int_0^z \pi \overline{B}_i(z') \frac{\partial^2 \mathcal{T}_i(z',z)}{\partial z' \partial z}\, dz' \\
&\qquad - \pi \overline{B}_i(z) \left.\frac{\partial \mathcal{T}_i(z',z)}{\partial z'}\right|_{z'=z} - \pi \overline{B}_i(z) \frac{\partial \mathcal{T}_i(0,z)}{\partial z} = 0. \quad (10.63)
\end{aligned}$$

Using similar reasoning, we can show that

$$\begin{aligned}
&\int_z^\infty \pi\left[\overline{B}_i(z') - \overline{B}_i(z)\right] \frac{\partial^2 \mathcal{T}_i(z,z')}{\partial z' \partial z}\, dz' - \int_z^\infty \pi \overline{B}_i(z') \frac{\partial^2 \mathcal{T}_i(z,z')}{\partial z' \partial z}\, dz' \\
&\qquad + \pi \overline{B}_i(z) \frac{\partial \mathcal{T}_i(z,\infty)}{\partial z} + \pi \overline{B}_i(z) \left.\frac{\partial \mathcal{T}_i(z,z')}{\partial z'}\right|_{z'=z} = 0. \quad (10.64)
\end{aligned}$$

Since the expressions on the left-hand sides of both (10.63) and (10.64) are equal to zero, we can subtract both expressions from inside the square brackets of (10.60) without changing the latter's validity. After considerable manipulation and cancellation of terms, we get

$$\boxed{\begin{aligned} \mathcal{H}(z) = \frac{1}{\rho(z)C_p} \Bigg\{ & \\ & - \left[F_i^{\uparrow}(0) - \Delta\tilde{\nu}_i \pi \overline{B}_i(z)\right] \frac{\partial \mathcal{T}_i(0,z)}{\partial z} && (A) \\ & + \left[F_i^{\downarrow}(\infty) - \Delta\tilde{\nu}_i \pi \overline{B}_i(z)\right] \frac{\partial \mathcal{T}_i(z,\infty)}{\partial z} && (B) \\ & - \Delta\tilde{\nu}_i \pi \int_z^{\infty} \left[\overline{B}_i(z') - \overline{B}_i(z)\right] \frac{\partial^2 \mathcal{T}_i(z,z')}{\partial z' \partial z}\, dz' && (C) \\ & - \Delta\tilde{\nu}_i \pi \int_0^{z} \left[\overline{B}_i(z') - \overline{B}_i(z)\right] \frac{\partial^2 \mathcal{T}_i(z',z)}{\partial z' \partial z}\, dz' && (D) \\ \Bigg\} . \end{aligned}} \tag{10.65}$$

Equation (10.65) certainly looks different than the mathematically equivalent (10.60), and it tells a different kind of story as well. Each of the partial derivatives represents the degree of *radiative coupling* between level $z$ and some other part of the column. It characterizes the degree to which radiation *emitted* by one component is *absorbed* by the the atmosphere at *z and vice versa.*

Thus, for each of the lines (A)–(D), the difference inside the square brackets represents the imbalance between (a) radiation emitted from the remote location and reabsorbed at level $z$, and (b) energy lost by emission from level $z$ and reabsorbed at the remote location. If the difference is positive, then the contribution to heating at level $z$ is positive; negative implies cooling. I will emphasize the word **exchange** wherever it appears below in order to reinforce this idea of a *two-way* process.

Let's now interpret each line in turn:

**Term (A)** represents net heating/cooling through radiative **exchange** with the lower boundary. In the thermal IR band, we can usually take the surface to be black, in which case we can replace $F_i^{\uparrow}(0)$ with $\Delta\tilde{\nu}_i \pi \overline{B}_i(T_s)$, where $T_s$ is the surface temperature. Since $T_s$ is usually greater than $T(z)$, Term A usually represents a heating term.

**Term (B)** represents heating/cooling through radiative **exchange** with the top of the atmosphere (TOA). If we're working in the thermal IR band, then $F_i^{\downarrow}(\infty)$ is usually taken to be zero, in which case the exchange is strictly one-way. In the LW band, therefore, term (B) describes *cooling to space.* If, on the other hand, we're working with solar radiation, then $F_i^{\downarrow}(\infty)$ represents the incident flux of solar radiation at the TOA, and $\overline{B}(z) = 0$, so that term (B) describes heating through the direct absorption of solar radiation.

**Terms (C) and (D)** collectively represent radiative **exchanges** between level $z$ and all other levels in the atmosphere $z'$. In order for the net effect of this exchange to be significant, there must be a large temperature difference between the two levels, *and* the radiative coupling (represented by the second derivatives of $\mathcal{T}$) must be strong. The coupling is strongest when the transmission $\mathcal{T}$ changes rapidly in response to changes in both $z$ and $z'$. In regions of the spectrum for which the atmosphere is strongly absorbing (e.g., the middle of the $CO_2$ 15 $\mu$m band), the coupling is strongest with levels $z'$ that are very close to $z$ and where the temperature difference is small; therefore the contribution of these wavelengths to local heating at $z$ is negligible.

Note further that (C) represents **exchanges** between $z$ and higher levels of the atmosphere; (D) represents **exchanges** with lower altitudes. In the middle of the troposphere, the temperature generally decreases in a quasi-linear fashion with increasing $z$. This implies that the cooling contribution by (C) will usually be offset to a large degree by heating from (D). The reverse applies in the stratosphere, where temperature tends to increase with height. It follows, therefore, that heating/cooling due to exhanges with other levels will be strongest where there is a minimum or maximum in $T(z)$. In particular, the tropopause level "sees" radiation arriving from the warmer stratosphere (term C) as well as from warmer levels of the troposphere (term D); hence, the tropopause experiences positive heating contributions from both (C) and (D).

In the longwave band, every term except (B) represents radiative exchanges between levels having temperatures falling somewhere in the range 190–310 K. The typical temperature *difference* between two levels that are strongly coupled is on the order of 10s of K or less. Furthermore, as noted above, terms (C) and (D) tend to partly cancel each other, except in the vicinity of the tropopause.

Term (B), by contrast, represents a direct loss of radiation to space, with no compensating return radiation. This term is therefore always negative, and it is quite often the largest term overall in the longwave radiation budget, especially at altitudes for which the atmosphere rapidly becomes more transparent with increasing height (e.g., due to the rapid narrowing of absorption lines). *To a very good approximation in many cases, radiative cooling profiles in the atmosphere can be estimated from term (B) alone.* This is called the *cooling-to-space approximation.*

## Model Atmospheres

It's nice to have an equation like (10.65) to tell you *how* heating or cooling at a particular level is physically related to profiles of temperature and band transmittance. But simply staring at it won't tell you much about the actual magnitude of the heating or cooling to expect at any particular level $z$. You have to write a program to numerically evaluate the various terms in the equation using suitable band transmittance models, and you then have to apply your program to a particular profile of atmospheric temperature, humidity, and trace gas composition.

Although you could, in principle, pull any old radiosonde sounding off the Internet and run your program on it (although some assumptions would still be required concerning ozone profiles and the like), atmospheric scientists usually like to start by running their radiative transfer codes on idealized profiles called *model atmospheres.* These don't represent actual observations but are designed to reflect typical atmospheric conditions for a particular location and season. The use of standard model atmospheres has

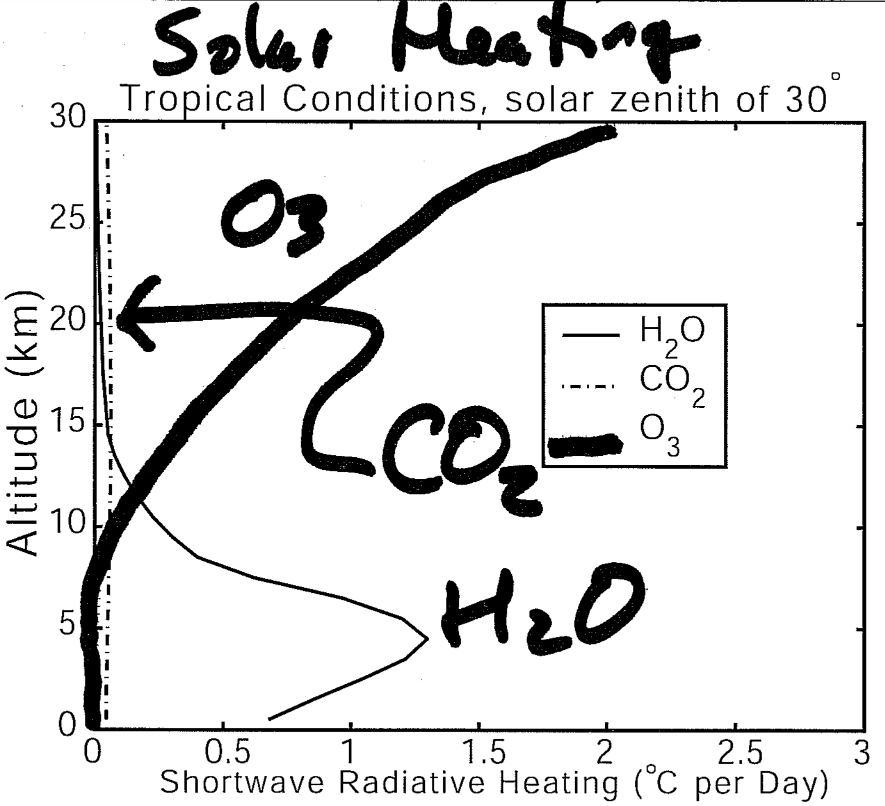

**Fig. 10.6:** Typical heating profiles due to solar absorption in a cloud-free tropical atmosphere, segregated according to the responsible atmospheric constituents ozone ($O_3$), carbon dioxide ($CO_2$), and water vapor ($H_2O$). *(Figure courtesy of S. Ackerman.)*

two advantages:

- It facilitates the comparison of different radiative transfer codes. If results differ significantly for two methods using the exact same model atmosphere, then at least one of them (perhaps both) must be doing something wrong!
- The results of a radiative transfer calculation for a model atmosphere can be regarded as typical, in some sense, for the region and season and question it represents, even if it is understood that actual conditions will vary from day to day and location to location. Thus a model atmosphere will usually be employed in a calculation any time an actual sounding is not available but you need representative radiative transfer results.

There are at least seven standard model atmospheres in circulation that radiation and remote sensing scientists tend to rely on for

[illegible]

the above purposes: (1) tropical, (2) midlatitude summer, (3) midlatitude winter, (4) subarctic summer, (5) subarctic winter, (6) arctic summer, and (7) arctic winter. Although the associated temperature and humidity profiles are highly idealized, they vary in the ways you'd expect: the tropical model atmosphere is warm and humid with a high, cold tropopause; the arctic winter atmosphere is extremely cold with low humidity and a very low tropopause, and the other five models fall somewhere in between these two extremes.

It's worth keeping in mind that variations in water vapor mixing ratio are often even more important for radiative heating profiles than are variations in the temperature profiles themselves. This is because of water vapor's relative opacity over much of the SW and LW bands.

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## Shortwave Heating

Figure 10.6 depicts heating profiles due to the absorption of solar radiation in a cloud-free tropical atmosphere. The heating rate is given in degrees per day. Separate profiles are shown for the three major contributors to atmospheric absorption in the solar band: water vapor, ozone, and carbon dioxide. The first two of these constituents are clearly the dominant absorbers in the solar band, with water vapor being responsible for the bulk of the absorption (up to 1.3 K/day) in the troposphere ($z < 15$ km in this model atmosphere) and ozone dominating in the stratosphere, with heating rates in excess of 2 K/day.[5] Carbon dioxide plays a minor role, being responsible for around 0.05 K/day of heating at all levels.

To first order, the differences between the shapes of the heating profiles for different constituents mirrors the difference in constituent mixing ratios:

- Water vapor is most abundant at the warm altitudes in the lower troposphere, because it condenses to liquid water or ice and precipitates out when it is transported to higher, colder altitudes.

- Ozone is concentrated primarily in the stratosphere, where it

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[5] The heating rate continues to increase with altitude above 30 km, peaking at an altitude high in the upper stratosphere.

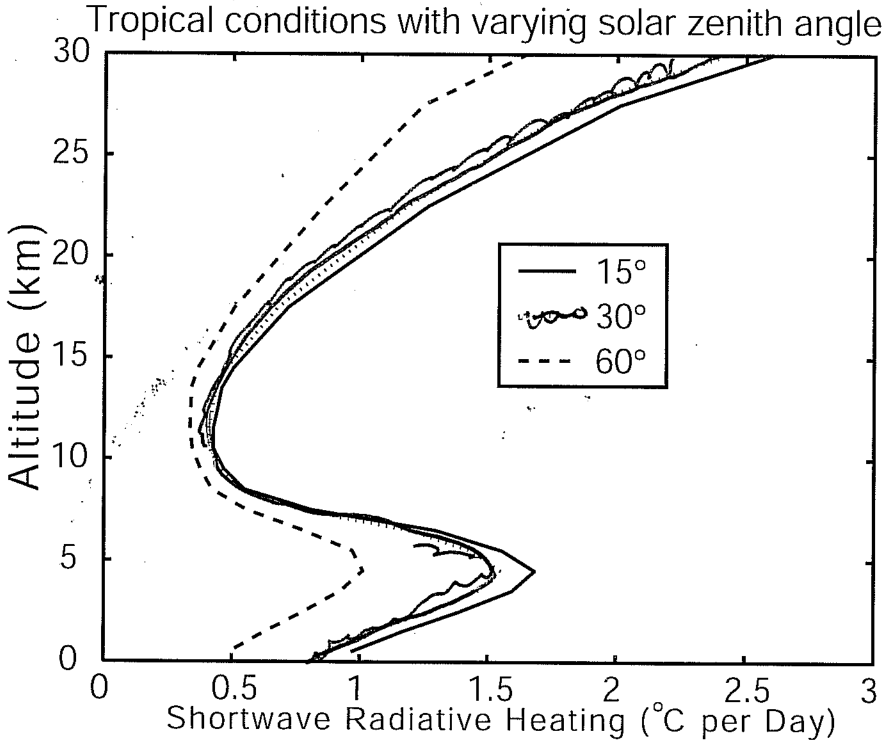

**Fig. 10.7:** Typical heating profiles due to solar absorption in a cloud-free tropical atmosphere at three different solar zenith angles. The heating profile for $\theta_s = 30°$ corresponds to the sum of the three profiles in Fig. 10.6. *(Figure courtesy of S. Ackerman.)*

is created by the action of UV-C radiation on molecular oxygen. In fact, the stratosphere itself, which is *defined* by the general increase of temperature with height starting around 10–15 km owes its very existence to the absorption of solar radiation by ozone. *If there were no molecular oxygen in the atmosphere, there would be no ozone layer. Without ozone, there would be very little solar heating of the middle atmosphere and therefore no stratosphere!*

- Carbon dioxide is evenly mixed throughout the atmosphere, because atmospheric sources and removal mechanisms for $CO_2$ operate very slowly in comparison to those for water vapor and ozone.

**Problem 10.2:** Actual water vapor mixing ratios in the standard tropical atmosphere continue to increase all the way to the surface.

Yet the heating rate associated with water vapor peaks at an altitude near 5 km and then decreases sharply below that level. Explain why.

Total shortwave heating rates are of course the sum of those contributed by the individual constituents. Profiles of total heating due to the absorption of solar radiation in a tropical atmosphere are shown in Fig. 10.7. The different profiles correspond to different solar zenith angles. Not surprisingly, when the sun is lower in the sky, overall heating rates are reduced.

Of course, the heating profiles described above include only contributions due to *direct* absorption of shortwave radiation. In the absence of clouds, solar radiation not absorbed by the atmosphere reaches the surface, where a fraction equal to one minus the surface albedo gets absorbed. Much of that energy *indirectly* heats the atmosphere by way of three mechanisms: (1) emission and reabsorption of longwave radiation (discussed below), (2) direct conduction of heat from the surface to the overlying air, followed by convective mixing, and (3) evaporation of water from the surface, followed by latent heat release in clouds.

## Longwave Cooling

In the longwave band, each part of the atmosphere is simultaneously emitting and absorbing radiation. Where absorption dominates, there is net heating; where emission dominates, there is net cooling. Profiles of heating/cooling in the longwave band are more difficult to interpret because all four of the terms in (10.65) are potentially significant.

Fig. 10.8 shows longwave heating profiles for a tropical atmosphere, again segregated according to the responsible constituent. Water vapor is actually represented here by two curves, one for the heating contribution due to its rotation/vibration bands near 6.3 $\mu$m and beyond 15 $\mu$m; the other due to the relatively weak but pervasive continuum component that dirties up the spectral "windows" between conventional absorption bands.

Here is a brief rundown of the major features seen in this plot:

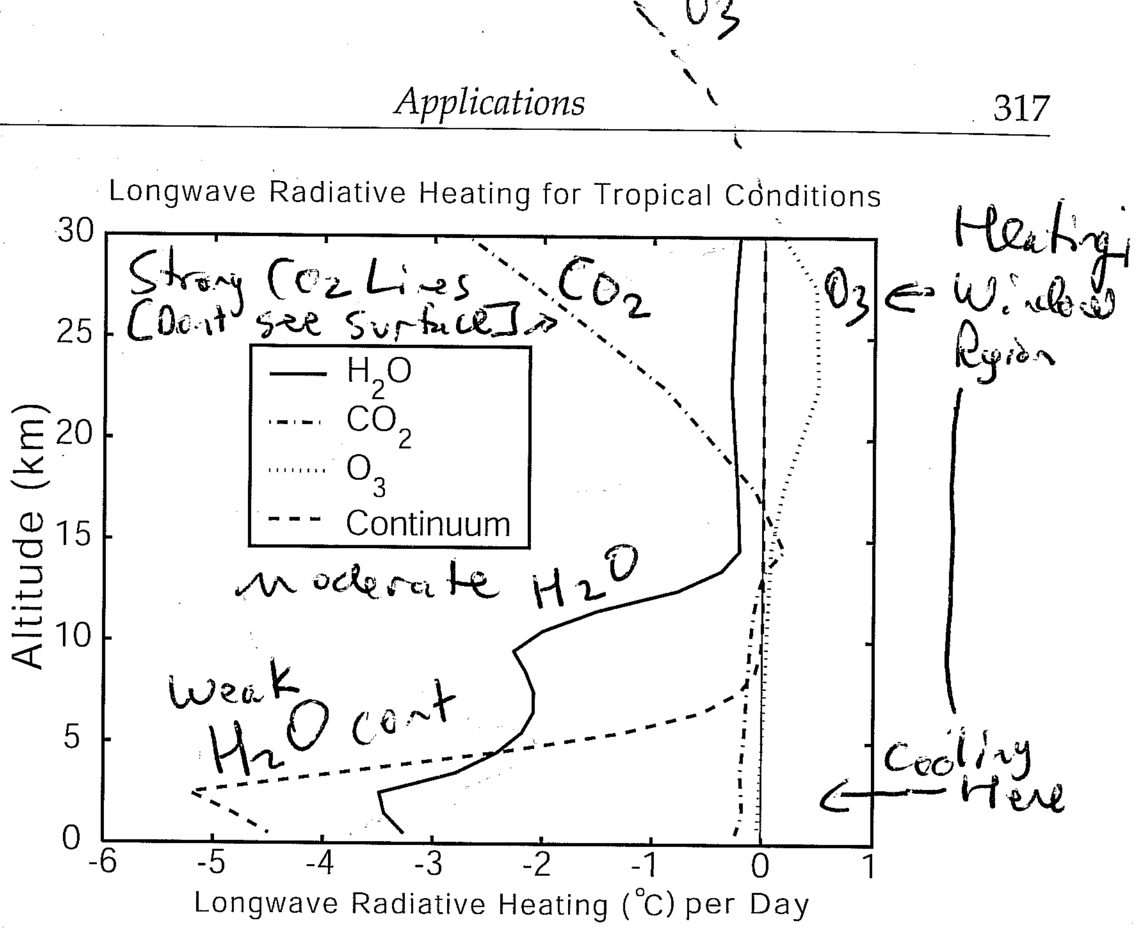

**Fig. 10.8:** Typical heating rate profiles due to longwave (thermal IR) radiative transfer in a cloud-free tropical atmosphere, segregated according to the responsible atmospheric constituents ozone ($O_3$), carbon dioxide ($CO_2$), water vapor resonant absorption ($H_2O$), and water vapor continuum. Negative values represent cooling. *(Figure courtesy of S. Ackerman, with modifications.)*

**$CO_2$** Because of the high opacity of the pressure broadened 15 $\mu$m band, carbon dioxide contributes rather little to net radiative heating in the troposphere. Radiation emitted at one level is reabsorbed at nearby level having almost the same temperature. Only at the tropopause (near 15 km), where the temperature profile has a minimum, is there a small amount of net heating. At higher altitudes, pressure broadening is much weaker and the band "opens up," allowing emitted radiation to escape to space with little compensating radiation downward from higher levels. This is of course the *cooling to space* previously discussed in connection with term (B) in (10.65).

**$H_2O$** Because water vapor is concentrated at low altitudes, the cooling-to-space effect kicks in strongly between 3 and 10 km altitude, with maximum cooling rates between 2 and 3.5 K/day. The two peaks in the profiles are associated with different absorption bands, the stronger of these being associated with the higher altitude peak. Above the tropopause near 15 km, there is very little water vapor present and so its contribution to cooling at those altitudes is very modest – about 0.2 K/day.

**$H_2O$ continuum** One of the unique characteristics of continuum absorption is that it is very sensitive to pressure. The mass absorption coefficient of water vapor falls off rapidly with height. Consequently, the atmosphere in the upper troposphere and above is effectively transparent in this spectral band, while lower altitudes see fairly strong absorption. Once again, cooling-to-space is the dominant radiative exchange term at the rather sharp transition between the two zones, with peak cooling of 5 K/day occurring at the relatively low altitude of 3 km. Above 10 km, the continuum contribution to cooling is essentially zero.

**$O_3$** Ozone is responsible for the sole instance of significant LW *warming* in the atmosphere below 30 km, with the peak warming of 0.5 K/day occurring between 20 and 30 km. This heating is due to the absorption at the base of the ozone layer of radiation emitted by the ground in the 9.6 $\mu$m band — i.e., term (A) in (10.65). If the plotted profiles in Fig. 10.8 were extended to higher altitude, you would see the heating due to term (A) disappear and get replaced with fairly strong cooling-to-space at the top of the ozone layer.

The above discussion applies to a tropical atmosphere. Different model atmospheres have different profiles of temperature and LW absorption and therefore different profiles of LW heating/cooling. Fig. 10.9 shows total LW heating profiles for a tropical, midlatitude summer, and subarctic summer atmosphere. Qualitatively they are similar; only the details differ, mainly due to differences in the water vapor contribution and in the gross features of the temperature profile.

Note that the LW radiative exchanges in the cloud-free atmosphere have a net cooling effect at almost all levels of the atmosphere, whereas the effect of SW absorption was exclusively warming. In fact, in the stratosphere, SW warming and LW cooling nearly

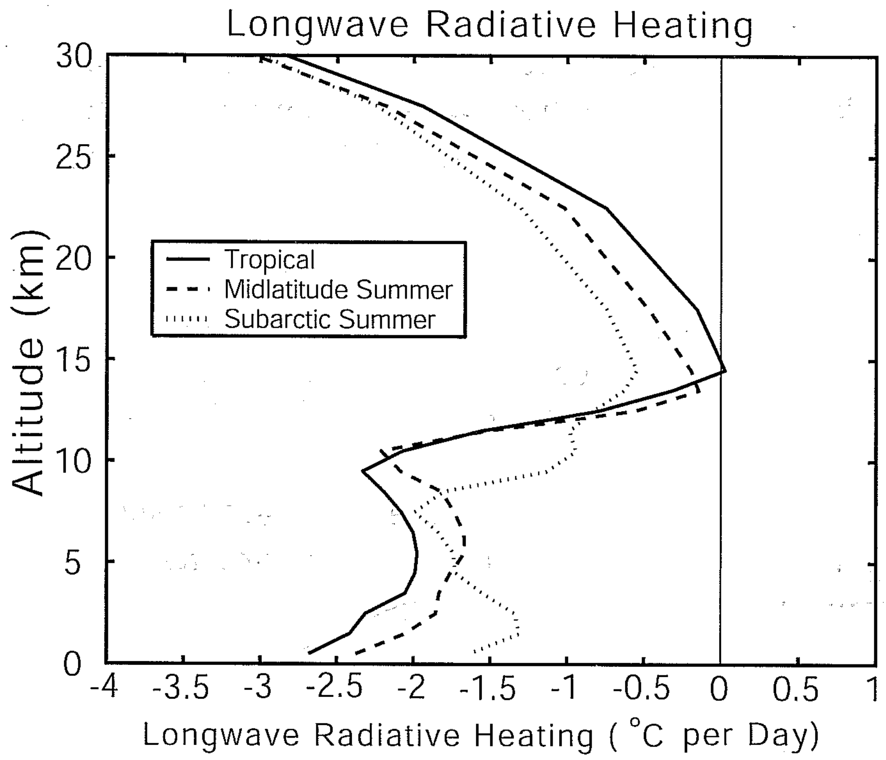

**Fig. 10.9:** Total longwave heating rate profiles for three different model atmospheres. *(Figure courtesy of S. Ackerman, with modifications.)*

cancel. This is because there are few other effective energy exchange mechanisms at work in the stratosphere; therefore the temperature of the stratosphere must necessarily approach that required for radiative equilibrium. In the troposphere, a variety of other processes are available to compensate for any radiative imbalance; therefore we don't expect the SW and LW heating profiles to be mirror images of one another.