$$F^{\uparrow} = \varepsilon\sigma T_s^4 = \sigma T_s^4 \ , \tag{6.39}$$

where, as before, we are taking the LW surface emissivity $\varepsilon \approx 1$.

The cooling rate of the ground is proportional to the net flux

$$F^{\text{net}} = F^{\uparrow} - F^{\downarrow} = \sigma T_s^4 - a_{\text{lw}}\sigma T_a^4 \ , \tag{6.40}$$

or

$$F^{\text{net}} = \sigma(T_s^4 - a_{\text{lw}} T_a^4) \ . \tag{6.41}$$

The effective value of a_{lw} for the cloud-free atmosphere ranges from approximately 0.7 in the wintertime arctic to approximately 0.95 in the tropics. This variation is driven primarily by the humidity of the atmosphere, as water vapor is a strong absorber of radiation over much of the thermal IR band. The corresponding range of effective atmospheric temperature T_a is 235 K (arctic winter) to 290 K (tropical), yielding typical clear-sky downwelling longwave fluxes $F^{\downarrow}$ ranging from a minimum near $\sim$120 W m^{-2} to a maximum of $\sim$380 W m^{-2}.

For a midlatitude winter situation, we may use $a_{\text{lw}} = 0.8$ and $T_a = 260$ K, and take the initial surface temperature to be $T_s = 275$ K. Plugging these values into (6.41) yields a positive (upward) net flux of 117 W m^{-2}. On level ground with no wind, very little of this heat loss from the ground is shared with the overlying air, so the surface temperature of the ground falls rapidly.

A crude estimate of the rate of the temperature fall may be had by noting that heat conduction in soil is rather slow, so that only the top few centimeters of soil experience the fluctuation of temperature associated with the diurnal (day-night) cycle. If we somewhat arbitrarily choose an effective depth over which to average the cooling as $\Delta Z = 5$ cm, and use a typical soil heat capacity (per volume) of $C \approx 2 \times 10^6$ J m^{-3}K^{-1}, then we have

$$\frac{dT}{dt} \approx \frac{-F^{\text{net}}}{C\Delta Z} \approx -4.2 \ \frac{\text{K}}{\text{hr}} \ . \tag{6.42}$$

You can see that it will not take long for the ground temperature to drop below freezing and, presumably, below the frost point of the overlying air, at which point frost will start to deposit (or sublime) directly onto the surface.

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The above calculation assumed a cloud-free atmosphere. What happens when we introduce a low-level opaque cloud deck whose temperature is only a few degrees below that of the surface? Taking $a_{\rm lw} = 1$ and $T_a = 270$ K, we now find a net surface flux of only 22 W m^{-2}, or less than a fifth of value for a clear sky. The cooling rate of the surface is reduced by the same ratio and is now only 0.8 K / hr. What a difference a cloud makes!

Of course, other processes, such as surface latent and sensible heat fluxes, while small (unless there is a significant breeze!), partially offset the radiational cooling predicted by the above simple analysis. Nevertheless, you should now be persuaded that radiation can have observable meteorological effects even on rather short time scales.

6.4.5 Radiative Cooling at Cloud Top

A similar analysis can be applied to the top of a continuous cloud layer, such as the low-level marine stratocumulus clouds that are common within midlatitude and subtropical high pressure zones over relatively cool ocean areas, such as the Eastern Pacific near California and Peru or in the vicinity of the Azores near North Africa and the Iberian Peninsula.

Although the processes involved in the steady-state maintenance of these persistent cloud sheets is complex and includes turbulent fluxes of sensible heat and moisture, it is at least possible to evaluate the potential role of radiative fluxes as one component in the overall balance.

To begin with, because the cloud layer is opaque to LW radiation and does not reflect appreciably at these wavelengths, we can consider it as having two radiating "surfaces", each with emissivity $\varepsilon \approx 1$. The base height $Z_{\rm base}$ for marine stratocumulus clouds is commonly near 300 m altitude, below which the air follows a dry adiabatic temperature profile of 9.8 K / km. This implies a temperature difference between cloud base and the ocean surface of only 3 K. Taking the surface temperature to be $T_s = 288$ K, we have $T_{\rm base} = 285$ K and a net LW flux at cloud base of

$$F_{\rm net,base} < \sigma(T_s^4 - T_{\rm base}^4) \approx 16 \;\; {\rm W\ m^{-2}}\ . \tag{6.43}$$

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The use of the '$<$' sign is intended to indicate that the above is an *upper bound* on the net flux at cloud base, because not all the radiation reaching cloud base originates at the surface; some is emitted at higher, slightly cooler altitudes on account of the high opacity of the atmosphere at some wavelengths.

Within the cloud, the moist adiabatic lapse rate of approximately 6 K/km prevails. Therefore, at cloud top, typically near 1 km altitude, the temperature is approximately 4 K colder than at cloud base, yielding $T_{top} \approx 281$ K. Because marine stratocumulus clouds typically occur in regions of high pressure and therefore subsiding air aloft, the atmosphere above cloud top is often warm but very dry. The downward flux of radiation from the overlying atmosphere may therefore be estimated as $a_{lw}\sigma T_a^4$, where we assume $T_a \approx 280$ K and $a_{lw} \approx 0.8$:

$$F_{net,top} \approx \sigma\left(T_{top}^4 - a_{lw} T_a^4\right) \approx 211 \ \text{W m}^{-2} . \tag{6.44}$$

In summary, we have substantial radiative cooling ($\sim$211 W m^{-2}) at cloud top but relatively little radiative warming ($\sim$16 W m^{-2}) at cloud base. There is therefore net radiative cooling of $\sim$195 W m^{-2} for the cloud layer as a whole. Note, by the way, that any time you cool a layer of air from above, you increase the lapse rate. Once the lapse rate exceeds the applicable adiabatic lapse rate (in this case the moist adiabatic lapse rate), you destabilize the layer, leading to vertical overturning. The same applies to warming from below. Consequently, the net radiative cooling experienced by the cloud ends up being distributed pretty much throughout the entire atmospheric boundary layer (or mixed layer), which extends from the surface to cloud top. In the present example, this represents $\sim$195 W m^{-2} of cooling distributed over a 1 km depth of air. We can compute the cooling rate as

$$\left.\frac{dT}{dt}\right._{rad} \approx \frac{F_{net,base} - F_{net,top}}{C_p \rho \Delta Z} \approx -14 \ \text{K/day} , \tag{6.45}$$

where $C_p = 1004\,\text{J kg}^{-1}\,\text{K}^{-1}$ is the heat capacity at constant pressure of air, $\rho \approx 1.2$ kg m^{-3} is the density of air near the surface, and $\Delta Z = 1$ km.

In reality, we know that the temperature in the marine boundary layer is close to a steady state, implying that there must be compensating inputs of energy in the form of surface latent and sensible