6.4.4 Nighttime Radiative Cooling

In the previous section, we discussed the criteria for radiative equilibrium in a simple two-level model, based on an assumed balance between the *long-term globally averaged* LW fluxes and the *long-term globally averaged* SW flux from the sun. Of course, the solar input at any location is not constant with time but varies from zero at night to a maximum at around noon, as well as varying with latitude and season. Also both SW and LW fluxes at the surface vary sharply with temperature, humidity, and cloud cover. Although it remains a very crude representation of reality, we can nevertheless invoke our two-level model to learn a few things about *short-term, local* radiative processes.

In particular, let us examine the basis for the well-known rule of thumb that dew or frost is far more likely to occur on clear nights than cloudy nights. Both dew and frost occur when the surface cools to below the dewpoint or frostpoint temperature of the air, while the air above the surface may remain at a significantly warmer temperature. The latter observation rules out direct thermal conduction (through physical contact of the air with the surface) as the reason for the cooling of the surface, as heat always conducts from warm to cold. The only other obvious candidate is radiation.

Because we are addressing a night-time situation, we don't need to worry about SW fluxes. Furthermore, we're not interested in establishing the criteria for radiative equilibrium but rather in estimating the surface cooling rate under conditions of disequilibrium. We therefore need only consider flux components 6 and 8 in Fig. 6.8, which we will refer to simply as $F^{\downarrow}$ and $F^{\uparrow}$ for the purposes of this discussion:

$$F^{\downarrow} = a_{\mathrm{lw}}\sigma T_a^4 \,, \tag{6.38}$$

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

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

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

or

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

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 $a_{\mathrm{lw}} = 0.8$ 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

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

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.

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 $a_{lw} = 1$ 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.

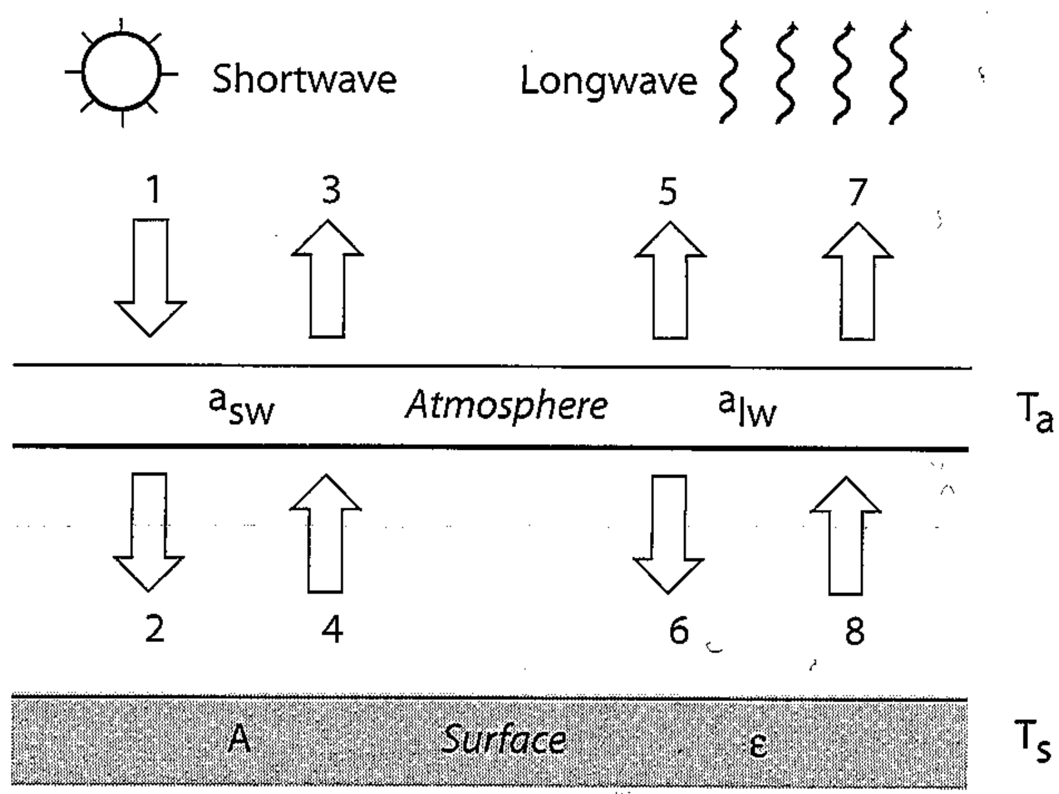

Fig. 6.8: Schematic depiction of the radiative coupling between the surface and a thin isothermal atmosphere.