Diurnal Cycle

A Betts, Atmospheric Research, Pittsford, VT, USA

Synopsis

The diurnal cycle over land is driven by solar heating in the daytime, and longwave cooling at night. In summer, the maximum temperature decreases with increasing cloud cover, because clouds reflect sunlight. In winter, the minimum temperature falls steeply under clear skies, because clouds reduce the longwave cooling to space. Over moist soils, increased evaporation reduces the diurnal temperature and humidity ranges. A few hours after sunrise, there is a transition when the nighttime stable layer is eroded by surface heating. Carbon dioxide shows a dawn maximum as nighttime respiration is trapped near the surface, and an afternoon minimum.

Introduction

Near the Earth's surface, many variables have a characteristic *diurnal or daily* cycle, driven by the diurnal cycle of the incoming solar radiation, which is zero at night and peaks at local noon. The atmosphere is relatively transparent to the short-wave radiation from the sun and relatively opaque to the thermal radiation from the Earth. As a result, the surface is warmed by a positive net radiation balance in the daytime in summer, and cooled by a negative radiation balance at night. The surface temperature oscillates almost sinusoidally between a minimum at sunrise and a maximum in the afternoon. This is referred to as the diurnal cycle of temperature. In warm seasons, the daily net radiation balance is positive, and the daily mean temperature is determined by the daily mean surface energy balance, which involves not only the short- and long-wave radiation components, but also heat transfers to the atmosphere.

The magnitude of this diurnal range of temperature is determined by many factors, which we shall discuss. Clouds have a large impact on the surface radiation balance and this differs between warm and cold seasons. The nature of the underlying surface is important, whether land or water, and so is the coupling to the atmosphere above. The phase change of water, particularly evaporation and condensation plays an important role in moderating the diurnal range of temperature, because of the large latent heat of vaporization. (In cold climates, the freezing and thawing of the soil are also important on the seasonal timescale.)

Over the ocean (and large lakes), the diurnal temperature range is small, because the incoming solar energy is mixed downward into the ocean 'mixed layer,' which is usually tens of meters deep. One day of solar heating will warm a layer of water 50 m deep to less than 0.1 K, because of its large thermal capacity. Only in light winds, when the downward mixing is small, does the diurnal range of sea surface temperature reach 1 K. On timescales longer than the diurnal, evaporation of water primarily balances the surface net radiation budget.

Over land, only a small fraction (<20%) of the net radiation at the surface is conducted downward in the daytime, or stored by warming trees on the surface, for example. As a result, the surface temperature rises rapidly after sunrise, until near balance is achieved between the net radiation and the direct transport of heat to the atmosphere (referred to as the *sensible heat flux*) and evaporation of water (or transpiration from plants), referred to as the *latent heat flux*. If the surface is a desert, then the daytime temperature rise is large, but if water is readily available for transpiration, the daytime rise of temperature is greatly reduced, because most of the net radiation goes into the latent heat of vaporization. The surface sensible and latent heat fluxes have a large diurnal cycle, with a peak near local noon, as they are driven primarily by the incoming solar radiation. The surface temperature peaks a little later in the afternoon, when the surface sensible heat flux goes negative as the surface cools.

Coupling of the Summer and Winter Diurnal Cycle to Clouds

The diurnal cycle over land is driven by the surface net radiation balance, which we may write

$$R_n = SW_n + LW_n \quad [1]$$

where the net radiation, R_n, is the sum of the net shortwave radiation, SW_n a heating term, and the net longwave radiation, LW_n a cooling term. The net shortwave flux is reduced by reflection at the surface, which we call the surface albedo, α_s; and by the reflection and absorption by the cloud fields above, which we call the effective cloud albedo (ECA). So we can write the reduction of the downward clear-sky flux, SW_{dn}(clear) in the form

$$SW_n = (1-\alpha_s)(1-\mathrm{ECA})SW_{dn}(\mathrm{clear}) \quad [2]$$

Grassland or crops have α_s in the range 0.15–0.2, but this rises to 0.6–0.8 with snow cover. The boreal forest has a lower α_s of order 0.1, and this rises to about 0.4–0.5 with snow on the canopy, and falls back to 0.2 once the canopy sheds the snow. This is because snow on the ground is largely in the shade of the trees. The ECA has an even wider range: zero with a clear sky by definition, and >0.9 in overcast conditions with heavy rain. So cloud cover is an important modulator of the diurnal cycle. We will illustrate this using Canadian Prairie data, which have long-term records of opaque or reflective cloud cover that have been calibrated to the ECA.

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Figure 1(a) shows that in summer (June–July–August) the amplitude of the diurnal cycle increases as daily mean opaque cloud cover falls from 95 to 5%. In summer, near-surface temperature is minimum at sunrise, but this falls only slightly when cloud cover is small. On the other hand, the afternoon temperature maximum increases steeply as cloud cover decreases, and SW_n increases. Cooled by the net outgoing longwave radiation, LW_n, the surface temperature falls at night from the afternoon maximum back to the sunrise minimum.

Figure 1(b) shows an inverted pattern for winter (December–January–February): it is coolest with near-clear-sky conditions. This shows that there are fundamental differences between the surface radiation balance with clouds between summer and winter. The sun is low in the sky in winter, and the Prairies have a high surface albedo with snow, so SW_n in eqns [1] and [2] becomes small. However, clouds also blanket the Earth and greatly reduce LW_n, the surface cooling to space, and this longwave effect of clouds becomes dominant in winter. We see that the diurnal range of temperature is very small with 95% cloud cover. As cloud cover decreases, the surface cooling increases and the temperature drops. The minimum temperature at sunrise falls the most, because in the daytime the LW_n cooling is partly offset by SW_n heating.

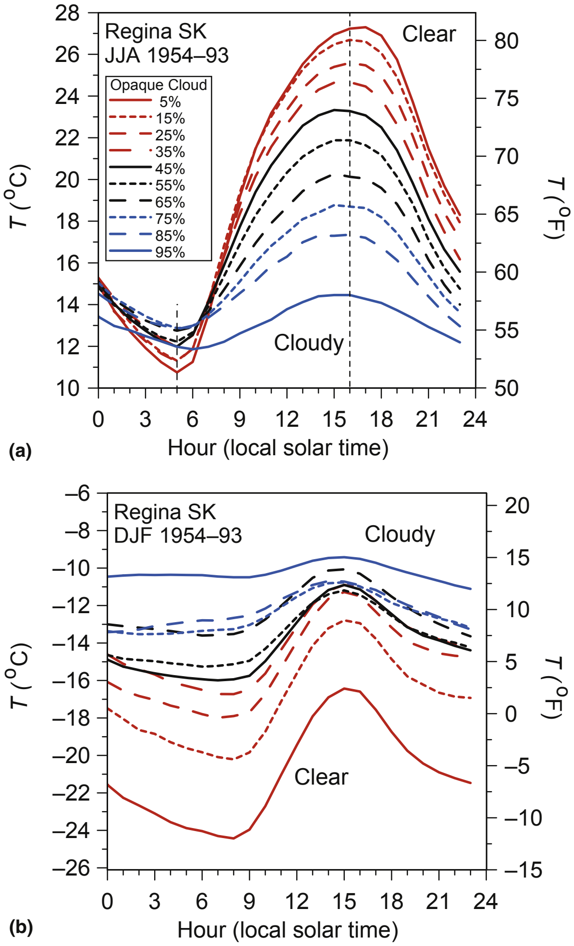

Figure 1 Dependence of diurnal cycle of temperature on opaque cloud cover in (a) summer and (b) winter for Regina, Saskatchewan, Canada.

Dependence of Diurnal Cycle on Evaporation

Figure 2 illustrates this diurnal variation using data from sunny days in midsummer during a 1987 field experiment (with acronym FIFE) conducted over grassland near Manhattan, Kansas. The panels on the left (from top to bottom) show net radiation, R_n, sensible heat flux, H, and latent heat flux, LE. The surface energy balance can be written as

$$R_n = H + LE + G \qquad [3]$$

where G is the storage in the ground and vegetation, which we do not show. In addition a small amount of energy goes into photosynthesis, which again we do not show. The time axis is local solar time, which is UTC-6 h. The data have been grouped and averaged based on the percent soil moisture (SM) in the first 10 cm of soil, so that there are three curves (each an average of about 10 days) representing dry, medium, and wet soils. The upper left panel shows that the mean net radiation on these sunny days is very similar. However because SM is a major control on evaporation, the partition of the net radiation into sensible and latent heat is very different. When the soil is wet, the latent heat flux (or 'evaporative energy' flux) is about three times the sensible heat flux, whereas when the soil is dry, these two fluxes are nearly equal. The panels on the right side show the response to the different surface forcing. The upper right panel shows the surface temperature (measured by an infrared radiation thermometer, mounted on a tower and pointed downward at the grass). Although R_n is the same, on days when the soil is dry and water is not readily available for evaporation, the surface gets very hot, as warm as 44 °C near noon. This warm surface temperature drives the large sensible heat flux H and heats the air above the surface. The diurnal range of the surface temperature is more than 20 °C on these days, while for the air at 2 m above the surface in the middle panel, the diurnal range is only 12 °C. As SM increases, the daily maximum surface and air temperature decrease. The upper two panels on the right are similar, except that the amplitude of the surface temperature is larger than that of the air temperature. Both are related to the sensible heat flux H. Note that the air temperature has a broad afternoon maximum, because H is upward as long as the surface is warmer than the air. The surface temperature falls below the air temperature only in late afternoon, H then changes sign, and at night the surface is cooler than the air. The lower right panel shows the diurnal cycle of relative humidity (RH) as a percentage. Over the wetter soils, the RH of the air at 2 m reaches 85% before sunrise, and falls in the daytime as the surface and air warm. The fall of RH is smallest on the days with the greatest evaporation, LE. When evaporation is reduced because the soil is dry, daytime RH falls as low as 30% and even at night, only reaches 72% at sunrise.

Coupling between the Surface Diurnal Cycle and the Atmospheric Mixed Layer

As the land surface is heated during the daytime, a dry convective boundary layer grows in depth. This is called the

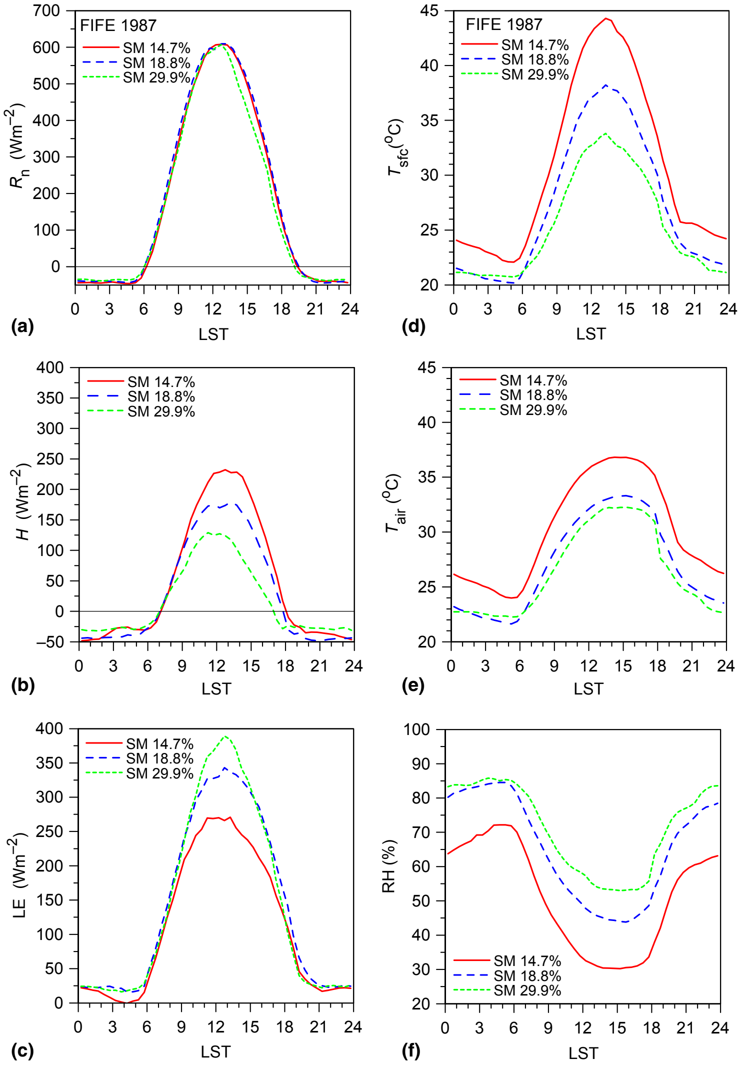

Figure 2 Diurnal cycle of net radiation (R_n), sensible heat flux (H), and latent heat flux (LE) (a–c); surface temperature, air temperature, and RH (d–f), stratified by percent volumetric SM in first 10 cm of the soil.

'mixed layer,' because the turbulent dry convection rapidly stirs the layer to one of near-neutral buoyancy and near-constant water vapor mixing ratio. The diurnal cycle of the surface and the mixed layer are tightly coupled. As a result the pre-existing atmospheric structure above the surface at sunrise has a considerable impact on the daytime diurnal cycle, as illustrated in the following figures using surface and sounding data collected over the boreal forest in Saskatchewan, Canada during the Boreal Ecosystem-Atmosphere Study (BOREAS) in 1994. Figure 3 shows the surface diurnal cycle for 2 days in spring. The upper panel shows for each day the temperature at two levels, an upper level T_U, which is at 21 m, about 5 m above the canopy of a jack pine forest, and a lower level about 5 m above the forest floor. On both days the surface cools strongly at night and rises steeply after sunrise with a greater diurnal range than the near-clear-sky case in Figure 1. The diurnal range under the canopy is larger than above it. At night on 26 May, the winds are lighter, and the near-surface nighttime boundary layer is more stable (see Figure 4). The air under the canopy becomes effectively decoupled from the atmosphere above and the stable temperature gradient across the canopy at night reaches 7 K. There is very little evaporation from either the forest, or the cold lakes at this time in spring. The lower panel shows RH measurements above the canopy. In the late afternoon, RH falls as low as 20% on 31 May. Before sunrise on this day, RH above the canopy reaches 90% as T_U falls to a minimum of 4 °C. RH was not measured below the canopy, but the temperatures there are cold enough to saturate the air in the hours before sunrise. The dew point is often used to estimate minimum nighttime temperatures at the surface.

The right-hand scale of the upper panel shows the corresponding dry potential temperature, which is defined as

$$\theta = (T + 273.15)(1000/p)^{.286} \quad [4]$$

where p is the surface pressure (here about 950 hPa, since the observation site is about 500 m above sea level). The potential

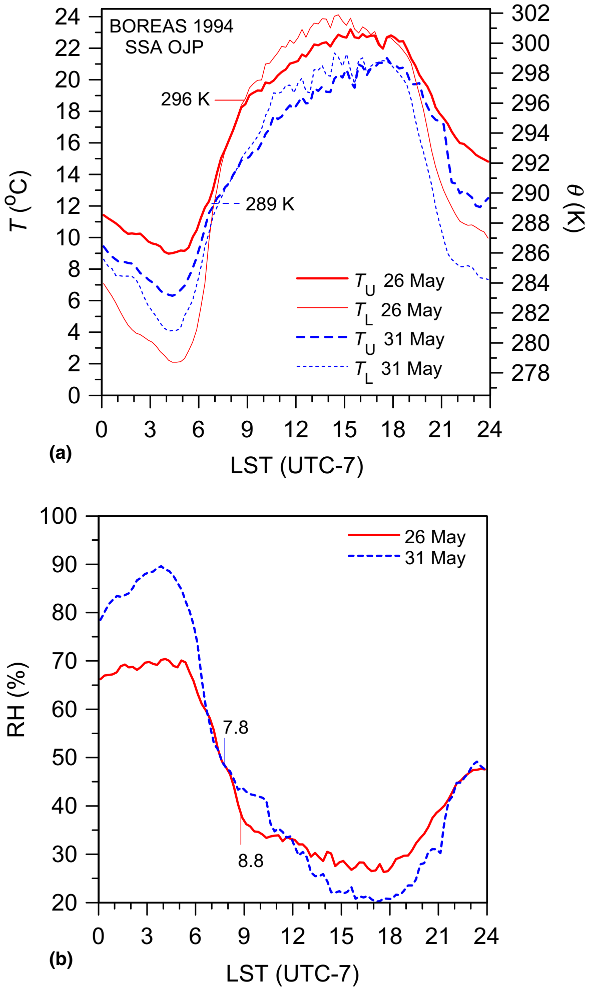

Figure 3 Diurnal cycle of temperature, above and below a boreal forest canopy (a), and of RH above the canopy (b) for 2 days in May 1994.

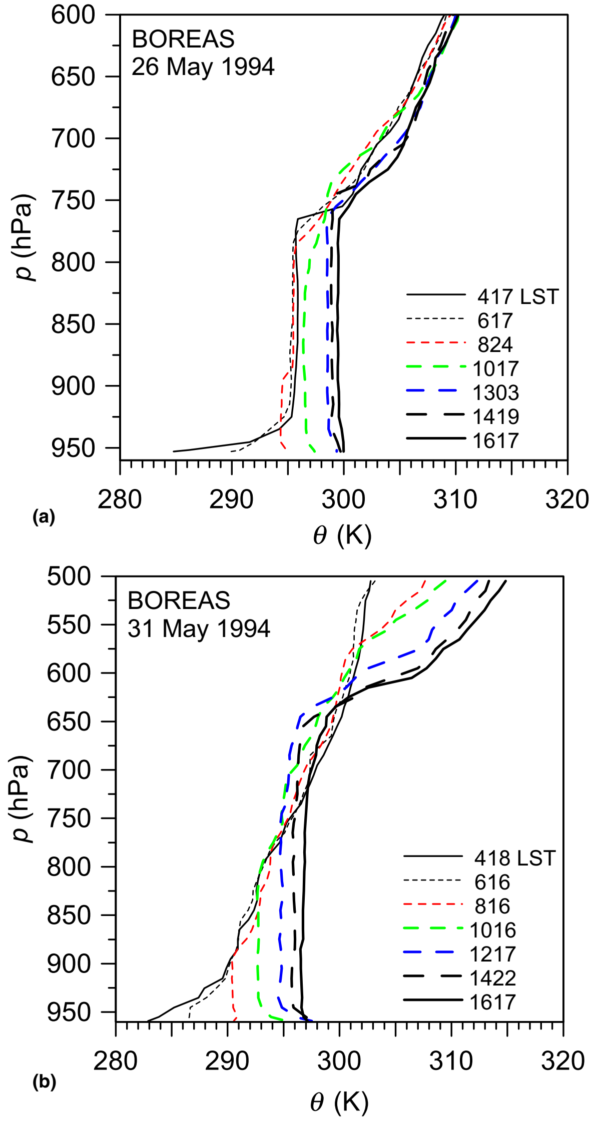

Figure 4 Profiles of potential temperature in the lower troposphere on (a) 26 May and (b) 31 May.

temperature, θ, is useful as a variable because it allows us to compare the surface and atmosphere above. During the daytime the boundary layer above the surface is mixed to almost constant potential temperature (see Figure 4). The strong radiative cooling of the surface at night generates a stable layer close to the ground, typically only a few hundred meters deep. About 3–4 h after sunrise, the surface has warmed enough to remove this stable surface layer and reconnect to a deeper layer. When this happens, the rate of rise of temperature and the rate of fall of RH decrease sharply. In Figure 3, this occurs on 26 May at a local time of 8.8 h, when θ reaches 296 K; while on 31 May, it occurs at 7.8 h, when $\theta = 289$ K, and on this day the change is smaller.

Figure 4 shows sequences of seven profiles of potential temperature in the lower troposphere, measured by rawinsonde ascents, nominally every 2 h from sunrise to late afternoon on the 2 days. The upper panel shows at sunrise (417 LST, solid) a cold (stable) surface layer only about 25 hPa deep (200 m), with a deep layer above of constant θ, which is the residual or 'fossil' mixed layer from the previous day. At the surface the temperature warms rapidly, as the surface sensible heat flux is trapped in this shallow surface layer. The profile at 0824 LST shows a mixed layer with $\theta = 294.5$ K to 890 hPa. Shortly afterward, when the surface potential temperature reaches $\theta = 296$ K, the new growing boundary layer merges with the deep residual mixed layer. From then on, the surface and mixed layer warm much more slowly, as seen in Figure 3. Even though H exceeds 300 W m^{-2} at all the forest sites for several hours around local noon (not shown), this large heat flux is distributed through a deep layer.

The lower panel shows the time sequence on 31 May. Note that at sunrise (solid), the profile is quite different than on 26 May. Instead of a deep layer of constant θ, produced by dry convection the previous day (a so-called dry adiabatic structure), there is a layer from 920 to 650 hPa in which

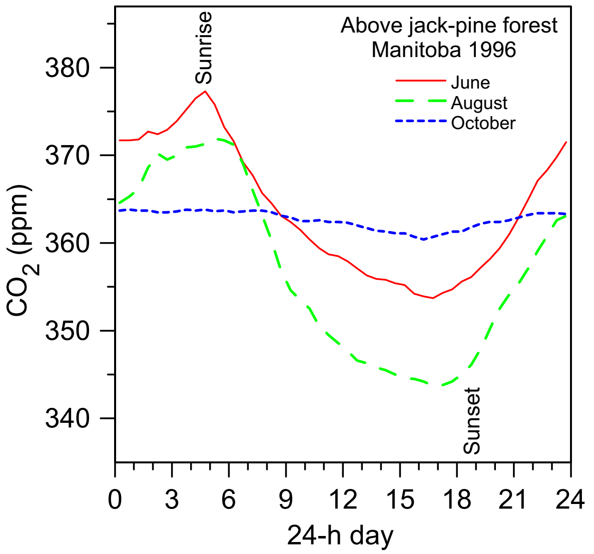

Figure 5 Monthly mean diurnal cycle of CO_2 for June, August, and October for a boreal jack pine site. Data from McCaughey, J.H., Dr., Queens University, Kingston, Ontario.

θ increases steadily with height. In fact, this layer was produced by showers the previous evening (and it has a so-called wet adiabatic structure). The change in slope of the early morning profile at 920 hPa is at $\theta = 289$ K, and hence we see on Figure 3, a change in the rate of warming, once the surface reaches this potential temperature. This change of slope is more dramatic on 26 May, because the change in the vertical profile is also greater. On 31 May, the mixed layer grows steadily all day until it is 300 hPa deep (about 3000 m) in the late afternoon. On both these days, there is some broken cumulus cover in the afternoon at the top of the mixed layer. The rapid warming on 31 May, that is seen between 500 and 600 hPa, is related to the lowering and change in structure of a powerful jet stream above, not by surface processes.

Diurnal Cycle of CO_2

The diurnal cycle of the solar radiation drives a diurnal cycle in CO_2 through photosynthesis and respiration in plants. Figure 5 shows the mean diurnal cycle over a young jack pine canopy (about 5–6 m tall) near Thompson, Manitoba from the 1996 BOREAS experiment for 3 months, June, August, and October. During the summer months, CO_2 decreases during the daylight hours as it is taken up in photosynthesis, and increases at night as it is released by respiration from both plants and soil. The amplitude of the diurnal cycle increases from June to August, but the monthly mean decreases as there is a net CO_2 uptake by the entire Northern Hemisphere. By October of this year however the diurnal cycle is very small, as temperatures have dropped low enough that both photosynthesis and respiration have almost ceased.

See also: **Clouds and Fog:** Climatology. **Numerical Models:** Clouds. **Radiation Transfer in the Atmosphere:** Radiation, Solar.

Further Reading

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