

Figure 2.4 Schematic representation of airflows in the outer layer, (a) The daytime mixed layer, and the associated vertical profiles of the sensible heat flux density (Q_H) and potential temperature (θ) within a mixed layer of depth h^* . (b) The corresponding conditions in the nocturnal stable layer. (Note—see p. 53 for the definition of θ .)

overlying warmer air and when they are repelled they transport heat downwards. This is referred to as *convective entrainment*.

At night, due to stability, turbulence is largely restricted to a shallow layer. Convection is entirely of mechanical origin. A true mixed layer is not present, but turbulent transfer removes heat from the lowest layer to form an inversion layer (Figure 2.4b). Turbulence may even become so weak that radiative flux divergence plays a role in shaping the temperature profile. At the top of the inversion layer turbulence may break down into wave-like oscillations.

2 SUB-SURFACE CLIMATES

(a) Soil heat flux (Q_g) and soil temperature (T_s)

Heat transfer and the thermal climate of soils are governed by four linked thermal properties: thermal conductivity, heat capacity, thermal diffusivity

and thermal admittance. As already noted, heat is conducted down into the soil by day and upwards at night. We can re-write equation 2.3 to express this sensible heat flux density (Q_G) as Fourier's law:

$$Q_G = -k_s \frac{\partial \bar{T}_s}{\partial z} \simeq -k_s \frac{(\bar{T}_2 - \bar{T}_1)}{(z_2 - z_1)} \quad (2.5)$$

where the subscripts refer to levels in the soil (see Figure 2.2). As before, the sign indicates the flux is in the direction of decreasing temperature, therefore by day when $\partial \bar{T}_s / \partial z$ is negative the equation gives a positive value of Q_G which is in accord with the flux convention adopted on p. 23; k_s is known as the *thermal conductivity* ($\text{Wm}^{-1}\text{K}^{-1}$), which is a measure of the ability to conduct heat. It is formally defined as the quantity of heat (J) flowing through unit cross-sectional area (m^2) of the substance in unit time (s), if perpendicular to it there exists a temperature gradient of 1 degree m^{-1} . Typical values for a range of natural materials are listed in Table 2.1; of these motionless air is the most notable because it is such a very poor conductor of heat (i.e. it is a good insulator).

Unfortunately k_s is not a simple constant for a given soil. It varies both with depth and with time. However if we restrict ourselves to bulk averages k_s depends upon the conductivity of the soil particles, the soil porosity, and the soil moisture content. Of these the soil moisture content is the only short-term variable for a given soil. The addition of moisture to an initially dry soil increases its conductivity (Figure 2.5a). This happens for two reasons. First, coating the soil particles increases the thermal contact between grains. Second, since the soil pore space is finite the addition of pore water must expel a similar amount of pore air. From Table 2.1 we can see that this means replacing soil air with a substance whose conductivity is more than an order of magnitude greater.

We have already introduced heat capacity (C) (p. 36). It relates to the ability of a substance to store heat and expresses the temperature change produced as a result of gaining or losing heat. The value of C_s for a soil can be calculated by evaluating the fractions of soil solid, water and air (Appendix A2, p. 375). The value for a given soil is strongly dependent, in an almost linear fashion, on the soil moisture content (Figure 2.5b). Adding water with a very high heat capacity excludes a proportionate volume of soil air of low heat capacity (see Table 2.1). The result is a reduction in the soil's thermal sensitivity.

The thermal diffusivity of a soil (k_{HS}) is its ability to diffuse thermal influences. It controls the speed at which temperature waves move and the depth of thermal influence of the active surface. From equations 2.3 and 2.5 we see that $k_{HS} = k_s / C_s$ which shows that thermal influences are directly proportional to the ability to conduct heat (k_s) but inversely proportional to the amount of heat necessary to effect temperature change (C_s). Thermal diffusivity may be viewed as a measure of the time required for temperature

Table 2.1 Thermal properties of natural materials

Material	Remarks	ρ Density (kg m^{-3} $\times 10^3$)	c Specific heat ($\text{J kg}^{-1} \text{K}^{-1}$ $\times 10^3$)	C Heat capacity ($\text{J m}^{-3} \text{K}^{-1}$ $\times 10^6$)	k Thermal conductivity ($\text{W m}^{-1} \text{K}^{-1}$)	κ Thermal diffusivity ($\text{m}^2 \text{s}^{-1}$ $\times 10^{-6}$)	μ Thermal admittance ($\text{J m}^{-2} \text{s}^{-1/2} \text{K}^{-1}$)
Sandy soil (40% pore space)	Dry	1.60	0.80	1.28	0.30	0.24	620
Clay soil	Saturated	2.00	1.48	2.96	2.20	0.74	2550
	Dry	1.60	0.89	1.42	0.25	0.18	600
Peat soil	Saturated	2.00	1.55	3.10	1.58	0.51	2210
	Dry	0.30	1.92	0.58	0.06	0.10	190
Snow	Saturated	1.10	3.65	4.02	0.50	0.12	1420
	Fresh	0.10	2.09	0.21	0.08	0.10	130
Ice	Old	0.48	2.09	0.84	0.42	0.40	595
	0°C, pure	0.92	2.10	1.93	2.24	1.16	2080
	4°C, still	1.00	4.18	4.18	0.57	0.14	1545
Water*	10°C, still	0.0012	1.01	0.0012	0.025	21.50	5
Air*	Turbulent	0.0012	1.01	0.0012	~125	~10 × 10 ⁶	390

*Properties depend on temperature, see Appendix A3.
Sources: van Wijk and de Vries (1963), List (1966).

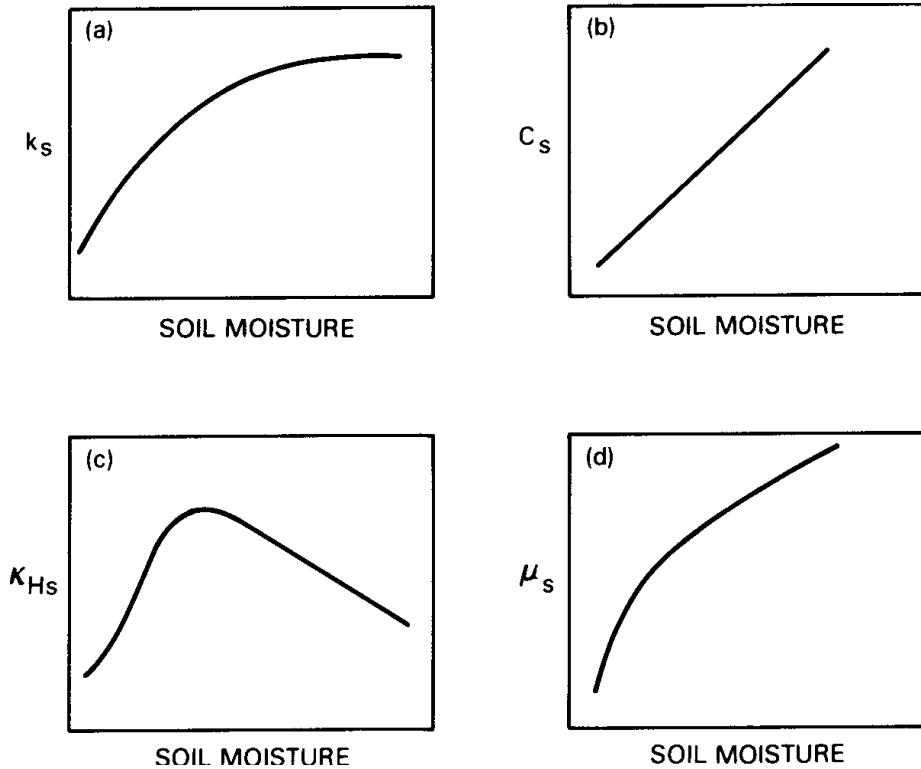


Figure 2.5 Relationship between soil moisture content: (a) thermal conductivity, (b) heat capacity, (c) thermal diffusivity and (d) thermal admittance for most soils.

changes to travel. For example, the daytime heat input will generate a temperature wave that travels rapidly and to considerable depth in soil where conductivity is high, but if it takes large amounts of heat to warm intermediate layers because of a high heat capacity it will be slowed and not penetrate as far. Typical values of μ_s are given in Table 2.1.

The value of μ_s is obviously affected by the same soil properties that influence k and C_s , especially soil moisture (Figure 2.5c). Note that adding moisture to a dry soil initially produces a sharp increase in μ_s by increasing thermal contact and expelling soil air (i.e. increasing k as in Figure 2.5a). However, in most soils beyond about 20% soil moisture content by volume μ_s begins to decline. This happens because whereas k levels off (Figure 2.5a) the value of C_s continues to increase at higher moisture contents (Figure 2.5b).

Soils with high diffusivities allow rapid penetration of surface temperature changes and permit these effects to involve a thick layer. Thus for the same heat input their temperature regimes are less extreme than for soils with low diffusivities. By day the surface heating is used to warm a thick layer of soil, and at night the surface cooling can be partially offset by drawing

upon heat from a similarly thick stratum. Soils with poor diffusivities concentrate their thermal exchanges only in the uppermost layer, and consequently experience relatively extreme diurnal temperature fluctuations. Therefore, in general a wet clay has a conservative thermal climate, whereas an almost dry peat is extreme.

This is a convenient juncture to introduce a related thermal property called the soil *thermal admittance* ($\mu_s = C_s \kappa_{Hs}^{1/2} = (k_s C_s)^{1/2}$). Technically this is a *surface* rather than a soil property. It is a measure of the ability of a surface to accept or release heat since it expresses the temperature change produced by a given heat flux change. This is why some materials with high μ (such as metals) initially feel cooler to the touch than those with low μ (such as wood) even though both are actually at the *same* room temperature of, say, 20°C (for values see Table 7.4). Your finger is at a temperature of about 30°C and the 10-degree temperature difference is rapidly sensed at a surface of high μ because heat transfer at the finger-material interface is excellent. Similarly, note the *apparent* difference of placing your bare foot on a clay tile bathroom floor compared with a bathmat or cork tile. On the other hand, if the materials being touched are at a higher temperature than your hand or foot, that with higher μ would initially feel warmer.

A surface also has an analogous atmospheric admittance ($\mu_a = C_a K_H^{1/2}$). The relative magnitude of the two properties are important in determining the sharing of sensible heat between the soil and the atmosphere since:

$$\mu_s / \mu_a = Q_G / Q_H$$

The amplitude of the surface temperature wave is closely linked to these properties. The larger these values are the easier it is for heat to be transported to or from the interface and the smaller will be the surface temperature variations. For a given site the value of μ_a is determined by the state of turbulence (i.e. the eddy diffusivity, K_H), and μ_s by the soil moisture (Figure 2.5d). For a given state of the atmosphere, sites with large μ_s (Table 2.1) will accept or release heat to or from soil storage with relative ease and hence will exhibit relatively small surface temperature changes through a day.

The course of soil temperature with time is very regular in comparison with any other atmospheric element. Typical soil temperature variations at a number of depths on a cloudless day are given in Figure 2.6a. The near-surface temperature variation is wave-like and agrees closely with that of the surface. The wave penetrates downward to lower depths, but in doing so its amplitude decreases, and the times of maximum and minimum temperature are lagged (shift to the right in time). Both features depend on k_s . The wave amplitude at any depth $(\Delta \bar{T}_s)_z$ is given by:

$$(\Delta \bar{T}_s)_z = (\Delta \bar{T})_0 e^{-z(\pi / \kappa_{Hs} P)^{1/2}} \quad (2.6)$$

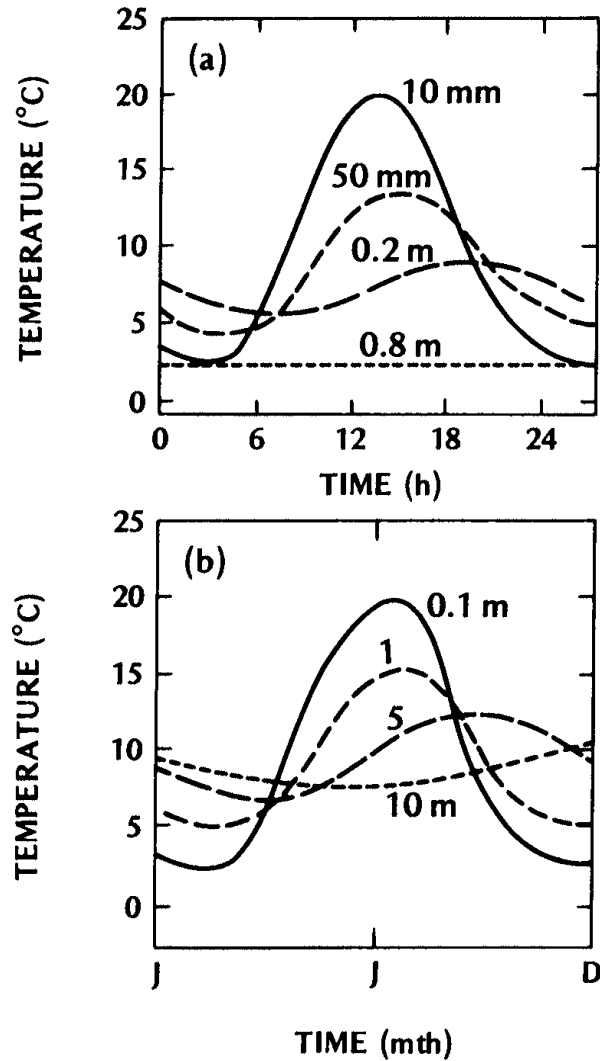


Figure 2.6 Generalized cycles of soil temperature at different depths for (a) daily and (b) annual periods.

where, $(\Delta \bar{T})_0$ —surface temperature wave amplitude, e —base of Napierian logarithms, P —wave period (s). This shows that the diurnal temperature range decreases exponentially with depth. In most soils the daily surface temperature wave is only discernible to a depth of about 0.75 m. In soils with low k_s values it is even less, indicating that flux convergence has extinguished Q_G in a thin near-surface layer. The time lag for the wave crest (maximum) and trough (minimum) to reach lower depths is given by:

$$(t_2 - t_1) = \frac{(z_2 - z_1)}{2} (P/\pi\kappa_{Hs})^{1/2} \quad (2.7)$$

where t_1 and t_2 are the times at which the wave crest or trough reaches depths z_1 and z_2 . Note that because of this time lag, at any given time the

soil may be cooling in its upper layers but warming at only a short distance beneath, and vice versa when the upper layers are warming (Figure 2.6a).

The annual soil temperature regime (Figure 2.6b) follows a wavelike pattern entirely analogous with the diurnal one. The wave period is of course dependent upon the annual rather than the daily solar cycle. With that adjustment, equations 2.6 and 2.7 still apply to the wave amplitude and lag with depth. With a longer period the wave amplitude decreases less rapidly with depth, and the depth of the affected layer is much greater than for the diurnal case. In fact the difference between the depth of penetration of the annual and daily waves is approximated by the square root of the ratio of their respective periods, i.e. $(3.15 \times 10^7 \text{s} / 8.64 \times 10^4 \text{s})^{1/2} \approx 19$. Therefore given the daily depth is 0.75 m in typical soils, the annual wave may penetrate to about 14 m. This would then be termed the depth of zero annual range. The temperature at this depth is sometimes used as a surrogate for the average annual air temperature of the site. This is based on the premise that long-term thermal equilibrium exists between the soil and the atmosphere. During the warm season soil temperatures decrease with depth and the associated downward heat flux builds up the soil's heat store. In the cold season the gradient is reversed and the store is gradually depleted. The spring and autumn are transitional periods when the soil temperature gradient reverses sign. These reversals (or 'turnovers') are important biological triggers to soil animals and insects. In the spring they may come out of hibernation, and/or move upwards towards the warmer surface layers. In the autumn they retreat to depths where soil warmth is more equable.

The effects of cloud on the diurnal soil temperature pattern are fairly obvious. With overcast skies absolute temperatures are lower by day but warmer at night, and the wave amplitude is smaller; variable cloudiness induces an irregular pattern upon the diurnal wave. Using the same argument as for the annual/daily comparison a 15 min temperature variation induced by cloud would only travel about one tenth of the distance of the daily wave, so typically it would not be registered below about 75 mm. Rainfall is capable of either increasing or decreasing soil temperatures depending upon the temperature of the rain in comparison with the soil. It is also capable of transporting heat as it percolates down through the soil. The effects of a snow or vegetation cover over the soil are dealt with in Chapters 3 and 4, respectively.

Appendix A2 (pp. 359 and 374) provides examples of the instruments for measuring soil heat flux and soil temperature in the field.

(b) Soil water flow (J) and soil moisture (S)

Soil moisture is usually expressed in one of two ways. *Soil moisture content (S)* is a measure of the actual water content, and is defined as the percentage

volume of a moist soil occupied by water. This is particularly pertinent in water balance studies (p. 31) where changes in mass are important. *Soil moisture potential* (ψ) on the other hand is an indirect measure of water content, and may be visualized as the energy necessary to extract water from the soil matrix. The units of ψ are those of negative pressure ($\text{Pa} = \text{mb} \times 10^{-2}$), which can also be expressed as a head of water displaced (1 m head of water = 1.0×10^4 Pa). This concept is of value in estimating the availability of water for plant use, and in calculating moisture movement. Methods for determining these two terms are given in Appendix A2 (p. 388).

The forces which bind soil water are related to the soil porosity and the soil water content (Figure 2.7). The forces are weakest in the case of open textured, wet soils, and greatest for compact dry soils. Thus at a given value of S the water potential is greatest for a clay, least for a coarse sand, and intermediate for a loam. Similarly in a given soil the potential increases as S decreases, but not in a linear fashion. It is relatively easy to extract moisture from a wet soil but as it dries out it becomes increasingly difficult to remove additional units. Figure 2.7 shows that in the range of water potentials which permit plants to extract soil water, sand has the least available water (7% vol.) and a silt loam the most (16% vol.).

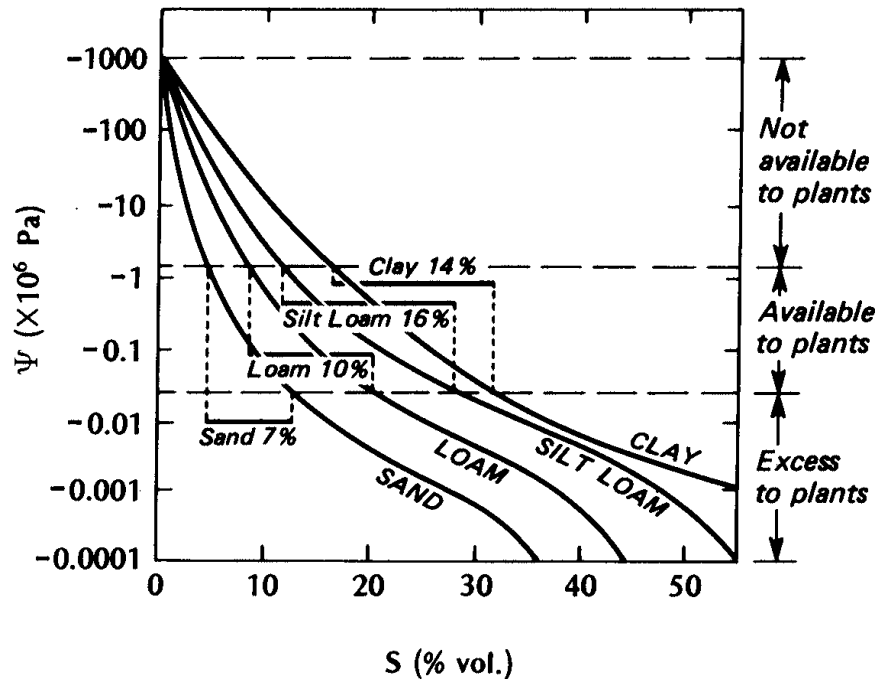


Figure 2.7 Relationship between soil moisture potential (ψ) and soil moisture content (S) in soils with different textures. Heavy horizontal bars show the volumetric water available to plants (modified after Buckman and Brady, 1960).

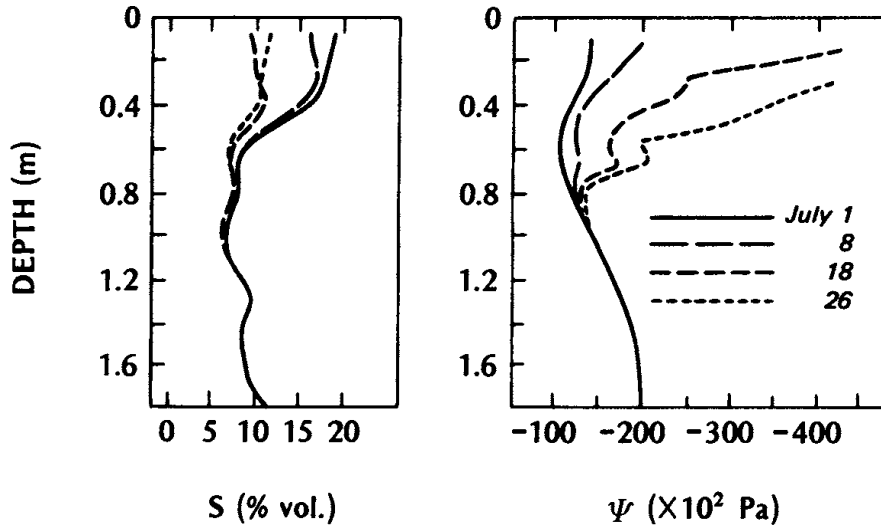


Figure 2.8 Profiles of soil moisture content (S) and soil moisture potential (ψ) in a sandy loam during a drying phase (after Rouse and Wilson, 1972).

Figure 2.8 conveniently illustrates these two concepts of soil moisture. It shows profiles of both S and ψ at the same site during a period of one month when the soil was almost continually drying-out. As S decreased throughout the period, the value of ψ increased. It is possible to utilize such changes in soil moisture content to estimate evapotranspiration losses to the atmosphere by integrating the area between successive profiles in time. In this case it appears as though the soil down to a depth of 0.8 m participated in the soil-air exchange.

The vertical flux of soil water (J) in the absence of percolating rain, is composed of both liquid (J_l) and vapour (J_v) flow. The movement of liquid moisture in a saturated or unsaturated soil may be considered analogous with the flux-gradient relationship for heat (equation 2.3). In this case the flux of water is related to the vertical water potential gradient by Darcy's Law:

$$J_1 = -K_f \frac{\partial \bar{\Psi}}{\partial z} \quad (2.8)$$

where K_f —hydraulic conductivity. The effect of evapotranspiration is to create a potential gradient which becomes greater than the opposing gravitational gradient and encourages the upward movement of water from low to high water potential. Unfortunately in unsaturated soils K_f is not a constant, but depends on both S and ψ as well as other soil factors.

The flow of water vapour (J_v) obeys the flux-gradient relationship of equation 2.4b:

$$J_v = -\kappa_{va} \frac{\partial \bar{p}_v}{\partial z}$$

where k —molecular diffusivity for water vapour (see Appendix A3, p. 393) and p^{va} —vapour density (p. 63) or water vapour concentration. The air in the pore spaces of moist soils is in close contact with soil water and is therefore commonly close to being saturated. The saturation vapour concentration is directly related to temperature (see Figure 2.15), being higher at higher temperatures. This creates a water vapour concentration gradient and a corresponding vapour flux from high to low soil temperatures. Thus noting the temperature distributions in Figures 2.2 and 2.6a, there is a tendency for vapour to flow down into the soil by day and up towards the surface at night. The nocturnal vapour flux in the soil commonly results in condensation upon the cold soil surface known as *distillation*. This is not the same process as dewfall (p. 67) which involves the turbulent transfer of atmospheric vapour down onto the surface.

3. SURFACE LAYER CLIMATES

(a) Lapse rates and stability

The dominant process in the lower atmosphere is convection (free, forced and mixed, see p. 16). A major control on the type and extent of convective activity is the vertical temperature structure as expressed in the concept of stability. Therefore it is helpful to explain stability before proceeding.

Consider a discrete parcel of air moving up through the atmosphere, and assume that it neither receives nor gives out heat to the surrounding air (such a parcel is said to be moving *adiabatically*). As it rises it encounters lower atmospheric pressure because the mass of air above it becomes progressively less. Thus the internal pressure of the parcel becomes greater relative to its surroundings and the parcel will tend to expand. To push away the surrounding air requires work and therefore energy. But the only energy available is the thermal energy of the parcel itself (since we are assuming no exchange with the surroundings), thus as the parcel rises it cools. In dry (unsaturated) air the rate of temperature change of such a parcel with height is the constant value of $9.8 \times 10^{-3} \text{ } ^\circ\text{Cm}^{-1}$ called the *dry adiabatic lapse rate* (Γ). If the parcel becomes saturated some vapour condenses into droplets thereby releasing latent heat (L) which reduces the rate of cooling, but the value is not constant. In most of the applications in this book (i.e. in the boundary layer below cloud base height) the dry adiabatic assumptions are approximately valid. Eventually of course the parcel will cease to rise and will impart its heat by mixing with the air at that level.

It is important however not to confuse the dry adiabatic lapse rate (Γ) with the *environmental lapse rate* (ELR). The former is the rate at which a dry parcel will cool if it is moved upward through the atmosphere, and also the rate at which it will warm if it moves down towards the ground (i.e.