Note to readers:

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The Physical Environment
A New Zealand Perspective

Edited by
Andrew Sturman and
Rachel Spronken-Smith

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The previous chapter has focused on regional and local climates and indicated how variable these are throughout New Zealand. This chapter discusses climate at an even smaller scale—that of microclimate. Consideration of site microclimate must begin with an understanding of the radiative and energy exchanges occurring at the surface, as it is these exchanges that result in the thermal and moisture characteristics of the site. These energy exchanges were discussed at the global scale in Chapter 4. It is surprising that despite the importance of such surface energetics in determining microclimate, relatively few environments in New Zealand have been studied with a view to specifying their full radiation budget and energy balance. This chapter discusses microclimates of several different surfaces and draws upon New Zealand examples where possible.

The radiation budget for a surface is given by:

\[ Q^* = K_\downarrow - K_\uparrow + L_\downarrow - L_\uparrow = K^* + L^* \]

where \( Q^* \) is the net all-wave radiation flux density (often referred to simply as net radiation); \( K_\downarrow \) is the incident short-wave radiation; \( K_\uparrow \) is the reflected short-wave radiation; \( K^* \) is the net short-wave radiation; \( L_\downarrow \) is the incident long-wave radiation; \( L_\uparrow \) is the emitted long-wave radiation; and \( L^* \) is the net long-wave radiation. The incident short-wave radiation, \( K_\downarrow \), is controlled by the azimuth and altitude of the Sun relative to the horizon and has a maximum at solar noon under clear skies. As discussed in Chapter 4, cloud and air pollution both decrease the amount of \( K_\downarrow \) received at a surface. At the microscale we must also consider factors such as shading by terrain, vegetation, and buildings, all of which can reduce \( K_\downarrow \). The reflected short-wave radiation, \( K_\uparrow \), depends on the magnitude of \( K_\downarrow \) and also on the short-wave reflectivity of the surface—called the albedo, \( \alpha \). Values of albedo for different surface types are given in Table 7.1.

The incident long-wave radiation emitted by the atmosphere \( (L_\downarrow) \) depends on the atmospheric temperature \( (T_a) \), emissivity \( (\varepsilon_a) \), and water vapour. If clouds are absent, then \( L_\downarrow \) is given by:
Table 7.1 Albedos and emissivities for a range of surface types.

<table>
<thead>
<tr>
<th>Surface</th>
<th>Albedo</th>
<th>Emissivity</th>
</tr>
</thead>
<tbody>
<tr>
<td>Snow</td>
<td>—old</td>
<td>0.40–</td>
</tr>
<tr>
<td></td>
<td>—fresh</td>
<td>0.95</td>
</tr>
<tr>
<td>Ice</td>
<td>—glacier</td>
<td>0.20–0.40</td>
</tr>
<tr>
<td>Grass</td>
<td>—long (1.0 m)</td>
<td>0.16–</td>
</tr>
<tr>
<td></td>
<td>—short (0.02 m)</td>
<td>0.26</td>
</tr>
<tr>
<td>Crops</td>
<td></td>
<td>0.18–0.25</td>
</tr>
<tr>
<td>Orchards</td>
<td></td>
<td>0.15–0.20</td>
</tr>
<tr>
<td>Forests</td>
<td>—deciduous—bare</td>
<td>0.15–</td>
</tr>
<tr>
<td></td>
<td>—leaved</td>
<td>0.20</td>
</tr>
<tr>
<td></td>
<td>—coniferous</td>
<td>0.05–0.15</td>
</tr>
<tr>
<td>Water</td>
<td></td>
<td>0.03–1.00</td>
</tr>
<tr>
<td>Urban</td>
<td></td>
<td>0.10–0.27</td>
</tr>
</tbody>
</table>

After Oke 1990

\[ L_\downarrow = e_a \sigma T_a^4 \]

where \( \sigma \) is the Stefan–Boltzmann proportionality constant \((5.67 \times 10^{-8} \text{ W m}^{-2} \text{K}^{-4})\). Since \( T_a \) and \( e_a \) do not change much during a day, \( L_\downarrow \) is relatively constant through the day unless cloud cover occurs, which increases \( L_\downarrow \) because clouds both absorb and re-emit \( L_\uparrow \) from the surface. The long-wave radiation emitted from a surface (\( L_\uparrow \)) similarly depends on surface temperature (\( T_o \)) and emissivity (\( e_o \)):

\[ L_\uparrow = e_o \sigma T_o^4 (1-e_o L_\downarrow) \]

The term in brackets accounts for the small amount of \( L_\downarrow \) that is reflected, but for many surfaces this term is small and can be neglected. Owing to the diurnal range of temperature, \( L_\uparrow \) varies considerably during the day, with a peak emission corresponding to the time of maximum solar heating, around solar noon.

The net all-wave radiation, \( Q^* \), is the most important energy flux because it represents the energy available to the environment. Typically \( Q^* \) is a surplus by day as net short-wave gain exceeds net long-wave loss, while at night \( Q^* \) is a deficit since there is no short-wave input (Figure 7.1).

The net all-wave radiation is the link between the radiation budget and the energy balance:

\[ Q^* = Q_{H} + Q_{E} + Q_{G} \]

where \( Q_{H} \) is the sensible heat flux density (often simply referred to as sensible heat); \( Q_{E} \) is the latent heat flux (often referred to as latent heat or evaporation, or
Figure 7.1 Radiative and energy fluxes for an ideal surface during (a) day and (b) night. (After Oke 1990)

During the daytime $Q^*$ is dissipated mainly through the convective fluxes, $Q_H$ and $Q_E$ (Figure 7.1). If water is freely available and the atmosphere is dry, creating a vapour gradient between the surface and the air, then $Q_E$ will be the dominant energy sink. This results in a low Bowen ratio, $\beta$, which is the ratio of $Q_H:Q_E$. As surface moisture decreases the magnitude of $Q_H$ increases and $\beta$ also increases. While $Q_E$ depends mainly on the vapour gradient, $Q_H$ depends on the temperature gradient between the surface and the air. At night convection is suppressed, so that $Q_H$ and $Q_E$ are small and may be directed towards the surface (i.e. there could be a temperature inversion, which occurs when the surface is cooler than the air above, and there may be dewfall). Consequently the conductive flux, $Q_G$, is directed upwards from the soil, replenishing the deficit of $Q^*$ (Figure 7.1).

**Microclimates of snow and ice**

Snow and ice surfaces have the ability to transmit short-wave radiation, so we must consider radiative and energy exchanges of a volume. There is an exponential decay of $K\downarrow$ with depth ($z$) into the snow or ice-pack (Figure 7.2). This rate of decay is given by Beer's Law:
\[ KJ \tau = KJ_o e^{-\alpha} \]

where \( KJ_o \) is the incident short-wave radiation at the surface, \( e \) is the base of natural logarithms, and \( \alpha \) is an extinction coefficient \((m^{-1})\). While Beer's Law strictly only applies to the transmission of individual wavelengths in a homogeneous medium, it has been used with success to describe the penetration of solar radiation into snow and ice, water, and crop canopies.

![Graph](image)

**Figure 7.2** Decay of incident short-wave radiation \((KJ)\) with depth in snow on the névé of the Franz Josef Glacier at 0900, 1200, 1500 and 1800 h on 2 March 1992. (Kelliher et al. 1996)

An important characteristic of snow and ice is their high albedo, \( \alpha \) (Table 7.1). With such high surface reflectivity much of the \( KJ \) is immediately reflected, resulting in a low energy input into the snow- or ice-pack. Another consequence of this high albedo is the potential for sunburn especially on sunny days and over fresh snow. Not only are people at risk of sunburn from incident solar radiation, they are also exposed to the significant upward flux, \( K\tau \), owing to the high albedo.

Although snow and ice are almost perfect radiators of long-wave radiation (owing to their high emissivity—see Table 7.1), because of their low surface temperature the resulting \( L\tau \) is relatively small. If the surface is melting, then \( T_o \) is zero \((273.2 \text{ K})\), and if we assume \( e = 1 \), then \( L\tau \) is 316 W m\(^{-2}\). The amount of \( L\tau \) will depend largely on cloud cover, and if the cloud base is warmer than the snow, there can actually be a significant \( L\tau \) and hence a positive \( L^\ast \).
Because of high albedo, which results in high losses of short-wave radiation, the net radiation, $Q^*$, has only a small daytime surplus. As for other surfaces, there is a nocturnal deficit, which means that on a daily basis $Q^*$ is small or possibly negative. Thus there is little energy available to channel into the turbulent fluxes. In studies of the energy balance of snow and ice surfaces the usual emphasis is on the amount of energy available for melting ($\Delta Q_M$), and the energy balance is given as:

$$Q^* + Q_R = Q_H + Q_E + \Delta Q_S + \Delta Q_M$$

where $Q_R$ is the precipitation heat flux and $\Delta Q_S$ is the net heat storage in the volume.

Limited research has been carried out measuring radiation budgets and energy balances over snow and ice in New Zealand. Often the focus is on the amount of ablation, with few studies considering the full radiation budget and energy balance. A study on the lower parts of the Franz Josef Glacier, South Westland, by Ishikawa et al. (1992) gives insight into seasonal differences in both radiative and energy-balance components and the links between this surface forcing and microclimatic variables. Figure 7.3c shows the clear seasonal differences in $K_f$, with solar radiation six times

![Figure 7.3](image-url)
larger in the summer. The diurnal and seasonal trends of $K_J$ are also closely linked to air temperature, as shown in Figure 7.3b. However, the albedo was similar between seasons, with an average of 0.38 for winter and 0.31 for summer. In winter $Q^*$ was a heat sink (~13%), with 80% of the energy for melting coming from sensible heat (Figure 7.4). In summer, when the net radiation surplus is larger, this energy source contributes more to melting (21%), but $Q_H$ is still the dominant heat source (55%). Latent heat transfer contributed similar amounts in winter (20%) and summer (25%).

Synoptic conditions are a very important control on the energy balance of snow and ice as illustrated in Figure 7.5. On the Ivory Glacier (a cirque glacier) in the Southern Alps, for summer periods $Q^*$ is an important source for melting during southerly circulation patterns, while the turbulent fluxes are relatively more important during northerly circulation patterns (Hay & Fitzharris 1988b).

Figure 7.4 (Left) Mean daily energy balance fluxes over the Franz Josef Glacier in June, August, October, December, and February. Ablation stake measurements are compared with heat balance estimates. (Ishikawa et al. 1992)

Figure 7.5 Daily glacier ablation and energy sources for melt by synoptic type. Percentages give relative size of each energy source. Top part of the figure shows days of each weather type (numbers) and standard deviation of daily ablation (bars). (Hay & Fitzharris 1988b)

### Microclimates of tussock grassland and pasture

The radiation budget for plant canopies is complicated because of the transmission of short-wave radiation into the canopy and the internal absorption, reflection, and transmission. As for snow and ice surfaces, the transmission of $K_J$ into the canopy is given by a form of Beer's Law:

$$K_J(t) = K_{J_0} e^{-A_{J}(t)}$$
where $A_L(z)$ is the leaf area cumulated from the top of the canopy down to the level $z$ which allows the above equation to be applied to plant canopies with different stand architectures. Because of the stand architecture and angle of solar incidence, as well as the radiative properties of leaves, the albedo of a vegetation stand is lower than the value for its individual leaves. Therefore, although most leaves have an albedo of about 0.30, the albedo of grass ranges from 0.16 to 0.26 (Table 7.1) and is largely dependent on height of the canopy. Albedo typically follows a U-shape pattern during daytime, with higher values at sunrise and sunset, and a minimum around solar noon. This is thought to be mainly due to the dependence of albedo on solar angle. When the sun is directly overhead, penetration into the canopy increases, causing greater solar trapping in the canopy and reducing the reflection from the stand. This relation is clearly shown in Figure 7.6 for tussock grassland albedo, which varies from 0.26 in the morning and evening down to 0.12 around solar noon (Brown & Fitzharris 1993). Other possible reasons for the daily variation in albedo include different albedos of leaves at different wavelengths, increases in reflection around sunrise and sunset because of dew on the canopy, and changes in crop physiology.

Despite the considerable variation in albedo, the effect on the total radiation budget is small, since the highest albedos occur at times of least $KJ$. Net long-wave radiation, $L^*$ is usually negative, owing to the relatively warm surface, resulting in $L^*$ exceeding $L\downarrow$. Consequently net radiation, $Q^*$, is a surplus throughout the stand by daytime, with especially large values near the top of the stand where absorption is greatest. At night $Q^*$ becomes negative.

As with snow and ice environments, we must consider volume exchanges for the energy balance due to the soil–plant–air interface. For the case of a vegetated surface we must add some terms into the energy balance:

$$Q^* = Q_{\downarrow} + Q_{\uparrow} + \Delta Q_S + \Delta Q_p.$$  

As well as a storage term ($\Delta Q_S$), which accounts for physical heat storage by substances in the system (e.g. absorption or release of heat by the air, soil, and plant biomass), there is biochemical heat storage ($\Delta Q_p$), which is due to plant photosynthesis.

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**Figure 7.6** Zenith-angle dependence of tussock albedo from measurements at Flagstaff, Otago, and Cass, Canterbury. Note that at low zenith-angles the sun is directly overhead. (Brown & Fitzharris 1993)
In practice $\Delta Q_p$ is relatively small ($< 16 \text{ W m}^{-2}$) compared with the other fluxes, so this term is often neglected. If the canopy is very low and/or sparsely vegetated, the storage term can be neglected and, rather than considering a volume, the energy balance is calculated for a surface (see page 114).

Given the vast extent and agricultural significance of tussock and pasture grasslands in New Zealand, it is indeed surprising that there has not been much research on the energetics of these surfaces. Many studies have focused on evapotranspiration, but few derive the full surface energy balance with links to climate variables. As noted above, the partitioning of net radiation into evapotranspiration depends largely on the availability of soil moisture and energy, the magnitude of the surface-to-air vapour gradient, and on wind. When considering plant canopies there is a further control on evapotranspiration—that of the physiological control by plants through stomatal activity. This control is termed canopy resistance ($r_c$). If water is freely available, the atmosphere is dry and windy, and there is a low canopy resistance, then $Q_E$ will be the dominant heat sink, followed by $Q_H$ and $Q_G$. This situation results in a low Bowen ratio (i.e. $\beta < 1$). If irrigated pasture is surrounded by dry grasslands, then advection of warm dry air across the pasture can provide an additional heat source for evaporation, resulting in evaporation rates that can exceed net radiation (i.e. $Q_E > Q^*$). Such a situation is referred to as the oasis effect and, as well as having very high $Q_E$, is characterised by a negative (downward-directed) $Q_H$ and hence a negative Bowen ratio. This oasis effect can also occur under synoptic-scale advection during warm dry föhn wind events.

The surface energy balance for tussock grassland during different synoptic conditions in summer was studied by Campbell (1989) and is shown in Figure 7.7. During a typical sunny (except for cloud around midday) summer day with warm temperatures, light winds, and a strong vapour gradient, we expect a high $Q_E$. However, tussock has a reasonably high canopy resistance (about 158 s m$^{-1}$), meaning that it can regulate water loss. Consequently, sensible heat can be the dominant energy sink through most of the day, with $Q_E$ only exceeding $Q_H$ (and $Q^*$) in the evening (Figure 7.7a). In Figure 7.7b cloud increases through the day, and although $Q_H$ peaks at the same time as

![Figure 7.7](image_url)
Microclimates and Earth-surface Energy Exchanges

$Q^*$, $Q_E$ peaks later because of a stronger vapour gradient later in the day. Figure 7.7c gives the energy balance under föhn conditions, which are cloudy, windy, and warm, with a very strong vapour gradient. $Q^*$ is greatly reduced and $Q_H$ is directed towards the surface in the morning and late afternoon, thus acting as an additional energy source allowing $Q_E$ to exceed $Q^*$ (i.e. there is a considerable oasis effect). If the tussock grassland is degraded (such as in the Mackenzie Basin, inland Canterbury), then under dry summer conditions Bowen ratios may reach as high as 5–8, which are typical of arid environments.

In contrast to tussock grassland, the surface energy balance for irrigated pasture in the Manawatu (Figure 7.8) shows $Q_E$ to be the dominant energy sink, with a small oasis effect occurring in the late afternoon. At this well-watered site the Bowen ratio was consistently less than unity during daytime hours.

![Figure 7.8](image)

**Figure 7.8** Surface energy balance fluxes on a clear day over an irrigated pasture in the Manawatu during April. (Sturman & Tapper 1996)

**Microclimates of wetlands—the paradox of the peat bog**

Intuitively we expect $Q_E$ to be the dominant heat sink in wetland areas, and in many instances this is true. However, Campbell and Williamson’s (1997) research shows that plants can have such high resistance to transpiration that we can observe Bowen ratios of 3–5—comparable to those observed in arid environments. The diurnal energy balance on a sunny summer’s day for a raised peat bog in the lower Hauraki Plain is shown in Figure 7.9. The surface is covered by an extremely dense canopy of *Empodisma minus* with an average height of 0.7 m. The spiky sedge *Baumea teretifolia* contributes about 15% of the biomass and *Sphagnum cristatum* occurs beneath the canopy, particularly in more open areas. The peat soil beneath the canopy was close to saturation, but the water table was 0.03–0.2 m below the surface. The energy balance for the peat bog shows that $Q_E$ is very low and remains relatively constant throughout daytime despite changes in the vapour gradient. Thus, despite the saturated surface,
evapotranspiration is very low because of a very high canopy resistance (in the range 150–600 s m\(^{-1}\)) and an extremely dense canopy preventing diffusion of water vapour from a moist peat surface. This large canopy resistance is thought to be a response of the plants to a nutrient-poor environment that has resulted in acidic soils.

![Figure 7.9](image-url)  
**Figure 7.9** Diurnal energy balance and Bowen ratios (dots) of a peat bog in the Hauraki Plain for (a) a clear-sky day and (b) a cloudy day in summer. (Campbell & Williamson 1997)

**Microclimates of crops**

The links between the surface radiation and energy balance and microclimate for a barley field are illustrated in Figure 7.10. This day was mostly sunny except for some cloud between 1100 and 1400 h, as shown by the dip in \( K_\parallel \). During the day \( Q^* \) was a surplus and closely linked to \( K_\parallel \). The surface albedo showed the characteristic U-shape pattern for vegetated surfaces (see above), with a solar noon value of about 0.21. \( L^* \) is always negative but small, probably because of evaporative cooling moderating leaf-surface temperatures and/or a relatively high atmospheric emissivity due to high vapour content. The daytime surplus of \( Q^* \) was mostly dissipated by \( Q_E \), the high rates of which were made possible because of an abundant supply of soil moisture and a low canopy resistance (about 60–70 s m\(^{-1}\)). Note that canopy resistance is low in the early morning because of dew on the crop, while late in the afternoon, as light intensity decreases and water stress increases, \( r_c \) increases as the stomata close. Because most of \( Q^* \) is used to evaporate water, there is little energy left to heat the air, canopy, or soil.

The profiles of climate parameters in the barley field for selected times throughout the day are shown in Figure 7.10(d–f). Owing to drag exerted on the wind by the crop, wind speed is reduced within the canopy, particularly where the foliage density is greatest, but there is a small 'jet' occurring in the more open stem area. By day the maximum net radiation absorption is near the top of the canopy and hence this is the level of maximum heating. Temperatures decrease with height both above and below
Figure 7.10 (a) Radiation budget, (b) energy balance, (c) canopy resistance, and profiles of (d) wind speed, (e) temperature, and (f) vapour pressure in and above a barley field in summer at Rothamsted, England. h represents crop height. (After Oke 1990)

this level, and so $Q_H$ is carried both into the air and down into the crop. Because of evaporative cooling at the surface an inversion is produced above the crop, resulting in a downward sensible heat flux from the air and hence a small oasis effect, whereby evaporation exceeds net radiation for a period in the late afternoon. At night longwave emission from the crop resulted in a temperature minimum just below the canopy crown, with a weak inversion above the canopy (particularly during clearer skies, e.g. 0100–0200 h). During the day vapour pressure decreases with distance from the crop, since the plants and soil are moisture sources, hence the large $Q_E$. At night under clear skies dewfall occurred, giving an inverted vapour pressure profile and hence a negative or downward latent heat flux.

**Microclimates of forests (exotic and indigenous)**

While similar in some respects to crops, forest canopies have several features that must be considered when studying their energetics and microclimates. Because of their
considerable biomass, we must account for energy stored in the canopy (\(\Delta Q_S\)). The arrangement of biomass can be very variable, depending on the number and nature of strata. The structure and height of the canopy means the surface can be very rough and can lead to greater radiation trapping. Finally, the stomatal resistance may be much higher than that for crops.

The main site of radiative exchange in orchards and forests is the canopy layer, with the trunk zone of lesser importance. \(K_L\) penetrates through the tree canopy, decaying with height according to Beer's Law. The amount of solar radiation reaching the floor depends on the height, density, and species of the stand, as well as on the angle of solar incidence, and may be as little as 5% of the above-canopy flux. Because of the considerable trapping potential of the canopy (depending on the species and time of year), forest albedos are relatively low (about 0.05–0.20; see Table 7.1). Hence \(K^*\) is relatively high for these surface types. As with other surfaces, above the canopy \(L_T\) (from both the floor and the canopy) usually exceeds \(L_J\), giving a negative net long-wave radiation budget. Consequently \(Q^*\) may be high compared with other surface types.

Some surface energetics have been studied for both indigenous and exotic forests in New Zealand (Kellieher et al. 1992a; Whitehead et al. 1994). The emphasis in such studies is on moisture and carbon dioxide fluxes rather than full energy balance estimates. In a mixed red beech (Nothofagus fusca) and silver beech (N. menziesii) forest at Maruia, Nelson, during a late summer day, \(Q_H\) accounts for most of the available energy, with a mean Bowen ratio of 1.23 (Table 7.2). This relatively high \(\beta\) is due to the small size of beech leaves, resulting in a high sensible heat transfer and relatively low evapotranspiration because of high canopy resistance. In contrast, in an exotic Pinus radiata forest in the central North Island, \(Q_E\) is the dominant energy sink, despite a higher canopy resistance to evapotranspiration (Table 7.2).

### Table 7.2 Turbulent energy fluxes, Bowen ratio (\(\beta\)) and canopy resistance for indigenous (Nothofagus) and exotic (Pinus radiata) forest.

<table>
<thead>
<tr>
<th>Forest species</th>
<th>(Q_H)</th>
<th>(Q_E)</th>
<th>(\beta)</th>
<th>Maximum canopy resistance (s m(^{-1}))</th>
</tr>
</thead>
<tbody>
<tr>
<td>Nothofagus spp.</td>
<td>6.31</td>
<td>5.13</td>
<td>1.23</td>
<td>340</td>
</tr>
<tr>
<td>Pinus radiata</td>
<td>4.75</td>
<td>5.70</td>
<td>0.83</td>
<td>450</td>
</tr>
</tbody>
</table>

After Kellieher et al. 1992a and Whitehead et al., 1994

A consequence of forest-stand architecture is substantial modification of local microclimate. Figure 7.11 shows a main canopy located at about 5–9 m height, and this is the primary site of radiation absorption, evapotranspiration, and drag on airflow. On a sunny day within a forest, below the main canopy temperatures are cooler, winds are weaker, and it is more humid. These profiles are very similar to those for a crop surface (Figure 7.10) except that the above-forest gradients are much weaker because of greater mixing over a rough surface, which diffuses atmospheric variables throughout a deep layer.
Microclimates of urban areas

The impact of humans on the landscape is seldom more striking than in urban areas. A consequence of the built environment is the modification of radiative, thermal, aerodynamic, and moisture characteristics of the surface. Because of the complexity of the urban surface we must adopt a building-air volume approach to consider the radiative and energy transfers. A difficulty in estimating the impact of urbanisation on climate is defining a rural reference from which to make comparisons. If the city is surrounded by a fairly uniform landscape type, the situation is more simple than if it is surrounded by a mosaic of semi-arid and irrigated fields. Depending on where the rural measurements are taken, there may be quite a difference in the nature of the urban effect.

Urban areas are typically characterised by a polluted urban boundary layer. The pollution results in attenuation of $K\downarrow$, with the amount of reduction in incident short-wave radiation depending on the nature and type of pollutants. In a large city $K\downarrow$ may be reduced by as much as 10–20%, but even for the case of Christchurch, owing to the high air pollution levels in winter, $K\downarrow$ is reduced by up to 30% (Tapper 1976). Albedo may be decreased in urban areas because of the canyon geometry, which results in greater radiational trapping, and also because most urban surfaces have a relatively low albedo compared with rural surfaces. Thus urban areas have an average albedo of about 0.15, which is lower than most rural surfaces except for forests and areas with dark soils. Therefore, although $K\downarrow$ is reduced in urban areas, the lower albedo means that $K^\uparrow$ is reduced and hence $K^*$ is only slightly lower for the city.

The warm, polluted urban boundary layer means that $L\downarrow$ is increased in urban areas, but because built materials in urban areas are often good conductors and storers of heat, urban surfaces are typically warmer than rural surfaces and hence $L^\uparrow$ is increased. Therefore again we can see that the impacts of urbanisation offset one another, and consequently $L^*$ is similar between urban and rural surfaces. So despite urbanisation radically altering components of the radiative balance, the net impact on $Q^*$ is small.
In urban areas there is an extra source of energy, $Q_F$—due to anthropogenic processes such as combustion—so that the urban surface energy balance is given as:

$$Q^* + Q_F = Q_H + Q_E + AQ_S + AQ_A$$

where $AQ_A$ is an advection term for net energy gain or loss due to sensible and latent heat transport. As discussed above, $Q^*$ may be similar for rural and urban surfaces. The magnitude of $Q_F$ depends largely on the average energy use by individuals (which in turn depends on the climate and the nature of the economy) and the population density. In Montreal in winter $Q_F$ can be as high as 153 W m$^{-2}$ (exceeding $Q^*$ input), while in more temperate climates (such as Vancouver) $Q_F$ is about 15 W m$^{-2}$ in summer and 23 W m$^{-2}$ in winter. In Christchurch in winter $Q_F$ is estimated to be about 5 W m$^{-2}$, which is relatively low compared with most Northern Hemisphere cities. This may be due to the relatively low building and population density of the city.

Because of the warm, rough urban surface and the predominance of built materials, which effectively ‘waterproof’ the surface, the available energy ($Q^* + Q_E$) is usually channelled into $Q_H$. The amount of $Q_E$ will depend largely on the amount of greenspace and water availability (e.g. irrigation). Figure 7.12 shows the diurnal surface energy balance for a suburb in Christchurch during both summer and winter. The relatively low evaporation in summer is thought to be due to the influence of a large warehouse and paved yard adjacent to the measurement tower. In winter the turbulent fluxes are very small because of a very stable atmosphere, so that most net radiation is channelled into the storage flux (which may also include an advective component).

![Figure 7.12](image)

**Figure 7.12** Diurnal energy balance for a mixed suburban and commercial neighbourhood in Christchurch during (a) summer and (b) winter. (Spronken-Smith 1998)

The alteration of the surface energy balance in urban areas results in several impacts on climatic variables at pedestrian level. The best-known impact is that of the urban heat island, whereby the air in urban areas is usually warmer than in the surrounding countryside. Figure 7.13 shows the spatial variation in temperature over Christchurch for a winter’s day at 2230 h. On this occasion urban temperatures were
up to 6.3°C warmer than rural temperatures. Owing to differential cooling rates between urban and rural areas, the urban heat island has a diurnal peak soon after sunset. Furthermore, the phenomenon is best developed under calm and clear conditions. As wind speed increases, the urban heat island decreases because the warm near-surface air is mixed into cooler air above the city. Clear skies allow maximum long-wave radiative transfer, which rapidly cools the countryside, but in the city canyon geometry effectively traps the long-wave radiation emitted by buildings, hence slowing the rate of cooling.

![Figure 7.13](image)

During the day urban air is usually drier than rural air because of less evapotranspiration. However, at night urban air can be slightly more moist. This is probably because there is greater dewfall in the countryside (depleting the moisture content of the lower atmosphere near the ground), while in urban areas there may be weak evaporation, reduced dewfall, anthropogenic vapour, and stagnation of air in urban canyons, resulting in more humid air. Thus, as well as being heat islands, cities may also be moisture islands at night.

In addition to influencing temperature and moisture, urbanisation has impacts on wind speed. Urban areas can both increase and decrease wind speeds. For example, if wind is blowing into a canyon system, a small jet may be formed along the street. Alternatively, if the wind is blowing across the canyon system, it may skim across the roofs, with little air entering the canyon and hence decreased speeds at pedestrian level.

The microclimate of urban areas is very complex because of the mosaic of surface types, which have different radiative, thermal, moisture, and aerodynamic characteristics. This complexity makes it difficult to generalise regarding the urban microclimate,
but nevertheless we can observe recurring patterns in the radiative and energy fluxes, as well as in climate parameters.

Summary

This chapter has discussed radiation and energy balances and resulting microclimates for several different surface types. While incomplete, Table 7.3 attempts to summarise the surface energetics for a range of New Zealand environments. One must be careful when interpreting such data, owing to the different study periods and techniques used to collect them. Nevertheless, some useful generalisations can be made from Table 7.3. The energy balance data exhibit marked seasonal trends, with $Q^*$ high during summer and very low (even negative for the glacier) during winter. As well as this temporal variation there is obvious spatial variation, since similar surface types may have very different energy partitioning (e.g. compare the two glaciers in summer in Table 7.3). Over glacial surfaces the turbulent fluxes are directed mainly towards the surface and

<table>
<thead>
<tr>
<th>Environment</th>
<th>Location</th>
<th>Source</th>
<th>$Q^*$</th>
<th>$Q_E/Q^*$</th>
<th>$Q_J/Q^*$</th>
<th>$Q_G/\text{(or } \Delta Q_S\text{)}$</th>
<th>$\beta$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Glacier</td>
<td>Franz Josef, Westland, Ishikawa et al. 1992</td>
<td>-0.52</td>
<td>1.62</td>
<td>6.31</td>
<td>3.90</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Glacier</td>
<td>Ivory Glacier, Westland, Hay &amp; Fitzharris 1988</td>
<td>6.6</td>
<td>-0.30</td>
<td>-0.58</td>
<td>1.9</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Tussock grassland (moist, healthy)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>— summer</td>
<td>Franz Josef, Westland, Ishikawa et al. 1992</td>
<td>11.8</td>
<td>0.13</td>
<td>0.89</td>
<td>-0.02</td>
<td>6.8</td>
<td></td>
</tr>
<tr>
<td>Tussock grassland (dry, degraded)</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>— summer</td>
<td>Mackenzie Basin, inland Canterbury, Oliphant (pers. comm.)</td>
<td>2.67</td>
<td>1.19</td>
<td>-0.15</td>
<td>-0.04</td>
<td>-0.13</td>
<td></td>
</tr>
<tr>
<td>— winter</td>
<td>McKenzie Basin, Canterbury, Oliphant (pers. comm.)</td>
<td>(4.66)</td>
<td>(0.64)</td>
<td>(0.26)</td>
<td>(0.10)</td>
<td>(0.41)</td>
<td></td>
</tr>
<tr>
<td>Peat bog</td>
<td>Hauraki Plains, Campbell &amp; Williamson 1997</td>
<td>19.8</td>
<td>0.20</td>
<td>0.78</td>
<td>0.02</td>
<td>3.89</td>
<td></td>
</tr>
<tr>
<td>Beech forest</td>
<td>Maruia, Nelson, Keilhauer et al. 1992</td>
<td>13.3</td>
<td>0.25</td>
<td>0.50</td>
<td>0.25</td>
<td>1.99</td>
<td></td>
</tr>
<tr>
<td>Pinus forest</td>
<td>Haupapa, central North Island, Whitehead et al. 1994</td>
<td>3.94</td>
<td>0.10</td>
<td>-0.03</td>
<td>0.93</td>
<td>-0.28</td>
<td></td>
</tr>
<tr>
<td>Urban</td>
<td>Christchurch, Sprouk-Smìth 1998</td>
<td>1.23</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>— winter</td>
<td></td>
<td>0.83</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Table 7.3 Summary of daily (and daytime in brackets) surface energy balance fluxes (MJ m$^{-2}$ d$^{-1}$) and ratios for New Zealand environments.
typical Bowen ratios range from 2 to 4. For tussock grassland and pasture, energy partitioning depends greatly on the health of the vegetation as well as the moisture status of the ground. The lowest Bowen ratios occur over well-watered pasture, while very high ratios occur over the degraded and dry tussock grassland of the Mackenzie Country. Over forest surfaces the Bowen ratio tends to be much lower, as evaporation from the canopy is of more importance in the surface energy balance. Over urban areas $Q_H$ tends to dominate the energy balance, except in winter under very stable conditions, which tend to suppress the turbulent fluxes and hence $\Delta Q_S$ increases in importance as an energy sink.

It is apparent that there is much variation in energy receipt and partitioning, and hence microclimate, among these surface types. The surface properties of importance in determining this energy partitioning include albedo, moisture content, and vegetation type (particularly the canopy resistance to vapour transfer). Another major control of the radiation and energy balance is that of synoptic conditions. In New Zealand's dynamic synoptic environment, energy partitioning can vary greatly, even on a daily basis, as different weather systems move over the area.

**Further reading**
