Planetary boundary layer

Lecture 7

Chap 6.1-6.4

7 Boundary Conditions and Surface Forcings

Homework

7.3 – Mariana

7.5 – Sara

7.6 – Cátia

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From molecular processes to turbulence…

Molecular conduction of heat, molecular diffusion of tracers, and molecular viscous transfer of momentum cause transport between the surface and the lowest millimeters of air. Once in the air, turbulence takes over to transport momentum, heat and other constituents to greater depths in the atmosphere. The molecular and turbulent transport processes work together as sketched in Fig 7.1.

Effective turbulent flux includes molecular diffusion

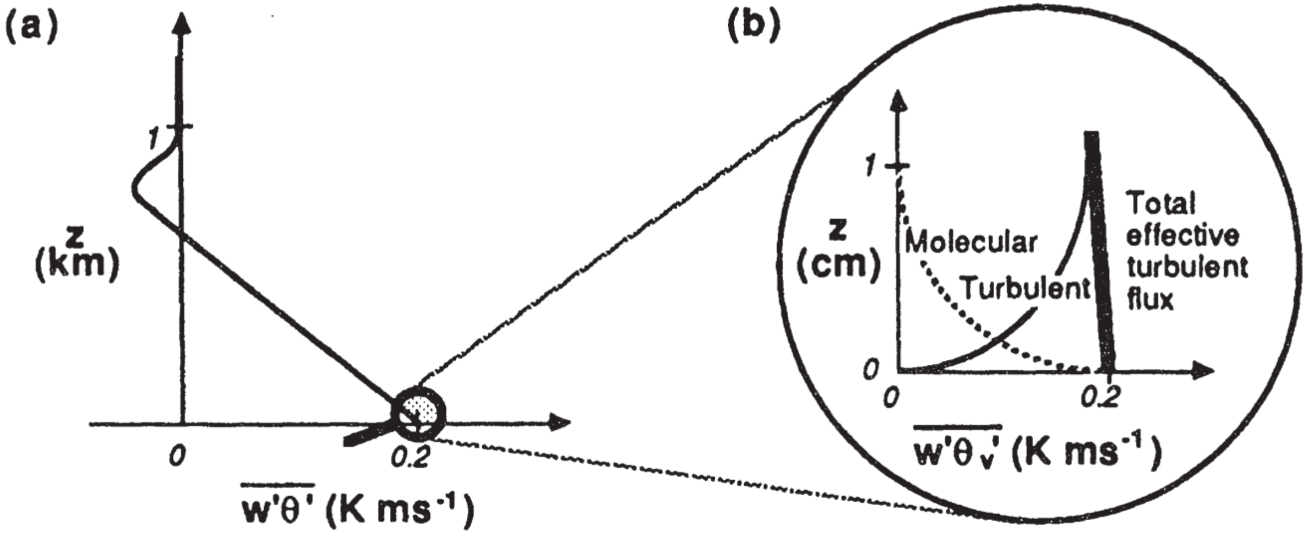


Fig. 7.1 (a) The effective turbulent heat flux using daytime convective conditions as an example, may be nonzero at the surface. (b) This effective flux, however, is the sum of the actual turbulent flux and the molecular flux.

Micro layer (1 mm) (Fourier/Fick law)

The thin layer of air in which molecular processes dominate is called the ***micro layer***. Within this layer, molecular transport, such as conduction of heat, can be described by:

$$Q_H = -\nu_\theta \frac{\partial T}{\partial z} \qquad (7.1)$$

where ν_θ is the molecular thermal diffusivity (on the order of $2\times10^{-5}\ m^2s^{-1}$ for air). For a typical kinematic heat flux of $0.2\ K{\cdot}m{\cdot}s^{-1}$, (7.1) tells us that a temperature gradient of $1\times10^4\ K{\cdot}m^{-1}$ is required. This corresponds to a 10 °C temperature difference across a micro layer 1 mm thick.

Effective flux is almost constant in the surface layer (z<50m)

From this point on, whenever we refer to a surface turbulent flux in this text, we are really implying a surface ***effective*** flux. In this way, we can ignore the molecular processes, and just use turbulence equations such as (3.5.3) with the effective flux. As shown in Chapters 1, 3 and 4, the effective flux varies by only a small portion of its magnitude in the lowest tens of meters of the BL. Thus, the turbulent flux measured at the standard "surface" instrument shelter (screen) height of 2 m provides a good approximation to the effective surface turbulent flux.

Surface energy balance

Q_H – *Sensible heat*
Q_E – *Latent heat*
Q_S^* – *Net radiation*
Q_G – *Soil heat flux*

Upward >0

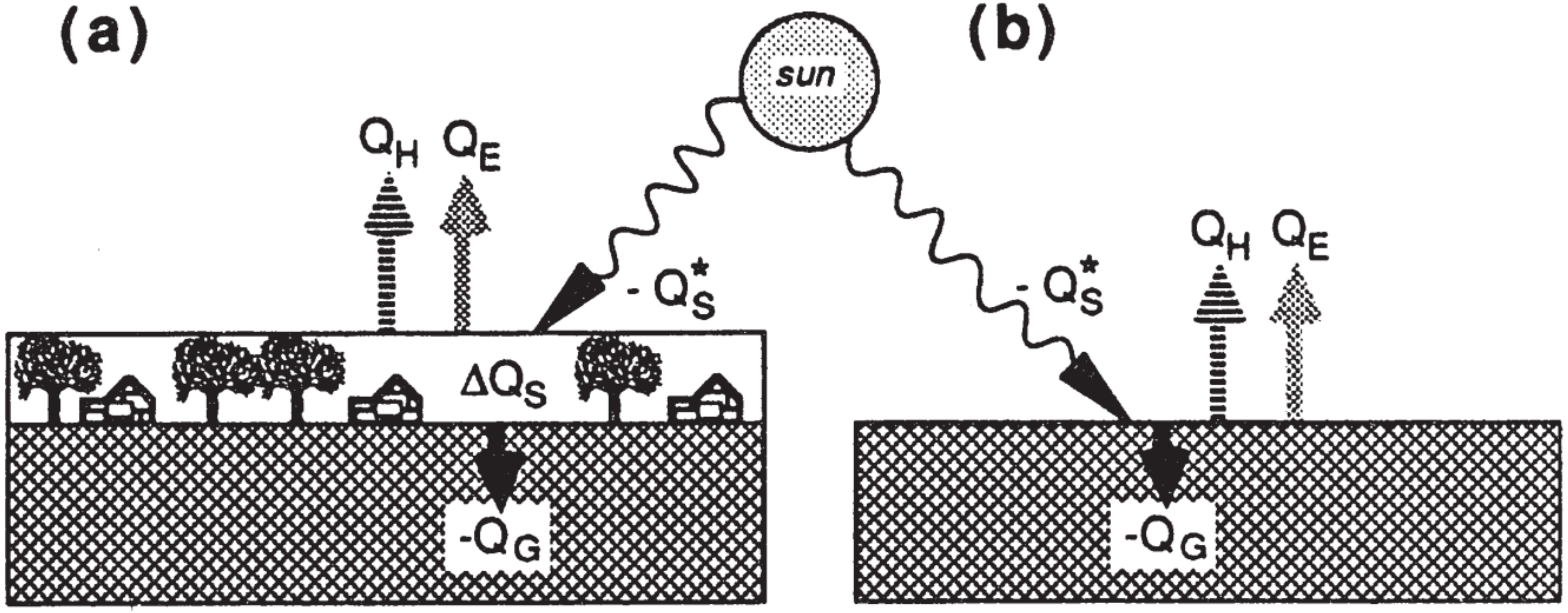


Fig. 7.2 Contributions to the surface energy balance (a) for a finite thickness box and (b) for an infinitesimally thin layer. $-Q_S^*$ is the net radiative contribution, Q_H is turbulent sensible heat flux, Q_E is turbulent latent heat flux, $-Q_G$ is molecular flux into the ground, and ΔQ_S is storage.

The external forcing is Q_S^* and all the other terms are response terms.

Thin surface layer

Very complex processes can occur within our imaginary layer (Geiger, 1965; Oke, 1978; Brutsaert, 1985), such as: radiation between leaves, plants, buildings, and animals; turbulent circulations different from those higher in the boundary layer; vertical variations of the sensible and latent heat flux associated with evaporation and condensation; and transpiration. Because of this complexity, we have employed the simplification of a layer into which the net effect of all of these processes can be lumped together as ΔQ_S.

Sometimes, we prefer to conceptually employ an infinitesimally thin layer, as sketched in Fig 7.2b. This is not a really layer, but a plane. The resulting ***surface heat budget*** is

$$-Q_s^* = Q_H + Q_E - Q_G \tag{7.2c}$$

Typical fluxes

Daytime land

Nightime land

Oasis: warm air over wet surface

Daytime sea

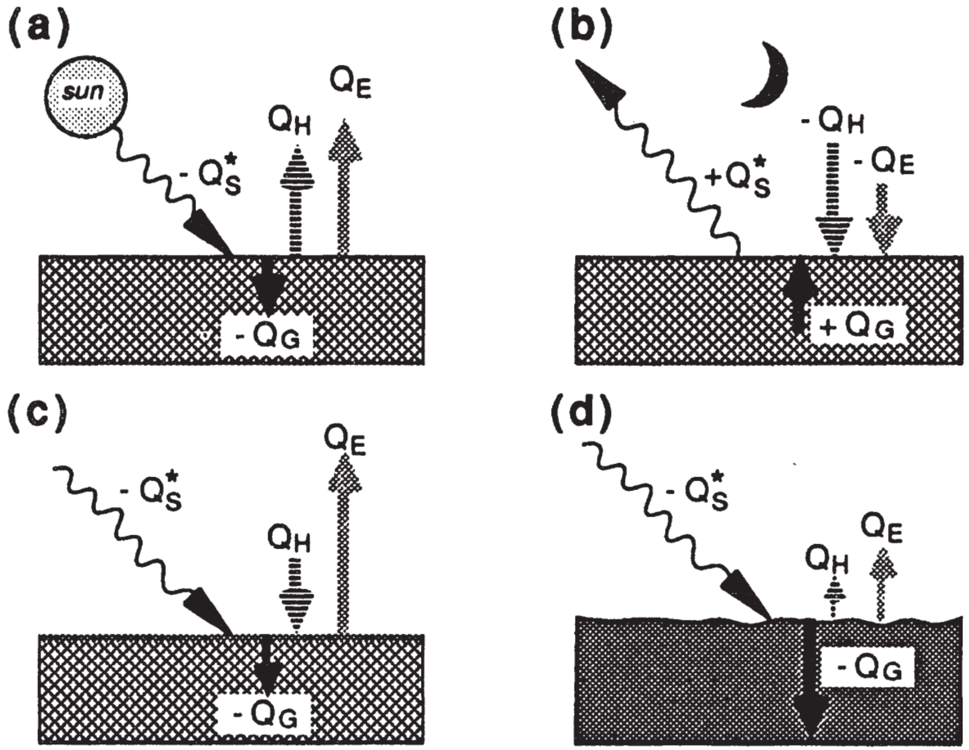


Fig. 7.3 Typical variation of terms of the surface energy balance for (a) daytime over land; (b) nighttime over land; (c) oasis effect of warm dry air advection over a moist surface; and (d) daytime over the sea with no advection. Arrow size indicates relative magnitude.

Diurnal cycle over irrigated land (clear sky)

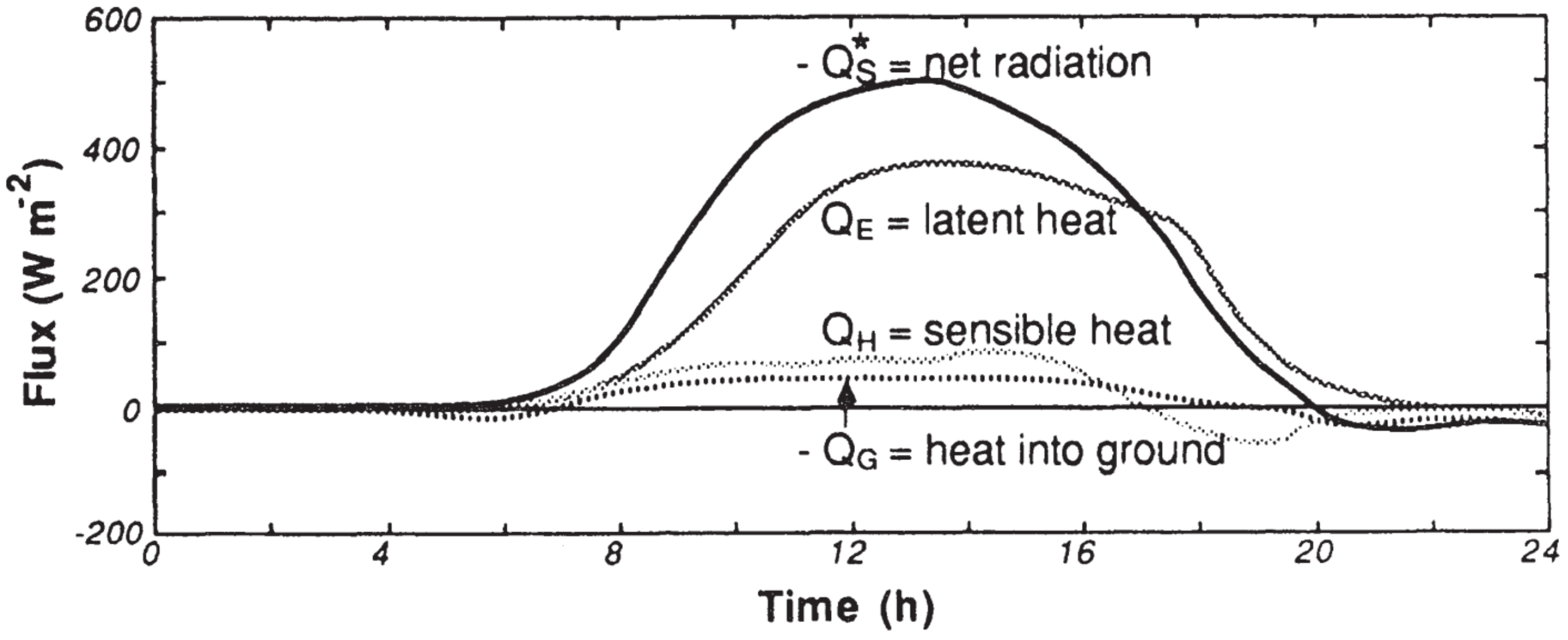


Fig. 7.4 Energy balance components for 25 July 1976, with cloudless skies at Pitt Meadows, Canada (49°N) over a 0.25 m tall stand of irrigated mixed orchard and rye grass (after Oke, 1978).

Q^*- net radiation

7.3 Radiation Budget

It is often convenient to split the net radiation term into four components:

$$Q^* = K\uparrow + K\downarrow + I\uparrow + I\downarrow \qquad (7.3)$$

where $K\uparrow$ = upwelling ***reflected*** short wave (solar) radiation

$K\downarrow$ = downwelling shortwave radiation ***transmitted*** through the air

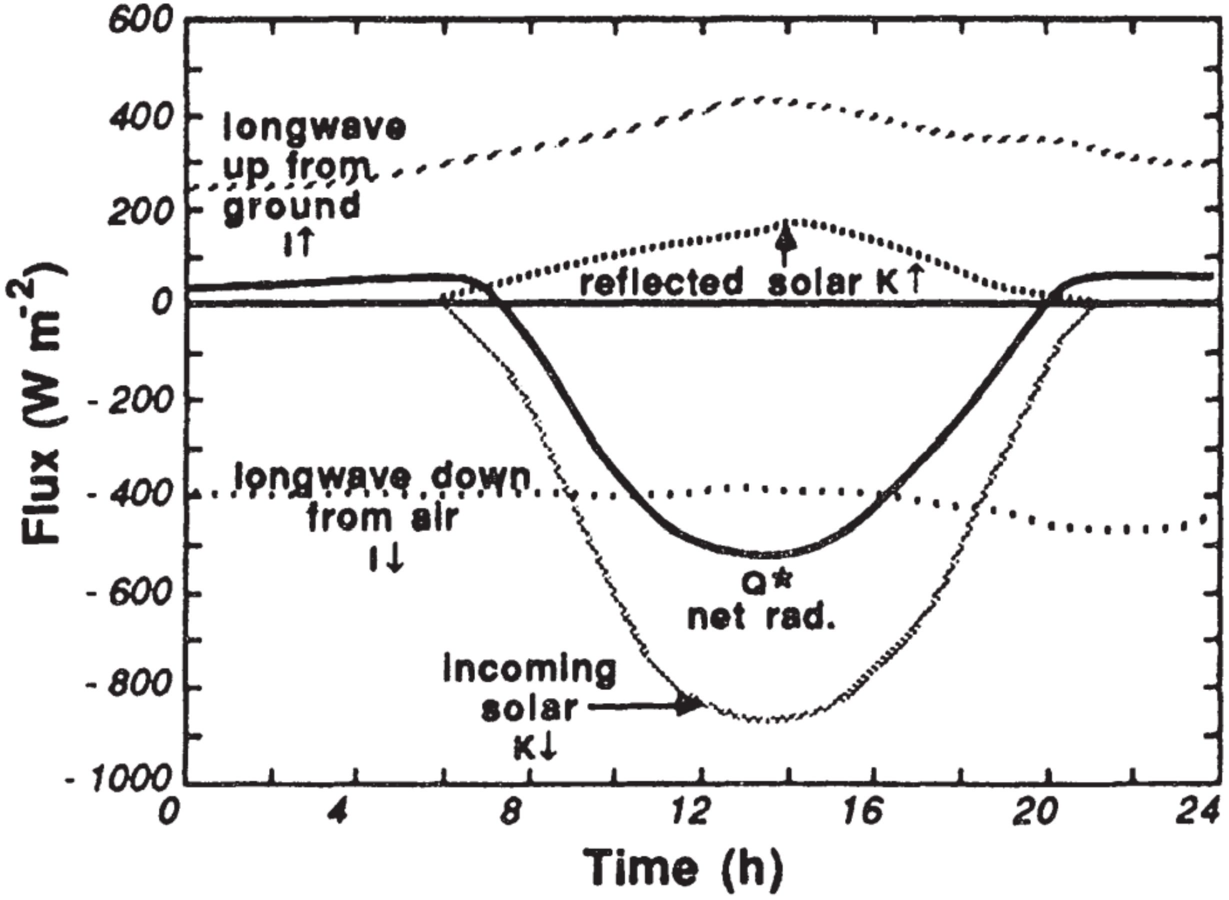
$I\uparrow$ = longwave (infrared, IR) radiation ***emitted*** up

$I\downarrow$ = longwave ***diffusive*** IR radiation down

Radiation diurnal cycle (clear sky)

Fig. 7.5

Surface radiation budget components for 30 July 1971, at Matador, Saskatchewan (50°N) over a 0.2 m stand of native grass. Cloudless skies in the morning, increasing clouds in the late afternoon and evening (after Ripley and Redmann, 1976).



Shortwave Radiation

The intensity of incoming solar radiation at the top of the atmosphere is called the ***solar irradiance***, S. Although it was formerly known as the ***solar constant***, this term is being used less frequently because of the realization that the solar irradiance is not constant — ranging from about -1360 to -1380 $W \cdot m^{-2}$. We will use a value of $S = -1370$ $W \cdot m^{-2}$ (Kyle, et al., 1985), or $S = -1.127$ $K \cdot m \cdot s^{-1}$ in kinematic units, where the density and specific heat of air in the boundary layer is used for the conversion to kinematic units.

Some of this radiation is attenuated by scattering, absorption, and reflection from clouds on the way down to the surface. When the sun is lower in the sky, the radiation will also be attenuated by its longer path through the atmosphere en route to the surface. Define T_K as the net sky ***transmissivity***, or the fraction of solar radiation that makes it to the surface. Define Ψ as the solar elevation angle; namely, the angle of the sun above the local horizon. One simple parameterization (Burridge and Gadd, 1974) for the transmissivity is:

$$T_K = (0.6 + 0.2 \sin \Psi) \cdot (1 - 0.4\, \sigma_{C_H}) \cdot (1 - 0.7\, \sigma_{C_M}) \cdot (1 - 0.4\, \sigma_{C_L}) \qquad (7.3.1a)$$

$\Psi(t) = solar\ elevation;$ σ_X represents the cloud-cover fraction

$$\Psi(t) = solar\ elevation = 90 - \nu$$

The solar elevation angle is also important because when it is less than 90°, the radiation that does reach the surface is spread out over a larger area, reducing the radiation per unit surface area by a factor of sin Ψ. The expression for downwelling radiation at the surface is approximately

$$K{\downarrow}_s = S \cdot T_K \cdot \sin \Psi \quad \text{for daytime (i.e., sin } \Psi \text{ positive)} \tag{7.3.1b}$$

$$= 0 \quad \text{for nighttime (i.e., sin } \Psi \text{ negative)}$$

Determination of the local elevation angle is a straightforward exercise in geometry (Zhang and Anthes, 1982), resulting in:

$$\sin \Psi = \sin \phi \sin \delta_s - \cos \phi \cos \delta_s \cos\left[\left(\frac{\pi\, t_{UTC}}{12}\right) - \lambda_e\right] \tag{7.3.1c}$$

where ϕ and λ_e are the latitude (positive north) and longitude (positive west) in radians,

albedo

δ_s is the ***solar declination angle*** (angle of the sun above the equator, in radians), and t_{UTC} is Coordinated Universal Time in hours. The solar declination angle is

$$\delta_s = \phi_r \cos\left[\frac{2\pi\,(d - d_r)}{d_y}\right] \qquad (7.3.1d)$$

where ϕ_r is the latitude of the Tropic of Cancer (23.45° = 0.409 radians), d is the number of the day of the year (e.g., October 27 = day 300), d_r is the day of the summer solstice (173), and d_y is the average number of days per year (365.25).

Define the ***albedo***, a, as the fraction of downwelling radiation at the surface that is reflected. The albedo varies from about 0.95 over fresh snow, 0.4 over light-colored dry soils, 0.2 over grass and many agriculture crops, 0.1 over coniferous forests, to 0.05 over dark wet soils. The upwelling (reflected) radiation is thus

$$K\uparrow_s = -\,a\,K\downarrow_s \qquad (7.3.1e)$$

The albedo of water not only varies with wave state, but is a strong function of sun angle (Krauss, 1972). When the sun is directly overhead over a smooth water surface, the albedo is about 0.05, while it increases to nearly 1.0 at low elevation angles.

Longwave

When clouds are present, much of the outgoing radiation can be balanced by downward radiation from the clouds. Low clouds are more effective at this than high clouds. For overcast clouds at all three levels, we might expect the net radiation to be approximately zero.

The net upward longwave radiation at the surface is sometimes approximated (Burridge and Gadd, 1974) by :

$$I^* = (0.08\ \mathrm{Kms^{-1}})\ (1 - 0.1\sigma_{C_H} - 0.3\,\sigma_{C_M} - 0.6\,\sigma_{C_L}) \qquad (7.3.2a)$$

This type of parameterization is known to be an oversimplification of the actual physics. Nevertheless, it is useful when detailed radiation parameterizations are not appropriate.

Fluxes at Interfaces (transport velocity)

The vertical flux, F_ξ, of any variable ξ is assumed to be driven by the difference in $\overline{\xi}$ across the interface

$$F_\xi = -U_T\,(\overline{\xi}_{top} - \overline{\xi}_{bottom}) \qquad (7.4a)$$

where U_T represents a ***transport velocity*** across that interface, and the $\overline{\xi}_{top}$ and $\overline{\xi}_{bottom}$ are the values just above and below the boundary. It can also be thought of as a ***conductivity***, because a given $\overline{\xi}$-difference (voltage potential) yields a greater flux (current) if the conductivity is greater.

The transport velocity is usually parameterized as a function of some measure of turbulence appropriate to the type of interface:

$$U_T = C_D \cdot \overline{M} \qquad \text{at } z = 0 \qquad (7.4b)$$

and

$$U_T = w_e \qquad \text{at } z = z_i \qquad (7.4c)$$

Surface Fluxes - Drag and Bulk Transfer Methods

Definitions. In 1916, G.I. Taylor suggested that a velocity squared law might be used to describe the drag of the atmosphere against the earth's surface. Using u_*^2 as a measure of surface stress associated with drag, we find that

$$u_*^2 = C_D \overline{M}^2 \qquad (7.4.1a)$$

For momentum transfer, C_D is called the ***drag coefficient***. Generically it is still a bulk transfer coefficient, and sometimes is written as C_M in the literature. The individual components of surface stress are correspondingly given by the ***drag laws***:

$$(\overline{u'w'})_s = - C_D \overline{M}\, \overline{U} \qquad (7.4.1b)$$

$$(\overline{v'w'})_s = - C_D \overline{M}\, \overline{V} \qquad (7.4.1c)$$

Heat and moisture

At first glance (7.4.1b) does not appear to follow the form of (7.4a), but it turns out that the proper form is followed because the wind speed below the surface is zero. Thus, $\overline{U} = \overline{U} - 0 = \overline{U}_{air} - \overline{U}_{ground} = \overline{U}_{top} - \overline{U}_{bottom}$. The three factors C_D, $\overline{M}$ and $\overline{U}$ should all correspond to the same height above the surface. Often 10 m is assumed as the standard height, if not otherwise specified.

Similar expressions can be used to parameterize surface heat and moisture fluxes:

$$(\overline{w'\theta'})_s = - C_H \overline{M} (\overline{\theta} - \theta_G) \tag{7.4.1d}$$

$$(\overline{w'q'})_s = - C_E \overline{M} (\overline{q} - q_G) \tag{7.4.1e}$$

Surface roughness is relevant

Table 7-2. Average values of drag coefficients (C_{DN}, for 10 m winds) over continents for neutral stability. Geostrophic drag coefficients (C_{GN}) for neutral stability over continents. After Garratt (1977).

Continent	C_{DN}	C_{GN}
North America	10.1×10^{-3}	1.89×10^{-3}
South America	26.6×10^{-3}	2.16×10^{-3}
Northern Africa	2.7×10^{-3}	1.03×10^{-3}
Southern Africa	12.9×10^{-3}	1.98×10^{-3}
Europe	6.8×10^{-3}	1.73×10^{-3}
U.S.S.R.	7.9×10^{-3}	1.83×10^{-3}
Asia (north of 20°N)	3.9×10^{-3}	1.31×10^{-3}
Asia (south of 20°N)	27.7×10^{-3}	2.18×10^{-3}
Australia	6.0×10^{-3}	1.50×10^{-3}

Small scale

On a smaller scale, one measure of roughness is the spacing density of individual obstacles or ***roughness elements***. For example, the leaves of many trees, plants, and crops can form a ***canopy*** elevated above the ground surface. If we imagine that a large box could be placed over one whole plant or tree that would just touch the top and sides of the plant, then the volume of this box represents the ***space*** taken by the plant. Of course, most of this space is filled by air between the leaves and branches. The total surface area of the plant, including the area of both sides of each leaf can theoretically be measured or estimated. The ***area density of roughness elements***, S_r, is defined as the plant surface-area divided by the space volume. A ***dimensionless canopy density***, C_*, can be defined by:

$$C_* = c_m S_r h^* \qquad (7.4.1g)$$

where c_m is the drag coefficient associated with an individual roughness element (c_m = 0.05 to 0.5 for typical plants and crops), and h^* is the average height of the canopy (Kondo and Kawanaka, 1986).

Atmospheric PBL over oceans

Over oceans, the drag laws are a bit easier to parameterize, because the roughness length associated with ocean wave height is a known function of surface stress or wind speed:

$$z_o = 0.015 \frac{u_*^2}{g} \qquad (7.4.1h)$$

which is known as ***Charnock's relation*** (1955). Stronger wind stress make higher waves, which results in a greater roughness length. The application of roughness length to bulk transfer is tied to the topic of measurement heights, which is discussed next.

Stability

Dependence on Stability. Statically unstable flows generally cause a greater transport rate across an interface than statically neutral flows, which in turn transport more than stable flows. Sometimes Richardson numbers are used as a measure of stability, while at other times $\zeta = z/L$ is used. The dimensionless wind shear, $\phi_M(\zeta)$, and lapse rate, $\phi_H(\zeta)$, can be used with surface-layer similarity theory (to be described in Chapter 9) to give stability correction terms [$\psi_M(\zeta)$ and $\psi_H(\zeta)$]. This yields:

$$C_D = k^2 \left[\ln\left(\frac{z}{z_o}\right) - \psi_M(\zeta) \right]^{-2} \qquad (7.4.11)$$

$$C_H = k^2 \left[\ln\left(\frac{z}{z_o}\right) - \psi_M(\zeta) \right]^{-1} \left[\ln\left(\frac{z}{z_o}\right) - \psi_H(\zeta) \right]^{-1} \qquad (7.4.1m)$$

It is usually assumed that $C_E = C_H$. Figs 7.9 and 7.10, based on the work of Louis (1979), Garratt (1977), Joffre (1982) and Greenhut (1982), show the variation of bulk transfer coefficients with stability, sensor height, and roughness.

Partitioning of Flux into Sensible and Latent Portions

$$(-Q^*_s + Q_G) = Q_H + Q_E$$

Bowen

$$\beta = \frac{Q_H}{Q_E} = \frac{\left(c_p\, \overline{w'\theta'_s}\right)}{\left(L_v\, \overline{w'q'_s}\right)} = \frac{\gamma\, \overline{w'\theta'_s}}{\overline{w'q'_s}}$$

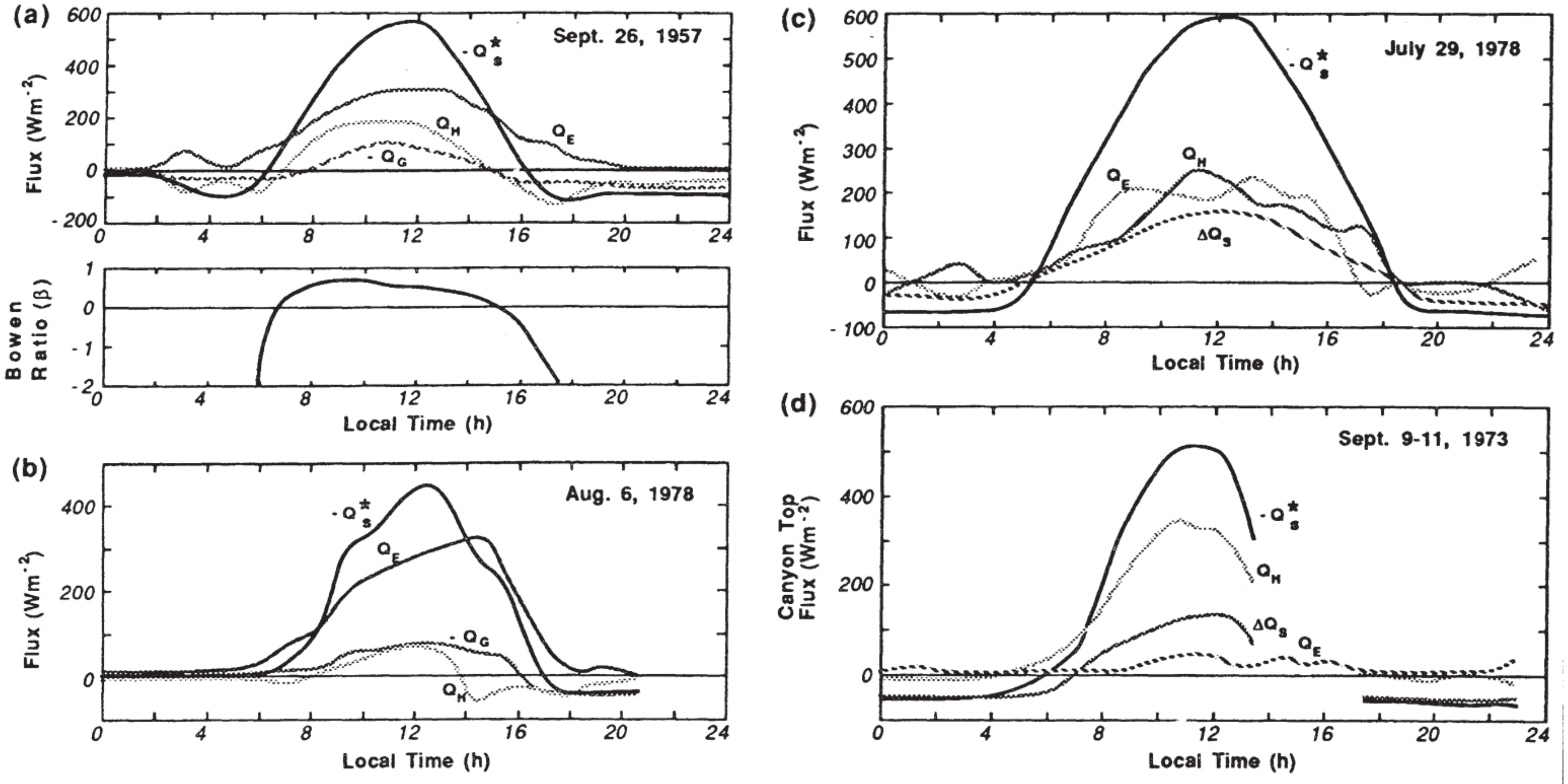


Fig. 7.12 (a) The diurnal variation of the energy balance and the Bowen ratio (β) of an irrigated field of grass at Hancock, Wisconsin (data from Tanner and Pelton 1960). (b) Energy balance of an irrigated suburban lawn (160 m²) in Vancouver, B.C. (after Oke, 1978). (c) Energy balance of a suburban area in Vancouver, B.C. (d) Energy balance of a complete urban canyon system. Exchanges are expressed as equivalent flux densities passing through the canyon top using mean hourly data for a 3-day period in September 1977 (after Oke 1978).

Bowen ratio

It was once suggested that if the Bowen ratio for a surface were known, then (7.5.1a) could be coupled with (7.5) to give:

$$Q_H = \frac{\beta(-Q^*_s + Q_G)}{(1+\beta)} \qquad (7.5.1b)$$

$$Q_E = \frac{(-Q^*_s + Q_G)}{(1+\beta)} \qquad (7.5.1c)$$

Attempts to use this approach have mostly failed, because the Bowen ratio usually varies with time and weather over each site. Furthermore, the ***evapotranspiration*** component of latent heat flux from plants is a complex function of the age, health, temperature and water stress of the plant. The pores, or stomates, of the plant open and close to regulate the life processes of the plant. Thus, the ***stomatel resistance*** to water flux, or transpiration, also varies.

Penman-Monteith

One way to include the evaporative cooling effects such as occurs during advection is via a correction term, F_w, added to the Priestly-Taylor parameterization for Q_E and subtracted from that for Q_H, to yield the ***Penman-Monteith*** (Penman, 1948; Monteith, 1965; deBruin and Holtslag, 1982) form:

$$Q_H = \frac{\left[\gamma(-Q_s^* + Q_G) - F_W\right]}{(X_G\, s_{cc} + \gamma)} \tag{7.5.3a}$$

$$Q_E = \frac{\left[X_G\, s_{cc}(-Q_s^* + Q_G) + F_W\right]}{(X_G\, s_{cc} + \gamma)} \tag{7.5.3b}$$

where X_G is like a relative humidity of the earth or plant surface. F_w is like a specific humidity flux, and is approximated by a bulk transfer law of the form: $F_w = C_E\, \overline{M}\, (X_G - X_s) \cdot \overline{q}_{sat}$, where X_s is the relative humidity of the air near the surface.

Ground heat flux

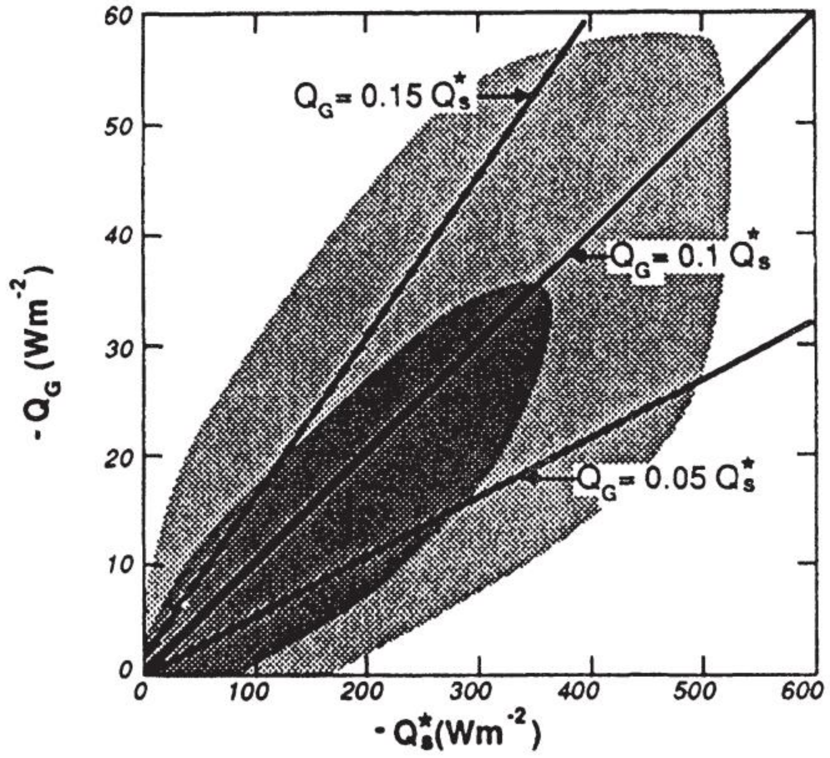


Fig. 7.16 Range of soil heat flux density Q_G plotted against net radiation for 1040 hourly values during daytime. Darker shadings indicate a higher density of data points. Lines indicate various approximations. (After DeBruin, personal communication).

Previous Homework

6.5 – Maria

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6.17 – Florian

6.27 – Jason