Chapter 2. Radiation, energy, and climate

2.1 Heat Budget of the atmosphere

Almost all energy affecting the earth is derived from solar radiation, which is of short wavelength (<4μm) due to the high temperature of the sun (6000K) (i.e., Wien's Law). The solar constant has a value of approximately 1366W m–2. The sun and the earth radiate almost as black bodies (Stefan's Law, $F = \sigma T^4$), whereas the atmospheric gases do not. Terrestrial radiation, from an equivalent black body, amounts to only about 270W m^{-2} due to its low radiating temperature (263K); this is infrared (longwave) radiation between 4 and 100μm. Water vapor and carbon dioxide are the major absorbing gases for infrared radiation, whereas the atmosphere is largely transparent to solar radiation (the greenhouse effect). Trace-gas increases are now augmenting the 'natural' greenhouse effect (33K). Solar radiation is lost by reflection, mainly from clouds, and by absorption (largely by water vapor). The planetary albedo is 31 percent; 49 percent of the extraterrestrial radiation reaches the surface. The atmosphere is heated primarily from the surface by the absorption of terrestrial infrared radiation and by turbulent heat transfer. Temperature usually decreases with height at an average rate of about 6.5°C/km in the troposphere. In the stratosphere and thermosphere, it increases with height due to the presence of radiation absorbing gases. The excess of net radiation in lower latitudes leads to a poleward energy transport from tropical latitudes by ocean currents and by the atmosphere. This is in the form of sensible heat (warm air masses/ocean water) and latent heat (atmospheric water vapor). Air temperature at any point is affected by the incoming solar radiation and other vertical energy exchanges, surface properties (slope, albedo, heat capacity), land and sea distribution and elevation, and also by horizontal advection due to air mass movements and ocean currents.

We can summarize the net effect of the transfers of energy in the earth– atmosphere system averaged over the globe and over an annual period: The incident solar radiation averaged over the globe is:

$$\text{Solar constant} \times \pi r^2 / 4\pi r^2$$

where r = radius of the earth and $4\pi r^2$ is the surface area of a sphere. This figure is approximately $342\ \mathrm{Wm}^{-2}$, or $11.109\ \mathrm{Jm}^{-2}\mathrm{yr}^{-1}$($10^9$J = 1GJ); for convenience, we will regard it as 100 units. Referring to Figure 2.1, incoming radiation is absorbed in the stratosphere (3 units), by ozone mainly, and 20 units are absorbed in the troposphere by carbon dioxide (1), water vapour (13), dust (3) and water droplets in clouds (3). Twenty units are reflected back to space from clouds, which cover about 62 percent of the earth's surface, on average. A further 9 units are similarly reflected from the surface and 3 units are returned by atmospheric scattering. The total reflected radiation is the ***planetary albedo*** (31 percent or 0.31). The remaining 49 units reach the earth either directly (Q = 28) or as diffuse radiation (q =21) transmitted via clouds or by downward scattering. The pattern of outgoing terrestrial radiation is quite different (see Figure 2.2). The black-body radiation, assuming a mean surface temperature of 288K, is equivalent to 114 units of infrared (longwave) radiation. This is possible because most of the outgoing radiation is reabsorbed by the atmosphere; the ***net*** loss of infrared radiation at the surface is only 19 units. These exchanges represent a time-averaged state for the whole globe. Recall that solar radiation affects only the sunlit hemisphere, where the incoming radiation exceeds $342\ \mathrm{Wm}^{-2}$. Conversely, no solar radiation is received by the night-time hemisphere. Infrared exchanges continue, however, due to the accumulated heat in the ground. Only about 12 units escape through the atmospheric window directly from the surface. The atmosphere itself radiates 57 units to space (48 from the emission by atmospheric water vapor and CO2 and 9 from cloud emission), giving a total of 69 units (L_u); the atmosphere in turn radiates 95 units back to the surface (L_d); thus,

$L_u + L_d = L_n$ is negative.

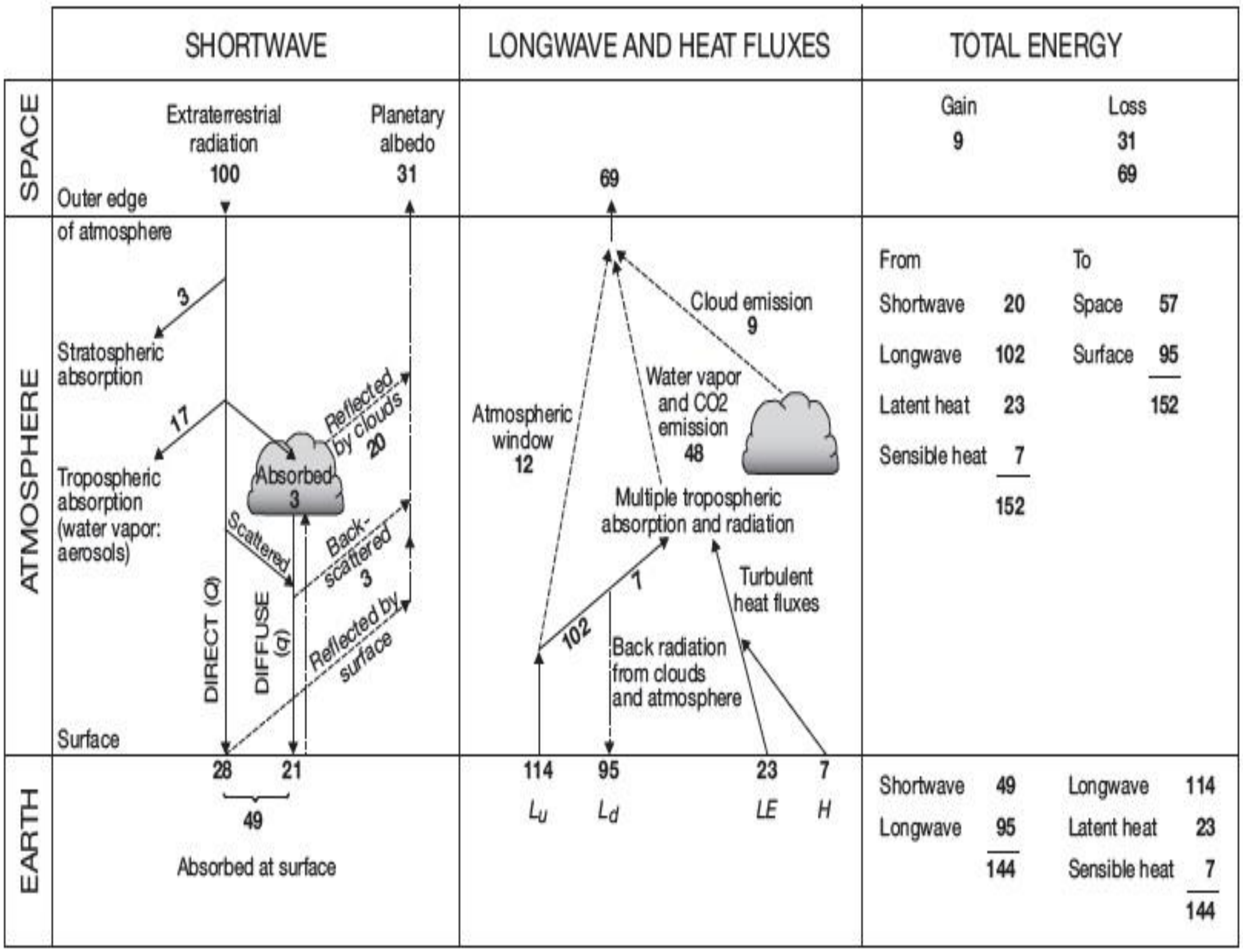

Figure 2.1 The balance of the atmospheric energy budget. The transfers are explained in the text. Solid lines indicate energy gains by the atmosphere and surface in the left-hand diagram and the troposphere in the right-hand diagram. The exchanges are referred to 100 units of incoming solar radiation at the top of the atmosphere (equal to 342 W m^{-2}).

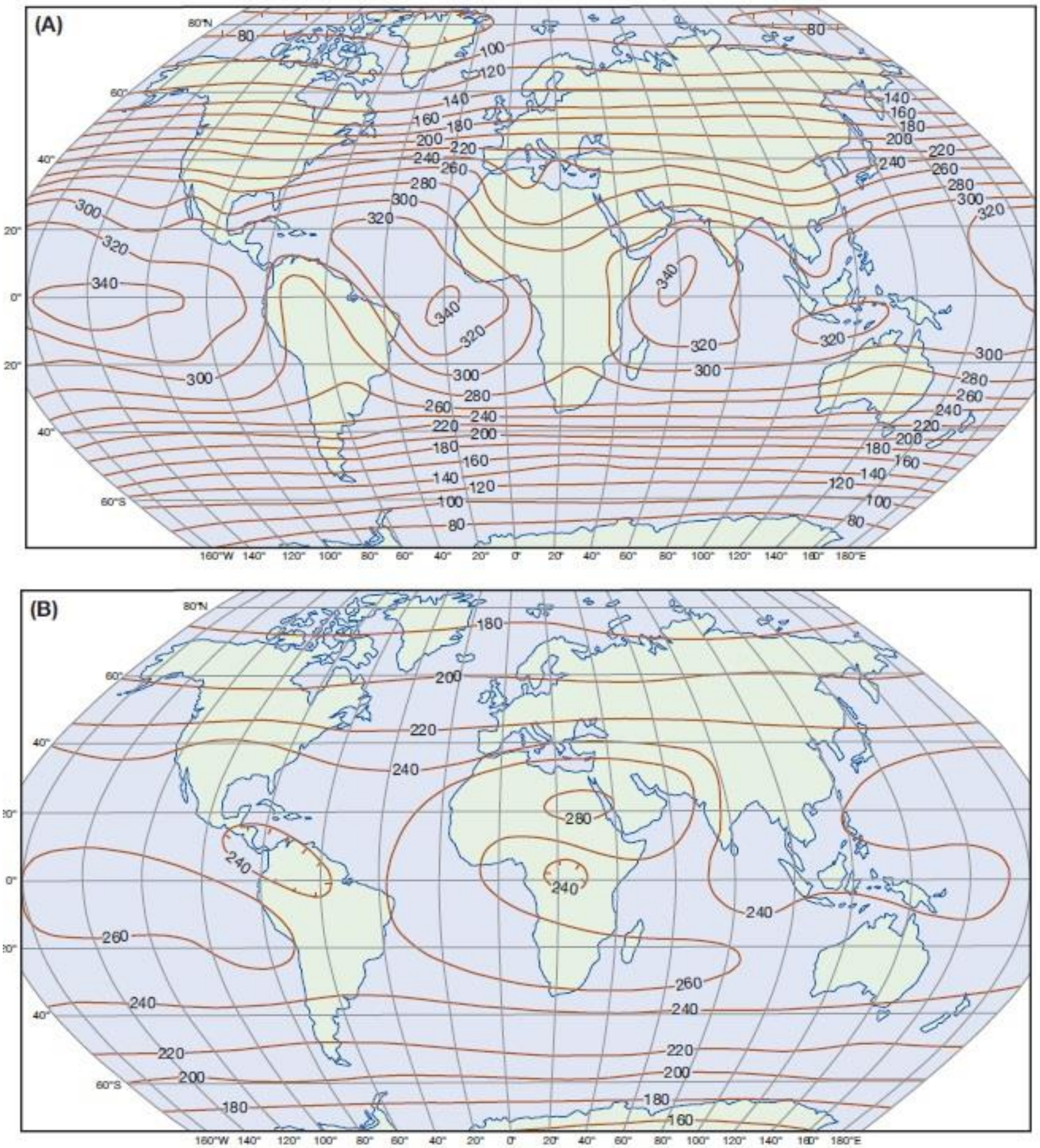

Figure 2.2 Planetary short and longwave radiation (W m-2): (A) Mean annual absorbed shortwave radiation for the period April 1979 to March 1987; B Mean annual net planetary longwave radiation (Ln) on a horizontal surface at the top of the atmosphere.

2.2 Energy flow representation

The exchanges and flows associated with energy inputs into the Earth- atmosphere system is represented by a series of symbolic equations. Use of the equations permits easy calculation once values are input. Shortwave solar radiation (K↓) reaching the surface is made up of the vertical radiation (S) and diffuse radiation (D):

$$K\downarrow = S + D$$

Some of the energy is reflected back to space (K↑) so that net shortwave radiation (K*) is the difference between the two:

$$K^* = K\downarrow - K\uparrow$$

Net longwave, terrestrial radiation (L*) comprises downward atmospheric radiation (L↓) less upward terrestrial radiation (L↑):

$$L^* = L\downarrow - L\uparrow$$

The amount of energy available at any surface is thus the sum of K* and L*. This is net all-wave radiation (Q*):

$$Q^* = K^* + L^*$$

which may also be given as:

$$Q^* = (K\downarrow - K\uparrow) + (L\downarrow - L\uparrow)$$

Q* may be positive or negative.
High positive values will occur during high sun periods when K↓ is at its maximum and atmospheric radiation, L↓, exceeds outgoing radiation, L↑. Negative values require outgoing values to be greater than incoming. This happens, for example, on clear nights when L↑ is larger than other values. On a long-term basis, Q* will vary with latitude and surface type.

2.2.1 The heat budget

Consider a column of the Earth's surface extending down to where vertical heat exchange no longer occurs (Figure 2.3). The net rate (G) at which heat in this column changes depends upon the following:

Net radiation $(K\downarrow - K\uparrow) + (L\downarrow - L\uparrow)$

Latent heat transfer (LE)
Sensible heat transfer (H)
Horizontal heat transfer (S)

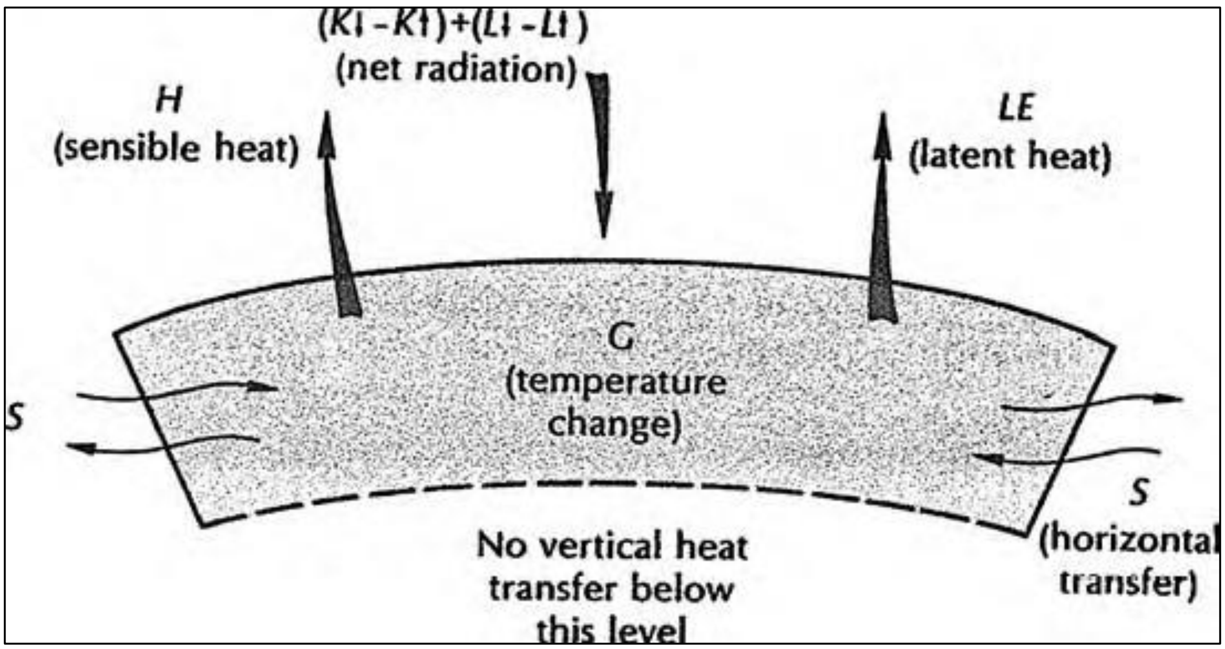

Figure (2.3): Model of energy transfer in the atmospheric system.

In symbolic from:

$$G = (K\downarrow - K\uparrow) + (L\downarrow - L\uparrow) - LE - H \pm S$$

Since

$$(K\downarrow - K\uparrow) + (L\downarrow - L\uparrow) = Q^*$$

then

$$G = Q^* - \mathrm{LE} - \mathrm{H} \pm \mathrm{S}$$

In terms of Q^*

$$Q^* = \mathrm{G} + \mathrm{LE} + \mathrm{H} \pm \mathrm{S}$$

The column will not experience a net change in temperature over an annual period; that is, it is neither gaining nor losing heat over that time, so G = 0 and can be dropped from the equation.

$$Q^* = \mathrm{LE} + \mathrm{H} \pm \mathrm{S}$$

This equation will apply to a mobile column, such as the oceans. On land, where subsurface flow of heat is negligible, S will be unimportant. The land heat budget becomes

$$Q^* = \mathrm{LE} + \mathrm{H}$$

The ratio between LE and H is given as the Bowen Ratio.