1. Introduction

1.1 Orientation:

Symbol and units

1.2 Atmospheric scales:

Atmosphere is characterized by phenomena whose space and time scales cover a very wide region:

(1) Space scales: Determined by typical size or wavelength

(2) Time scales: Determined by typical lifetime or period

Most atmospheric phenomena are associated with motions and range from small-scale turbulence up to jet streams and planetary waves [Handout: Fig. 1.1 after Smagorinsky, 1974].

In reality continuum of scales. Following classification is possible:

(1) Micro-scale: 10^{-2} to 10^{3} m

(2) Local scale: 10^{2} to 5×10^{4} m

(3) Meso-scale: 10^{4} to 2×10^{5} m

(4) Macro-scale: 10^{5} to 10^{8} m

Interaction between atmosphere and the earth's surface at the micro- and local scales are focus of this course.

Influence of surface is limited to the lowest 10 km (troposhere). Over time periods of about one day this influence is restricted to a much shallower zone known as the planetary boundary layer (PBL). This layer is characterized by well developed mixing generated by frictional drag and convective heating at the surface. Its height is not constant with time but a function of time of the day (large, about 1 to 2 km, by day and small at night).

Other atmospheric layers include [Handout: Fig. 1.2 from Oke, 1987]:

(1) Turbulent surface layer (characterized by intense small-scale turbulence)

(2) Roughness sub-layer (height related to size of surface roughness)

(3) Laminar boundary layer (in direct contact with the surface)

1.3 Concepts:

Traditionally climatology has been descriptive, *i.e.* concerned with description of distribution of parameters such as temperature, humidity, etc.

Descriptive parameter: **Temperature**, **humidity**

→ Measures of fundamental quantity: Gauge **thermal energy** and **water status**

→ Tied to: **Energy** and **water** **cycle** of earth-atmosphere system

Relationship between energy flow and climate can be illustrated using the First Law of Thermodynamics:

Energy can be neither created nor destroyed,
only converted from one form to another

Two possibilities exist:

(1) Energy Input = Energy Output

(2) Energy Input = Energy Output + Energy Storage Change

In (1) there is no change in the net energy status of the system through which the energy has passed. The system may have energy or may have changed the type of the output energy. (1) is only valid if values are integrated over long time period, say 1 year (*e.g.* mean Earth surface temperature remains constant). On shorter time scales (2) applies where energy is accumulated in or depleted from system storage (*e.g.* daytime/nighttime storage/release of heat energy in urban fabric). The system can have a wide range of scales, be simple or very complex.

Energy in Earth-atmosphere system:

Energy of importance to climatology exists as:

Radiant, thermal, kinetic or **potential energy**

The exchange of energy within the earth-atmosphere system is possible in 3 modes:

Conduction, convection and **radiation**

Water in soil-atmosphere system:

In analogy, water flow is conserved at all times but it may be found in three different states:

Vapour, liquid or **solid**

Transport can be in a number of modes:

Convection, precipitation, percolation or **runoff**

1.4 Radiation, energy and water balance:

Atmospheric and hydrospheric processes are best analysed within the framework of energy flow systems.

Boundary layer climates operate on time scales of less than one day: Diurnal energy regime at ideal site (flat, homogeneous, moist, bare or low vegetation, cloudless conditions) to ensure that processes are spatially uniform and restricted to the vertical dimension.

Radiation balance (units are in W m^{-2})

Shortwave radiation (or solar radiation): (0.15–3 μm; see definition of radiation spectrum, GE2219 - Lecture 1)

$$K\uparrow = K\downarrow(\alpha)$$
$$K^* = K\downarrow(1-\alpha)$$

where α - albedo

Longwave radiation (or terrestrial radiation): (3-100 μm)

$$L\uparrow = \varepsilon_0 \sigma T_0^4 = f(T_0, \varepsilon_0)$$
$$L\downarrow = \varepsilon_a \sigma T_a^4 = f(T_a, \varepsilon_a) \qquad \text{for cloudless skies}$$
$$L\uparrow = \varepsilon_0 \sigma T_0^4 + \underbrace{L\downarrow(1-\varepsilon_0)}_{\text{reflected incoming}}$$

where subscripts *a* and *0* stand for atmosphere and surface, respectively, T – temperature, ε – emissivity and σ – Stefan Boltzman constant (5.67 10^{-8} W m^{-2} K^{-4})

Net-radiation balance:

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

Explanations and examples [Handouts: Figs. 1.8, 1.9, 1.11 and 1.12 from Oke, 1987]:

$K\downarrow$: In the ideal case (Fig. 1.9) the incoming solar radiation is controlled by the azimuth and zenith angles of the sun relative to the horizon, with a maximum at local solar noon. On average 75% of $K\downarrow$ is received as direct (S) and 25% as diffuse (D) radiation (but depends strongly on cloud amount and composition of atmosphere).

$K\uparrow$: Shortwave radiation reflected off the surface depends on $K\downarrow$ and the albedo, α. Since α is relatively constant with time, $K\uparrow$ is a mirror image of $K\downarrow$.

$L\downarrow$: Incoming longwave radiation, emitted by the atmosphere in the absence of clouds, depends upon atmospheric temperature, T_a, and emissivity, ε_a (~ 0.7), in accord with Stefan-Boltzmann law. Neither of these quantities changes a lot with time and $L\downarrow$ is almost constant through the day (300 – 400 W m^{-2}).

$L\uparrow$: The outgoing longwave radiation from the surface is governed by surface temperature, T_0, and emissivity, ε_0 (= 0-1). If ε_0 is less than unity, material is a grey body. Second term on r.h.s. accounts for the amount of $L\downarrow$ that is reflected. Adjustment is usually very small because T_a and ε_a are smaller than corresponding surface values.

- The net radiation (**Q^***) is the most important energy exchange because it represents the limit to the available energy source or sink.
- Typically Q^* involves a daytime surface radiant surplus when net shortwave gain (**K^***) exceeds net longwave loss (**L^***). At night, Q^* is negative in the absence of solar radiation.
- Differences in Q^* between different type of surfaces primarily depend on the values of T_0 and α.

Surface energy balance

The net radiation flux is the end result of the radiation budget and provides the basic input into the surface energy balance:

$$Q^* = Q_H + Q_E + Q_G$$

$\boldsymbol{Q_H}$: Sensible heat flux [W m^{-2}]

$\boldsymbol{Q_E}$: Latent heat flux [W m^{-2}]

$\boldsymbol{Q_G}$: Ground (storage) heat flux [W m^{-2}]

The sign convention is that non-radiative fluxes directed away from the surface are positive. Thus the terms on r.h.s of above equation are positive when they represent a loss of heat for the surface and negative when they are gains. During daytime Q_E and Q_H are usually losses and Q_G a gain at the surface and the opposite holds at night. [Handout: Fig. 1.11 from Oke, 1987]

Daytime: Q^* is dissipated by Q_G, Q_E and Q_H, depending on surface characteristics and availability of moisture. Free convection enhances transport away from surface.

Nighttime: Q^* loss is replenished by conduction upwards from soil (Q_G). Turbulent transport is dampened by stable stratification of atmosphere.

Partitioning of radiative surplus or deficit is governed by nature of surface and the relative ability of the surface and atmosphere to transport heat: [Handout: Fig. 8.10 from Oke, 1987]

Mass balance: [Handouts: Figs. 1.13 and 1.14 from Oke, 1987]

Similar to energy, cycling of mass (e.g. water, CO_2, etc.) can be analysed within the same balance framework:

$$p = E + \Delta r + \Delta S$$

p : Precipitation [mm]

E : Evapotranspiration (loss of water to the air from all sources) [kg m^{-2} s^{-1}]

Δr : Net runoff [mm]

ΔS : Net change in soil moisture (storage) content

Linkage to energy balance through evaporation term:

$$Q_E = L_v E$$

The mass flow (E) can be converted to an energy flow (Q_E) by multiplication with latent heat of vaporization (L_v = 2.453 MJ kg^{-1} at 20 degC)

These notes are based on a course taught by T.R. Oke and M. Roth in 1999 at UBC (Geography)