

CHAPTER 7

EVAPORATION AND ENERGY BALANCE

Most of the water taken up by plants is lost to the atmosphere by transpiration. Transpiration, the loss of water vapor originating as liquid inside the plant tissue, is a special case of evaporation to be discussed later. We'll first consider evaporation in general, be it from an ocean, a soil, a water reservoir, a parking lot, or a community of vegetation. To simplify the discussion, all these objects or complex of objects are referred to as a surface, at least initially.

Energy Supply, Transfer, and Evaporation

Evaporation is best considered in terms of energy supply and the transport of water and energy. Since the heat of vaporization for water is so high (2.44 kJ g^{-1} or 583 cal g^{-1} at 25°C), evaporation is closely dependent on energy supply, solar radiation being the immediate or ultimate energy source under natural conditions. The fate of solar radiation is shown in Fig. 1. Radiation

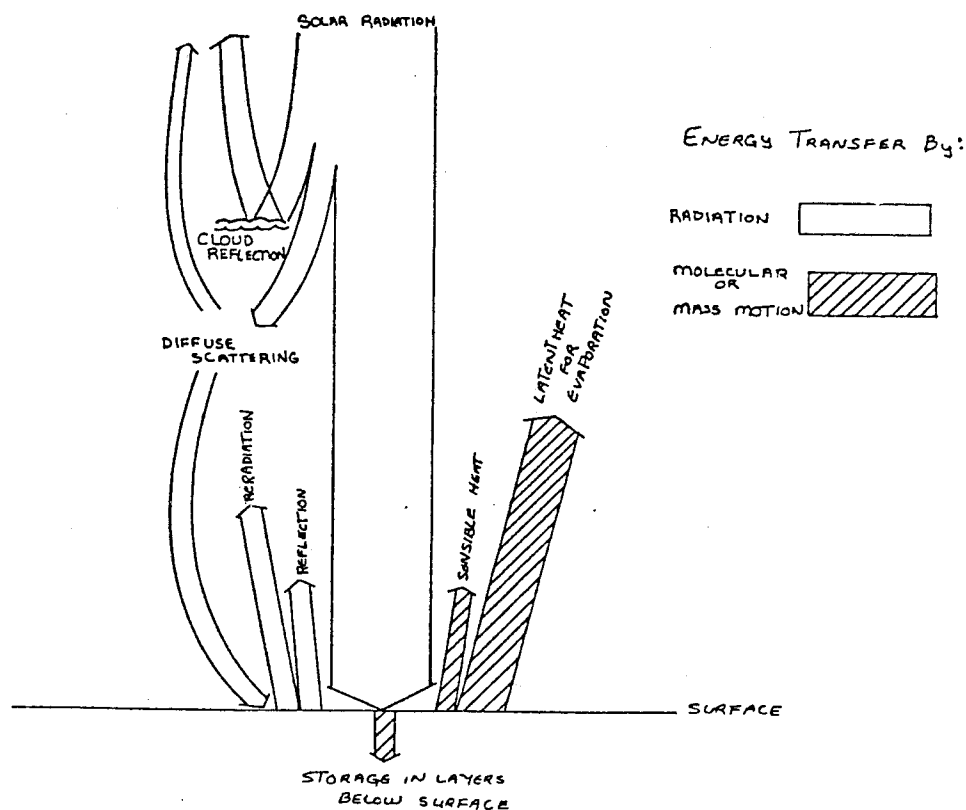


Figure 1. Diagrammatic representation of fate of energy of solar radiation at mid day. The width of the arrows approximates the proportion of energy transferred.

of any wavelength carries energy. Therefore, radiations of all wavelengths are pertinent, not just those within a narrow band of the spectrum, such as the visible range. When radiation strikes a surface, a part of it may be reflected, a part of it may pass through the surface, and the remainder of it absorbed. At the same time, the surface is emitting its own radiation. This happens even if the surface appears dark or black, as long as the temperature of the surface is above absolute zero (0°K). The radiation emitted, however, is of wavelengths longer than red (infrared and longer) and invisible to the eye, if the temperature of the surface is not much above ambient. A substantial part of the radiant energy absorbed by a surface may be dissipated by re-radiation. The net radiation of a surface (R_n) is the difference between the incoming (incident) and the outgoing (reflected, transmitted, and reradiated or radiated) radiations.

Energy is exchanged by several other processes in addition to transfer as electromagnetic waves. If a surface is wet and there is evaporation, energy would be consumed at a rate proportional to the rate of evaporation (E), or more specifically, at a rate lE , where l is the heat of vaporization. In meteorology, lE is referred to as the rate of latent heat transfer. The word latent is used because heat per se is not transferred, rather, heat or energy is consumed in the evaporation of water, and water vapor is what is really transferred. If the surface is hotter than the air, it will warm the air immediately adjacent to the surface. Air movement then provides another means of transferring energy. Energy is removed from the surface by the movement of this hot air layer away from the surface and the replacement of the hot air by colder air from above. This transfer of energy involves heat that can be felt, and is therefore termed sensible heat transfer and given the symbol H . Another means of removing energy is to store it away. Theoretically, a surface is infinitesimally thin. The layers of material below the surface, however, would have a capacity to store heat by warming up. This warming up represents a transfer of energy away from the surface to the layers below and can be represented by the symbol S , signifying the rate of physical storage. Still another means of consuming energy is to store it away in chemical bonds of compounds metabolically synthesized by organisms such as plants present on the surface. This rate of chemical (or metabolic) storage is given the symbol M here.

The foregoing discussion referred mostly to the transfer of energy away from the surface. It should be clear that transfer can also occur in the opposite direction, to the surface. The more common convention is to take incoming energy to the surface as positive, and outgoing as negative. By the law of conservation of energy, at any time the energy received by a surface must equal to that lost, used, and stored by the surface. That is, the income must balance the outgo. Or, stated as an equation,

$$R_n + lE + H + S + M = 0$$

(1)

All terms in the above energy balance equation are rates, expressed as energy transferred per unit area per unit of time (energy flux). Again, each term can be either positive or negative. R_n is positive during the day but usually negative at night. λE would be negative if water is evaporating (using up energy), but positive if water is condensing as dew on the surface (yielding energy). H would be positive when the surface is colder than the air above, and negative when the surface is hotter than the air above. Similarly, S would be positive if the surface is colder than the layers below the surface; the layers below would yield heat to the surface and cool down. On the other hand, S would be negative if the surface is warmer and heat moves from the surface to the layers below and warming them up. The storage flux (S , either positive or negative) can be large if temperature of the surface is changing rapidly, but negligible relative to the other terms in the energy balance equation if the temperature is changing very slowly. Usually, M (due to photosynthesis, using light energy, or to respiration, yielding energy as heat) is quite small relative to the other terms and, therefore, is neglected. Equation (1) is then simplified to

$$R_n + \lambda E + H + S \cong 0 \quad (2)$$

R_n is determined mostly by angle of the sun, cloudiness of the sky, and optical properties of the surface, and is not directly affected by λE and H . If S is small in magnitude relative to the other terms in Eq. (2), a change in λE would bring about an opposite change in H , and vice versa, while R_n may remain little affected. How λE changes are discussed below.

Water Vapor Transfer and Heat Transfer

For a given positive R_n , what determines how much of a portion of the energy received should go to evaporation and how much to sensible heat transfer? Put it another way, what determines the ratio $H/\lambda E$ for a given R_n ? It turns out that the answer is the ratio of the water-vapor-concentration gradient and the temperature gradient from the surface to the air above. To see how this is the case, we describe the transport of water vapor and of sensible heat in terms of transport equations.

Water vapor transfer from a large and relatively uniform field or body of water is predominantly upward, with the atmosphere acting as a sink of almost unlimited size. In many cases, heat transfer to or from such surfaces is also mainly in the vertical direction. Therefore, both water and heat transfers can be treated as one-dimensional (vertical) transport processes. Using versions of the general transport equation, we write expressions for flux of water vapor (W) and for flux of heat (H):

$$W = \rho_a K_w \frac{dq}{dz} \quad (3)$$

$$H = c_p \rho_a K_h \frac{dT}{dz} \quad (4)$$

Here, W is in g of water transferred per m^2 of surface area per second; and H is joule of heat transferred per m^2 of surface area per second; ρ_a is air density in $g\ m^{-3}$; c_p is specific heat capacity of the air in $J\ g^{-1},\ deg^{-1}$, K_w and K_h are respectively eddy transfer coefficients for water vapor and for heat in $m^2\ s^{-1}$, q is water vapor concentration in g of water per g of air, T is temperature in degrees K, and z is height above the surface in m. The appropriate driving forces for water vapor and heat transport are respectively gradients of water vapor concentration $\frac{dq}{dz}$ and gradients of

temperature $\frac{dT}{dz}$. For discrete height increments, the gradients are $\frac{\Delta q}{\Delta z}$ and $\frac{\Delta T}{\Delta z}$.

The eddy transfer coefficients K_w and K_h reflect largely the turbulent mixing of the air. The more turbulent air is, the faster is the mixing and the faster the net transport of the water vapor would be from where its concentration is relatively high to where its concentration is relatively low. Similar consideration applies to the transfer of sensible heat, from where the temperature is relatively high to where the temperature is relatively low. Needless to say, the eddy transfer coefficients are highly dependent on wind velocity. An increase in wind velocity increases the coefficients. Therefore, K_w and K_h are not constants and usually vary markedly with time and conditions. Note that fluxes in Eqs. (3) and (4) are not related to the driving forces by a minus sign, as done in other transport equations. The reason is that here we have defined outgoing as negative, making the minus sign unnecessary.

Equations (3) and (4) are not analogous in form since the flux in one case is water while in the other it is energy. In order to apply the energy balance approach, we convert water flux to energy flux by recognizing that under steady-state conditions, the vertical flux of water vapor (W) at any height above the surface must be the same as the rate of water evaporation per unit area at the surface (E). The assumption of steady-state conditions is similar to the application of the equation of continuity in treating flow problems in pipes. For the steady-state situation, W at height z equals E at the surface. That is, for every gram of water transferred at height z , 1 g of water must be evaporated at the surface, or, $W = E$. To convert water vapor transfer to energy transfer, we start with Eq. (3) and recognize that the evaporation of 1 g of water requires l joule of energy,

$$lE = l \rho_a K_w \frac{dq}{dz} \quad (5)$$

In Eqs. (4) and (5), l , ρ_a and c_p are nearly constants and are used to give the equations dimensional consistency. The important terms in these equations to focus our attention on are the fluxes (H and lE), eddy transfer coefficients (K_h and K_w), and the gradients $\frac{dq}{dz}$ and $\frac{dT}{dz}$.

It turns out that, during the day over open areas, mass flow of air dominates over molecular diffusion as the mechanism of transport. Mass flow is the result of wind motion or thermal convection. Thermal convection refers to air motion set up by the tendency of warmer and lighter (because of expansion induced by higher temperature) air masses to float upward and of cooler and denser air masses to sink. Needless to say, such convection occurs only if the lower air layer is warmer and lighter than the layer above it. With mass flow, any material in a volume of air is carried along with the air. Hence, there is little or no differentiation, as far as transport is concerned, among the substances contained in that air; be they heat, water vapor, or air pollutants. It follows that the eddy transfer coefficients for heat and water vapor are similar or identical ($K_h = K_w$, known as Reynold's analogy). In some nights with no wind and little thermal convection (surface air cooler and, therefore, denser than the air above, resulting in a stable air mass), transport is largely limited to molecular diffusion. Then the equality between K_h and K_w will not hold. In fact, they should no longer be termed eddy transfer coefficients.

Examining Eqs. (4) and (5) and keeping in mind that l , ρ_a , and c_p are nearly constant and that $K_w = K_h$, it can be seen that when H and lE are outgoing, the partition of outgoing energy (which equals incoming energy, R_n , according to Eq. (2)) between H and lE is determined by the size of $\Delta q/\Delta z$ relative to $\Delta T/\Delta z$. In other words, how much of the energy received by the surface goes to sensible heat transfer and how much goes to evaporating water are determined by the respective driving "force" gradients. The importance of the temperature and water vapor concentration gradients in affecting transport is further seen in Fig. 2.

An example of daily changes in the terms of Eq. (2) is given in Fig. 3. In the arid climate where the data were collected, a substantial part of the energy for evaporation from well watered fields comes from sensible heat. That is, H is substantially positive for most part of the day. As the field runs out of water 27 days after the irrigation, stomata of the crop closed substantially and lE was reduced. Consequently, H changes to negative and a portion of the energy received as net radiation is dissipated as sensible heat.

Water vapor concentration at a wet surface is dependent on surface temperature. Temperature of liquid water has a marked effect on the water vapor concentration in equilibrium with it, as shown in Fig. 4. The very thin layer of air immediately adjacent to the water or wet surface is essentially saturated with water vapor. The warmer the water or wet surface, the higher is the

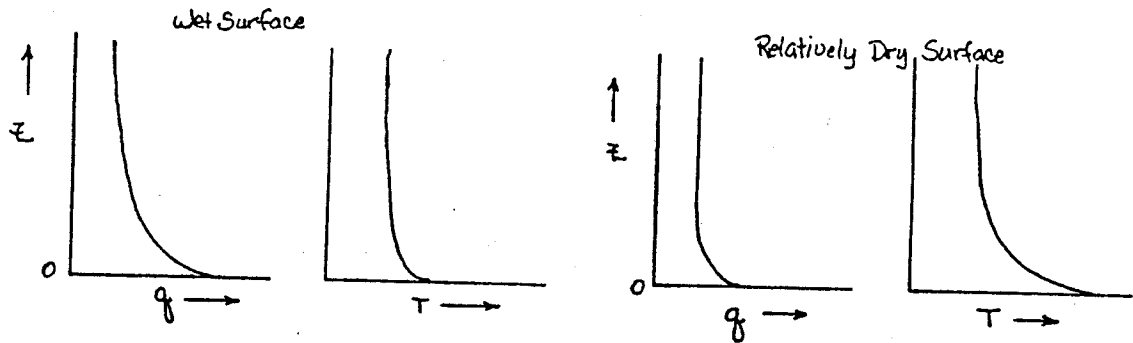


Figure 2. Conceptual depictions of vertical (z direction) profiles of water vapor concentration in the air and air temperature over a wet surface as contrasted with those over a relatively dry surface, under high radiation. The surface-air boundary is at $z = 0$. The large $\Delta q/\Delta z$ and small $\Delta T/\Delta z$ over the wet surface indicate that most of the R_n received is dissipated as latent heat. The small $\Delta q/\Delta z$ and large $\Delta T/\Delta z$ over the relatively dry surface indicate that most of the R_n is dissipated as sensible heat.

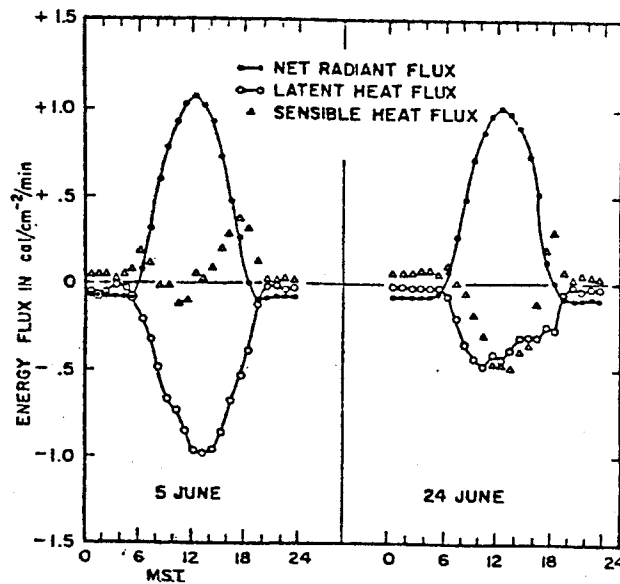


Figure 3. Principal components of the hourly energy balance of irrigated alfalfa at Phoenix (Arizona) in 1964, 8 and 27 days after irrigation. Reproduced from C. H. M. van Bavel (1967). Changes in canopy resistance to water loss from alfalfa induced by soil water depletion. *Agr. Meteorol.* 4:165-176.

water vapor concentration of the air adjacent to the surface. It is important to keep in mind that when a wet surface cools down or warms up, not only is the temperature gradient to the air above affected, but also the water vapor concentration gradient.

Bowen Ratio

The ratio of H to λE is known as the Bowen ratio, β . From Eqs. (4) and (5),

$$\beta = \frac{H}{\lambda E} = \frac{c_p \Delta T / \Delta z}{\lambda \Delta q / \Delta z}$$

We are dealing with the same height interval for the transport of heat and water, so Δz 's cancel,

$$\beta = \frac{H}{\lambda E} = \frac{c_p \Delta T}{\lambda \Delta q} \tag{6}$$

Equation (6) shows that if the temperature and humidity differences between two heights are known (these heights can be chosen for the convenience of taking measurements), and steady-state conditions are approached, β is easily calculated. If R_n and S are also known, λE (hence evaporation or evapotranspiration) and H can be estimated utilizing Eqs. (2) and (6). This estimation of evaporation from Δq , ΔT , R_n , and S is called the Bowen ratio method and is quite accurate when necessary conditions are met. [Question: Knowing how the equations come about and the assumptions involved, what are some of the necessary conditions?]

It is hoped that the preceding material not only provides information on evaporation but also demonstrates how the application of some relatively simple physical principles, along with some reasonable assumptions, can condense rather complex natural phenomena to basic essentials. The concepts developed have wide applicability and predictive value. For example, with appropriate modifications, they can be applied to the movement of airborne pollutants, to CO_2 assimilation by a vegetative surface, and to heat balance (and hence temperature regulation) in animals.

Reference

Rosenberg, N. J., B. L. Blad, and S. B. Verma. 1983. Microclimate: the Biological Environment. John Wiley and Sons, New York.

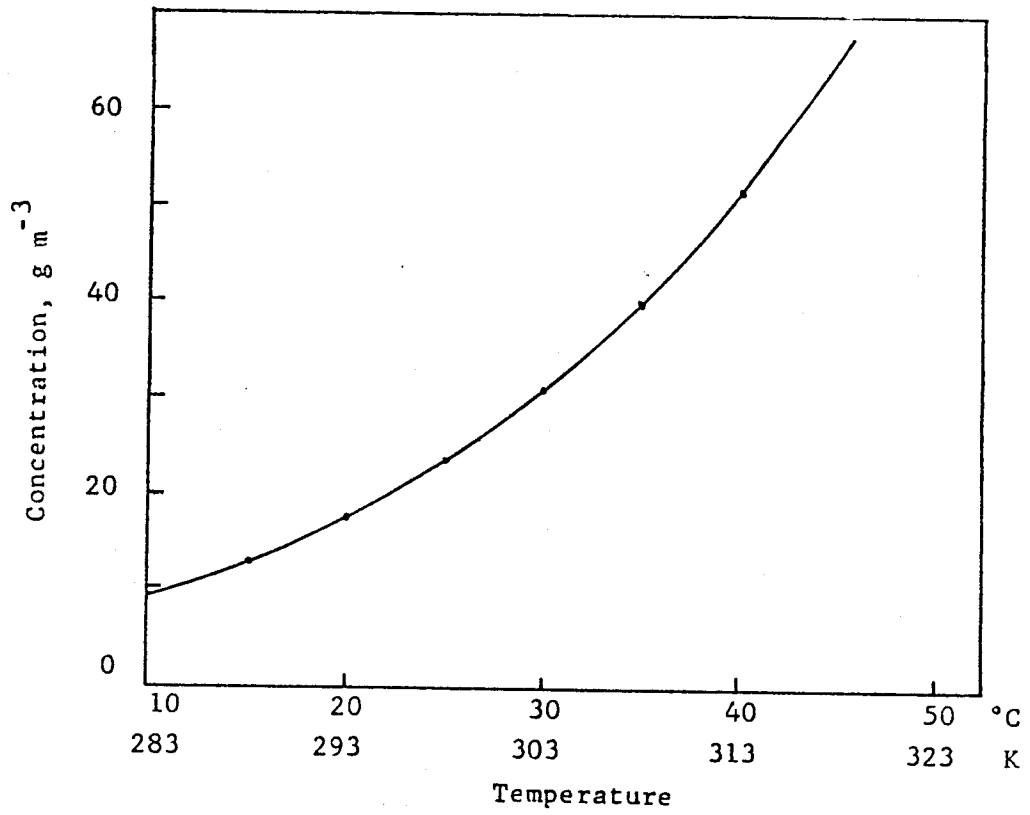


Figure 4. Saturation water vapor concentration in air as affected by temperature.