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Evaporation and surface temperature

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(Presidential address: delivered 20 June 1979)

1. Beginnings

According to the Oxford dictionary, the terms 'meteorology' and 'meteorologist' came into use about 1620 but the development of meteorology as science had to wait for much later developments in physics, particularly in heat and thermodynamics, during the late eighteenth and nineteenth centuries. Throughout the seventeenth century, observing and recording the weather provided an amiable hobby for country parsons, and excursions into experimental meteorology were rare. They also seem somewhat naive when compared with contemporary progress in astronomy for example. The mathematical knowledge of Edmund Halley, applied to the orbits of celestial bodies, was generations ahead of his contribution to the physics of evaporation as recorded in the *Philosophical Transactions* of the Royal Society. The volume for 1687 contains a paper entitled 'An Estimate of the Quantity of Vapour raised out of the Sea by the warmth of the Sun; derived from an Experiment shown before the Royal Society at one of their late Meetings.' Halley wrote:

'That the great quantity of aqueous Vapour contained in the Medium of the Air is very considerable seems most evident from the great Rains and Suns which are sometimes observed to fall . . . but in what proportions these Vapours rise . . . has not, that I know of, been any where well examined tho it seems to be one of the most necessary Ingredients of a real and Philosophical Meteorology, and as such to deserve the consideration of this Honourable Society.'

We now recognize that the physics of natural evaporation is a necessary ingredient not only of meteorology but of oceanography and hydrology too, so I hope the subject will seem appropriate for a Presidential Address to *this* honourable Society.

Unfortunately, I am prevented from repeating Halley's experiment by the fire regulations of Imperial College. He took a pan of water about 4 inches deep and 8 inches diameter, placed it on burning coal and raised the temperature 'to the same degree of heat which is observed to be that of the air in our hottest summers, the thermometer nicely showing it'. The weight of water decreased by 233 grains in 2 hours equivalent to about 1/10 of an inch in 12 hours. This figure happens to be about right for summer evaporation in Britain but Halley's main inference was that the evaporation from the Mediterranean must be about 5280×10^6 tons per day. Even in those days, extrapolation by laboratory meteorologists was sometimes rather far-fetched.

Halley's fire simulated the heating power of the sun. The ventilation of his pan appears to have been uncontrolled, but he was aware of the significance of air movement and referred to 'the Winds whereby the Surface of the Water is licked up sometimes faster than it exhales by the heat of the Sun'. Aristotle made the same point, so the subject had not made much progress in more than two thousand years.

Progress continued to be slow. One of Halley's successors as Astronomer Royal, speaking at the British Association in 1862, referred to the 'desperate science of meteorology' and G. J. Symons (1867), apparently amused rather than offended by the epithet, went one better by suggesting that evaporation was 'the most desperate art of the desperate science of meteorology'. One of the main reasons for despair was the lack of techniques for measuring evaporation over natural surfaces. Monthly rates of evaporation from small pans were recorded as early as 1772, and Victorian meteorologists seem to have vied with each other in devising ingenious gadgets for measuring evaporation from water and wet surfaces. Symons wrote, 'I hate evaporators', and attempting to settle the matter once and for all, he designed the tank still in use by our Meteorological Office.

But the physics of natural evaporation remained obscure and when Sir Napier Shaw published his *Manual of Meteorology* as recently as 1926, he wrote: 'Evaporation . . . has not yet reached its proper position in the discussion of the physical processes of the sequence of weather, partly because these processes have not yet been quantitatively explored and partly also because the greater amount of evaporation takes place in *an entirely uncontrolled manner* . . . on so vast a scale as to make insignificant anything that can be measured in a gauge.' The lack of knowledge revealed by the surprising phrase I have italicized was soon dispelled by the work of Bowen (1926) who showed how the partitioning of sensible and latent heat at a water surface could be determined from gradients of temperature and humidity; and somewhat later, Thornwaite and Holzman (1939) estimated evaporation from pasture by measuring gradients of vapour pressure and by determining a transfer coefficient from the wind profile.

The thermodynamic and aerodynamic aspects of evaporation were not fully reconciled until, in 1948, Howard Penman published a paper which has become one of the major classics of microclimatology; 'Natural evaporation from open water, bare soil and grass.' The Penman formula was soon adopted by hydrologists and irrigation engineers, but meteorologists were more cautious. With some justification, they were concerned about the poor exposure of the evaporation tanks which surrounded the deep brick-lined 'Bear Pit' at Rothamsted (see Penman 1948, Fig. 7); and they were also critical of the relatively large number of empirical constants which were needed to determine net radiation, for example, in the days before instruments were available to measure this quantity. A comment by Pasquill (1950) was probably characteristic of the attitude of the meteorological establishment 30 years ago. He wrote: '. . . in view of the restricted empirical basis of the method and the special circumstances of the test, it seems doubtful that the method can be applied with confidence outside the circumstances of the test.' It is also significant that the first major textbook on micrometeorology, published by Sutton in 1953, confines Penman's work to one short descriptive paragraph.

Over the past 30 years the value of Penman's contribution to the study of natural evaporation has been immense, both in the assessment of regional water resources and in the use of these resources for irrigation. The subject deserves a much more thorough review than is feasible within the scope of a Presidential Address, even in its extended written form. I shall confine myself to showing first how the analysis on which the Penman formula is based incorporates a number of fundamental principles relating the evaporation from a wet surface to its equilibrium temperature. I shall then try to demonstrate how the theory can be applied to a number of diverse systems.

2. Definitions

When water evaporates at the interface between a wet surface and the atmosphere, the temperature at each point on the surface tends to an equilibrium value at which the local *loss* of latent heat is balanced by the net *supply* of heat by processes such as radiation, convection and conduction. In micrometeorology it is conventional to regard the net flux density of radiation R_n as the main component of energy supply, balanced by losses of latent heat (λE), sensible heat (C)* and by storage in the substrate (G). With this convention, the heat balance of a surface is written as

$$R_n = \lambda E + C + G \quad . \quad . \quad . \quad . \quad (1)$$

Whereas the terms on the right-hand side of this equation are strong functions of surface temperature, R_n is a weak function of that temperature, especially during the day when R_n is dominated by its short-wave components. During daylight also, G is usually a small fraction of $\lambda E + C$. So to a good first approximation, the adjustment of surface temperature to achieve thermal equilibrium can be regarded as a process which makes $\lambda E(T_0) + C(T_0)$ tend to the value $H = R_n - G$ regarded as independent of temperature.

The equilibrium temperature achieved by a surface and corresponding values of $\lambda E(T_0)$ and $C(T_0)$ depend on the aerodynamic as well as the thermodynamic behaviour of the system containing the surface, so we must consider how to formalize the atmospheric transport of heat and water vapour in a simple and very general way.

When the transport of an entity depends on diffusion (e.g. when it is driven by molecular processes or by turbulence), Fick's First Law can be invoked to describe the flux in a specified direction as the product of a transport coefficient and a gradient of potential in that direction. The appropriate potential is the amount of entity present in unit mass of air, which is conveniently described as a specificity ($\mathscr{S}$). Within the boundary layer characteristic of a uniform surface (water, soil or vegetation) we are usually concerned with vertical transport expressed in terms of the flux of an entity at height z per unit area of ground surface, or $F(z)$. This flux can be treated as a current proportional to the potential gradient $\partial\mathscr{S}/\partial z$, and the constant of proportionality is then an 'Austauschkoeffizient' as used in the classical literature of turbulence (Sutton 1953). The coefficient is commonly replaced by the product of air density $\rho(z)$ and the more familiar diffusion coefficient $K(z)$ which has the same dimensions as the coefficients of molecular diffusivity viz. L^2T^{-1}. The vertical flux density of an entity can therefore be expressed as the product of three factors

$$F(z) = -\rho(z)K(z)\partial\mathscr{S}/\partial z \quad . \quad . \quad . \quad (2)$$

Because $\partial\mathscr{S}/\partial z$ cannot be measured directly, it is convenient to integrate Eq. (2) between heights z_1 and z_2 assuming that $F(z)$ has a constant value F between these heights. It follows that

$$F = \{\mathscr{S}(z_1) - \mathscr{S}(z_2)\} \Big/ \int_{z_1}^{z_2} \{\rho(z)K(z)\}^{-1}\, dz \quad . \quad . \quad (3)$$

Within a few metres of the earth's surface, ρ changes with height much less rapidly than K so it is appropriate to replace $\rho(z)$ by a weighted mean density $\bar{\rho}$ defined by

$$\bar{\rho}^{-1} = \int \{\rho(z)K(z)\}^{-1}\, dz \Big/ \int K(z)^{-1}\, dz \quad . \quad . \quad (4)$$

Substituting Eq. (4) in Eq. (3) then gives

$$F = \bar{\rho}\{\mathscr{S}(z_1) - \mathscr{S}(z_2)\} \Big/ \int K(z)^{-1}\, dz \quad . \quad . \quad (5)$$

* Many workers use H for sensible heat but this symbol is more appropriate for the enthalpy flux $\lambda E + C$ and will be so used in this paper. For the latent heat of vaporization, I prefer λ to L which is often needed for length or leaf area index in the analysis of evaporation from vegetation.

This equation is often treated as an analogue of Ohm's Law with the numerator of the right-hand side representing a potential difference and the denominator a resistance. In several texts (e.g. McIntosh and Thom 1969; Monteith 1975; Campbell 1977) the potential driving the diffusion process is inexactly defined as the volume concentration of the entity, i.e. $\rho(z)\mathscr{S}(z)$ rather than the product of $\bar{\rho}$ and the specificity $\mathscr{S}(z)$. In practice, $\rho(z)$ is almost invariably assigned an arbitrary value independent of height so that the distinction between the two definitions is largely academic in analysis. (Experimentally, however, it is important to take account of the way in which gradients of heat and/or water vapour modify the gradient of density – see, for example, Webb, Pearman and Leuning 1980).

Equation (5) shows that the diffusive flux of any entity can be expressed as the product of the difference in specificity of the entity between two points and the quantity $\bar{\rho}/r$ where $r = \int K(z)^{-1}\,dz$ is a resistance. This quantity has dimensions $ML^{-2}T^{-1}$; it can be regarded as the mass of air involved in the exchange of an entity with unit area of a surface in unit time. In the analysis which follows, transfer of sensible heat will be related to a resistance r_H and to differences in the specificity c_pT where c_p is the specific heat of air at constant pressure. Similarly, the transfer of latent heat will be related to a resistance r_v and a difference in the specificity λq where λ is the latent heat of vaporization and q is the specific humidity of air (weight of water vapour per unit mass of moist air).

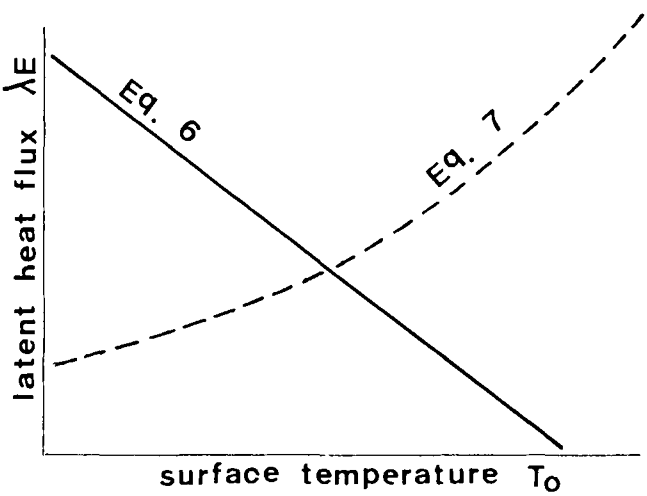

Figure 1. Dependence of latent heat flux on surface temperature as expressed by the heat balance equation (Eq. (6)) and the aerodynamic equation (Eq. (7)).

The general problem of partitioning available energy into sensible and latent heat can now be illustrated in a simple graph (Fig. 1) representing the two ways in which latent heat flux can be expressed as a function of the temperature of a wet surface. The heat balance equation can be written in the general form

$$\lambda E = H - C = H - \bar{\rho}c_p(T_0 - T)/r_H \qquad (6)$$

where r_H is the resistance between the surface and a convenient reference point at which air temperature is T. The enthalpy input H is $R_n - G$ in Eq. (1).

The aerodynamic equation for the same system can be written in the form

$$\lambda E = \bar{\rho}\lambda\{q_s(T_0) - q\}/r_v \qquad (7)$$

where q_s is the saturation value of the specific humidity at T_0 and r_v is a vapour transfer resistance between the surface and a reference point with specific humidity q.

Given the input of heat to the system H, the pair of variables (T, q) defining the state of air passing over the surface, and the pair of transfer resistances, it is possible to find the values of λE and T_0 which satisfy Eqs. (6) and (7). Figure 1 illustrates the general scheme and specific cases are considered later.

In meteorological problems the values of λE and T_0 derived by solving Eqs. (6) and (7) are not exact when H is a weak function of temperature through its dependence on the exchange of long-wave radiation at the surface. A more exact solution can be readily obtained by writing

$$R_{ni} = R_n - \sigma(T_0^4 - T^4)$$

where R_{ni} is an 'isothermal' flux density – the net radiant flux which the surface would receive if T_0 were equal to the reference temperature T; and σ is Stefan's constant. For small values of $T_0 - T$

$$\sigma(T_0^4 - T^4) \simeq 4\sigma(T_0 - T)T^3 \equiv \bar{\rho}c_p(T_0 - T)/r_R \qquad . \qquad . \qquad (8)$$

where $r_R = 4\sigma T^3/\bar{\rho}c_p$ can be regarded as a resistance to (long-wave) radiative transfer. Equation (1) now becomes

$$R_{ni} = \lambda E + \bar{\rho}c_p(T_0 - T)(r_H^{-1} + r_R^{-1}) + G \qquad . \qquad . \qquad (9)$$

It follows that for any solution of Eqs. (6) and (7), the dependence of R_n on T_0 can be allowed for by substituting R_{ni} for R_n and by replacing r_H by the (smaller) resistance offered by r_H and r_R in parallel, i.e. $(r_H^{-1} + r_R^{-1})^{-1}$. At 20°C, $r_R \simeq 200\,\mathrm{s\,m^{-1}}$ and is usually substantially larger than r_H.

The analytical solution of Eqs. (6) and (7) and a number of special cases will now be considered.

3. Solutions

Meteorologists enjoy posting through the atmosphere 'parcels' of air of indeterminate size and shape and this procedure can be used to find the evaporation from a surface whose temperature is not known *a priori*. When a parcel is kept near the ground at constant pressure, its thermodynamic state can be represented by the amount of sensible and latent heat it contains, i.e. by $(c_pT, \lambda q)$ or simply by (T,q). This state is represented by a point such as X in Fig. 2.

The curve SS′ is the locus of a saturated parcel and has a slope $\partial\{\lambda q_s(T)\}/\partial(c_pT)$. For the range of temperatures relevant to evaporation at the earth's surface, the quantity $c_p/\lambda = \gamma$ can be assigned a constant value of $4{\cdot}08 \times 10^{-4}\,\mathrm{K^{-1}}$. The quantity $\partial q_s(T)/\partial T = \Delta$ increases from $2{\cdot}81 \times 10^{-4}\,\mathrm{K^{-1}}$ at 0°C to $16{\cdot}6 \times 10^{-4}\,\mathrm{K^{-1}}$ at 30°C. The slope of SS′ which is Δ/γ therefore increases from 0·69 to 4·08 over the same temperature range.*

The unsaturated parcel at X can be saturated by evaporating water into it either adiabatically or diabatically. In an *adiabatic* process the increase in latent heat content is balanced by the decrease in sensible heat content so the locus of such a process in Fig. 2 is a line whose slope is -1, i.e. XY. To a good approximation, the coordinates of Y are $(c_pT', \lambda q_s(T'))$ where T' is the thermodynamic wet-bulb temperature of the parcel. The heat used for evaporation per unit mass of air is $c_p(T - T')$ and from the geometry of Fig. 2

* In most literature concerned with the Penman formula, vapour pressure is used in place of specific humidity. Δ and γ then have units of $\mathrm{mb\,K^{-1}}$ for example, but the ratio Δ/γ on which the analysis depends has almost the same value as here.

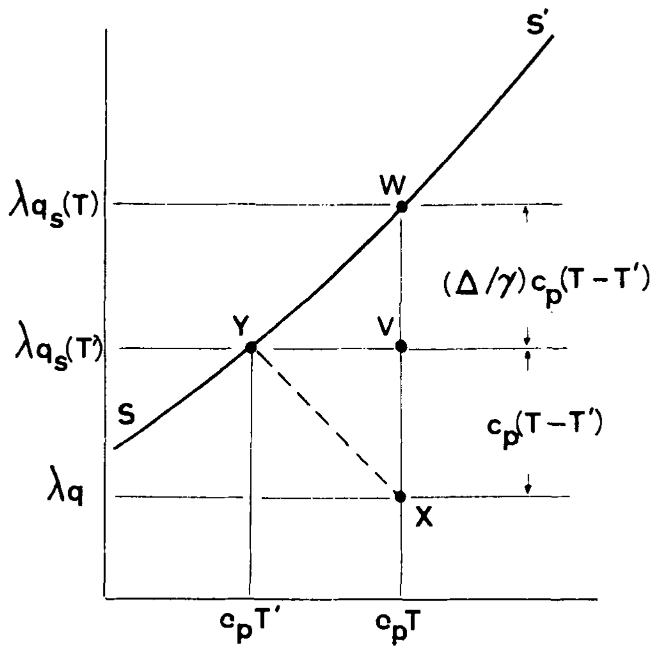

Figure 2. Adiabatic cooling of a parcel of air from X (c_pT, λq) to Y (c_pT', $\lambda q_s(T')$). SS' is the relation between λq and c_pT for saturated air and has a slope Δ. Since XY has a slope of -1,

$$XV = c_p(T-T')$$

and

$$\begin{aligned} WX &= \lambda\{q_s(T)-q\} \\ &= (1+\Delta/\gamma)c_p(T-T') \\ &= (\Delta+\gamma)\lambda(T-T') \quad \text{(see Eq. (10))} \end{aligned}$$

it can readily be shown that

$$c_p(T-T') = c_p\{q_s(T)-q\}/(\Delta_1+\gamma) \quad . \quad . \quad (10)$$

where Δ_1 is evaluated at an appropriate temperature between T and T'.

If the process of evaporation is *diabatic*, the enthalpy of the parcel changes by an amount Q, say. When the parcel becomes saturated, its temperature T'' will be greater or less than T' depending on whether Q is positive or negative. In this case the total amount of heat used for evaporation can be treated as the sum of two components: the heat available from adiabatic cooling to the wet-bulb temperature as before; and the heat available from the partitioning of Q between latent and sensible heat exchanges keeping the parcel saturated on the route YZ (Fig. 3). This ratio is the slope of SS', i.e. Δ/γ where Δ must be evaluated at an appropriate temperature between T' and the final temperature of the parcel T_0. This value will be identified as Δ_2. It follows that a fraction $\Delta_2/(\Delta_2+\gamma)$ of Q should be assigned to latent heat and the complementary fraction $\gamma/(\Delta_2+\gamma)$ to sensible heat. The total input of latent heat per unit mass of air is therefore

$$\Delta_2 Q/(\Delta_2+\gamma)+c_p(T-T')$$

When the difference between surface temperature and air temperature is not large, $\Delta_2 \simeq \Delta_1 = \Delta'$, say; and from Eq. (8), the two components of latent heat can be combined in the form

$$[\Delta' Q+c_p\{q_s(T)-q\}]/(\Delta'+\gamma).$$

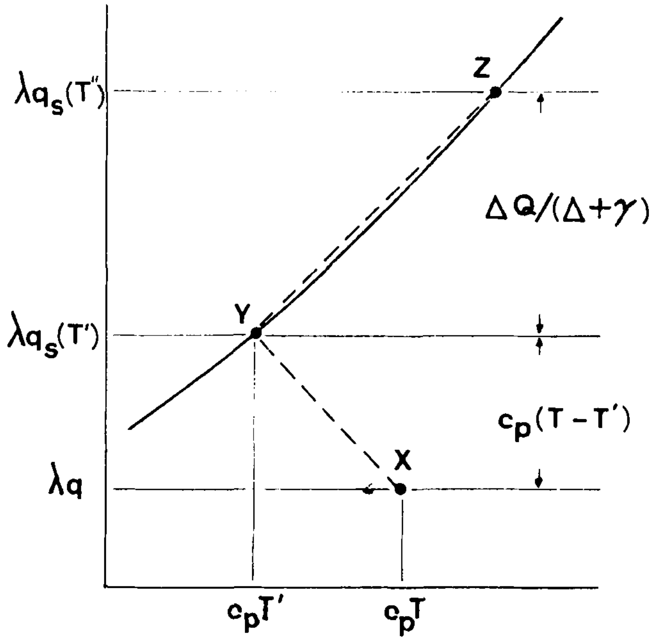

Figure 3. Adiabatic cooling of a parcel of air (XY) followed by heating in a state of saturation (YZ) with an increase of enthalpy Q.

This expression is not exact because it takes no account of change in the heat content of water vapour during evaporation; nor is it fully explicit because the temperature for evaluating Δ' is not known *a priori*. Nevertheless, when the difference between T and T' is only a few degrees K, it provides a close approximation to the partitioning of Q between sensible and latent heat and a more exact value could be obtained by iteration if needed. However, this extension of parcel theory is much less important in practice than the introduction of parameters which allow the rate of exchange at a surface to be evaluated.

Suppose a parcel containing an entity whose specificity is $\mathscr{S}_1$ initially gains or loses an amount of that entity on contact with a surface so that the specificity of the parcel changes to $\mathscr{S}_2$. The rate of exchange is given by the specificity difference $\mathscr{S}_2-\mathscr{S}_1$ (gram of entity per gram of air) multiplied by $\bar{\rho}/r$ (gram of air $m^{-2}\,s^{-1}$) where r is an appropriate resistance. It follows that Fig. 2 can be used to analyse the process of exchange at a surface provided the ordinate (specificity of latent heat) is multiplied by $\bar{\rho}/r_V$ and the abscissa (specificity of sensible heat) is multiplied by $\bar{\rho}/r_H$. The geometry of the figure remains unchanged but the slope of SS′ becomes Δ/γ^* where $\gamma^* = \gamma(r_V/r_H)$.

If air making contact with a surface were cooled adiabatically, the rate of loss of latent heat per unit area of the surface would be $\bar{\rho}c_p(T-T')/r_H$. If the same parcel of air gained enthalpy at a rate H, latent heat would be added at a rate $\Delta' H/(\Delta'+\gamma^*)$. The total flux density of latent heat can therefore be written

$$\lambda E = \frac{\Delta' H}{\Delta'+\gamma^*} + \frac{\bar{\rho}c_p(T-T')}{r_H} \quad . \quad . \quad . \quad (11a)$$

$$= \frac{\Delta' H + \bar{\rho}c_p\{q_s(T)-q\}/r_H}{\Delta'+\gamma^*} \quad . \quad . \quad . \quad (11b)$$

The corresponding sensible heat flux is

$$C = \frac{\gamma^* H}{\Delta' + \gamma^*} - \frac{\bar{\rho} c_p (T - T')}{r_H} \quad . \quad . \quad . \quad (12)$$

and the equilibrium surface temperature is

$$\begin{aligned} T_0 &= T + C r_H / \bar{\rho} c_p \\ &= T' + \gamma^* r_H H / \bar{\rho} c_p (\Delta' + \gamma^*) \quad . \quad . \quad . \quad (13) \end{aligned}$$

Equation (11b) has the same general form as the original Penman formula in which the saturation deficit was defined in terms of vapour pressure and the resistance r_H was incorporated in an empirical wind function. Several extensions of the equation will now be examined.

Case I. Relative humidity at surface is $h = 1$ so that the specific humidity of air in contact with the surface is $q_0 = q_s(T_0)$. This specification is appropriate for pure water or surfaces such as soil or foliage thoroughly wetted by rain. To a good approximation (see Robinson 1966), $r_v = r_H$ and $\gamma^* = \gamma$.

Case II. Relative humidity at surface is $h < 1$ so that $q_0 = h q_s(T_0)$. This specification is appropriate for water containing a substantial amount of dissolved salt or a layer of uniformly wet porous material. From the principles of physical chemistry, the value of h can be calculated as a function of the salt concentration or from the distribution of pore sizes. The analysis represented by Fig. 2 remains valid but $q_s(T)$ must be replaced by $h q_s(T)$ and Δ' by $\Delta' h$. It follows that the rate of latent heat loss is

$$\lambda E = \frac{h\Delta' H + \bar{\rho} c_p \{h q_s(T) - q\} / r_H}{h\Delta' + \gamma} \quad . \quad . \quad (14)$$

and the corresponding surface temperature is

$$T_0 = T + \frac{\gamma r_H H}{\bar{\rho} c_p (h\Delta' + \gamma)} - \frac{\{h\, q_s(T) - q\}}{(h\Delta' + \gamma)} \quad . \quad . \quad (15)$$

Case III. Specified surface resistance to vapour transfer. When sensible and latent heat exchange occur at a complex surface it is possible for r_V to exceed r_H by an amount r_s, say. The most common example is transpiration through the stomatal pores of a leaf. Then Eq. (11) is valid with

$$\gamma^* = \gamma(r_V / r_H) = \gamma(1 + r_s / r_H) \quad . \quad . \quad . \quad (16)$$

and with the essential proviso that the sources of sensible and latent heat *must have the same temperature* T_0.†

Case IV. Specified saturation deficit or wet bulb depression at the surface. For this case it is convenient to write the specific humidity deficit at a reference height, $q_s(T) - q$ as δq and the corresponding deficit at the surface as δq_0. The analysis illustrated by Fig. 2 can be extended by taking the air which is saturated at T_0 and heating it adiabatically until it has a humidity deficit of δq_0. Then it can be shown that

$$\lambda E = \frac{\Delta' H + \bar{\rho} c_p (\delta q - \delta q_0) / r_H}{\Delta' + \gamma} \quad . \quad . \quad . \quad (17)$$

† Equation (11b) with γ^* given by Eq. (16) is sometimes referred to as the Penman–Monteith equation in the mistaken belief that I derived it (Monteith 1965). In fact, Penman (1953) developed a formally identical equation for single leaves whereas my own analysis was applied to evaporation from a canopy. Covey (1959) may have been the first to evaluate surface resistances for vegetation and Rijtema (1965) interpreted the term in Eq. (9) in almost exactly the same way as I did.

The same equation can be derived by identifying δq_0 as the difference in specific humidity across a resistance r_s so that

$$\lambda E = \bar{\rho}\lambda\, \delta q_0 / r_s \quad . \quad . \quad . \quad . \quad (18)$$

Substitution of Eq. (18) in Eq. (11) then gives Eq. 17. An equation of this type was derived by Slatyer and McIlroy (1961) who used wet-bulb depression as a variable in place of δq_0. The equation has little practical value as it stands because δq_0 is rarely measurable. However, the equilibrium situation $\delta q = \delta q_0$ leads directly to the next and last case.

Case V. Atmosphere in equilibrium with underlying surface. It can be argued from general principles (Priestley and Taylor 1972; McNaughton 1976) that the gradient of the humidity deficit $\partial(\delta q)/\partial z$ should tend to zero as an air mass moves over a surface from which water is evaporating at a finite rate, irrespective of the value of δq at the surface. It is possible to write

$$\frac{\partial(\delta q)}{\partial z} \equiv \frac{\partial\{q_s(T)-q\}}{\partial z} \equiv \frac{\Delta\partial T}{\partial z} - \frac{\partial q}{\partial z} \propto \Delta C - \gamma\lambda E \quad . \quad . \quad (19)$$

The quantity δq is therefore a form of specificity associated with the flux $\Delta C - \gamma\lambda E$. Convergence of this flux will tend to decrease the gradient of δq until a limit is reached at which $\partial(\delta q)/\partial z = 0$, and $C/\lambda E = \gamma/\Delta$ so that

$$\lambda E = \Delta H/(\Delta+\gamma) \quad . \quad . \quad . \quad (20)$$

The fact that this equation usually underestimates observed rates of evaporation is discussed later.

4. Applications

(*a*) *Water: the problem of heat storage*

As the Penman formula was first derived to relate the evaporation from open water surfaces to climatological records, it may seem paradoxical that the formula is not commonly used to estimate the loss of water from reservoirs and lakes, or from the oceans for that matter. The problem is that for any body of water exceeding a metre or so in depth, the amplitude of the annual cycle of evaporation is significantly damped by the heat stored when water temperature rises in spring and released when the water cools later in the year. In an application of the Penman formula to determine evaporation from the Kempton Park reservoirs near London (mean depth about 7 m), Lapworth (1965) showed that heat storage reduced the rate of evaporation in early summer by about 2 cm per month or 25%; and for Lake Tiberias (mean depth 24 m), the heat storage reported by Stanhill (1969) must reduce summer evaporation by about 7·5 cm per month.

To estimate heat storage in water as a variable in the Penman formula, temperature must be recorded as a function of depth, but when the surface temperature alone has been measured, evaporation can usually be found directly from the product of a vertical gradient of specific humidity and an appropriate wind function. The same procedure is employed to estimate evaporation from the sea where the local heat balance is affected by the disposition of currents as well as by storage.

The apparently simple nature of water surfaces has encouraged climatologists to adopt the loss of water from containers of various shapes and sizes as standard estimates of evaporation; and for over a century many unproductive attempts have been made to determine the 'best' design of pans and tanks (cf. Shaw 1885; WMO 1966).

Evaporimeters can be used to estimate the rate of water loss from nearby lakes or from well watered crops, provided they have been calibrated to establish an appropriate 'pan factor' – the ratio of lake (or crop) evaporation to pan evaporation over the same period. This empirical procedure works well enough for many practical purposes when the pan factor is almost constant or has a systematic annual trend more or less insensitive to differences of weather from year to year. However, there is little point in trying to use the loss of water from a pan to estimate evaporation from a region where vegetation is periodically short of water. As the actual rate of evaporation from the area decreases, the atmosphere becomes drier and pan evaporation increases. This type of negative correlation is a major disadvantage of all pan and tank measurements and was well displayed by the measurements reported by Richards (1979) for the dry summer of 1976 (Fig. 4). Perhaps this is why Buchan (1867) wrote that 'There is no class of observation which shows such diversity, we may almost say contrariety of results, as those made by different observers on evaporation.'

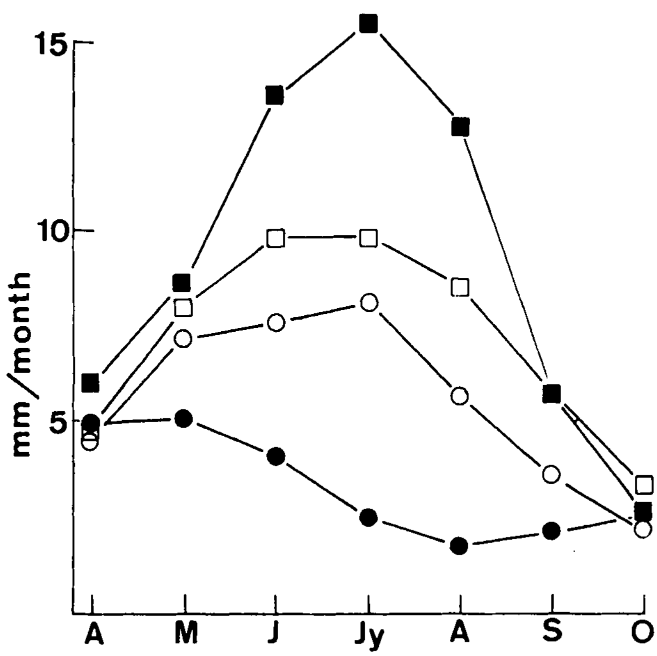

Figure 4. Monthly evaporation rates at Cardington (from Richards 1979)

	open water	grass
1976	■	●
1970–1975	□	○

It is time this comment was taken more seriously. In parts of the world, such as Europe, where climatological records are well maintained and there is a good network of radiation stations, the daily measurement of evaporation from pans and tanks is a time-wasting anachronism, made redundant by the versatility of Penman-type formulae for estimating water loss from specific types of vegetation or from a whole region.

(*b*) *Bare soil: progressive drying*

When bare soil is thoroughly wetted, the soil surface behaves like water in so far as the relative humidity of air in contact with the surface is 100%. The rate of evaporation can therefore be calculated from Eq. (11), and as a matter of observation, is usually very close to the rate for adjacent short vegetation, despite differences in radiative and aerodynamic properties.

Loss of water from the soil surface establishes a gradient of water potential (as measured by the free energy of the liquid) which drives water towards the surface from deeper, wetter layers. (There is a concurrent flux of vapour determined mainly by the direction of the temperature gradient and initially negligible.) This process cannot continue indefinitely because the conductivity of soil for water decreases very rapidly as it dries and it is usually only a few days before the rate of evaporation becomes limited by the upward diffusion of liquid water towards the surface. In this (the so-called 'second stage' of evaporation) equilibrium between the flux of water in the soil and the flux of vapour in the atmosphere is achieved by a progressive downward adjustment of the relative humidity h at the surface.

In theoretical treatments (e.g. Rose 1968), it is usually assumed that h drops instantaneously from its initial value of unity during the first stage of drying to a fixed value of h_0 at the onset of the second stage where h_0 is determined by atmospheric factors. The theory predicts that the subsequent rate of evaporation is inversely proportional to the square root of elapsed time. Although this prediction conforms with the observed behaviour of a drying soil, it is not consistent with meteorological principles as expressed in Eq. (14) where a constant value of h implies a *constant* evaporation rate. This inconsistency is resolved in the analysis which follows, by abandoning the details of water movement in the soil in favour of a somewhat crude model of drying.

We shall assume that the evaporation of water takes place from wet soil below a dry soil layer of increasing thickness, treated as isothermal. If the layer has a resistance r_s to the diffusion of water vapour

$$E = \bar{\rho}\{q_s(T)-q\}/r_s \quad . \quad . \quad . \quad (21)$$

The depth of the dry layer will depend on the total amount of water lost from the soil after second-stage drying begins and the dependence is assumed to have the form

$$r_s = m\int_0^t E\,dt \quad . \quad . \quad . \quad . \quad (22)$$

where the constant m is implicitly a function of the liquid and gaseous diffusivities of soil. It is convenient to express the rate of evaporation as a fraction of the evaporation rate when $r_s = 0$ initially and from Eqs. (11) and (16) it follows that

$$\frac{E}{E_0} = \frac{\Delta'+\gamma}{\Delta'+\gamma(1+r_s/r_H)} \quad . \quad . \quad . \quad (23)$$

where r_H is the aerodynamic boundary layer resistance for heat transfer between the soil surface and air at a reference height at which temperature and humidity are assumed to be independent of the state of the soil surface. (Screen height is often chosen, somewhat arbitrarily, in calculations of this type.) Then from Eqs. (21), (22) and (23),

$$E\{\int E\,dt+A/E_0\} = A \quad . \quad . \quad . \quad (24)$$

where

$$A = r_H(\Delta'+\gamma)E_0/\gamma m \quad . \quad . \quad . \quad (25)$$

The solution of Eq. (24) satisfying the conditions $E = E_0$, $\int E\,dt = 0$ when $t = 0$, is

$$E = \{2t/A+1/E_0^2\}^{-\frac{1}{2}} \quad . \quad . \quad . \quad (26)$$

so that

$$E/E_0 = \{2tE_0^2/A+1\}^{-\frac{1}{2}} \quad . \quad . \quad . \quad (26a)$$

and $$\int E\,dt = \{2At + A^2/E_0^2\}^{\frac{1}{2}} - A/E_0 \quad . \quad . \quad . \quad (27)$$

Equations (26) and (27) are consistent with measurements of evaporation from drying soil and have the merit that when $t = 0$, E is not infinite as in the standard soil physics treatment but has a constant value E_0 determined by the state of the atmosphere.

This simple analysis takes no account of diurnal changes of E_0 and h_0 or of temperature gradients in the dry soil layer which substantially modify the effective value of r_s as estimated by Fuchs and Tanner (1967) for example. Van Bavel and Hillel (1976) outlined a much more detailed and exact analysis of soil evaporation in specified weather, based on a computer simulation. Figure 5 shows the values of evaporation rate and surface temperature amplitude which they estimated for a loam soil exposed to hot dry weather (Phoenix, Arizona) and to a cooler more humid climate (Binghampton, New York). In the second stage of drying, the rate of evaporation was nearly proportional to $t^{-\frac{1}{2}}$ but for the same value of t was faster at Phoenix (consistent with the higher value of E_0 and therefore of A).

As attempts are now being made to estimate soil water content by remote sensing of surface temperature from aircraft or from satellites, it is relevant to note that the amplitude in Fig. 5 appears to increase abruptly at the onset of second-stage drying and is not otherwise a sensitive index of moisture content. In the temperate climate case it is even doubtful whether the small change in amplitude could be detected radiometrically.

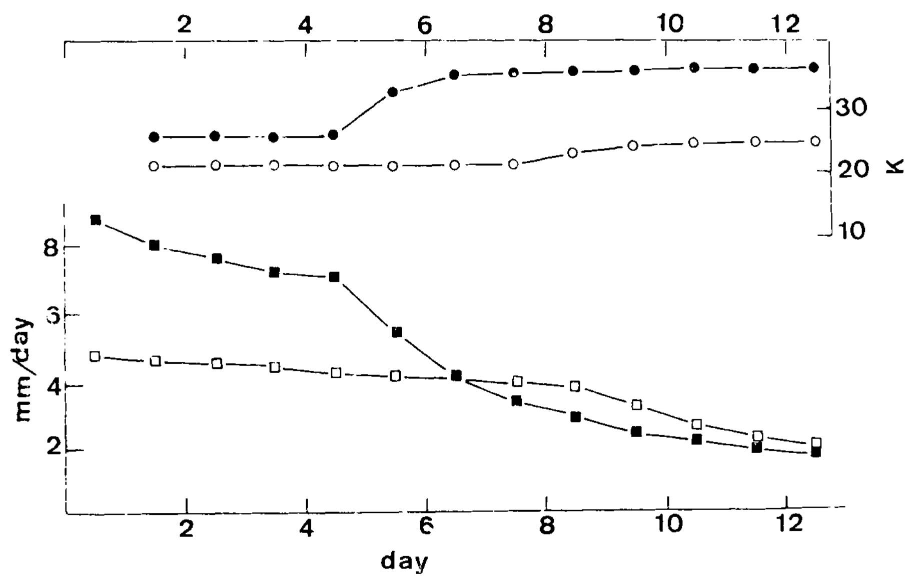

Figure 5. Evaporation from a loam soil and surface temperature amplitude calculated for a hot dry climate (full points) and for a cooler, more humid climate (open points). The soil is assumed to have a depth of 1·1 m, saturated throughout on day zero (from van Bavel and Hillel 1976).

(c) *Leaves: the significance of stomata*

The leaves of most land plants consist of cells supplied with water by an elaborate plumbing system and assembled in such a way that evaporation from cell walls keeps the air spaces between them almost saturated with vapour even when the leaf is transpiring. Transpiration is the loss of water through stomatal pores whose opening and closing is controlled by the hydraulic behaviour of guard cells acting as valves. In response to light, stomata open and water vapour escapes from within the leaf, but the degree of opening is reduced when the supply of water from the soil reservoir is restricted or when the temperature of a leaf is too high or low. Attention has recently been focused on the tendency of stomata to close when the demand for transpiration is increased, for example, by exposing the leaf to drier air. This is commonly known as the 'saturation deficit' effect but there is no reason to suppose that stomatal cells respond to the humidity of the atmosphere *per se*. When stomata are closed, leaves continue to lose water much more slowly through a thin waxy cuticle.

For many species the spacing between stomata is of the order of 0·1 mm whereas the boundary layer attached to a leaf surface, whether laminar or turbulent, usually has a thickness of several millimetres. The resistances of individual pores therefore behave as if they were wired in parallel with each other and with the cuticle which they perforate and this compound physiological resistance of the leaf surface r_s can be treated as if it were placed in series with the boundary layer resistance r_a (Cowan 1972). Ignoring small differences between the boundary layer resistances for heat and water vapour, it follows that with $r_H \simeq r_a$, $r_v \simeq r_a + r_s$, we have $\gamma^* = \gamma(1 + r_s/r_a)$.

For many species the minimum value of r_s for each leaf surface containing stomata is of the order of 100 s m^{-1} (Körner, Scheel and Bauer 1979) whereas cuticular resistances are between 2000 and 4000 s m^{-1}. Although the resistance of the cuticle can often be neglected because it is so much larger than the stomatal resistance, the layer of cuticular wax can be damaged by wind in such a way that evaporation through the cuticle becomes a significant fraction of total water loss (Grace 1977). Loss through the cuticle may also be significant in hot, dry environments during the night, or during the day when stomata are closed by water stress.

At wind speeds commonly encountered in the field, say from 1 to 3 m s^{-1}, r_a is usually less than the equivalent resistance for a smooth flat plate, presumably because of the effects of surface roughness, peripheral irregularities and fluttering (Monteith 1980). In the field, moreover, it is extremely difficult to quantify the effects of free convection at low wind speeds and of turbulence in the wake of neighbouring leaves. A boundary layer resistance of 40 s m^{-1} will be used in the calculations which follow, appropriate for one side of a leaf with a characteristic dimension of 5 cm exposed to a windspeed of 0·5 m s^{-1}.

Figure 6 shows the thermodynamic and aerodynamic relations for a leaf assuming that the stomatal resistance (of each surface) has values of 100, 200 and 400 s m^{-1} and that short-wave irradiances are 600, 300 and 150 W m^{-2}. When the reflection coefficient of the leaf is 0·25 and net long-wave loss is 100 W m^{-2}, the corresponding values of net radiation flux are 350, 125 and 12 W m^{-2}. The straight bold lines on the Fig. 5 define the heat balance for different values of S (or R_n). The dashed lines define the aerodynamic relation when air temperature is 20°C and specific humidity is 11 g kg^{-1} ($\sim$15 mb).

The graph shows that when S = 600 W m^{-2} (about $\frac{2}{3}$ of the maximum June value in Britain) the leaf would be 5·8 K hotter than the surrounding air in the absence of transpiration (point X); but with stomata wide open (r_s = 100 s m^{-1}), the excess is restricted to 2·7 K (Y). In dull light, the leaf would be 1·4 K cooler than the air if the stomata stayed open (W). In practice, stomata would tend to close in dull light in such a way that $1/r_s$ was approximately proportional to the irradiance. So a resistance of 100 s m^{-1} at S = 600 W m^{-2} would increase to about 400 s m^{-1} at S = 150 W m^{-2} and the corresponding cooling by transpiration would be 0·4 K (Z).

The process of transpiration and the tendency of stomata to open in response to light both have the effect of stabilizing the difference between leaf and air temperature. For temperate plants there is no clear advantage in this behaviour because leaf temperature is often below the optimum for growth. In tropical climates transpiration is an effective but

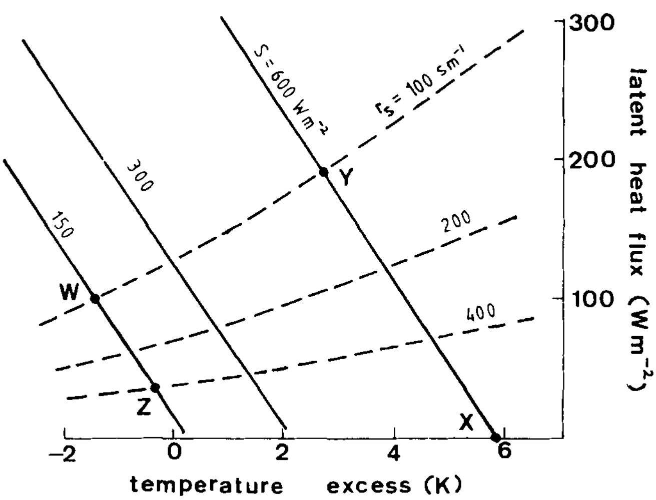

Figure 6. Loss of latent heat from a leaf as a function of the difference between leaf and air temperature. Full lines represent the heat balance equation for different values of solar irradiance S; broken lines represent the aerodynamic equation for different values of the stomatal resistance r_s. The x axis corresponds to $r_s = \infty$, i.e. no evaporative cooling. For interpretation, see text.

costly method of keeping leaves cool in an environment where rainfall is erratic, and stomata may operate in a much more complex way to optimize the amount of carbon dioxide assimilated per unit of water lost (Cowan and Farquhar 1977).

The values of leaf temperature and transpiration rate determined by Fig. 6 could be estimated directly from Eqs. (11) and (13). Because the value of Δ' in these equations increases rapidly with temperature, the excess temperature of a leaf will decrease as air temperature increases provided all the other variables in the system stay unchanged. Linacre (1964) suggested that when air temperature is below 32 °C, leaves tend to be hotter than the surrounding air and *vice versa*. Priestley and Taylor (1972) pointed out that this observation was consistent with the concept of an equilibrium evaporation rate of $1{\cdot}26\,\Delta H/\lambda(\Delta+\gamma)$ (see 4(*h*)) and Priestley (1966) cited climatological records to support the hypothesis that in a region with adequate water for evaporation, air temperature is limited to a maximum of about 32 °C. This hypothesis has not been disproved but there is abundant evidence from the literature that there is a wide range of air temperatures over which leaves may be a few degrees warmer or cooler than the air passing over them. Large temperature gradients also exist across leaves as a consequence of local differences in transfer coefficients (see Monteith 1975, for examples).

(*d*) *Vegetation: complexity in the canopy*

The exchange of heat and water vapour between a stand of vegetation and the atmosphere is a much more complex process than the corresponding exchange at the surfaces of individual leaves. Nevertheless, the same basic principles are valid: for every element of foliage throughout the canopy, fluxes of sensible and latent heat are limited by different resistances r_H and r_V but originate from surfaces which have the same temperature. Moreover, the distribution of temperature throughout the canopy adjusts to an equilibrium value at which the total loss of sensible and latent heat balances the net heat input. Equations (11) and (13) are therefore applicable with a surface resistance r_c for the whole canopy defined as $r_V - r_H$. Assuming similarity of heat and momentum transfer, the value of r_H in near-neutral conditions can be found from

$$r_H = \int_{z_0}^{z-d} dz/[K_M(z)] = u(z)/u_*^2 \quad . \quad . \quad (28)$$

where $K_M(z)$ is a momentum transfer coefficient at height z and other symbols have their usual significance. The formula can be extended to account for buoyancy (Heilman and Kanemasu 1976). Within a canopy, however, there are likely to be systematic vertical differences in the distribution of sources and sinks for heat, water vapour and momentum. It is therefore not evident *a priori* whether r_c can be regarded as a physiological resistance depending mainly on stomatal components or whether it contains a significant aerodynamic element.

The behaviour of r_c has been examined experimentally and theoretically. In the field, the diurnal and seasonal behaviour of r_c has been determined by measuring the rate of evaporation from vegetation by some independent method. Working with barley, Monteith, Szeicz and Waggoner (1965) found that r_c was independent of windspeed (implying a negligible aerodynamic component) and was close to a value estimated for all the component leaves treated as parallel resistors i.e. r_s/L where L is a leaf area index and r_s is a mean stomatal resistance. Similar agreement has now been demonstrated for a range of crops, e.g. for sorghum (Szeicz, van Bavel and Takami 1973), for beans (Black, Tanner and Gardner 1970) for sugar beet (Brown and Rosenberg 1977), and for Douglas Fir (Tan and Black 1976).

Further circumstantial evidence for the physiological significance of r_c is provided by its diurnal and seasonal behaviour. Like stomatal resistance, r_c usually increases during the day, particularly when the soil is dry, and increases with the average age of leaves. Figure 7 is a particularly good example of the consistent diurnal and seasonal behaviour of r_c in a pine forest.

The theoretical justification for regarding r_c as a bulk stomatal resistance was examined analytically by Cowan (1968) and Thom (1975). They drew attention to the fact that the appropriate value of r_H to use in Eq. (11) is not u/u_* but $u/u_*^2 + A/u_*$ where A is a constant with a value of about 4 for arable crops. The additional term is an empirical correction for the fact that the bluff body forces associated with part of the momentum transfer to leaves have no analogue in heat (or vapour) transfer. When r_c is calculated from Eq. (11) without including the correction, the consequent error may be positive or negative but does not usually exceed the error arising from the measurement of variables in the equation. The very detailed analysis described by Shuttleworth (1976) further substantiates the validity of identifying r_c as a physiological resistance, except in the special case where a canopy is partly wetted by rain.

The distinction between aerodynamic and physiological resistances has proved to be particularly useful in predicting the loss of water from foliage fully wetted by rain. From Eq. (11), the rate of evaporation from 'wet' foliage expressed as a multiple of the transpiration rate from 'dry' foliage in the same weather is

$$\frac{\Delta' + \gamma(1 + r_c/r_a)}{\Delta' + \gamma} = 1 + \left(\frac{\gamma r_c/r_a}{\Delta' + \gamma}\right) \quad . \quad . \quad . \quad (29)$$

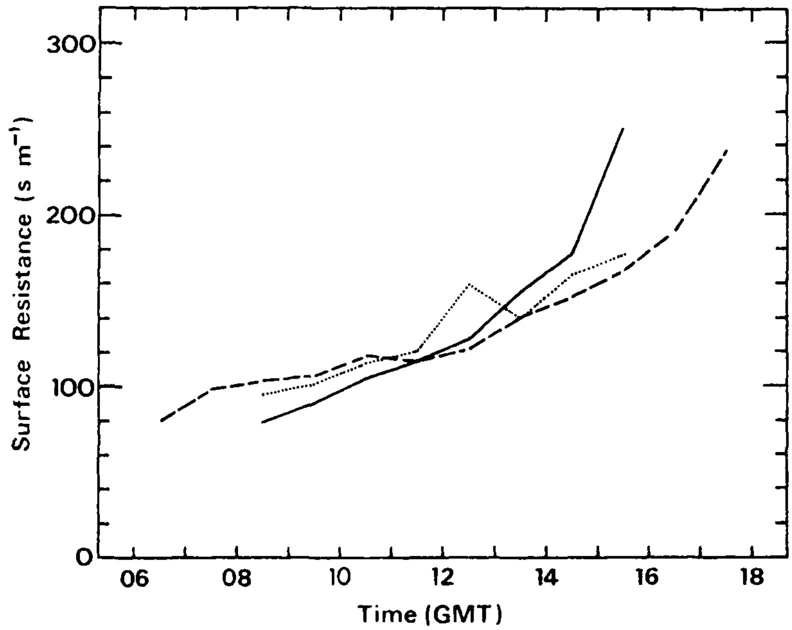

Figure 7. Diurnal variation of median surface resistance of a pine forest at Thetford, Norfolk, 1972–1974.
.... March (9 days)
------ June (28 days)
——— September/October (9 days)
(from Gash and Stewart 1975).

For a well-watered arable crop with $r_c \simeq r_a = 50\,\mathrm{s\,m^{-1}}$, say, and with $\Delta' = 9{\cdot}0 \times 10^{-4}\,\mathrm{K^{-1}}$ at 20 °C, the term in square brackets is about 0·3. Since arable crops intercept relatively little rain, the increased rate of evaporation is barely significant in terms of seasonal water loss. For a coniferous forest, however, the roughness of the surface ensures that r_a is small, often less than $10\,\mathrm{s\,m^{-1}}$ whereas the minimum value of r_c is somewhat larger than for an arable crop, say $100\,\mathrm{s\,m^{-1}}$. On this basis, evaporation from wet foliage should proceed at least four times faster than from the same foliage when dry.

About 15 years ago I suggested that 'the rapid loss of intercepted rain after a shower may be an important factor in the water economy of trees growing in a dry region and in the yield of water from forested catchments' (Monteith 1965). This assertion was inconsistent with much of the evidence reviewed by Penman (1963), but Rutter (1965) demonstrated that the evaporation of water from a wetted pine branch was much faster than transpiration from the same branch when it was dry.

Work at the Institute of Hydrology has now demonstrated the importance of evaporation from wet forests on a catchment scale. Clarke and Newson (1978) compared annual evaporation (as rainfall minus runoff) from catchments, mainly under hill pasture, at the head of the River Wye, and from neighbouring catchments, mainly under coniferous forest at the head of the Severn. For pasture, the mean evaporation from 1974 to 1976 was 42 cm and for forest it was 69 cm. (The difference was even greater in the relatively wet year of 1974 than in the two drier years which followed so depth of rooting was not the main cause.) The authors point out that few catchments in Britain have large proportions of forest land at present but if forest areas increase as planned 'the resultant reduction in water yield may then be worthy of consideration'.

During the period when water is evaporating from a wet forest, very large amounts of energy are needed to sustain fluxes of latent heat substantially greater than the net income of radiant energy. The additional energy can be accounted for on a mesoscale by condensation in cloud (Thom and Oliver 1977; Shuttleworth and Calder 1979).

Despite its versatility, Eq. (11) has not been widely adopted in practical hydrology or agronomy and has not found favour with WMO and FAO consultants (see Doorenbos and Pruitt 1975, for example). However, it has recently become a central component of the very detailed evaporation calculations now performed by the Meteorological Office, taking account of differences in vegetation type and soil water regime over the whole of England and Wales. One of the few practical cases which cannot be handled by Eq. (11) is the evaporation from a row crop with incomplete ground cover. This is a particularly complex system for heat and mass transfer but useful empirical schemes have been proposed by Ritchie (1971) and by Tanner and Jury (1976).

(*e*) *Glasshouse cooling: does foliage benefit?*

Cooling by evaporation of water is a recognized method of reducing the temperature of air in buildings. In a recent analysis for glasshouses, Landsberg, White and Thorpe (1979) predicted that a drop of 8 to 12 K could be achieved even in strong sunshine. However, a decrease of dry-bulb temperature by evaporation must always be accompanied by an increase of wet-bulb temperature, a reduction in the transpiration rate, and therefore an increase in the excess of foliage temperature over the temperature of the surrounding air. Equation (13) can be manipulated to show how far the apparent advantage of evaporative cooling in a greenhouse is likely to be offset by the thermal consequences of raising the humidity.

Suppose that air forced into a greenhouse by a fan passes over a set of wet pads which saturate it. At maximum efficiency the process will be adiabatic and the air will be reduced to wet-bulb temperature T'; but if the pads acquire a significant amount of heat by conduction or by radiation from their surroundings the temperature of the air entering the glasshouse will be greater than T'.

In general, if the external air has a temperature T and a specific humidity q, it follows from Eq. (13) that the temperature of the cooled air will be

$$T_i = T - \{q_s(T) - q\}/(\Delta' + \gamma_1^*) \qquad (30)$$

where γ_1^* is a modified psychrometer constant for the pads. The lower limit of $T_i = T'$ is achieved when $\gamma_1^* = \gamma = c_p/\lambda$. The specific humidity of air entering the house is assumed to be $q_s(T_i)$.

Within the house the mean temperature of air will differ from T_i because of the input of heat from radiation and the exchange of heat through the walls. The radiative input is determined by the net radiant flux R absorbed by vegetation and other surfaces and this flux is partitioned between sensible and latent heat transfer to the atmosphere of the house. For the purposes of this analysis, the increase in house temperature ascribed to the sensible heat component will be taken as αR where α is a constant. Then provided the rate of heat transfer through the walls is negligible compared with the rate of heat transfer by ventilation (Case A), the mean air temperature in the house will be $\bar{T} = T_i + \alpha R$. At the other extreme, if heat transfer through the walls is much faster than the transfer by ventilation, the mean temperature will be $\bar{T} = T+\alpha R$ (Case B).

The mean surface temperature of foliage is given by Eq. (13) as

$$\bar{T}_f = \bar{T} + \frac{\gamma_2^* r_H R}{\bar{\rho} c_p(\Delta' + \gamma_2^*)} - \frac{q_s(\bar{T}) - q}{\Delta' + \gamma_2^*} \quad . \quad . \quad (31)$$

where γ_2^* $(= \gamma r_V/r_H)$ is the modified psychrometer constant for the foliage.

In Case A the decrease in foliage temperature as a result of evaporative cooling (δT_f) is the difference between $\bar{T}_f$ when $\bar{T} = T+\alpha R$ and when $\bar{T} = T_i+\alpha R$ or

$$\delta T_f = (T+\alpha R)-(T_i+\alpha R)+\frac{q_s(T+\alpha R)-q}{\Delta'+\gamma_2^*}+\frac{q_s(T_i+\alpha R)-q_s(T_i)}{\Delta'+\gamma_2^*} \quad . \quad (32)$$

Substituting for $T-T_i$ from Eq. (31) gives

$$\delta T_f = (T-T_i)\{1-(\Delta'+\gamma_1^*)/(\Delta'+\gamma_2^*)\} \quad . \quad . \quad (33)$$

where the term in curly brackets may be regarded as a figure of merit for the system. It is clear that the mean temperature of the foliage will be decreased by the action of an evaporative cooling system only when $\gamma_1^* < \gamma_2^*$, i.e. when the value of r_V/r_H for the wet pads is less than the value for foliage.

Suppose, for example, that $r_V = r_H$ for the cooling system (maximum cooling) and that $r_V = 3r_H$ for a freely transpiring crop. Then if Δ' is evaluated at 20°C, $\delta T_f = 0{\cdot}38\,(T-T_i)$. In this case, over 60% of the thermal advantage volume from the cooling system is lost by the concomitant increase of humidity.

Case B is even less favourable for the reduction of leaf temperature: the air temperature in the house is not lowered by evaporation because of rapid heat transfer through the walls but the specific humidity is raised from q to $q_s(T_i)$. It follows that the decrease in leaf temperature by evaporative cooling is

$$\delta T_f = \frac{q_s(T+\alpha R)-q}{\Delta'+\gamma_2^*}+\frac{q_s(T+\alpha R)-q_s(T_i)}{\Delta'+\gamma_2^*}$$

$$= \frac{q-q_s(T_i)}{\Delta'+\gamma_2^*} = -(T-T_i)\left(\frac{\gamma_1^*}{\Delta'+\gamma_2^*}\right) \quad . \quad . \quad (34)$$

The negative sign implies that leaf temperature is always increased by the action of an evaporative cooler in Case B.

Any real system will operate in a regime between Cases A and B and it is likely that foliage temperature is often little changed by evaporative cooling. However, dry surfaces of pots and staging will be cooled and in an extreme climate this could benefit the root system. A more important advantage is likely to be the decrease in transpiration rate which could substantially reduce the water stress on plants in a dry environment and thereby increase their rate of growth. Moreover, the tendency for stomata to open wider in more humid air (see 4(*c*)) may partly offset the thermal disadvantage of a higher wet-bulb temperature.

(*f*) *Man: survival by sweating*

In principle the type of analysis represented by Fig. 1 can be used to determine the mean skin temperature of a man (or any other animal) in a specified environment and the rate at which water evaporates from the skin surface when sweat glands are present and active. However, an extension of the figure is needed to take account of the physiological link between sweat rate and skin temperature.

In most circumstances, metabolism is a major source of heat so that the heat balance equation for an animal in a steady state can be written as

$$M+R = \lambda E+\bar{\rho}c_p(\bar{T}_s-T)/r_H \quad . \quad . \quad . \quad (35)$$

It is convenient to define M as the rate of production of metabolic heat per unit area of body surface reduced by the loss of heat by evaporation in the respiratory tract which is an inevitable consequence of breathing. For man in an atmosphere with a specific humidity of about $6\,\mathrm{g\,kg^{-1}}$ (10 mbar) M is about 92% of the gross rate of heat production (Campbell 1977). $\bar{T}_s$ is the mean temperature of the skin and r_H the resistance for sensible heat transfer from the skin to the ambient air. We shall consider the simplest case of a man without clothing so that $r_H = r_a$ is a boundary layer resistance which can be estimated by treating the body as a cylinder exposed to a uniform airstream (Monteith 1975).

The aerodynamic equation for a naked man is

$$\lambda E = \lambda\bar{\rho}\{q_s(\bar{T})-q\}/(r_a+r_s) \quad . \quad . \quad . \quad (36)$$

assuming that the boundary layer resistances for heat and vapour transfer are the same and putting r_s for the resistance to the diffusion of water vapour in the skin. For adults, this resistance has a maximum value of about $8\,\mathrm{ks\,m^{-1}}$ (Kerslake 1972) when the temperature of the skin is too low to induce sweating and the corresponding value of λE is usually about 10% of the minimum metabolic rate. In principle, the minimum value of r_s is zero, achieved when the whole body surface is covered with sweat.

The physiological link between sweat rate W and T_s can be represented by the empirical relation

$$W = M+h(T_s-T_0), \quad T_s > T_0 \quad . \quad . \quad . \quad (37)$$

where T_0 is the mean skin temperature needed to induce sweating and h has the dimensions of a heat transfer coefficient. Representative values of T_0 and h are 34 °C and $120\,\mathrm{W\,m^{-2}\,K^{-1}}$ (Hatch 1963).

Equations (35) to (38) are general enough to be applied to the whole range of environmental temperatures at which the human body can survive but we are concerned here only with the hotter part of the range in which the regulation of body temperature is achieved by changes in sweat rate. In this regime, the value of M is usually between 100 and $200\,\mathrm{W\,m^{-2}}$ but values up to $750\,\mathrm{W\,m^{-2}}$ have been recorded during strenuous exercise (Durnin and Passmore 1967).

Equations (35) to (37) can now be solved on a heat balance diagram. In Fig. 8(a) the bold line represents Eq. (35) and the two dashed lines correspond to Eq. (36) with maximum and minimum values of r_s. The dotted line represents Eq. (37). The point X is a solution which satisfies the equations of heat balance and of sweat production. In this state the rate at which sweat is excreted by the sweat glands is exactly matched by the rate at which it evaporates so that there is no accumulation of liquid sweat on the skin surface. The value of r_s can be determined from Eq. (36).

In Fig. 8(b) corresponding to a faster metabolic rate, the point X lies above the line corresponding to evaporation from a body completely covered with sweat and in this case it would be necessary for the body surface temperature to rise to the point Y where the heat balance and aerodynamic equations are satisfied simultaneously. An excess of sweat would therefore be produced (corresponding to ZY) and would collect on the skin surface before dribbling off.

When the body is effectively a wet cylinder with $r_s = 0$ and $\gamma^* = \gamma$, the surface tempera-

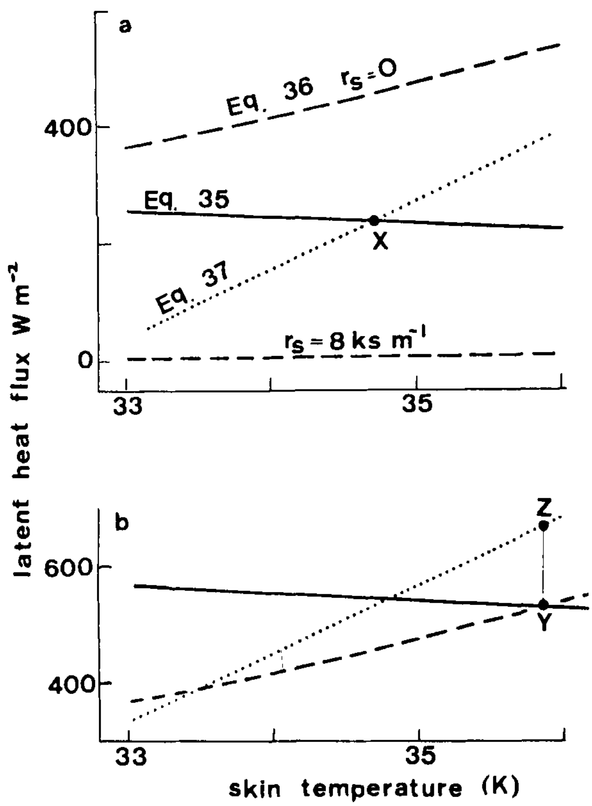

Figure 8. Relationships between latent heat loss by the evaporation of sweat and skin surface temperature for a naked man. Assumed conditions are:

(a) T(air) = 30 °C, q = 19 g kg^{-1}, r_H = 100 s m^{-1}, M = 150 W m^{-2}, R_n = 150 W m^{-2}. The three equations are satisfied at the point X for which E = 240 W m^{-2}, T_s = 34·7 °C, r_s = 90 s m^{-1}.

(b) M = 450 W m^{-2}. Eqs. (35) and (36) are satisfied at Y where E = 530 W m^{-2}, T_s = 35·7 °C, r_s = 0. Excess sweat (ZY) is equivalent to 170 g m^{-2} h^{-1}.

ture is given by Eq. (13) which can be written in the form

$$\bar{T}_s = T' + \gamma r_H(R+M)/\{\rho c_p(\Delta' + \gamma)\} \quad . \quad . \quad (38)$$

Wet-bulb temperature is therefore an appropriate index of heat stress in an environment where the second term in Eq. (39) can be treated as constant. This point was discussed by Brunt (1943) in a previous Presidential Address.

The fastest rate at which sweat production can be sustained is about 1 to 1½ litres per hour and in an environment where this could evaporate completely, the loss of latent heat would be about $500\,W\,m^{-2}$. The loss of sensible heat in a wind speed of $2\,m\,s^{-1}$ is about $12\,W\,m^{-2}$ per degree difference in temperature between the skin surface and the air. Evaporation of sweat therefore enables man to engage in heavy work in a cool environment with little radiative load (e.g. $M = 500$, $R = 200$, $\lambda E = 500$, $C = 200\,W\,m^{-2}$); or simply to survive in a hot environment with strong sunshine (e.g. $M = 150$, $R = 400$, $\lambda E = 500$, $C = 50\,W\,m^{-2}$). In the tropics, and particularly in the humid tropics, heavy work cannot be sustained for long periods out of doors and thermodynamic restrictions on human activity have an important bearing on patterns of crop husbandry in the absence of mechanization.

The simple analysis outlined above can be extended to clothed subjects or to hairy animals. It will always be unrealistic, however, to assume that sweat rate and temperature are uniform over the whole body surface. The more difficult case of a non-uniform surface has been examined by Kerslake (1972).

(g) *Regional evaporation: (i) appropriate parametrization*

We have considered applications of the Penman formula to a number of discrete and well-defined systems. A much more difficult problem is the determination of sensible and latent heat fluxes over an inhomogeneous region as might be needed, for example, by a hydrologist concerned with the water balance of a catchment or by a meteorologist modelling the behaviour of the planetary boundary layer.

When the original formula was used to estimate annual evaporation over the British Isles (Penman 1950), figures for regions roughly the size of counties agreed well with catchment records of rainfall and runoff. In hindsight, the agreement was partly fortuitous because evaporation must have been overestimated during drought and underestimated in rainy weather (Thom and Oliver 1977). Moreover, the empirical radiation equations used in the formula were not valid over the whole country (Monteith 1966).

In atmospheric modelling the main problem has been to specify the effective wetness of different types of surface and three forms of parametrization have been attempted.

Relative humidity. The evaporation equation which Gadd and Keers (1970) used in a 10-level model of the atmosphere is formally identical to Eq. (11) with a value of $h = 0{\cdot}8$ assumed for land surfaces. A fixed relative humidity is appropriate for bare soil when the water content of the surface layer is constant or for a large area of very salt water such as Lake Eyre (Penman 1955). It is not appropriate for a drying soil and should not be used for vegetation because a comparison of Eqs. (11) and (14) shows that h must be a function both of weather and of the surface resistance.

Relative evaporation. If the rate of potential evaporation E_0 is estimated for a surface by assuming $h = 1$ and $r_V = r_H$, the actual rate can be expressed as $E = fE_0$ where f is a time-dependent factor. In Penman's original analysis, E_0 was determined for water in a tank and the value of f for turf with a similar exposure ranged from 0·6 in winter to 0·8 in summer, provided transpiration was not restricted by the availability of water to roots.

Later work showed that f is almost constant when the soil water deficit d is less than a critical value d_1. When $d > d_1$, f decreases approximately linearly with increasing d, reaching zero at a limit d_2 (Ritchie, Burnett and Henderson 1972). Priestley and Taylor (1972) drew attention to the usefulness and limitations of this way of summarizing the behaviour of vegetation subjected to water stress but there has been a tendency to disregard the extent to which d_1 and d_2 depend on the weather, soil physical characteristics, and type of vegetation.

This form of parametrization has been used in the 11-layer model of the general circulation developed in the Meteorological Office (Saker 1975). Although it is appropriate for vegetation, it does not describe the behaviour of bare soil for which the relationship discussed in section 4(*b*) implies that when $d > d_1$, f is inversely proportional to $d - d_1$. Moreover, the value of d_1 for a given type of soil will usually be an order of magnitude less than the value for a stand of vegetation growing in the same soil, typically about 1 cm rather than 10 cm.

Surface resistance. Fixed values of surface resistance have recently been incorporated in meso-scale models of the atmosphere as developed, for example, by Carpenter (1977), who found that the partitioning of radiant energy at the ground was extremely sensitive to the value of this parameter. In the Meteorological Office Rainfall and Evaporation Calculation System, thankfully abbreviated to MORECS (Wales-Smith and Arnott 1979), a maximum evaporation rate for different types of vegetation is calculated by adopting appropriate minimum values of surface resistance which change seasonally. When the soil water deficit exceeds a preset limit, the evaporation rate is multiplied by a factor f depending on a soil water deficit d as already described. Values of evaporation predicted by MORECS are consistent with measurements of changes in soil water content with a neutron probe but the scheme is a rather inelegant hybrid and could possibly be improved by incorporating direct relationship between surface resistance and soil water deficit or water potential as described by Russell (1980) and Federer (1979) for example. However, because of the very complex nature of water flow through plants and of the response of stomata to a restricted supply of water, the specification of surface wetness for atmospheric models is not likely to improve rapidly and there is no immediate hope of replacing the rather arbitrary empirical expressions in current use by more fundamental relationships. I therefore see little point in constructing physically elaborate but morphologically unrealistic models of vegetation as proposed by Deardorff (1978).

(*h*) *Regional evaporation: (ii) the concept of equilibrium*

Because surface resistance remains a diagnostic rather than a prognostic parameter, other methods have been sought for estimating regional evaporation from climatological records. One ingenious scheme has evolved from the proposal by Bouchet (1963) that actual and potential transpiration be treated as complementary quantities in the following sense. Over a region with dimensions of the order of 10 to 100 km, suppose that the rate of actual evaporation E is less than the maximum rate E_0 which would be obtained in the same weather if water were freely available over the whole region. Then putting $E = E_0 - \delta E$, the quantity E is proportional to the additional amount of energy which is available for heating the air so that the saturation deficit will increase with the size of δE, other factors being equal. Suppose this drier air now passes over some surface within the region which does have an unlimited supply of water for evaporation. Bouchet postulated that the rate of evaporation from the surface would exceed E_0 by δE so that $E_0 + \delta E = E^*$ say. In practice, the relationship between E, E_0 and E^* must depend on many factors such as the relative value of surface resistances for 'wet' and 'dry' surfaces and the size of the 'wet' area within the region.

Eliminating δE from the two relationships in the last paragraph gives the rate of actual evaporation as

$$E = 2E_0 - E^* \quad . \quad . \quad . \quad (39)$$

so that

$$E_0 = (E + E^*)/2$$

The derivation of this equation appears to be based on intuition but it finds support in measurements of the type plotted in Fig. 4 where, in the exceptionally dry summer of 1976, the mean of the evaporation rate from grass (E) and from a tank ($?E^*$) is approximately equal to the average value of $(E+E^*)/2$ in the preceding 6 years.

Another strand now enters the argument in the specification of an 'equilibrium' rate of evaporation. According to McNaughton (1976) (see section 3), air passing over a region of uniform wetness as specified by a constant surface resistance, achieves equilibrium with the surface when the gradient of saturation deficit vanishes so that the evaporation rate becomes $\Delta H/\lambda(\Delta+\gamma)$. The evidence reviewed by Priestley and Taylor (1972) suggests that the evaporation from oceans, from bare soil, and from vegetation is often about 20 to 30% more than this value. They therefore expressed the equilibrium evaporation rate as

$$E = \alpha\Delta H/\lambda(\Delta+\gamma) \quad . \quad . \quad . \quad (40)$$

and settled on a 'best' value of $\alpha = 1{\cdot}26$.

A whole series of papers published in the past decade reports measurements of evaporation from wet or well-watered surfaces consistent with $\alpha = 1{\cdot}26$ to within a few percent. The degree of consistency is surprisingly good for daily measurements of evaporation (Davies and Allen 1973; Stewart and Rouse 1977) and is even more remarkable for hourly measurements (Rouse and Stewart 1972). Unfortunately, these experimental tests of the Priestley-Taylor formula have not been matched by any theoretical explanation of why α should be 1·26 (or thereabouts) rather than 1·00 as predicted for the behaviour of a planetary boundary layer, treated as a closed system for heat and vapour exchange apart from the input of radiant energy at the ground. The absorption of radiant energy by aerosol, as estimated by Glazier, Monteith and Unsworth (1976) for anticyclonic weather in the English Midlands, could account for values of α between 1·1 and 1·3; and the entrainment of dry air through an inversion may also be implicated. However, there seems no reason why either of these processes should make α *constant*.

An even more serious objection to the general application of the Priestley–Taylor formula is the fact that it takes no account of the aerodynamic properties and physiological behaviour of the surface. Shuttleworth and Calder (1979) strongly emphasized this point in a paper reviewing measurements of evaporation from two pine forests at sites in eastern England (Thetford) and Wales (Plynlimon). They distinguished days when the foliage was wetted by rain from 'dry' days. Averaged over several years, the mean values of α for the wet days were 1·50 (Thetford) and 9·69 (Plynlimon); corresponding 'dry' day values were 0·62 and 0·74. The authors concluded by stating their belief 'that evapotranspiration in general and forest evapotranspiration in particular are processes subject to physical and physiological control; and that the laws of physics are not influenced by human desire for computational simplicity'.

Coniferous forest may be one of the exceptions which proves the Priestley–Taylor rule, but I believe there are likely to be many unpublished sets of observations in which α has not conformed to a narrow range of values close to 1·26. It would certainly be premature – and could be a retrograde step – to regard the observed value of α divided by 1·26 as an 'index of aridity' (Priestley and Taylor, loc. cit; Thompson, 1975).

Returning to Bouchet's attempt to relate actual to potential evaporation, Morton (1975) suggested that E_0, the rate of evaporation from an extensive well-watered surface could be identified as $\alpha\Delta H/\lambda(\Delta+\gamma)$. He proceeded to develop a somewhat cumbersome method of estimating E^* from a function which included an arbitrary transfer coefficient independent of wind speed and an advective energy term both to be determined by 'calibration' against measurements from a catchment. A simple and more rational system was proposed by Brutsaert and Strickler (1979) who suggested that E^* can be determined from the equation for open-water evaporation, i.e. Eq. (11) with $\gamma^* = \gamma$. The rate of actual evaporation is then given by

$$\lambda E = \frac{(2\alpha-1)\Delta' H - \bar{\rho}c_p\{q_s(T)-q\}/r_H}{\Delta+\gamma} \qquad (41)$$

Rates of daily evaporation estimated by this formula agreed with the loss of water from a rural catchment in eastern Holland during the dry summer of 1976 and critical tests are needed in other environments. The scheme suggested by Brutsaert and Strickler has the merit of elegance and simplicity but its foundations need strengthening. Apart from the uncertainty which surrounds the value of α and its physical significance, Bouchet's hypothesis of complementarity between actual and potential rates of evaporation needs to be substantiated by an appropriate model of the planetary boundary layer. In developing such a model, it would be essential to take account of the system of positive feed-back by which dry air may cause partial closure of stomata and a reduction in the input of water vapour from vegetation to the atmosphere. The effects of stomatal closure in dry air have been demonstrated by Calder (1978) but it is not yet clear how the physiological link between r_s and humidity can be reconciled with the physical link represented by Eq. (18), for example.

5. Envoi

The evaporation of water and the associated diffusion of water vapour and heat are processes of profound terrestrial importance. According to estimates by Budyko (1974) and others, latent heat accounts for about 90% of the annual heat transfer from the ocean to the atmosphere and 50% of the transfer from land. Atmospheric cycles of evaporation and condensation are intimately linked with the hydrological cycles of rainfall and runoff and with the global distribution of food crops and forests. The transport of latent heat is equally important for larger animals, including man, which cannot survive unless the temperature of the body core is held within a narrow range. When air temperature is within or above this range, sweating, panting and similar processes are essential for dissipating energy from metabolism and from the interception of radiation. Man also exploits the physics of evaporation in diverse activities from simple household operations like drying clothes and cooking to the generation and transmission of power and many industrial processes.

In all these systems (climatological, hydrological, ecological, domestic and industrial) one of the central thermodynamic problems is the partitioning of available energy between sensible and latent heat. In the last analysis, partitioning depends on the ratio of Δ to γ which is determined by the physical properties of water as liquid and as vapour. I believe that the nature and distribution of life on our planet depends just as much on the value of Δ/γ as on the magnitude of the Solar Constant and other prestigious parameters of geophysics. The study of natural evaporation is indeed one of the most necessary Ingredients of a real and Philosophical Meteorology.

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