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EVAPORATION AND ENVIRONMENT

BY J. L. MONTEITH

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A turgid leaf exposed to bright sunshine can transpire an amount of water several times its own weight during a summer day. Rapid evaporation is sustained by a supply of heat from the atmosphere and by a movement of water within the plant preventing the desiccation of leaf tissue. In analysis, the need for energy and the need for water have often been dissociated. Meteorologists investigating the energetics of transpiration have assumed that leaves behave like pieces of wet, green blotting paper, and plant physiologists have demonstrated mechanisms for the conduction of water at arbitrary rates unrelated to the physics of the environment.

This paper describes progress towards a reconciliation of parallel concepts in meteorology and physiology. The path for the diffusion of water vapour from leaf cells to the free atmosphere is divided into two parts, one determined primarily by the size and distribution of stomata, and the other by wind speed and the aerodynamic properties of the plant surface. Diffusive resistances for single leaves and for plant communities are established from measurements in the laboratory and in the field and are then used:

- (i) to predict relative rates of evaporation from leaves with wet and dry surfaces;
- (ii) to investigate the dependence of transpiration rate on wind speed and surface roughness;
- (iii) to demonstrate that the relation between transpiration rate and leaf area is governed by stomatal closure in leaves well shaded from sunlight;
- (iv) to calculate maximum rates of transpiration for different crops and climates.

A final section on the advection of dry air stresses the importance of physiological restraint on the rate of transpiration from an irrigated field surrounded by dry land.

FIRST PRINCIPLES

The evaporation of water is like a commercial transaction in which a wet surface sells water vapour to its environment in exchange for heat. For each gram of water evaporated at 20° C., the surface demands 585 calories of heat and several forms of payment are acceptable. When

water evaporates from plants, the environment can supply heat by solar radiation, by turbulent transfer from the atmosphere, or by conduction from the soil. A few elementary principles of thermodynamics are needed to relate the state of the environment to the rate of evaporation from a free water surface, an isolated leaf, or from a crop canopy.

In problems of natural evaporation, the state of a given mass of air can be described by its temperature T and its vapour pressure e . The total heat content of the air is the sum of a sensible heat content, depending on temperature, and a latent heat content, depending on vapour pressure. When temperature changes by δT and vapour pressure by δe , approximations discussed by Brunt (1939) lead to the relation

$$\frac{\delta e}{\text{change in latent heat}} = \frac{\gamma \delta T}{\text{change in sensible heat}}, \tag{1}$$

where γ , a function of the physical properties of dry air, is $0.66 \text{ mb. } ^\circ \text{C.}^{-1}$ at 20°C. and 1 atm. pressure.

Consider a mass of air isolated from all external sources of heat. Any spontaneous change in latent heat content must be balanced by an equal and opposite change in sensible heat content. In particular, when the air is unsaturated and contains a small amount of liquid water, the water will evaporate (increasing the latent heat content of the air) and the air will cool (decreasing the sensible heat content of the air). The process stops when the air becomes saturated at a temperature T' , known as the wet-bulb temperature, and from equation 1, the equation for the redistribution of energy is

$$e_s(T') - e = \gamma(T - T'), \tag{2}$$

where $e_s(T')$ is the saturation vapour pressure at T' .

Further increase of vapour pressure can be produced only by adding an amount of heat Q that is shared between an increase of sensible heat (increasing T) and an increase of latent heat (increasing e). While liquid water remains in the system, small changes are related by

$$\delta e_s = \Delta \delta T, \tag{3}$$

where Δ , the change of saturation vapour pressure with temperature, increases with temperature. (At 0°C. , $\Delta = 0.5 \text{ mb. } ^\circ \text{C.}^{-1}$; at 25°C. $\Delta = 1.9 \text{ mb. } ^\circ \text{C.}^{-1}$.) Combining equations 1 and 3,

$$\text{increase of latent heat} = \frac{\Delta Q}{\Delta + \gamma}, \tag{4a}$$

$$\text{increase of sensible heat} = \frac{\gamma Q}{\Delta + \gamma}. \tag{4b}$$

The rate of evaporation from a surface can be calculated from the rate of turbulent transfer of air. If air at temperature T is moving over a surface at temperature T' , then 1 cm.^3 of unsaturated air at temperature T receives $\rho c(T - T')$ calories of heat from the surface. If E is the rate of evaporation in grams per cubic centimetres of air per second, then the rate at which air supplies heat is λE .

where D is the wet-bulb depression, $T - T'$, and E_1 is the evaporation rate from a free water surface.

When water evaporates from a surface at temperature T at a rate E , the heat balance is

from equation (4a), where

Fig. 1 shows how the evaporation rate E is determined by combining equations (5) and (6). The initial state of the air is at point W (unsaturated vapour pressure e and temperature T). The final state is at point X (saturated vapour pressure $e_s(T')$ and temperature T').

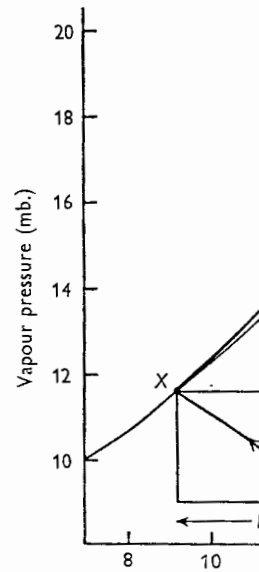


Fig. 1. Graphical representation of the relationship between vapour pressure and temperature.

can supply heat by solar sphere, or by conduction thermodynamics are needed of evaporation from a free anopy.

of a given mass of air can pressure e . The total heat at content, depending on ding on vapour pressure. pressure by δe , approximation

$$\frac{\delta T}{\text{sensible heat}}, \quad (1)$$

dry air, is 0.66 mb. °C.⁻¹

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t T' . oduced only by adding an ncrease of sensible heat ncreasing e). While liquid related by

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ressure with temperature, 0.5 mb. °C.⁻¹; at 25° C. nd 3,

$$\frac{\Delta Q}{\Delta + \gamma}, \quad (4a)$$

$$\frac{\gamma Q}{\Delta + \gamma}, \quad (4b)$$

The rate of evaporation from a wet surface with temperature T' can now be calculated from the rate of increase in the latent heat content of surrounding air at temperature T . If the specific heat of air is ρc (cal. cm.⁻³), then 1 cm.³ of unsaturated air cooling from T to T' °C. will supply $\rho c(T - T')$ calories of heat for evaporation. If r_a is the time in which 1 cm.³ of air exchanges heat with 1 cm.² of surface, $1/r_a$ is a rate of exchange in cubic centimetres of air per square centimetre of surface per second. The rate at which air supplies heat to the surface is

$$\lambda E_1 = \frac{\rho c(T - T')}{r_a} = \frac{\rho c D}{r_a}, \quad (5)$$

where D is the wet-bulb depression, λ is the latent heat of water (cal. g.⁻¹) and E_1 is the evaporation rate (g. cm.⁻² sec.⁻¹).

When water evaporates into *saturated* air, latent heat must be taken up at a rate

$$\lambda E_2 = \frac{\Delta H}{\Delta + \gamma} \quad (6)$$

from equation (4a), where H (cal. cm.⁻² sec.⁻¹) is the *rate* of heat supply.

Fig. 1 shows how the evaporation from a wet surface may be found by combining equations (5) and (6). In this example, air at 13° C. and 9 mb. vapour pressure (point W) passes over a wet surface at 16° C. and 18.2 mb. saturated vapour pressure (point Y). In terms of thermodynamics, the

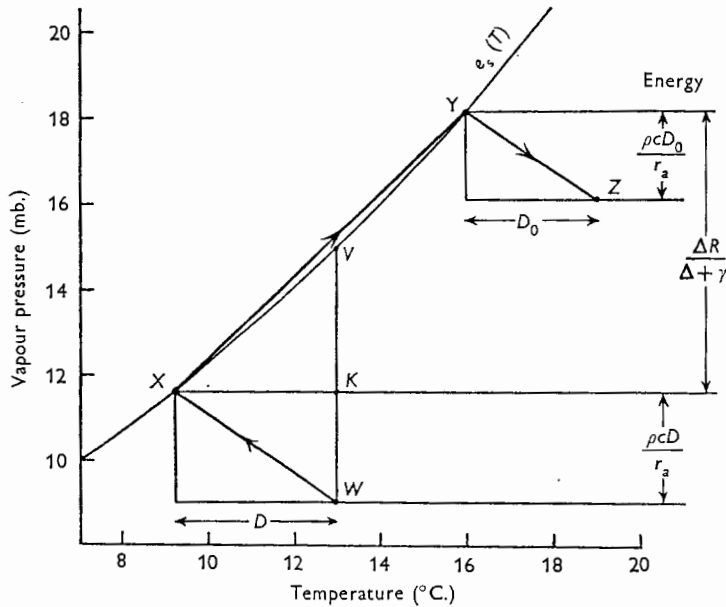


Fig. 1. Graphical representation of evaporation formulae (see text).

amount of energy needed to change air from state W to state Y is completely independent of the path chosen from W to Y and of the infinite possibilities the route WXY is the most revealing. Starting from W , the air is saturated by cooling to its wet-bulb temperature (9.2°C.) along a line WX with slope $-\gamma$ and the heat released for evaporation is λE_1 (equation 5). Maintaining saturation, the air is now heated from state X to state Y where X and Y are both on the saturation vapour pressure curve $e_s(T)$. Approximating to this curve by the chord XY with slope Δ , the corresponding latent heat of evaporation is λE_2 (equation 6). The temperature-dependent function Δ is evaluated at the mean of the wet-bulb temperature of the air and the surface temperature.

The total latent heat of evaporation is

$$\lambda E = \lambda(E_1 + E_2) = \frac{\Delta H}{\Delta + \gamma} + \frac{\rho c D}{r_a}. \quad (7)$$

For air at W , the relation between the wet-bulb depression D and the saturation vapour pressure deficit is given by

$$\frac{e_s(T) - e}{D} = \frac{\gamma W}{KX} = \frac{\gamma K}{KX} + \frac{KW}{KX} = \Delta' + \gamma,$$

where Δ' is evaluated at the mean of wet- and dry-bulb air temperatures. Assuming $\Delta \approx \Delta'$, equation (7) may be written

$$\lambda E = \frac{\Delta H + \rho c \{e_s(T) - e\} / r_a}{\Delta + \gamma}. \quad (8)$$

Equation (8) was first derived by Penman (1948) and equation (7) is a variant preferred by Slatyer & McIlroy (1961). Neither equation is valid for a surface where the vapour pressure is less than the saturation vapour pressure at surface temperature. For example, the water in a leaf evaporates at the surfaces of cell walls surrounding sub-stomatal cavities, and reaches the outer surface of the leaf by molecular diffusion through stomata and through the cuticle. When a leaf has a uniform temperature T_o , the vapour pressure of air in contact with wet cell walls is usually very close to the saturation vapour pressure $e_s(T_o)$. At a dry leaf surface, the vapour pressure e_o is always less than $e_s(T_o)$ during transpiration, and surface air is never saturated. In Fig. 1, the unsaturated air in contact with a leaf is given a wet-bulb depression D_o and its state is represented by point Z . To reach Z from Y , the air is de-saturated by condensation of water vapour and latent heat is released at a rate

$$-\lambda E_3 = \frac{\rho c D_o}{r_a}.$$

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The total rate of evaporation for the path *WZ* is

$$\lambda E = \lambda(E_1 + E_2 + E_3) = \frac{\Delta H}{\Delta + \gamma} + \frac{\rho c(D - D_0)}{r_a} \tag{9}$$

Some theoretical implications of this expression were examined by Slatyer & McIlroy but its practical exploitation is severely limited by the difficulty of measuring or calculating the wet-bulb depression of air at a leaf surface. This difficulty may be resolved by eliminating surface parameters as follows.

Anticipating later discussion, the same value of r_a is valid for the diffusion of heat and water vapour. When e_o is the vapour pressure at a leaf surface, the transpiration rate is proportional to $(e_o - e)/r_a$ and, by analogy with Ohm's law, $(e_o - e)$ is a potential difference maintaining a current or flux of water vapour across an external diffusion resistance r_a in the air surrounding a leaf. Similarly, the rate of diffusion *within* a leaf is proportional to $\{e_s(T_o) - e_o\}/r_l$, where r_l is the internal resistance of the leaf. This resistance is provided by stomatal pores, by the cuticle, and by the cell walls during severe water stress (see the contribution by Meidner, p. 185). In a turgid leaf with ample water for transpiration, the resistance of cell walls is negligible and the cuticle presents a resistance much larger than the parallel resistance of stomata. The total leaf resistance is almost equal to the resistance of the stomata alone and is governed by stomatal sizes and populations (Milthorpe, 1962).

When the rates of diffusion within the leaf and in the external air are equal

$$E \left(\frac{\gamma \lambda}{\rho c} \right) = \frac{e_s(T_o) - e_o}{r_l} = \frac{e_o - e}{r_a} \tag{10}$$

where the constant $\gamma \lambda / (\rho c)$ preserves the consistency of units, and by rearranging terms

$$e_s(T_o) - e = \left(1 + \frac{r_l}{r_a} \right) (e_o - e) \tag{11}$$

Equation (11) shows that a leaf with internal resistance r_l can be treated formally as a free water surface if the vapour pressure difference $(e_o - e)$ is replaced by $\{e_s(T_o) - e\} / (1 + r_l/r_a)$. Alternatively, the psychrometer constant γ can be replaced by a modified value

$$\gamma^* = \gamma \left(1 + \frac{r_l}{r_a} \right) \tag{12}$$

Returning to equation (8), the latent heat of transpiration from a leaf becomes

$$\lambda E = \frac{\Delta H + \rho c \{e_s(T) - e\} / r_a}{\Delta + \gamma^*}, \quad (13)$$

as derived by Penman (1953) with somewhat different symbolism.

The alternative equations (9) and (13) show how the rate of evaporation from a wet surface or from a leaf depends on four weather parameters: net heat from external sources, usually radiation; temperature, humidity, and wind speed. When the leaf resistance is constant, evaporation increases (i) linearly with radiation, (ii) linearly with saturation deficit or wet-bulb depression, (iii) with air temperature when radiation and saturation deficit are constant. The variation with wind speed is more complex. An increase of wind speed increases the rate of exchange of air between the environment and the surface and so increases $1/r_a$. From equation (9), evaporation rate increases with wind speed when $D > D_0$. This condition is *always* satisfied by a free water surface evaporating into unsaturated air

$$(D > 0, \quad D_0 = 0)$$

and will *sometimes* be satisfied by a leaf. Evaporation rate is independent of wind speed when $D = D_0$, i.e. when water evaporates into saturated air ($D = D_0 = 0$) or when wet-bulb depression is the same at the leaf surface and in the surrounding air. Then the evaporation rate is $\lambda E = \Delta H / (\Delta + \gamma)$. Finally, the rate of transpiration from a leaf will decrease with increasing wind speed when $D < D_0$.

Physically, the dependence of transpiration rate on wind speed depends on relative changes of surface temperature and the rate of exchange ($1/r_a$). Provided H is positive, an increase of wind speed always decreases leaf temperature. By itself, this cooling tends to decrease losses of sensible and latent heat, but this effect is compensated by an increase in ($1/r_a$). When this increased rate of exchange more than compensates for the decrease of temperature, the loss of sensible heat is forced at the expense of latent heat and transpiration rate decreases with wind speed. Conversely, when an increase in the rate of exchange fails to compensate for cooling, the loss of latent heat is forced at the expense of sensible heat and transpiration increases with wind speed.

EXTERNAL RESISTANCE

(i) *Single leaves*

When a potential difference of one volt is maintained across an electrical resistance, the value of the resistance in ohms is the time in seconds for the

passage of unit charge. A volume of air to exchange 1 This time depends partly o geometry of the surface. T different size, it is conver meters (d/Kr_a) and (ud/ν) diameter of a circular disk o circular diffusion coefficient f omatic viscosity ($\text{cm}^2 \text{sec}^{-1}$) molecular diffusion coeffi

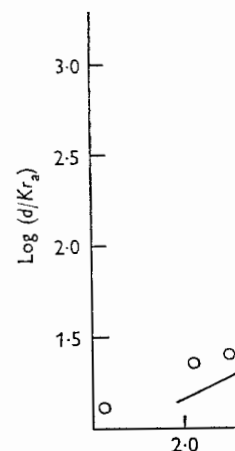


Fig. 2. Non-dimensional exchange coefficient. Both scales are logarithmic. ●, KU', Kuiper (1961); RA, Rasch

number and is the ratio of thickness (Kr_a) of a laminar boundary layer. When K is the diffusion coefficient, u is the wind speed, and ν is the kinematic viscosity, the Nusselt number is proportional to the ratio of inertial to viscous forces. When the surface is a flat plate, the Nusselt number is proportional to 2×10^4 , viscous forces predominate but when the number exceeds this value, turbulence occurs (Fishenden & Saunders, 1957). The full potential for evaporation from squares

passage of unit charge. As a diffusive resistance, r_a is the time for unit volume of air to exchange heat (or water vapour) with unit area of surface. This time depends partly on the speed of the air-stream and partly on the geometry of the surface. To compare measurements of r_a for surfaces of different size, it is convenient to work with two non-dimensional parameters (d/Kr_a) and (ud/ν) where d = characteristic length of surface, e.g. diameter of a circular disk or width of a narrow rectangle (cm.); K = molecular diffusion coefficient for heat or water vapour ($\text{cm}^2 \text{sec}^{-1}$); ν = kinematic viscosity ($\text{cm}^2 \text{sec}^{-1}$); u = wind speed (cm. sec^{-1}). When K is the molecular diffusion coefficient for heat transfer, (d/Kr_a) is a Nusselt

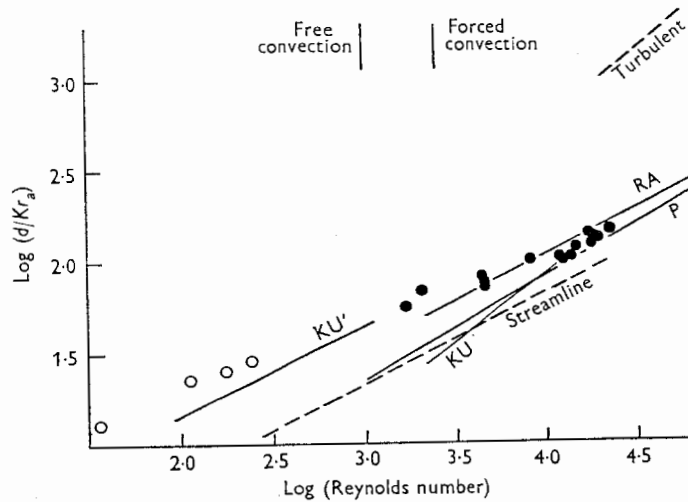


Fig. 2. Non-dimensional exchange parameter (d/Kr_a) as function of Reynolds number. Both scales are logarithmic. ●, Martin (1943); ○, Slatyer & Bierhuizen (1964); KU and KU', Kuiper (1961); RA, Raschke (1956); P, Powell (1942).

number and is the ratio of the characteristic length d to the effective thickness (Kr_a) of a laminar boundary layer for the molecular diffusion of heat. When K is the diffusion coefficient for water vapour, (d/Kr_a) is an analogous number for vapour transfer. The Reynolds number (ud/ν) is proportional to the ratio of inertial to viscous forces in the air near the surface. When the surface is a flat plate and the Reynolds number is less than 2×10^4 , viscous forces predominate and maintain laminar streamline flow, but when the number exceeds 2×10^4 the flow tends to become turbulent (Fishenden & Saunders, 1957). The dashed lines in Fig. 2 show an abrupt increase in Nusselt number, and hence a decrease in r_a , with the onset of turbulence. The full points are calculated from measurements of the evaporation from squares of filter paper (Martin, 1943), and the lines P

and KU show the mean of wind-tunnel measurements by Powell (1942) and Kuiper (1961). The line RA was derived from measurements by Raschke (1956) on the heat exchange of an *Alocasia* leaf. All three sets of measurements (and a fourth set by de Vries & Venema (1954) omitted for clarity) are close to the streamline relationship for flat plates and (d/Kr_a) is an order of magnitude less than for turbulent flow at the same Reynolds number.

On the left-hand side of the diagram, open circles are recent measurements by Slatyer & Bierhuizen (1964) at wind speeds less than 3 cm. sec.⁻¹, and KU' is the mean of measurements by Kuiper at wind speeds less than 20 cm. sec.⁻¹. Kuiper attributed the discontinuity between KU and KU' to the transition from streamline to turbulent flow with increasing wind speed. It is unlikely that this transition occurred at such a small Reynolds number (2×10^3) or that the Nusselt number decreased by an order of magnitude after the onset of turbulence. More plausibly, the two sets of measurements at slow wind speeds (KU' and open circles) are characteristic of a régime of free convection where transfer processes are governed by a downward flow of relatively cool dense air from the surface of a transpiring leaf, or by the upward movement of relatively warm air when the leaf is heated by radiation.

When Raschke measured the exchange of heat and water vapour at the surface of leaves in the open air, the diffusion resistances for heat and vapour were identical within the limits of experimental error. Powell found that the resistance for the diffusion of water vapour from wet surfaces in a wind tunnel was 0.9 of the diffusion resistance for heat. This small difference is negligible in most problems of heat balance and evaporation. Slatyer & Bierhuizen report mean values of r_a for heat that are only 0.4 of the resistance for vapour transfer at the same wind speed, and this much larger difference can be attributed to the exchange of heat by buoyancy in the régime of free convection. During the day, at levels in vegetation where the wind speed exceeds 50 cm. sec.⁻¹, the exchange of heat, water vapour, and carbon dioxide at leaf surfaces is governed by the laws of forced convection and, in principle, all three rates of transfer can be calculated from the same external resistance.

Reviewing all the measurements summarized in Fig. 2, the line RA is probably the most representative of leaves in the open air. This line gives the relation

$$\begin{aligned} r_a &= 3.4 d(ud/\nu)^{-0.5} \\ &= 1.3 \sqrt{d/u}, \end{aligned} \quad (14)$$

whereas the results of de Vries & Venema give

$$r_a = 20 d(ud/\nu)^{-0.7}. \quad (15)$$

For Reynolds numbers below similar values of r_a lying between two similar surfaces in parallel for a single surface, and I (14) for a range of wind speed

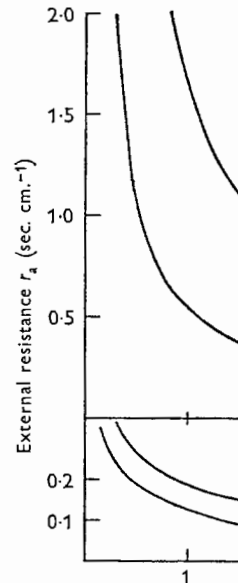


Fig. 3. (a) External resistance of leaf surface; (b) external resistance of crop canopy at 2 m. above zero plane d' (equivalent to d).

(ii)

Diffusion by atmospheric surfaces, measured diffusion profiles of vapour pressure (Rider, 1951; Rider, 1954). measured at several heights to give a straight line intercept (Long (1960) showed that the

where u and e are measured at crop, and k ($=0.41$) is a coefficient

measurements by Powell (1942) based from measurements by *casia* leaf. All three sets of Venema (1954) omitted for flat plates and (d/Kr_a) is an the same Reynolds number. circles are recent measure- speeds less than 3 cm. sec.⁻¹, per at wind speeds less than uity between KU and KU' flow with increasing wind ed at such a small Reynolds decreased by an order of e plausibly, the two sets of open circles) are character- sfer processes are governed air from the surface of a of relatively warm air when

eat and water vapour at the on resistances for heat and xperimental error. Powell of water vapour from wet on resistance for heat. This of heat balance and evapora- s of r_a for heat that are only same wind speed, and this the exchange of heat by during the day, at levels in sec.⁻¹, the exchange of heat, ces is governed by the laws ee rates of transfer can be

ed in Fig. 2, the line RA is he open air. This line gives

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$$(15)$$

For Reynolds numbers between 10^3 and 3×10^4 , the two equations give similar values of r_a lying between 0.2 and 0.02 sec. cm.⁻¹. Treating a leaf as two similar surfaces in parallel, external resistance is half the resistance for a single surface, and Fig. 3a shows values calculated from equation (14) for a range of wind speeds and leaf widths.

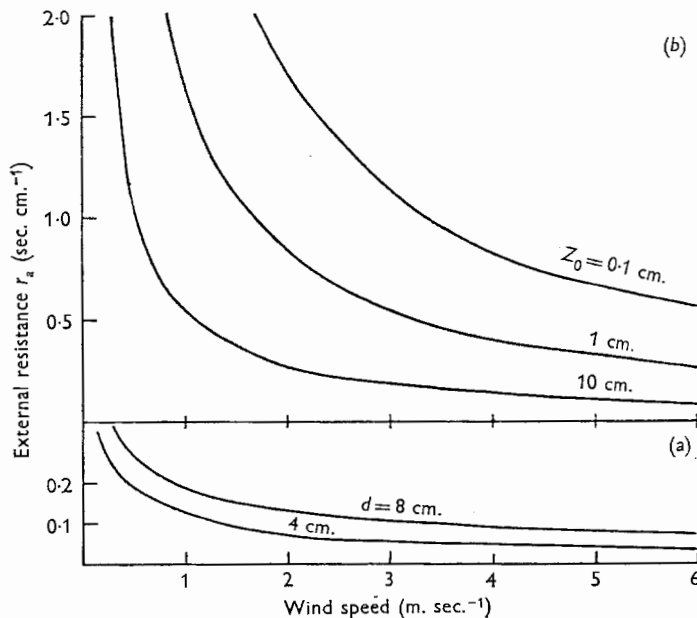


Fig. 3. (a) External resistance of single leaves with characteristic length d (equation 14); (b) external resistance of crop canopy with roughness constant z and wind speed measured at 2 m. above zero plane d' (equation (17b)).

(ii) *Uniform crop surfaces*

Diffusion by atmospheric turbulence governs processes of exchange between the atmosphere and the surface of a crop. Over extensive, uniform surfaces, measured diffusion coefficients of water vapour and momentum are equal over a wide range of atmospheric stability (Priestley, 1959) and profiles of vapour pressure and wind speed are the same shape (Robinson & Rider, 1951; Rider, 1954). When vapour pressure e and wind speed u are measured at several heights *above* a crop canopy, e can be plotted against u to give a straight line intercepting the axis $u = 0$ at $e = e_0$. Penman & Long (1960) showed that the latent heat of evaporation is

$$\lambda E = \frac{k^2 \rho c (e_0 - e) u}{\{\ln z(-d')/z_0\}^2}, \tag{16}$$

where u and e are measured at the same convenient height z above the crop, and k ($= 0.41$) is a constant of proportionality. The zero plane height

d' is chosen to make wind speed proportional to $\ln(z-d')$, and the roughness parameter z_0 is found by plotting u against $\ln(z-d')$ to give an intercept $\ln z_0$ on the axis $u = 0$. By definition, the external or aerodynamic resistance between the crop surface and the air at height z is then

$$r_a = \frac{\rho c}{\gamma \lambda} \frac{e_0 - e}{E} \quad (17a)$$

$$= \frac{\{\ln(z-d')/z_0\}^2}{k^2 u}, \quad (17b)$$

and the variation of r_a with roughness and wind speed is shown in Fig. 3*b*. Businger (1956) incorporated equation (17*b*) in Penman's equation for evaporation and the success of this modification was later confirmed by Tanner & Pelton (1960*a*).

When the roughness length of a surface is constant, its aerodynamic resistance is inversely proportional to wind speed. In practice, many crops get smoother, and roughness decreases as wind speed increases (Monteith, 1963*a*; Stoller & Lemon, 1963). For these surfaces, there is a range of wind speeds within which the resistance is almost constant.

The derivation of an aerodynamic resistance from a logarithmic wind profile is strictly valid only when the change of temperature with height is close to the dry adiabatic lapse rate. In practice, this usually means that the surface is only a fraction of a degree Centigrade hotter or cooler than the air at screen height. When the temperature difference is several degrees and the wind speed is a few metres per second or less, the rate of mixing of the air by buoyancy is comparable with mixing by mechanical turbulence. When temperature decreases with height so that the air is unstable, the logarithmic wind profile exaggerates the increase of wind with height, and r_a is overestimated when a value of surface roughness derived in neutral conditions is used in equation (17*b*). Conversely, when temperature increases with height and the air is stable, r_a is underestimated. Resistances can be corrected for buoyancy from functions of the wind and temperature gradients (Monteith, 1963*a*) or from fluxes of sensible heat and momentum (Monteith, Szeicz & Yabuki, 1964).

Table 2 contains values of r_a for several types of vegetation, calculated where possible from the ratio of a vapour pressure gradient to an evaporation rate. When the air is unstable this ratio gives a 'true' resistance, smaller than the value estimated from wind speed and surface roughness. For example, the smoothest surface in Table 2 is short grass with

$$z_0 = 0.1 \text{ cm.}$$

In neutral conditions, the about 1 sec. cm.^{-1} , but the is only 0.5 sec. cm.^{-1} .

For several farm crops, t 0.5 sec. cm.^{-1} . It is notor profiles over forests, and cu roughness length. A rough only 500 cm. tall (Baumgar 0.06 sec. cm.^{-1} is consisten estimated transpiration rate

The diffusion resistance of sizes and populations with shape of the stomatal wall. of 0.4 sec. cm.^{-1} for sunflo a figure of 1.4 sec. cm.^{-1} fo resistance of 0.8 sec. cm.^{-1} total leaf resistance tend to stomatal path alone.

There are three ways of rates. The most fundamer supplementary measuremer e_0 . Given these measurer written as

In a second method avoidi leaf surface, the rate of tra rate of evaporation E_p from exposed in the same air st surface with surface temper

Elimination of r_a then give:

$$r_l = \frac{f}{\gamma}$$

.NT

to $\ln(z-d')$, and the roughness against $\ln(z-d')$ to give an external or aerodynamic resistance at height z is then

$$(17a)$$

$$(17b)$$

and speed is shown in Fig. 3b. Penman's equation for transpiration was later confirmed by

is constant, its aerodynamic resistance is constant. In practice, many crops show that wind speed increases with height, and on rough surfaces, there is a range of wind speeds, most constant.

Resistance from a logarithmic wind profile and temperature with height is constant. In practice, this usually means that the air is at a temperature higher or lower than the surface temperature difference is several degrees or less, the rate of transpiration with mixing by mechanical turbulence at height so that the air is well mixed. It overestimates the increase of wind speed with height. The value of surface roughness is constant on (17b). Conversely, when wind speed is stable, r_a is underestimated. In functions of the wind and temperature from fluxes of sensible heat (Penman, 1964).

Penman's equation, calculated from the vapour pressure gradient to an evaporation rate gives a 'true' resistance, which is constant and surface roughness. In short grass with

In neutral conditions, the resistance calculated from equation (17b) is about 1 sec. cm.⁻¹, but the 'true' value of r_a derived from equation (17a) is only 0.5 sec. cm.⁻¹.

For several farm crops, the aerodynamic resistance lies between 0.2 and 0.5 sec. cm.⁻¹. It is notoriously difficult to measure representative wind profiles over forests, and curious values emerge from attempts to determine roughness length. A roughness of $z_0 = 290$ cm. seems large for pine trees only 500 cm. tall (Baumgartner, 1956), but the corresponding resistance of 0.06 sec. cm.⁻¹ is consistent with measured vapour pressure gradients and estimated transpiration rates.

INTERNAL RESISTANCE

(i) *Single leaves*

The diffusion resistance of leaf stomata can be calculated from stomatal sizes and populations with the help of simplifying assumptions about the shape of the stomatal wall. Penman & Schofield (1951) derived a resistance of 0.4 sec. cm.⁻¹ for sunflower leaves and Penman & Long (1960) quote a figure of 1.4 sec. cm.⁻¹ for wheat. Bange (1953) calculated a minimum resistance of 0.8 sec. cm.⁻¹ for *Zebrina* leaves. Experimental values of total leaf resistance tend to be larger than these theoretical values for the stomatal path alone.

There are three ways of estimating leaf resistance r_l from transpiration rates. The most fundamental is also the most difficult because it needs supplementary measurements of surface temperature T_o and vapour pressure e_o . Given these measurements, r_l is calculated from equation 10 rewritten as

$$r_l = \left(\frac{\rho c}{\gamma \lambda} \right) \frac{e_s(T_o) - e_o}{E}. \quad (10a)$$

In a second method avoiding the measurement of vapour pressure at the leaf surface, the rate of transpiration E from a leaf is compared with the rate of evaporation E_p from a piece of wet filter paper of the same size exposed in the same air stream. If the paper behaves like a free water surface with surface temperature T_p ,

$$E = \frac{\rho c}{\gamma \lambda} \frac{e_s(T_o) - e}{r_l + r_a}, \quad (18a)$$

$$E_p = \frac{\rho c}{\gamma \lambda} \frac{e_s(T_p) - e}{r_a}. \quad (18b)$$

Elimination of r_a then gives

$$r_l = \frac{\rho c}{\gamma \lambda} \left[\frac{e_s(T_o) - e}{E} - \frac{e_s(T_p) - e}{E_p} \right]. \quad (19)$$

In a simpler third method, r_a is calculated from leaf size and wind speed using equation (14) or Fig. 3*a*. When the reflectivity of the leaf and a similar piece of wet filter paper are equal, both surfaces absorb the same amount of radiation and equations (12) and (13) give

$$E/E_p = \frac{\Delta + \gamma}{\Delta + \gamma^*}, \quad (20)$$

$$r_i = r_a \left(1 + \frac{\Delta}{\gamma} \right) \left(\frac{E_p}{E} - 1 \right). \quad (21)$$

When stomata are fully open in bright light, the leaf resistances of several species lie between 1 and 3 sec. cm.⁻¹ (Table 1), an order of magnitude greater than the external resistance at wind speeds met in the field.

Table 1. Leaf resistances estimated from transpiration rates

Species	r_i (sec. cm. ⁻¹)	Reference
Turnip, sugar beet	1.5-1.7	Gaastra (1959)
Bean, tomato	2.3-3.3	Kuiper (1961)
<i>Xanthium</i>	0.8-1.7	Mellor <i>et al.</i> (1964)
Cotton	0.9-1.3	Slatyer & Bierhuizen (1964)
Barley	1.0-2.0	Armstrong (unpublished)

(ii) Uniform crop surfaces

The external resistance of a crop can be found from wind, temperature and humidity profiles above the crop. When these are the same shape, plotting temperature (and vapour pressure) against wind speed at a series of heights gives a straight line from which $T = T_0$ at $u = 0$ can be read (and similarly $e = e_0$). Then by analogy with the equation defining the internal resistance of a single leaf, the surface resistance of a crop canopy may be written

$$r_s = \frac{\rho c}{\gamma \lambda} \left\{ \frac{e_s(T_0) - e_0}{E} \right\}. \quad (10b)$$

Although it will seldom be possible to identify T_0 and e_0 with conditions at a specific level within the canopy, the parameters r_a and r_s allow the equations of heat and vapour flux derived for a single leaf to be applied to a plant community. In particular, the rate of transpiration from a crop in given weather can be calculated from equation (13).

When evaporation from the soil surface is negligible, r_s is expected to be a plant parameter revealing diurnal and seasonal changes of stomatal resistance. When evaporation from the soil is comparable with transpiration, r_s combines the resistance of crop leaves with the capillary resistance

to the diffusion of water inversely proportional to r_s (Covey (1959)).

Resistances were estimated and were incorporated and photosynthesis (Mon Tanner (1963)). He object sources and sinks of heat, v levels within a crop canopy are determined from mea profiles are the same sha gences of flux or by effe estimated from measure are smallest. It has also b parameter because it incl leaves and air moving thro should change with wind show no such effect.

Table 2 summarizes val tion rates and profiles us O'Neill), or from compone canopy using equation (13 the mean resistance for radiation. At Kew and O The resistance for Camb profiles randomly timed o established that the ratio o tion from a wet surface at t ing surface resistance is d Trickett (1964) calculated t ments of surface temperat

Minimum values of surf roughly a third to a quarter (Table 1). Oversimplifyin crop is equivalent to the s parallel, each layer contain

The surface resistance c cm.⁻¹ in mid-June to 0.7 s the total leaf area index (le from natural senescence an attacked by mildew. The i

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Reference

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 Kuiper (1961)
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to the diffusion of water vapour in soil pores. The surface resistance is inversely proportional to a wetness parameter $\lambda E / \{e_s(T_o) - e_o\}$ proposed by Covey (1959).

Resistances were estimated from evaporation measurements at Rothamsted and were incorporated in a theoretical model of crop transpiration and photosynthesis (Monteith, 1963*a*) that was adversely criticized by Tanner (1963). He objected that the derivation of r_s was invalid when the sources and sinks of heat, water vapour, and momentum were set at different levels within a crop canopy, but this criticism is irrelevant when r_s (and r_a) are determined from measurements *above* the canopy where established profiles are the same shape. When profiles are distorted by small divergences of flux or by effects of buoyancy, resistances can sometimes be estimated from measurements very near the surface where these anomalies are smallest. It has also been argued that r_s is not a purely physiological parameter because it includes an external resistance between transpiring leaves and air moving through the canopy. If so, this external component should change with wind speed but recent measurements at Rothamsted show no such effect.

Table 2 summarizes values of surface resistance calculated from evaporation rates and profiles using equation (10*b*) (Rothamsted, Kew, Davis, O'Neill), or from components of the energy balance and weather above the canopy using equation (13) (Hancock, Munich). Most of these values are the mean resistance for each hour during daylight, weighted by net radiation. At Kew and O'Neill, only 3 or 4 hr. were available each day. The resistance for Cambridge is an unweighted mean from a set of 23 profiles randomly timed over 5 days. From these profiles, Pasquill (1950) established that the ratio of actual transpiration rate to potential transpiration from a wet surface at the same temperature was 0.41, and a corresponding surface resistance is derived from equation (11). Linacre, Palmer & Trickett (1964) calculated the resistance of crops under glass from measurements of surface temperature.

Minimum values of surface resistance are between 0.3 and 0.5 sec. cm.⁻¹, roughly a third to a quarter of the minimum resistance for individual leaves (Table 1). Oversimplifying, the effective surface resistance of a transpiring crop is equivalent to the stomatal resistance of three or four leaf layers in parallel, each layer containing unit leaf area.

The surface resistance of barley increased from a minimum of 0.3 sec. cm.⁻¹ in mid-June to 0.7 sec. cm.⁻¹ at the end of July. During this period, the total leaf area index (leaves and sheaths) decreased from 10 to 6, partly from natural senescence and partly because many of the lower leaves were attacked by mildew. The increase of resistance with time may be caused by

a decrease in total leaf area leaves, or by a combination soil moisture deficit increases was probably too small to r Eliminating differences i Bavel (1964) found that d *vulgaris* var. *sudanensis*) 75 difference reveals an increa from a neighbouring stand with the development of i resistance in leaves that ha

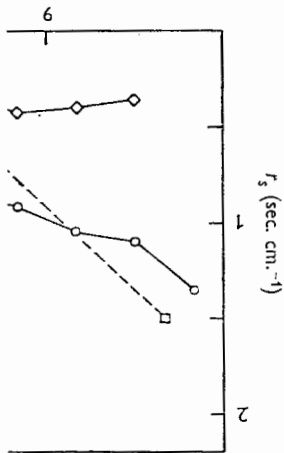


Fig. 4. Diurnal variation of crop resistance (rs) in leaves of crop (Time), \diamond , 28 May 1960; \square , 28 May 1960; \circ , 28 May 1960.

The increase of surface correlated with a decrease in August when the soil moisture decreased to 15 sec. cm^{-1} . From the fraction of net radiation that the soil water content. Diurnal changes of seasonal changes. Because the loss of water from leaves increase of surface resistance

Table 2. Weather and resistances

Surface	Height (cm.)	Site	Dates	Reference	Evapora- tion E (mm. $\text{hr}^{-1} \times$ 100)	Net heat H/A ($\text{min}.$ $\text{hr}^{-1} \times$ 100)	Mean air tempera- ture ($^{\circ}\text{C}.$)	Mean vapour pressure (mb.)	Mean wind speed ($\text{m}.$ sec^{-1})	R_0 ($\text{cm}.$)	Resistances (sec. cm^{-1})						
											r_a	r_l	r_s ($\Delta E/H$) $\times 100$				
Grass																	
Timothy and meadow fescue	60-70	Rothamsted	18. v. 1961	Monreith (1963a)	25	33	12.2	9.6	2	4	0.2	0.37	0.5	76			
Rough pasture with some clover	5-10	Cambridge	8-12. v. 1947	Pasquill (1950)	—	—	—	—	3	0.7	0.4	—	0.5	—			
Lawn	1	Kew	21. vi. 1949	Robinson & Rider (1951)	20	41	21.5	13.0	3	0.1	0.5	0.83	4.1	49			
Rye-grass	—	Davis, California	28. v. 1960	Monreith (1963b)	{ 48 35 30 }	60	21.6	13.2	2	5 (?)	0.3	0.87	0.5	80			
Natural prairie	6	O'Neill, Nebraska	1. vii. 1960	Lettau & Davidson (1957)	{ 37 31 30 }	60	30.6	16.3	7	5.0	0.3	1.32	2.1	138			
Alfalfa-brome mixture	15	Hancock, Wisconsin	25. viii. 1953 31. viii. 1953 24. ix. 1957	Tanner & Pelton (1960b)	{ 17 10 38 }	58	27.8	20.0	8	0.9	0.2	0.84	7.0	67			
Barley																	
Proctor	60 } 105 }	Rothamsted	{ 26. vi. 1963 23. vii. 1963	Szeicz (unpublished)	28 25	28	15.1 21.0	11.2 17.3	4 2	5 13	0.3 0.2	0.64 0.72	0.3 0.7	100 78			
Beans	95 } 95 } 135 }	Rothamsted	{ 15. vi. 1961 21. vi. 1961 25. vii. 1961	Monreith (1963b)	{ 27 25 15 }	29	17.0 17.9 20.9	12.8 11.6 14.9	1 1 1	9 7 3	0.3 0.2 0.2	0.62 0.60 0.82	0.5 1.1 2.3	93 63 47			
Rice																	
Paddy field	—	—	—	Covey (1959) from Kameski (1956)	—	—	—	—	—	—	—	—	0.5	—			
Pine forest	550	Munich	28. vi. - 7. vii. 1952	Baumgartner (1956)	60	90	27.7	12.3	2	290 (?)	0.06	0.62	0.9	67			
Glasshouse crops																	
Beans	18 } 18 } 18 }	Griffith, Australia	—	Linacre et al. (1964)	—	—	—	—	—	—	—	{ 0.2 0.2 0.3 }	0.3 0.3 0.8	— — —			

gradually closing in response to this stress. This interpretation is consistent with an increase of surface resistance when transpiration rate is increased. When the mean rate was $0.48 \text{ mm. hr.}^{-1}$, the resistance was $0.5 \text{ sec. cm.}^{-1}$. Three days later, hot dry winds forced the rate of evaporation to $0.84 \text{ mm. hr.}^{-1}$, and the corresponding increase of resistance to $1.1 \text{ sec. cm.}^{-1}$ may be a response of self-preservation stimulated by an increase in the water deficit of leaf tissue. The decrease of resistance after sunrise may demonstrate a slow opening of stomata in response to increasing light intensity; or an increase of permeability in plant roots as the soil gets warmer near the surface.

EVAPORATION FROM WET AND DRY SURFACES

The rate of evaporation E_0 from a surface thoroughly wetted by rain or dew can be calculated from equation (8). If E is the rate of transpiration in the same weather when the surface is dry, the relative transpiration rate is

$$\frac{E}{E_0} = \frac{\Delta + \gamma}{\Delta + \gamma(1 + r_s/r_a)}, \quad (20a)$$

and this fraction depends only on air temperature and the ratio r_s/r_a . (Equation (20a) is valid when wet and dry leaves reflect the same fraction of solar radiation, a condition usually satisfied unless the sun is very near the horizon.) Fig. 5 shows how the relative transpiration rate changes with resistance. Taking a constant minimum leaf resistance of 2 sec. cm.^{-1} , the ratio r_s/r_a ranges from 5 to 20 when r_a ranges from 0.4 to $0.1 \text{ sec. cm.}^{-1}$.

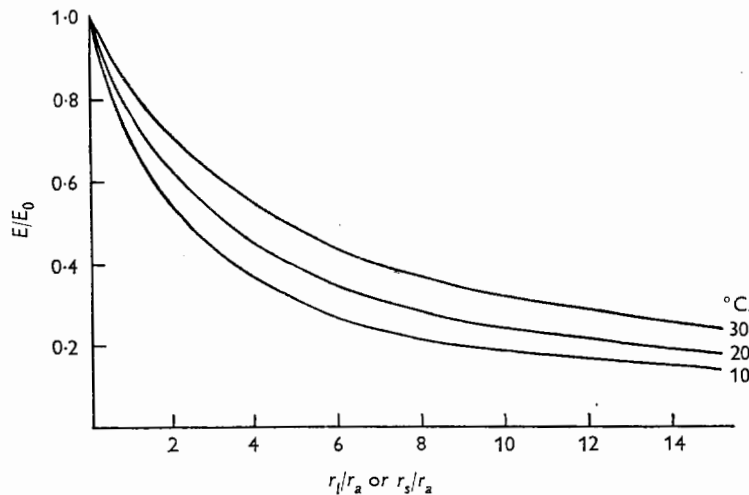


Fig. 5. Change of relative evaporation rate E/E_0 with ratio r_l/r_a or r_s/r_a and with temperature.

At 20°C . the relative transpiration rate is about 0.25. The absolute rate of transpiration is $0.12 \text{ mm. hr.}^{-1}$. The stomatal resistance and is 10 sec. cm.^{-1} . In still air when r_a is very large, the control of transpiration is not by stomata. As shown by Frankenberg (1960) who found that the rate of transpiration is almost linearly with stomatal resistance. The relative transpiration rates were independent of stomatal resistance.

Ehlig & Gardner (1964) found that at 16 atm. , transpiration from glass decreased to about 20% of the rate from soil. Because this decrease is not linear, the results of these and similar field conditions.

When *field crops* are supplied with water, r_s/r_a is between 1 and 2, and the relative transpiration rate is between 0.76 and 0.62. Frankenberg (1960), who found that the evaporation from the soil is the same as the evaporation from the leaves when the leaves were wetted by rain. The relative transpiration rate from grass, potato, etc. is about 0.8 E_0 .

McIlroy & Angus (1964) found that the evaporation from grass in a field with irrigation, a result predicted by equation (17b), at a mean r_a of $0.7 \text{ sec. cm.}^{-1}$, for *neutral* conditions during most daylight hours. In stable air, the resistance may often have reached a maximum. In frequent showers, the surface resistance may often have reached a maximum. Assuming a maximum value of r_s between E and E_0 would be 10 sec. cm.^{-1} . Similar measurements reported by Frankenberg (1960) are interpreted in the same way.

By contrast, the ratio r_s/r_a is about one-fifth of that in a field with irrigation and exposed to the same weather. In a shower may be an important factor in dry regions, and in the y

This interpretation is consistent when transpiration rate is hr.^{-1} , the resistance was forced the rate of evaporation; increase of resistance to evaporation stimulated by an decrease of resistance after tomata in response to inability in plant roots as the

DRY SURFACES

roughly wetted by rain or the rate of transpiration in relative transpiration rate is

$$(20a)$$

perature and the ratio r_s/r_a . es reflect the same fraction unless the sun is very near spiration rate changes with sistance of 2 sec. cm.^{-1} , the from 0.4 to $0.1 \text{ sec. cm.}^{-1}$.

At 20°C . the relative transpiration rate lies between 14 and 40%. The absolute rate of transpiration for *single leaves* is sensitive to changes in stomatal resistance and is relatively independent of wind speed. In very still air when r_a is very large, E approaches E_0 and then the physiological control of transpiration is negligible. This was illustrated by Bange (1953), who found that the rate of transpiration from *Zebrina* leaves increased almost linearly with stomatal width in 'still' air, whereas, in 'wind', transpiration rates were independent of stomatal width between 10 and 20μ .

Ehlig & Gardner (1964) found that when soil moisture potential reached 16 atm., transpiration from pepper, sunflower and cotton grown under glass decreased to about 20% of the maximum rate of transpiration in wet soil. Because this decrease depends on the ratio r_s/r_a and not on r_s alone, the results of these and similar experiments are not necessarily relevant to field conditions.

When *field crops* are supplied with adequate water, the minimum ratio r_s/r_a is between 1 and 2, and at 20°C . the corresponding ratio of E/E_0 is between 0.76 and 0.62. This range is consistent with measurements by Frankenberger (1960), who calculated the ratio of actual transpiration to the evaporation from the same crop in similar weather immediately after the leaves were wetted by rain. On many rainless days, the rate of transpiration from grass, potatoes, sugar beet and rye was between 0.6 and $0.8 E_0$.

McIlroy & Angus (1964) were unable to detect an increase in the rate of evaporation from grass immediately after the leaves were wetted by irrigation, a result predictable as follows. The grass was mown to a height of 6–10 cm. and the roughness length would be near $z_0 = 1 \text{ cm}$. From equation (17b), at a mean wind speed of 2.5 m. sec.^{-1} , the value of r_a is $0.7 \text{ sec. cm.}^{-1}$, for *neutral* conditions, but internal evidence shows that during most daylight hours air temperature increased with height. In this stable air, the resistance must have exceeded $0.7 \text{ sec. cm.}^{-1}$ (p. 214) and may often have reached $1\text{--}2 \text{ sec. cm.}^{-1}$. Because irrigation was very frequent, the surface resistance was probably about $0.3 \text{ sec. cm.}^{-1}$. Assuming a maximum value of $r_s = 0.3$, Fig. 5 shows that the difference between E and E_0 would be a few per cent at most and too small to detect. Similar measurements reported by McMillan & Burgy (1960) can be interpreted in the same way.

By contrast, the ratio r_s/r_a for a pine forest is 15 and the transpiration rate is about one-fifth of the evaporation rate from leaves wetted by rain and exposed to the same weather. The rapid loss of intercepted rain after a shower may be an important factor in the water economy of trees growing in dry regions, and in the yield of water from forested catchments.

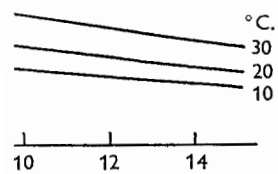


Fig. 5. E/E_0 with ratio r_s/r_a or r_i/r_a

LATENT HEAT AND AVAILABLE ENERGY

With values of r_s and r_a established for a range of crops, maximum rates of transpiration will now be related to prevailing weather. It is convenient to write equation (13) non-dimensionally as

$$\frac{\lambda E}{H} = \frac{(\Delta/\gamma) + r_i/r_a}{(\Delta/\gamma) + (r_s + r_a)/r_a}, \quad (22)$$

where

$$r_i = \rho c \frac{e_s(T) - e}{\gamma H} \quad (23)$$

is a property of the atmosphere with the dimensions of a diffusive resistance. When $r_i = r_s + r_a$, $\lambda E = H$, showing that r_i is a critical value of $(r_s + r_a)$ at which the net supply of heat from external sources is used exclusively to evaporate water. Then the flux of sensible heat is zero and the system is isothermal because it contains no gradients of temperature. This theorem identifies r_i as an 'isothermal resistance' proportional to the ratio of saturation deficit to net available heat. Table 2 shows that this ratio increases from a minimum of about 0.4 sec. cm.⁻¹ in the relatively humid climate of south-east England to a maximum of about 1.3 sec. cm.⁻¹ in the much drier climate of the western U.S.A.

(i) Dependence of $\lambda E/H$ on r_a

To find how $\lambda E/H$ changes with external resistance, equation (22) is differentiated with respect to r_a assuming constant radiation, temperature and humidity. The evaporation rate is independent of r_a when

$$r_s = \left(1 + \frac{\gamma}{\Delta}\right) r_i \quad (24)$$

and substitution of this ratio in equation (22) gives

$$\frac{\lambda E}{H} = \frac{\Delta}{\Delta + \gamma}$$

Fig. 6 shows this result graphically for weather with $r_i = 0.6$ sec. cm.⁻¹, $\Delta = 1$ mb. °C.⁻¹. At $r_s = (1 + \gamma/\Delta) r_i = 1$ sec. cm.⁻¹, the evaporation rate is independent of r_a and hence of wind speed and surface roughness. From equation (24), the Bowen ratio (sensible heat flux:latent heat flux) is then (γ/Δ) . When r_s exceeds 1 sec. cm.⁻¹, the evaporation rate decreases with increasing wind speed and roughness, and the Bowen ratio is more

than (γ/Δ) . Conversely, $\lambda E/H$ increases with wind speed and surface roughness when $r_s < (1 + \gamma/\Delta) r_i$.

In laboratory studies of small without intense radiation rates increase with wind speed. Salisbury & Raschke (1961)

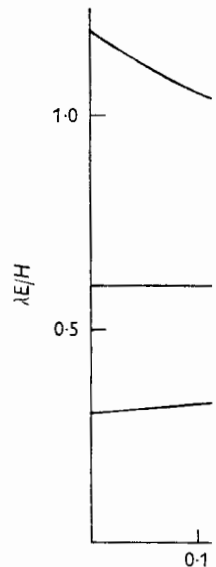


Fig. 6. Fraction of net heat used for evaporation when $\Delta = 1$ mb.

source of light (about 1 cal. cm.⁻² min.⁻¹) and only 0.8 mb. The calculation shows that r_s exceeded $(1 + \gamma/\Delta) r_i$ by a factor of two when wind speed was 10 m.p.h.

The dependence of transpiration on wind speed is established in the field because of the isothermal resistance theorem. Ladefoged (1963) found that transpiration is uncorrelated with wind speed in the isothermal resistance theorem. This result is confirmed and, from equation (24), is shown in Table 2.

BLE ENERGY

of crops, maximum rates of weather. It is convenient to

$$\frac{\lambda E}{H} = \frac{r_a}{r_s + r_a} \quad (22)$$

$$\frac{\lambda E}{H} = \frac{r_i}{r_s + r_i} \quad (23)$$

mensions of a diffusive re- that r_i is a critical value of m external sources is used sensible heat is zero and theadients of temperature. This ce' proportional to the ratio. ble 2 shows that this ratio n.⁻¹ in the relatively humid f about 1.3 sec. cm.⁻¹ in the

on r_a resistance, equation (22) is stant radiation, temperature ndent of r_a when

$$\frac{\lambda E}{H} = \frac{r_i}{r_s + r_i} \quad (24)$$

gives

er with $r_i = 0.6$ sec. cm.⁻¹, sec. cm.⁻¹, the evaporation eed and surface roughness. heat flux: latent heat flux) is e evaporation rate decreases nd the Bowen ratio is more

than (γ/Δ) . Conversely, when r_s is less than 1 sec. cm.⁻¹, evaporation increases with wind speed and roughness and the Bowen ratio is less than (γ/Δ) .

In laboratory studies of transpiration, r_i is usually large because H is small without intense radiation. Then $r_s < (1 + \gamma/\Delta) r_i$ and transpiration rates increase with wind speed. To demonstrate the opposite effect, Mellor, Salisbury & Raschke (1964) exposed leaves of *Xanthium* to an intense

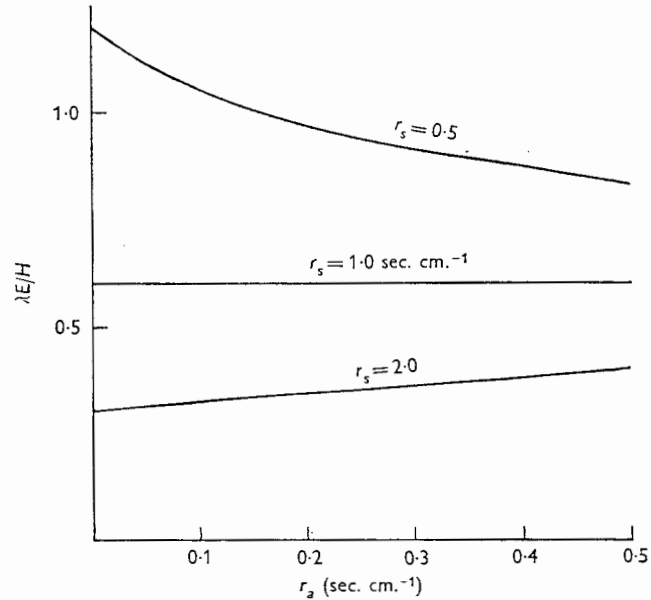


Fig. 6. Fraction of net heat used for evaporation as function of r_a and r_s (or r_i), assuming $\Delta = 1$ mb. ° C.⁻¹ and $r_i = 0.6$ sec. cm.⁻¹.

source of light (about 1 cal. cm.⁻² min.⁻¹) in air with a saturation deficit of only 0.8 mb. The calculated value of r_i is only 0.2 sec. cm.⁻¹, and presumably r_s exceeded $(1 + \gamma/\Delta) r_i$. The transpiration rate decreased by a factor of two when wind speed increased from 0.2 to 4 m. sec.⁻¹.

The dependence of transpiration rate on wind speed is difficult to establish in the field because r_s , r_i and temperature all vary diurnally. Ladefoged (1963) found that the rate of transpiration from beech trees was uncorrelated with wind speed between 1 and 7 m. sec.⁻¹. Very roughly, the isothermal resistance corresponding to mean weather was 0.3 sec. cm.⁻¹ and, from equation (24), $r_s = 0.5$ sec. cm.⁻¹, consistent with values in Table 2.

(ii) *Dependence of $\lambda E/H$ on r_s*

Because Δ and r_a appear both in the numerator and in the denominator of equation (22), $(\lambda E/H)$ is insensitive to changes of wind speed and temperature. This leaves only two variables, r_s and r_i , to determine the fraction of net available heat used for evaporation. The values of $(\lambda E/H)$ in Table 2 show a clear inverse correlation with r_s alone and Fig. 7 shows that they fit a straight line

$$\frac{\lambda E}{H} = \left(\log \frac{25}{r_s} \right)^{\frac{1}{2}}. \quad (25)$$

There are two reasons why $(\lambda E/H)$ should appear independent of r_i . First, saturation deficit decreases, and so r_i decreases, when air blows from a hot, dry surface (large r_s) to a relatively cool, moist surface (small r_s). This process introduces a correlation between the equilibrium values of r_i and r_s some distance from the boundary. Near the boundary, there is a

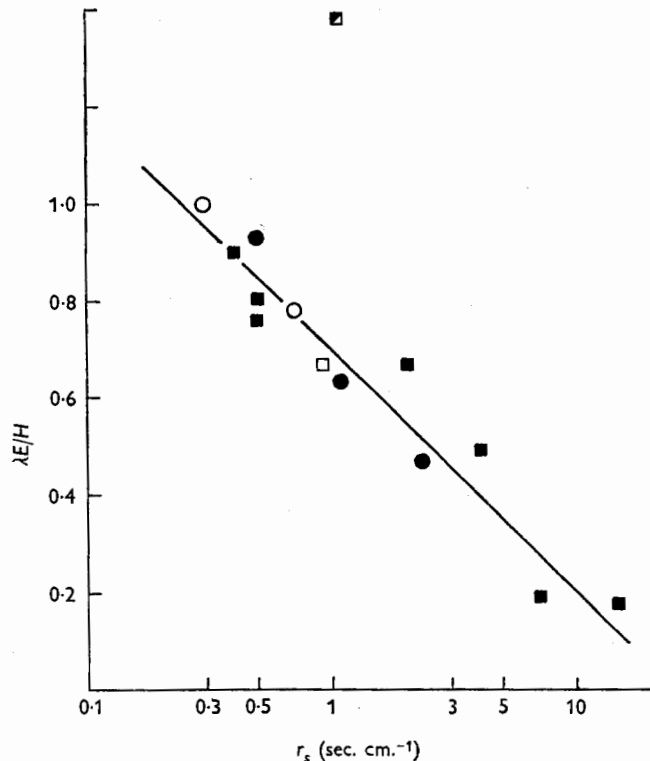


Fig. 7. Fraction of net heat used for evaporation as function of surface resistance r_s . Measurements from Table 2. ■, Grass; □, grass, Davis, 1 June 1960; □, pine forest; ●, beans; ○, barley.

lack of equilibrium illustrated 1 June 1960, and examined an obvious reason for the good agreement: r_i is much smaller than the

The numerical constant in equation (25) is based on measurements that establish that surfaces have their own characteristics which can be used cautiously to estimate transpiration from measurements of canopy resistance that uses water conservation. This constant has never been measured directly. Transpiration was about 12 $\text{mm} \text{ day}^{-1}$ and $r_s = 15 \text{ sec. cm.}^{-1}$.

TRANSPI

In previous sections, E was the sum of transpiration and evaporation from the surface. The vertical flux increases with distance from the surface. Saito (1962) and was calculated from a model. This model will not determine the energy balance determining the energy available for visible radiation (in the range 0.4 to 0.7 μm).

To relate the attenuation of radiation by foliage of a crop is divided into layers. The unit leaf area index can be described by the distance from the ground plane or, more simply, by the fraction of radiation passing straight through. The fraction ranges from 0.4 for crops with more leaves to 0.8 for crops with more leaves (Saito 1965). In each layer, a fraction of radiation is absorbed by leaves whose transmittance is t_n in the n th layer below the top of the

but the mean intensity of radiation intercepted is

When a crop with $s = 0.4$ and a visible intensity of 0.4 c

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 the equilibrium values of r_i
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lack of equilibrium illustrated by the anomalous point for Davis on 1 June 1960, and examined later in more detail. A second and more obvious reason for the good correlation of $(\lambda E/H)$ and r_s is that the range of r_i is much smaller than the range of r_s .

The numerical constant in equation (25) may depend on the choice of measurements that establish it and further work may show that different surfaces have their own characteristic constants. Meanwhile, the equation can be used cautiously to estimate the internal resistance of closed crop canopies from measurements of $(\lambda E/H)$. For example, pineapple is a crop that uses water conservatively for transpiration but its internal resistance has never been measured. Ekern (1964) found that the latent heat of transpiration was about 12% of net radiation and, if equation (25) is valid, $r_s = 15 \text{ sec. cm.}^{-1}$.

TRANSPIRATION AND LEAF AREA

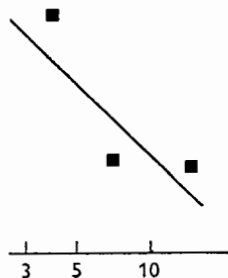
In previous sections, E was the total flux of water vapour from a field crop, the sum of transpiration from all the leaves intercepting radiation and evaporation from the surface of the underlying soil. Within the canopy, the vertical flux increases with height at a rate that was measured in wheat by Saito (1962) and was calculated by Penman & Long (1960) from a simple model. This model will now be extended to include a profile of net radiation determining the energy available for transpiration and a profile of visible radiation (in the range $0.4\text{--}0.7\mu$) determining stomatal aperture.

To relate the attenuation of radiation to the distribution of leaves, the foliage of a crop is divided into a set of horizontal leaf layers, each containing unit leaf area index. Within each layer, the distribution of leaves can be described by the distribution of angles with respect to a horizontal plane or, more simply, by a bulk parameter s which is the fraction of radiation passing straight through the layer without striking a leaf. This fraction ranges from 0.4 for crops with horizontal leaves (kale, clover) to 0.8 for crops with more nearly vertical leaves (cereals, grasses) (Monteith, 1965). In each layer, a fraction $(1-s)$ of incident radiation is intercepted by leaves whose transmission coefficient is τ . It can be shown that, in the n th layer below the top of the canopy, the intensity on a *horizontal* surface is

$$I_n = \{s + (1-s)\tau\}^{n-1} I_0,$$

but the mean intensity of radiation on *leaf* surfaces is only $(1-s)I_n$, the amount of radiation intercepted by unit leaf area.

When a crop with $s = 0.8$, $\tau = 0.1$ is exposed to bright sunshine with a visible intensity of $0.4 \text{ cal. cm.}^{-2} \text{ min.}^{-1}$, the mean intensity of light on



function of surface resistance r_s
 Davis, 1 June 1960; □, pine forest;

leaf surfaces is $0.08 \text{ cal. cm.}^{-2} \text{ min.}^{-1}$ in the first layer (with unit leaf area index) and $0.07 \text{ cal. cm.}^{-2} \text{ min.}^{-1}$ in the second layer. Gaastra (1959) and Kuiper (1961) found that the leaf resistances of several field crops increased rapidly with decreasing light intensity below $0.08 \text{ cal. cm.}^{-2} \text{ min.}^{-1}$ because stomata began to close. Comparison of these figures suggests that the amount of visible light reaching crop leaves in the field may often fail to reach the intensity needed for maximum stomatal opening.

This hypothesis will now be tested from profiles of radiation, temperature, humidity and wind speed available from a field of barley on 23 July 1963 (Long, Monteith, Penman & Szeicz, 1964). The transmission coefficient for visible radiation was 0.07 and s was 0.8 . The resistance of

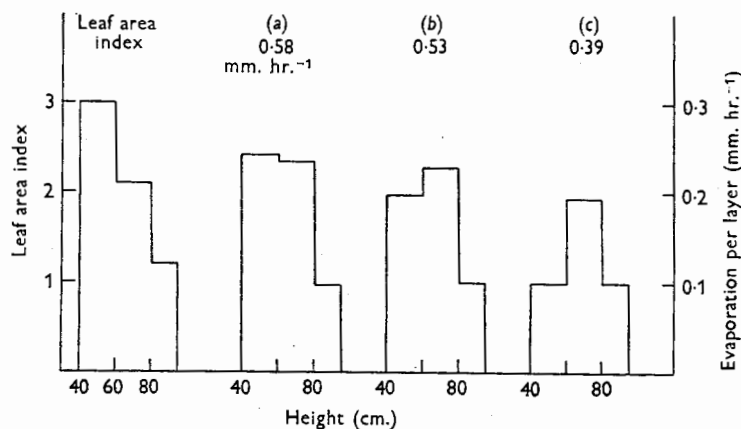


Fig. 8. Calculated transpiration from a stand of barley, 23 July 1963, in three layers, 40 to 60, 60 to 80, 80 to 95 cm.: (a) with T , e , u , r_s constant; (b) with T , e , u varying with height, r_s constant; (c) with T , e , u , r_s varying with height. Figures above each section are total transpiration.

barley leaves was assumed equal to the resistance of Kuiper's bean leaves at the same light intensity. Between 09.00 and 16.00 hours G.M.T., the measured mean rate of transpiration was $0.37 \text{ mm. hr.}^{-1}$, and Fig. 8 shows how estimates of this amount are improved as more of the relevant physical parameters are taken into account. In all the estimates the measured change of net radiation with height is included, but, in the first, values of temperature, humidity, and wind speed at a height of 2 m. were assumed to exist at all levels and all stomata were assumed to be fully open. For these conditions, the transpiration rate was $0.58 \text{ mm. hr.}^{-1}$ and the largest contribution came from the lowest layer of leaves, between 40 and 60 cm. Putting in the measured values of temperature, humidity and wind, but continuing to assume fully open stomata, the total transpiration calculated is $0.53 \text{ mm. hr.}^{-1}$, still too big, with the lowest layer contributing 20% less

than before. Finally, put from measurements of light in good agreement with the

Fig. 9 illustrates a more 0.8 . The transmission of light and 0.25 for both solar and similar in shape. Above total net radiation and 0.45 cal. values were assumed constant $r_i = 0.6 \text{ sec. cm.}^{-1}$, $r_a = 0.8$ that transpiration rates re

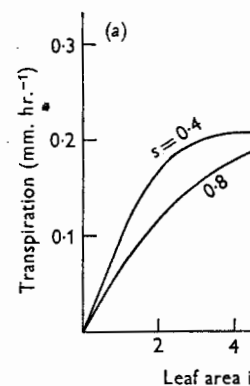
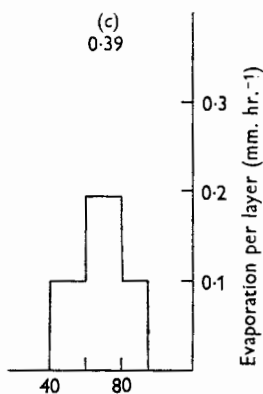


Fig. 9. (a) Calculated transpiration area index; (b) measured evaporation plotted as function of visible radiation (Metelerkamp, 1963, 1964.)

$s = 0.4$ but increase beyond side of the figure shows constant (Metelerkamp (1963, 1964) on the differences in leaf arrangement s). In particular, an increase beyond a leaf area index for a crop with $s = 0.8$. Transpiration from the soil is negligible evaporation would be large index.

st layer (with unit leaf area l layer. Gaastra (1959) and several field crops increased low 0.08 cal. cm.⁻² min.⁻¹ f these figures suggests that s in the field may often fail stomatal opening. profiles of radiation, tempera- a field of barley on 23 July (1964). The transmission co- was 0.8. The resistance of



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than before. Finally, putting in values of stomatal resistance estimated from measurements of light intensity, the total flux was 0.39 mm. hr.⁻¹, in good agreement with the measurement for the whole canopy.

Fig. 9 illustrates a more general relationship for crops with $s = 0.4$ and 0.8 . The transmission of leaves was assumed to be 0.05 for visible radiation and 0.25 for both solar and net radiation because their profiles are usually similar in shape. Above the canopy, intensities were 0.6 cal. cm.⁻² min.⁻¹ net radiation and 0.45 cal. cm.⁻² min.⁻¹ visible radiation, and the following values were assumed constant throughout the canopy: $\Delta = 1$ mb. °C.⁻¹, $r_i = 0.6$ sec. cm.⁻¹, $r_a = 0.2$ sec. cm.⁻¹. The left-hand side of Fig. 9 shows that transpiration rates reach a maximum at a leaf area index of 3 when

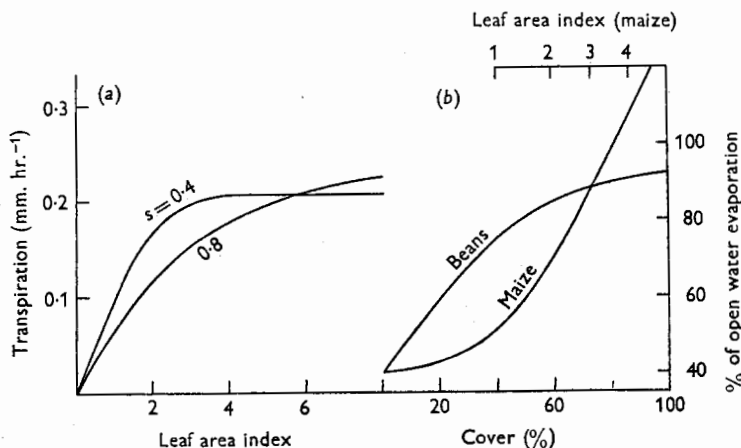


Fig. 9. (a) Calculated transpiration for crops with $s = 0.4$ and 0.8 , as function of leaf area index; (b) measured evaporation (soil and plants) as a fraction of open water evaporation plotted as function of visual estimate of cover or of leaf area index. (After Cackett & Metelerkamp, 1963, 1964.)

$s = 0.4$ but increase beyond an index of 6 when $s = 0.8$. The right-hand side of the figure shows comparable measurements by Cackett & Metelerkamp (1963, 1964) on beans and maize. The two curves may reflect differences in leaf arrangement between beans (small s) and maize (large s). In particular, an increase in the rate of transpiration from maize beyond a leaf area index of 4 is consistent with the theoretical prediction for a crop with $s = 0.8$. The calculations for Fig. 9a assume that evaporation from the soil is negligible. If the soil were wet, the total rate of evaporation would be larger and would increase less rapidly with leaf area index.

POTENTIAL TRANSPIRATION

Much of the evidence reviewed by Penman (1963) supports his hypothesis that vegetation with its roots in moist soil and with leaves forming a complete canopy transpires at a maximum or potential rate determined primarily by weather. From discussion in the last section, a canopy is 'complete' when the transpiration of its lowest leaves is restricted by stomatal closure. When leaves are predominantly horizontal, the minimum leaf area needed to form a complete canopy may be somewhat less than the minimum area at which the soil appears to be fully shaded (Fig. 9). Cereals and other crops with leaves more nearly vertical may not transpire at their maximum rate until the leaf area exceeds the minimum for full shading. Results in Table 2 imply that farm crops satisfying the conditions for potential transpiration have a minimum surface resistance between 0.3 and 0.5 sec. cm.⁻¹. On the slender evidence of one set of observations, the minimum resistance of a pine forest is about 0.9 sec. cm.⁻¹.

Sets of resistances were chosen to estimate the seasonal variation of potential transpiration from three contrasting surfaces in two climates: very short grass, a tall farm crop, and a pine forest, in south-east England and in California. Table 3 shows the choice of roughness length, surface resistance and reflectivity. Rationalizing differences between surfaces, the roughness length increases by a factor of 50 from short grass to tall crop, and by a further factor of 50 from crop to forest. The surface resistance for the forest is twice the resistance for grass and crop. Heat stored in the soil and in the products of photosynthesis was assumed to be 10% of net radiation so that H is 90% of net radiation. Differences in net radiation absorbed by pine forests and by short, green vegetation were calculated from reflexion coefficients for solar radiation given by Ångström (1925).

Assuming that weather at London Airport and Kew Observatory is representative of the Thames Valley, records were taken from the Monthly Weather Report for 1963, a year in which the distribution of solar radiation during summer was close to the long term average for the decade 1954-63. Net radiation absorbed during daylight was estimated empirically from solar radiation and cloudiness (Monteith, 1965). For the much drier climate of the Sacramento Valley in northern California, corresponding records from a U.S. Weather Bureau Summary were available for 1961 and radiation figures for Davis were kindly supplied by W. O. Pruitt and H. B. Schultz. Fig. 10 shows the seasonal variation of monthly transpiration in the Thames Valley and annual totals are in Table 3.

Relative transpiration rates depend on the relative values of r_s , r_a and r_i . Moving from the short grass to the tall crop, r_a decreases (because z_0

increases) whereas r_s and surface resistance is less transpiration rate increases with crop to the forest, there is reflectivity decreases) and only slightly smaller than

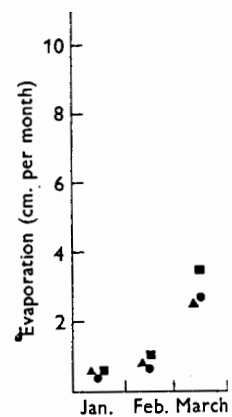


Fig. 10. Estimated annual variation of monthly transpiration in the Thames Valley. Some symbols are displaced horizontally. □, short grass; ○, tall crop; △, pine forest.

Table 3. Assumed

	reflectivity %	z_0 (cm.)
Short grass	25	0.1
Tall crop	25	5.0
Pine forest	15	250
Short grass	25	0.1
Tall crop	25	5.0
Pine forest	15	250

250 cm. increases the transpiration rate. Reflectivity *per se* increases the transpiration rate of surface resistance more than the increase in roughness and makes transpiration of a tall crop and only slightly greater than that of a pine forest.

The estimate predicts that a year will use about 95% of the potential transpiration rate of a pine forest.

RATION

1963) supports his hypothesis with leaves forming a canopy. The transpiration rate determined previously in the last section, a canopy is formed. The transpiration from the lowest leaves is restricted by the surface resistance. If the leaves are nearly horizontal, the minimum transpiration rate may be somewhat less than the rate from a fully shaded (Fig. 9) surface. The surface resistance may not transpire as much as needed. The minimum for full canopy is satisfied by the conditions of surface resistance between 0.3 and 1.0 sec. cm.⁻¹.

The seasonal variation of transpiration from various surfaces in two climates: forest, in south-east England and from short grass to tall crop, in California. The surface resistance for forest is assumed to be 10% of net radiation. Differences in net radiation between surfaces were calculated from the monthly distribution of solar radiation given by Ångström (1925). The data for Kew Observatory were taken from the Monthly distribution of solar radiation for the decade 1954-63, estimated empirically from 1965). For the much drier climate in California, corresponding data were available for 1961 and 1962, and applied by W. O. Pruitt and others (1965). The monthly transpiration rates are given in Table 3.

The relative values of r_s , r_a and r_i are given in Table 3. r_a decreases (because z_0

increases) whereas r_s and r_i stay constant in a given climate. Because the surface resistance is less than the critical value $(1 + \gamma/\Delta) r_i$ (p. 222), transpiration rate increases with increasing roughness. Moving from the tall crop to the forest, there is a further decrease of r_a , a decrease of r_i (because reflectivity decreases) and an increase of r_s . The stomatal resistance is now only slightly smaller than $(1 + \gamma/\Delta) r_i$ so the increase of roughness from 5 to

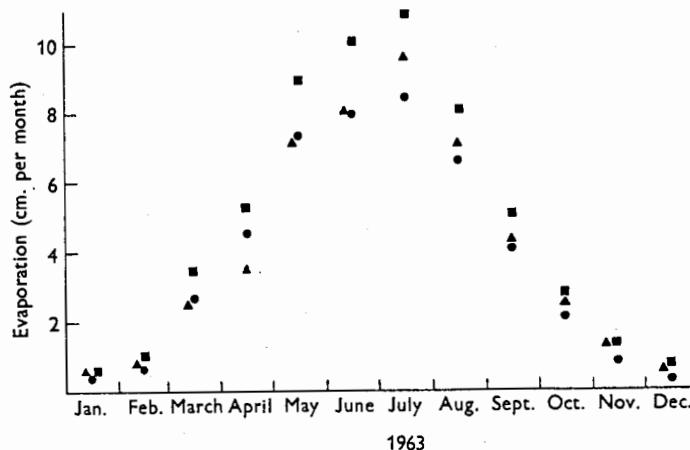


Fig. 10. Estimated annual variation of potential transpiration in Thames Valley, 1963. Some symbols are displaced horizontally to avoid overlapping. ●, Short grass; ■, tall crop; ▲, pine forest.

Table 3. Assumed crop parameters and annual transpiration

	reflectivity %	z_0 (cm.)	r_i (sec. cm. ⁻¹)	r_a (sec. cm. ⁻¹)	r_s (sec. cm. ⁻¹)	Δ (γ)	E (cm. yr. ⁻¹)	$\lambda E/H$ (%)
Thames Valley								
Short grass	25	0.1	0.8	1.1	0.5	1.3	47	76
Tall crop	25	5.0	0.8	0.36	0.5	1.3	58	95
Pine forest	15	250	0.7	0.025	1.0	1.3	48	69
Sacramento Valley								
Short grass	25	0.1	1.1	1.1	0.5	2.4	148	96
Tall crop	25	5.0	1.1	0.36	0.5	2.4	187	117
Pine forest	15	250	0.9	0.025	1.0	2.4	182	100

250 cm. increases the transpiration rate very little. The decrease of reflectivity *per se* increases transpiration by about 12%. Finally, the increase of surface resistance more than outweighs the changes in reflectivity and roughness and makes transpiration from the forest less than from a tall crop and only slightly greater than from short grass.

The estimate predicts that, in the Thames Valley, vegetation with the roughness of a tall crop and with access to soil moisture throughout the year will use about 95% of available heat to transpire water. This pre-

diction is supported by measurements during summer at Rothamsted, where net radiation was close to the latent heat of transpiration from cereals, sugar beet, and beans. Elsewhere, net radiation was within 20% of latent heat losses from grass (House, Rider & Tugwell, 1960), irrigated alfalfa-brome (Tanner & Lemon, 1962), Bermuda grass (*Cynodon* sp.: van Bavel & Harris, 1962), maize (Graham & King, 1961), and rice (Uchijima, 1961). By inference from Fig. 7, all these crops have a similar minimum surface resistance in the range 0.2–0.5 sec. cm.⁻¹ and, as a first approximation to guide the application of water by irrigation, the energy for potential transpiration is equal to the net income of radiation. Environments in which this rule breaks down are examined in a final section.

ADVECTION

When air with a water vapour content χ g. cm.⁻³ moves horizontally with wind speed u cm. sec.⁻¹, water vapour is transported at a rate $u\chi$ g. cm.⁻² sec.⁻¹. This transport is an 'advective' flow, present throughout the

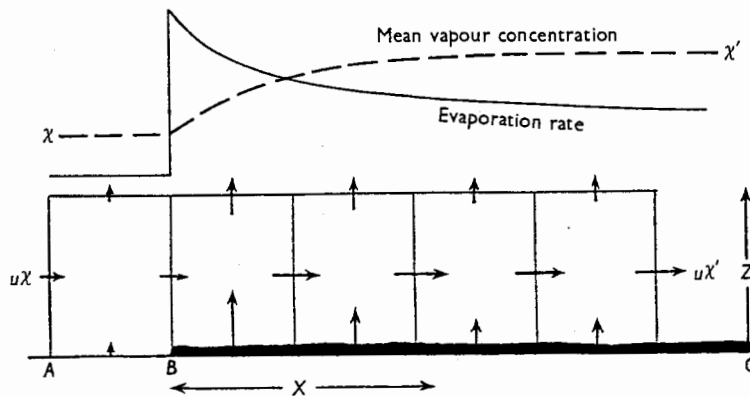


Fig. 11. Diagram representing the advection of water vapour and change in evaporation rate when air blows from a relatively dry surface AB to a moist surface BC . Fluxes of water vapour are proportional to lengths of arrows.

atmosphere because u and χ are always finite. In most meteorological estimates of evaporation, the vertical flux of water vapour a few metres above the surface is assumed equal to the flux at the surface. This assumption is not valid when $u\chi$ changes horizontally. For example, when air moves from an underlying dry surface to a moister surface, the concentration of water vapour increases and the increase in the horizontal flux of water vapour is often loosely referred to as 'local advection'. In Fig. 11, AB is a section of dry surface (e.g. bare soil) and BC is a wet surface (e.g. an irrigated crop). The atmosphere is represented in two dimensions by a

layer of height Z divide water vapour are represent flux. To satisfy the conservation entering any vertical section arrows.

The box on the extreme surface where χ is the mean surface and height Z . Here equal to the flux leaving the surface therefore, be equal to the air begins to move over the surface so χ increases and the horizontal flux in the box show the increase of water vapour, the vertical flux rate at the surface. Motion rate tends to decrease and is re-established when the new concentration χ' . In the fifth box, the flux at Z is once more equal to the flux at the surface.

The maximum height Z of the boundary layer is the depth of the boundary layer. The maximum height Z of the boundary layer is the depth of the boundary layer. Pruitt (1962) suggest that for a neutral atmosphere 95% of the evaporation is from the boundary layer.

Local rates of evaporation are determined by local radiation and from temperature. Penman's equation (13) for the boundary layer vapour pressure profile is (1964) found that the mean evaporation rate was only 73% of the evaporation rate from the irrigated lucerne in Utah.

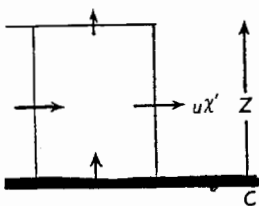
The leaves of an irrigated crop are cooler than the air because the available energy from net radiation is used for the downward transfer of sensible heat. This is one of the effects of the cause of anomalous evaporation between heat balance and evaporation.

summer at Rothamsted, heat of transpiration from radiation was within 20% (Tugwell, 1960), irrigated da grass (*Cynodon* sp.: van 1961), and rice (Uchijima, s have a similar minimum and, as a first approximation, the energy for potential diation. Environments in final section.

χ moves horizontally with speed u at a rate $u\chi$ g. cm.⁻², present throughout the

concentration χ

evaporation rate



vapour and change in evaporation rate over a moist surface BC. Fluxes of

3. In most meteorological conditions, water vapour a few metres above the surface. This assumption is valid. For example, when air passes over a moister surface, the concentration in the horizontal flux of water vapour 'advected'. In Fig. 11, the surface BC is a wet surface (e.g. shown in two dimensions by a

layer of height Z divided into arbitrary vertical sections, and fluxes of water vapour are represented by arrows whose length is proportional to flux. To satisfy the conservation of water vapour, the total length of arrows entering any vertical section or box must equal the total length of departing arrows.

The box on the extreme left shows a state of equilibrium over the dry surface where χ is the mean concentration of water vapour between the surface and height Z . Here, the advective flux $u\chi$ entering from the left is equal to the flux leaving on the right and the vertical flux at height Z must, therefore, be equal to the rate of evaporation from the surface. When the air begins to move over the wetter surface, the evaporation rate increases, so χ increases and the horizontal arrows entering and leaving the second box show the increase of advective flux. To satisfy the conservation of vapour, the vertical flux at height Z must now be less than the evaporation rate at the surface. Moving towards the right of the diagram, the evaporation rate tends to decrease as the air becomes moister and equilibrium is re-established when the mean concentration of vapour reaches a new value χ' . In the fifth box, the horizontal flux is constant at $u\chi'$ and the vertical flux at Z is once more equal to the rate of evaporation at the surface.

The maximum height Z' at which the vertical flux is equal to the evaporation rate is the depth of a boundary layer characteristic of the surface and increasing with downwind distance X . The relationship between Z' and X has not yet been established by experiment but measurements by Dyer & Pruitt (1962) suggest that Z'/X is much less than $1/50$. Theoretical analysis for a neutral atmosphere (Dyer, 1962) predicts that the vertical flux will be 95% of the evaporation rate at a height $Z = X/500$.

Local rates of evaporation at the surface can be estimated from net radiation and from temperature, humidity, and wind speed measured *within* the boundary layer. When weather is measured *above* this layer, Penman's equation (13) underestimates evaporation rates because the vapour pressure profile is not logarithmic. Abdel-Aziz, Taylor & Ashcroft (1964) found that the mean rate of transpiration calculated from equation (13) was only 73% of the actual rate of transpiration from a field of irrigated lucerne in Utah.

The leaves of an irrigated crop surrounded by dry land are often cooler than the air because the latent heat of transpiration exceeds the energy available from net radiation and the deficit of energy is made good by a downward transfer of sensible heat from warmer air to cooler surface. This is one of the *effects* of the advection of dry air over a wet surface, not a *cause* of anomalous evaporation rates as some writers imply. The relation between heat balance and advection was well illustrated by measurements

of transpiration from a field of irrigated Bermuda grass 15 m. square. When the surrounding land was wetted by rain, latent heat of evaporation λE was about 0.8 of net radiation but, during 12 rainless days, the transpiration rate increased as the surroundings became drier until λE was almost twice the net radiation. On a much larger scale, daily values of λE for cotton were 1.75 times net radiation although the upwind fetch over a uniformly cropped and irrigated area exceeded 10 miles. Both these experiments were quoted by Tanner & Lemon (1962).

Downwind from the boundary of a wet area, λE decreases as the air becomes cooler and moister. The rate of decrease was predicted by Philip (1959), who solved equations for the diffusion of heat and water vapour from a wet surface ($r_s = 0$). In a numerical example (Rider & Philip, 1960), λE was 1.24, 0.98, 0.78 and 0.42 times net radiation at downwind distances of 1, 10, 10^2 and 10^5 m. Rider, Philip & Bradley (1963) measured changes of χ with height (to 1.5 m.) and with distance (to 18 m.) over irrigated grassland when the upwind fetch over dry tarmac was 350 m. Measuring the divergence of horizontal flux, they found that the ratio of latent heat to the net supply of radiation decreased from 3.7 averaged over the first metre of grass to 1.7 averaged over the first 16 m., but these figures were not compared directly with estimates from theory.

A surprisingly small change of transpiration rate was reported for irrigated clover by Millar (1964). Downwind from the edge of the field, the transpiration rate, measured with a set of lysimeters, decreased by 10% from 14 to 470 cm., and was constant from 5 to 14 m. At mid-day, plants near the boundary began to wilt and the turgidity of leaves increased downwind to 14 m. at least. Other workers have shown that transpiration is often restricted by the rate at which plant roots extract moisture from the soil. Even when soil moisture tension is only 1 atm., the surface resistance of grass and maize begins to increase when the rate of transpiration exceeds 4 mm. day⁻¹ (Makkink & van Heemst, 1956; Denmead & Shaw, 1962). Associated physiological symptoms are loss of leaf turgidity, stomatal closure, and possibly some drying of cell walls. When crops are grown in an arid environment, transpiration rates often exceed 4 mm. day⁻¹ and surface resistance increases with transpiration rate as shown in Fig. 4. The correlation of transpiration rate and leaf turgidity in Miller's experiment suggests that the downwind change of evaporation rate expected for a wet surface was largely offset by the downwind change of surface resistance. Progress here needs a technique for measuring leaf resistances directly and a modification of theory to include a surface resistance that changes with transpiration rate.

- ABDEL-AZIS, M. H., TAYLOR, ÅNGSTRÖM, A. (1925). *Geogr*
 BANGE, G. G. J. (1953). *Acta*
 BAUMGARTNER, A. (1956). *B*
 BAVEL, C. H. M. VAN & HARI
 BRUNT, D. (1939). *Physical*
 Press.
 BUSINGER, J. A. (1956). *Neth*
 CACKETT, K. E. & METELERK
 CACKETT, K. E. & METELERK
 COVEY, W. G. (1959). Ph.I
 Texas.
 DENMEAD, O. T. & SHAW, R.
 DYER, A. J. (1962). Second A
 SC-80334, University of
 DYER, A. J. & PRUITT, W. C
 Contract DA-36-039-SC
 EHLIG, C. F. & GARDNER, W
 EKERN, P. C. (1964). Cyclo
 Institute, Hawaii.
 FISHENDEN, M. & SAUNDER
 University Press.
 FRANKENBERGER, E. (1960).
 FRITSCHEN, L. J. & VAN BAVE
 GAASTRA, P. (1959). *Meded.*
 GRAHAM, W. G. & KING, K.
 HALSTEAD, M. H. (1954). *Jol*
 326.
 HOUSE, G. J., RIDER, N. E. 8
 215.
 KANESEKI, S. (1956). *J. agric*
 KUIPER, P. J. C. (1961). *Me*
 LADEFOGED, K. (1963). *Phys*
 LETTAU, H. H. & DAVIDSON,
 vol. 2. London: Pergam
 LINACRE, E. T., PALMER, J. F
 LONG, I. F., MONTEITH, J. L
 17, 97.
 MCLROY, I. C. & ANGUS, D.
 McMILLAN, W. D. & BURG
 MAKKINK, G. G. & VAN HEE
 MARTIN, E. (1943). *Publ. Ca*
 MELLOR, R. S., SALISBURY, F
 MILLAR, B. D. (1964). *Aust.*
 MILTHORPE, F. L. (1962). In
 MONTEITH, J. L. (1963a). In
 York: Academic Press.
 MONTEITH, J. L. (1963b).
 DA-36-039-SC-80334, U
 MONTEITH, J. L. (1965). *Am*
 MONTEITH, J. L. (1965). *We*

REFERENCES

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- ABDEL-AZIS, M. H., TAYLOR, S. A. & ASHCROFT, G. L. (1964). *Agron. J.* **56**, 139.
- ÅNGSTRÖM, A. (1925). *Geogr. Ann., Stockh.*, **7**, 323.
- BANGE, G. G. J. (1953). *Acta bot. neerl.* **2**, 255.
- BAUMGARTNER, A. (1956). *Ber. dtsh. Wetterdienstes*, Nr. 28, Bd. 5.
- BAVEL, C. H. M. VAN & HARRIS, D. G. (1962). *Agron. J.* **54**, 319.
- BRUNT, D. (1939). *Physical and Dynamical Meteorology*. Cambridge University Press.
- BUSINGER, J. A. (1956). *Neth. J. agric. Sci.* **4**, 77.
- CACKETT, K. E. & METELERKAMP, H. R. R. (1963). *Rhod. J. agric. Res.* **1**, 18.
- CACKETT, K. E. & METELERKAMP, H. R. R. (1964). *Rhod. J. agric. Res.* **2**, 35.
- COVEY, W. G. (1959). Ph.D. Thesis, Agricultural and Mechanical College of Texas.
- DENMEAD, O. T. & SHAW, R. H. (1962). *Agron. J.* **45**, 385.
- DYER, A. J. (1962). Second Annual Report under USAEPG Contract DA-36-039-SC-80334, University of California, Davis.
- DYER, A. J. & PRUITT, W. O. (1962). Second Annual Report under USAEPG Contract DA-36-039-SC-80334, University of California, Davis.
- EHLIG, C. F. & GARDNER, W. R. (1964). *Agron. J.* **56**, 127.
- EKERN, P. C. (1964). Cyclostyled Research Report of the Pineapple Research Institute, Hawaii.
- FISHENDEN, M. & SAUNDERS, O. A. (1957). *Heat Transfer*. London: Oxford University Press.
- FRANKENBERGER, E. (1960). *Ber. dtsh. Wetterdienstes*, Nr. 73, Bd. 10.
- FRITSCHEN, L. J. & VAN BAVEL, C. H. M. (1964). *Agron. J.* **56**, 201.
- GAASTRA, P. (1959). *Meded. LandbHoogesch., Wageningen*, **59**, 1.
- GRAHAM, W. G. & KING, K. M. (1961). *Proc. Soil Sci. Soc. Amer.* **25**, 158.
- HALSTEAD, M. H. (1954). *Johns Hopkins University, Publications in Climatology*, **7**, 326.
- HOUSE, G. J., RIDER, N. E. & TUGWELL, C. P. (1960). *Quart. J. R. met. Soc.* **86**, 215.
- KANESEKI, S. (1956). *J. agric. Met., Tokyo*, **12**, 30.
- KUIPER, P. J. C. (1961). *Meded. LandbHoogesch., Wageningen*, **61**, 1.
- LADEFOGED, K. (1963). *Physiol. Plant.* **16**, 378.
- LETTAU, H. H. & DAVIDSON, B. (1957). *Exploring the Atmosphere's First Mile*, vol. 2. London: Pergamon Press.
- LINACRE, E. T., PALMER, J. H. & TRICKETT, E. S. (1964). *Agric. Met.* **1**, 165.
- LONG, I. F., MONTEITH, J. L., PENMAN, H. L. & SZEICZ, G. (1964). *Met. Rdsch.* **17**, 97.
- MCLROY, I. C. & ANGUS, D. E. (1964). *Agric. Met.* **1**, 201.
- MCMILLAN, W. D. & BURG, R. H. (1960). *J. geophys. Res.* **65**, 2389.
- MAKKINK, G. G. & VAN HEEMST, H. D. J. (1956). *Neth. J. agric. Sci.* **4**, 1.
- MARTIN, E. (1943). *Publ. Carneg. Instn.* **550**, 1.
- MELLOR, R. S., SALISBURY, F. B. & RASCHKE, K. (1964). *Planta*, **61**, 56.
- MILLAR, B. D. (1964). *Aust. J. agric. Res.* **15**, 85.
- MILTHORPE, F. L. (1962). In *Arid Zone Research*, vol. 16. Paris: UNESCO.
- MONTEITH, J. L. (1963a). In *Environmental Control of Plant Growth*, p. 95. New York: Academic Press.
- MONTEITH, J. L. (1963b). Second Annual Report under USAEPG Contract DA-36-039-SC-80334, University of California, Davis.
- MONTEITH, J. L. (1965). *Ann. Bot., Lond.* (in the Press).
- MONTEITH, J. L. (1965). *Weather* (in preparation).

- MONTEITH, J. L., SZEICZ, G. & YABUKI, K. (1964). *J. appl. Ecol.* **1**, 321.
- PASQUILL, F. (1950). Air Ministry, Met. Res. Cttee, M.R.P. 579.
- PENMAN, H. L. (1948). *Proc. roy. Soc. A*, **202**, 143.
- PENMAN, H. L. (1953). *Rep. 13th int. hort. Congr.* **2**, 913.
- PENMAN, H. L. (1963). *Vegetation and Hydrology*. Farnham Royal, England: Technical Communication No. 53, Commonwealth Bureau of Soils.
- PENMAN, H. L. & LONG, I. F. (1960). *Quart. J. R. met. Soc.* **86**, 16.
- PENMAN, H. L. & SCHOFIELD, R. K. (1951). *Symp. Soc. exp. Biol.* **5**, 115.
- PHILIP, J. R. (1959). *J. Met.* **16**, 535.
- POWELL, R. W. (1942). *Trans. Instn chem. Engrs, Lond.*, **18**, 136.
- PRIESTLEY, C. H. B. (1959). *Turbulent Transfer in the Lower Atmosphere*. Chicago: University of Chicago Press.
- RASCHKE, K. (1956). *Arch. Met., Wien*, **7B**, 240.
- RIDER, N. E. (1954). *Phil. Trans. A*, **246**, 481.
- RIDER, N. E. & PHILIP, J. R. (1960). *Ass. Int. de Hydrologique Scient.* **53**, 421.
- RIDER, N. E., PHILIP, J. R. & BRADLEY, E. F. (1963). *Quart. J. R. met. Soc.* **89**, 507.
- ROBINSON, G. D. & RIDER, N. E. (1951). *Quart. J. R. met. Soc.* **77**, 375.
- SAITO, T. (1962). *J. agric. Met., Tokyo*, **17**, 101.
- SLATYER, R. O. & BIERHUIZEN, J. F. (1964). *Aust. J. biol. Sci.* **17**, 115.
- SLATYER, R. O. & McILROY, I. C. (1961). *Practical Microclimatology*. Paris: UNESCO.
- STOLLER, J. & LEMON, E. R. (1963). U.S.D.A. Agricultural Res. Service, Production Research Report, no. 72.
- TANNER, C. B. (1963). In *Environmental Control of Plant Growth*, p. 141. New York: Academic Press.
- TANNER, C. B. & LEMON, E. R. (1962). *Agron. J.* **54**, 207.
- TANNER, C. B. & PELTON, W. L. (1960a). *J. geophys. Res.* **65**, 3391.
- TANNER, C. B. & PELTON, W. L. (1960b). *Soils Bulletin*, **2**. Department of Soils, University of Wisconsin, Madison, Wis.
- UCHIJIMA, Z. (1961). *Bull. nat. Inst. agric. Sci., Tokyo*, **A**, no. 8, 243.
- VRIES, D. A. DE & VENEMA, H. J. (1954). *Veg. acta Geobot.* **5**, 225.