Turbulence was concentrated around 35,000 ft, with an even more pronounced second maximum around 55,000 ft. The lower of these maxima has long been familiar from middle latitude work, but evidence for the upper has only recently been supplied by U2 flights (Reiter 1962) and is believed to arise from mesoscale disturbances extending over vertical distances of the order of 3,000 ft and horizontal distances of some hundreds of miles (Barbé 1958, Sawyer 1961).

A first test of turbulence parameters has been made for a sample of 14 flights with well-defined persistent layers, measuring 5,000 ft on the average. Although in individual cases large vertical shear values occurred at the base of turbulent layers, the average shear profile was fairly uniform. By contrast the static stability had a clear minimum near the centre of the turbulent layers where also the static stability appeared to decrease individually, due to cold air advection which is not compensated by subsidence. Individual Ri values of seven cases with strong turbulence were close to or even well below 1 near the disturbed layers.

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ROUTINE MEASUREMENT OF CROP EVAPORATION

by I. McIlroy

In introducing the speaker, the Chairman, Dr. C.H.B. Priestley, emphasized the far-reaching importance of evaporation measurement, not only for agriculture and hydrology but ultimately, on a world-wide network basis, in further development of forecasting methods.

Mr. I.C. McIlroy, of C.S.I.R.O. Division of Meteorological Physics, then discussed what he regarded as the most promising of a very limited number of approaches available for reliable routine estimation of evaporation, applicable to most natural surfaces but in particular to irrigated crops, with which it had already been very successful.

A rigorous derivation of this so-called combination method was then given. For space reasons this has had to be abbreviated here, both by the earlier introduction of certain assumptions and approximations necessary to achieve simplicity in the final working formula adopted for evaporation, and by the omission of any discussion of the errors introduced thereby.

The method combines two types of basic relationship, that of proportionality between the average rate of flow of a quantity and the associated concentration gradient and that of balance between the energy flows to and from a surface. The equations for vertical transfer through the atmosphere, of sensible heat and of the latent heat associated with evaporation respectively, can be written as:

$$H = h \ \Delta T \qquad \ldots(1)$$

and

$$LE = L \frac{h}{c_p} \Delta q$$

$$\simeq \frac{h}{\gamma} \Delta q \qquad \ldots(2)$$

where H and E are the upward fluxes of sensible heat and of water vapour, and L is the latent heat of evaporation, while ΔT and Δq are the corresponding differences of temperature and specific humidity, from the earth's effective boundary to a standard small height above it. h is a heat transfer coefficient, or conductance applicable to the air layer concerned, while $\frac{h}{c_p}$ and $\frac{h}{\gamma}$ (with c_p the specific heat of air at constant pressure, and $\gamma \simeq c_p/L$ the psychrometric constant in terms of specific humidity) are corresponding conductances for water vapour and latent heat respectively – on the assumption that near the ground the mode of transfer of heat is similar to that of vapour.

Energy balance at the earth's boundary can be expressed as

$$R = G + H + LE \qquad \ldots(3)$$

where R is the net radiation received at the surface and G is the heat flux into the ground (including that used in warming any vegetation present).

Combining these equations with an expansion of Δq in terms of the corresponding ΔT_w and ΔD, where T_w is wet bulb temperature and D (= $T - T_w$) is wet bulb depression, gives rise to a simple but generally quite accurate expression for latent heat flux.

$$LE = \frac{s}{s + \gamma} (R - G) + h (D - D_o) \qquad \ldots(4)$$

where s is nearly enough the slope of the saturation specific humidity curve, strictly at average wet bulb temperature but generally taken at dry bulb temperature at the same height as D, i.e. at the top of the layer; while D_o is the wet bulb depression at zero height, i.e. at the bottom of the layer.

All the quantities involved here are directly and easily measurable except h and D_o. Where wet bulb depression itself is not measured, its place in (4) can be taken by an equivalent expression in terms of temperature and some other index of atmospheric dryness, such as vapour pressure, relative humidity or dewpoint, depending on the type of instrument available.

In the case of h, which is strictly a complex function of windspeed, surface roughness and atmospheric stability, we are compelled in our present state of knowledge to use an empirical approximation,

$$h = a (b + u) \qquad \ldots(5)$$

where u is windspeed measured at some standard height, such as 0.5 m above the crop, for convenience the same as that at which T and D are also measured; while a and b are constants, to be determined initially for each new crop and soil type, by means of a sufficient period of comparison with an independent measure of LE, i.e. from a lysimeter temporarily installed for that purpose.

In practice D_o is sometimes small enough to neglect – always for free-water, or fully wet bare soil or plant cover, and quite often for only moderately well-watered vegetation. We then speak of potential evaporation, i.e. the rate obtaining from any given surface under conditions of completely unrestricted water supply. If we further limit outselves to daily or longer averages from soil or vegetated surfaces, G can also be neglected, at the cost of some increase in scatter. This gives a very simple working formula for average potential evaporation, E_p:

$$LE_p = \frac{s}{s + \gamma} R + h D \qquad \ldots (6)$$

However, where sufficient drying out of the soil occurs between rainfalls or irrigations, the leaves of a crop cease to behave at all like saturated surfaces, and D_o can become highly significant. Nevertheless, even in this case a simple and apparently reliable working formula can still be obtained, by using certain further assumptions to relate D_o to a more readily determinable soil parameter.

The first two assumptions are very reasonable, namely that temperature gradients within the leaves are small and that the internal surfaces from which the vapour is actually emitted remain effectively saturated. D_o can then be shown to vary inversely with the internal conductance, h_i, of the vapour diffusion path from these surfaces to the air just outside the leaves. It will also depend directly on the evaporation rate itself, and indirectly on the level of temperature.

The resulting overall formula for actual evaporation, E_a, as distinct from potential evaporation, becomes

$$E_a = \frac{E_p}{1 + \frac{\gamma}{s + \gamma} ({}^{h}/h_i)} \qquad \ldots (7)$$

where $\frac{\gamma}{s + \gamma}$ is strictly to be taken at leaf temperature, but in practice air temperature appears near enough; while h_i is itself no more readily determinable than D_o, at least as yet. However, with most species, average h_i tends to be governed to a large extent by the plants themselves, by varying the degree of closure of their stomata so as to prevent water loss from the leaves at a rate much greater than can be supplied by the roots. Thus the third assumption, which is by no means clear-cut but seems to have worked well so far, is that average h_i responds essentially to the average soil moisture content, M, in the root zone of the vegetation concerned.

The latter can be measured from time to time, by a variety of techniques, and in between can be estimated well enough by book-keeping methods – subtracting each day's calculated moisture loss by evaporation and adding in each rainfall or irrigation (less runoff or deep drainage). Although the relationship between h_i and M is undoubtedly complex, once again it suffices to use a linear approximation,

$$h_i = c\,(M - M_c) \qquad \ldots (8)$$

where c and M_c are constants for a given soil and rooting behaviour, with M_c representing a critical moisture content at which the rate of extraction by the plant has become unimportant from the practical point of view.

Mr. McIlroy presented a number of results from lysimeter comparisons, with a relatively infrequently irrigated lucerne crop at Tempe, Arizona, showing that equations (6) and (5), with properly determined constants in (5), together represent very satisfactorily the crop evaporation under potential conditions, which applied throughout most of the season; and that furthermore, when on two occasions drying out became severe enough for significant departures from the potential rate, the actual rate was well described by equations (7) and (8). This was particularly significant for the second drying cycle, since the constants used in (8) had been determined from data for the first cycle only.

These results were supported by similar ones for pasture at Aspendale, and elsewhere. In conclusion Mr. McIlroy recommended that authorities concerned with evaporation give immediate and serious consideration to the setting up of "combination method" stations. At present the constants a and b in equation (5) and, where growing conditions warrant this, c and M_c in (8), need to be determined separately for each new crop and soil. However, after a certain amount of spade-work is done this requirement should ease; and in any case, once each initial calibration is completed the lysimeter or other reference equipment can be moved on, while for routine operation thereafter the combination method itself requires only four simple measurements – of net radiation, windspeed (for determining h), temperature (for determining $\frac{s}{s+\gamma}$, from tables), and any convenient atmospheric moisture parameter – plus in some cases an occasional check of soil moisture content.

The talk was illustrated by several slides showing equipment used for measurement of elements in equation (4) in a field of cotton.

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