GEOG 321 - Reading Package Lectures 29-31

WATER BALANCE OF LAND SURFACES

We may write the water balance of a land-atmosphere interface:

$$p = E + f + \Delta r + \Delta S \qquad \star \tag{20.1}$$

where p is *precipitation*, created by either to cloud droplets or crystals that grow to a size where they can no longer be held in suspension and they fall to the Earth as rain, snow or hail. Near the surface water may also be deposited by direct condensation or sublimation as dew, hoar frost, and rime, or be impacted as fog-drip. Over land areas p is typically greater than E.

Utilizing energy provided by the energy balance water is *evaporated* from open water and the soil, and is *transpired* from vegetation. The composite loss of water to the air from all sources is called the *evapotranspiration* (E). Evapotranspiration consists of evaporation of free surface water (e.g. puddles), and soil pore water, and water transpired from vegetation.

f is the *infiltration* to deeper soil layers. Normally f is positive due to gravity. Although less common f can be negative, for example, where a ground water table intersects a hillslope leading to spring seepage. Infiltration is not easily determined so for practical purposes it is better to consider a column which extends from the surface to a depth where vertical moisture movement is absent (i.e. where $f \to 0$)

Δr is *net run-off* (i.e. the net change in runoff over a distance). This term may have a positive or negative sign. It is positive if more water leaves than arrives as is normally the case on sloping land. Δr is negative when surface flow leads to accumulation of water such as when lake levels rise.

ΔS is the net change in *soil moisture* content. Equation 1.19 shows how the water storage in the system is dependent upon the water input which is usually mainly p, and the water output via E and Δr. Input could also be supplied by irrigation which would require an additional term on the left-hand side of equation 20.1.

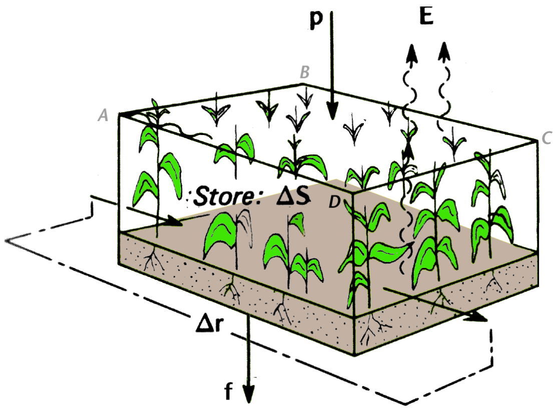

Figure 1: Schematic depiction of fluxes involved in the water balance of a soil-plant-air volume.

If advection is present the equation should include ΔA to allow for net horizontal moisture exchange.

The time scale over which equation 20.1 is valid proves awkward if we are properly to integrate it with the surface energy balance on time periods of a day or less. This arises because the input/output processes are fundamentally different in nature. Precipitation usually occurs in discrete, short-period bursts, whereas evaporation is a continuous and variable function. Thus, for example, for a flat surface, during periods with no precipitation water input is zero but the soil moisture store is being almost continually depleted by evapotranspiration. In these circumstances equation 20.1 effectively reduces to:

$$E = \Delta S \tag{20.2}$$

because Δr is negligible on level terrain. Therefore, unlike the annual situation where net water storage is zero, on the short time-scale ΔS is non-zero and very important. However, over relatively long periods (e.g. one week or more) all the terms can be evaluated to provide a budget estimate.

Further, on the spatial scale the water balance is commonly studied for a whole catchment basin, a lake or a glacier. Precipitation (p) is measured with standard rain gauges arranged in a suitable network for spatial sampling, and net runoff (Δr) by hydrologic stream gauging at the boundaries of the system. Therefore if either soil moisture change (S) or evaporation (E) are evaluated the budget is obtained, providing deep drainage to or from the system can be neglected. Soil moisture change can be measured by a regular programme of measurement using the simple gravimetric or tensiometer methods, or the neutron-scattering technique (see Lectures 11-13).

Connectivity of energy and water balance. The common term in the water and energy balance equations is evaporation. The fluxes of mass (E) and energy (Q_E) associated with evaporation are linked by the relation:

$$Q_E = L_v \, E \qquad \star \tag{20.3}$$

where the units of E are kg m^{-2} s^{-1} (mass transport through unit surface area in unit time). For hydrologic purposes it is convenient to express the mass flux density E in terms of an equivalent depth of water (millimetres) over the period of concern (usually an hour or a day). They are then consistent with the normal units of precipitation. A simple conversion of 1 mm evaporation = 2.45 $\mathrm{MJ\,m}^{-2}$ can be applied.

Condensation operates in the reverse sense by adding to both the mass and energy stores. Melting and freezing are energetically less significant, but still of importance especially in soil climate.

WATER BALANCE OF PLANT CANOPIES

The water balances equation outlined above refers to the vertical fluxes passing through a plane (ABCD Figure 1), including the storage and advection terms if their volumetric values are converted to fluxes per unit horizontal area. Such a simplification is adequate to make generalizations about a vegetation community, but it ignores exchanges within the volume and gives no insight into the internal workings of the stand climate.

Inside the canopy volume there are significant air, soil and plant flows leading to the re-distribution of water, and that there are a number of localized sites where temporary water storage occurs. Figure 1 is an attempt to schematize these features in the case of a plant stand where horizontal exchanges (in the air, along the ground, and within the soil) can be neglected. These conditions are met in the middle of an extensive plant community or crop growing on level terrain.

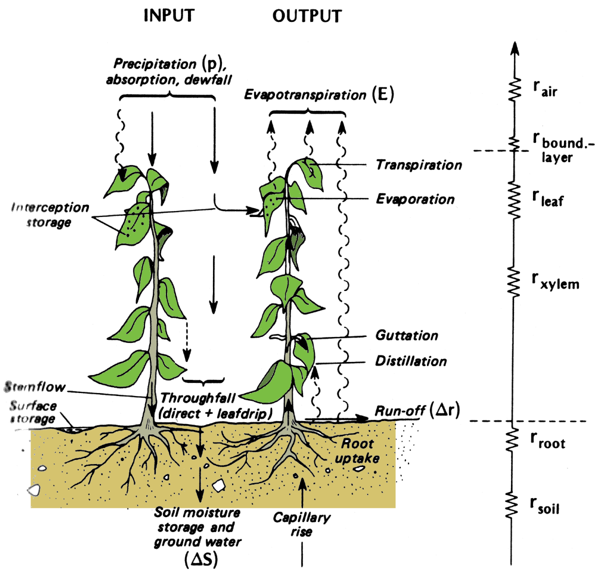

Figure 2: The water balance and internal flows of water in a soil-plant-atmosphere system. At the right is an electrical analogue of the flow of water from the soil moisture store to the atmospheric sink via the plant system.

Precipitation interception. In the absence of irrigation the primary water inputs to the system are rain, snow, fog, dew and frost from the atmosphere, and to a lesser extent soil water rising from below. Rain and snow entering the system is either intercepted by the foliage or falls directly through openings to the ground. The intercepted water contributes to a form of storage, which is further fed by the impaction of fog droplets, condensation (termed *dewfall* if the water comes from the air above, the *distillation* if it comes from the soil), and any plant water exuded onto the leaf surface through the leaf cuticle (a process known as *guttation*). The efficiency of precipitation interception depends both upon the nature and amount of the precipitation, and the vegetation characteristics such as the stand architecture, density and the area of foliage. For an initially dry canopy the interception efficiency is high in the early stages of rain or snowfall, or if the amount of input is small. Thus a high proportion of the water is retained by the canopy, but eventually a threshold storage capacity is surpassed and thereafter the efficiency declines. The excess water finds its way to the ground either as a result of leaf drip or by running down the stems.

Forests are able to retain a large proportion of their precipitation as interception storage. The canopy of a deciduous forest can intercept 10 - 25% of the total annual precipitation, and the range for coniferous forests is 15 40%.

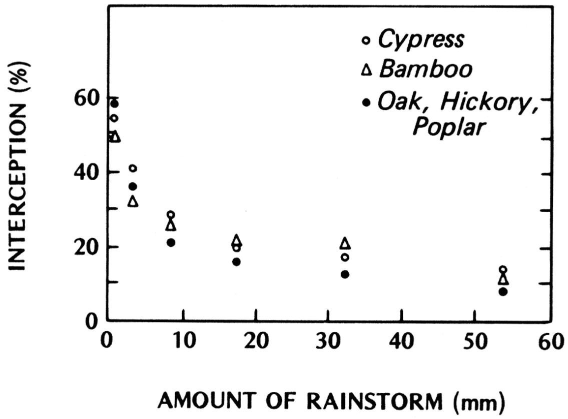

Figure 3: The relation between the rainfall interception efficiency of tropical and temperature forests and the amount of rain precipitated by a storm (after Pereira, 1973).

In general snowfall interception efficiency is similar to that for rainfall. For an individual storm the efficiency can be almost 100% in the early stages but when the maximum storage capacity is exceeded the efficiency drops. Figure 3 illustrates this fact for a range of tropical and temperate forests. It also shows that interception depends more upon the nature of the rainstorm than upon the tree species. The maximum storage capacity for rain is from 0.5 to 2 mm for all forest types, and from 2 to 6 mm for snow.

Interception is important in the water balance of forests because it is capable of 'short-circuiting' the availability of water to the ecosystem. The direct evaporation of intercepted water (rain, snow, dew or fog) can constitute the major portion of evapotranspiration (i.e. exceeding transpiration). This water therefore does not enter the soil moisture store for use by trees and understorey, nor is it released as runoff to streams. The importance of wetted-leaf losses depends on the rainfall climatology of the site including the nature of the storms and their seasonality.

Transport of water within the volume. The water incident upon the ground's surface (direct precipitation + leaf drip + stemflow) either infiltrates into the soil or remains as surface storage in the form of puddles. That which percolates downward enters the soil moisture store along with any deep soil water seeping upward by capillary action. Capillary rise can be substantial if the water table is not too deep and usually occurs as a response to moisture depletion in the upper layers brought about by the vegetative or soil evaporative demand. The root uptake to satisfy this demand can be seen as both the redistribution of moisture to the plant water store (supplemented by small amounts absorbed through leaves), and as the start of the process leading to transpiration.

The evaporative flux is sustained by depleting all of the four water stores (i.e. interception storage, ground surface storage, soil moisture storage and plant water storage). During the growing season the main loss is via the plant system through plant water movement and transpiration.

The rate of water flow through the system is a function of the vapour concentration difference between the air and the leaf, and the water potential difference between the leaf and the soil, and is regulated by a series of soil, plant and air resistances (Figure 2).

In the soil the flow is dependent upon the amount of water present since this determines the hydraulic conductivity of the soil (see reading package, Lecture 11-13) the inverse of which is the resistance offered by the soil to moisture extraction (r_{soil}). The plant uptake is governed by the extent of root development (r_{root}) and the ease of internal sap movement by the vascular system of the xylem (r_{xylem}). Diffusion within the leaf (r_{leaf}) includes the resistances of the mesophyll, the stomate (see below) and the cuticle and to the air involves the resistances of the boundary layer and the turbulent layers above (r_b and r_a). Water is also lost from the system by evaporation from the soil and from the exterior surfaces of the vegetation. The soil losses may either originate from surface puddles, or from soil layers near the surface. The exterior vegetation losses occur from surfaces wetted by precipitation interception, condensation and guttation.

PHOTOSYNTHESIS AND RESPIRATION

Plant growth is intimately tied to the supply of solar radiation (specifically in the PAR range) and carbon dioxide (CO_2) through the processes of photosynthesis and respiration.

Gross Photosynthesis:

$$\mathrm{CO_2 + H_20 + h\nu \rightarrow (CH_2O) + O_2} \qquad (20.4)$$

Respiration:

$$\mathrm{(CH_2O) + O_2 \rightarrow CO_2 + H_20 + Energy} \qquad (20.5)$$

Growth depends upon the excess of dry matter gain by the assimilation of CO_2 in gross photosynthesis (P), less the CO_2 loss via respiration (R_e). Thus the net rate of CO_2 assimilation, or the net rate of photosynthesis (ΔP) is:

$$\Delta P = P - R_e \tag{20.6}$$

each having the units of moles of CO_2 per unit area and unit time (mol $\mathrm{m^{-2}\ s^{-1}}$).

Direct energy exchange due to net photosynthesis. The rate at which heat is stored by net photosynthesis, ΔQ_P, is:

$$\Delta Q_P = \Delta P\,\phi \tag{20.7}$$

where ϕ - heat of assimilation of carbon, which is $469\,\mathrm{kJ\,mol^{-1}}$, or about $3.2\,\mathrm{W\,m^{-2}}$ per $\mathrm{g\,m^{-2}\,h^{-1}}$ of CO_2 assimilation.

Typically, by day P is greater than R and therefore ΔP is positive (i.e., a vegetated surfaces is a net CO_2 sink). Maximum values of ΔP for vegetated surfaces are dependent upon species but lie in the range from 2 to 5 $\mathrm{g\,CO_2\,m^{-2}\,h^{-1}}$, so that the largest values of ΔQ_P are typically 6 to 16 $\mathrm{W\,m^{-2}}$. At night R is unopposed by P (which requires solar radiation) so that ΔP is negative (i.e. the crop is a net CO_2 source), and since maximum crop respiration values are about 1 $\mathrm{g\,CO_2\,m^{-2}\,h^{-1}}$ it follows that the maximum nocturnal value of ΔQ_P is about -3 $\mathrm{W\,m^{-2}}$. In comparison with most other terms of the energy balance, ΔQ_P is so small that it is often neglected in energy balance considerations. Much more relevant in the surface energy balance of a land-atmosphere interface is the stomatal control of transpiration.

Stomatal control of transpiration. The daytime flux of CO_2 to the leaves is supplied by the atmosphere. The passageway between the atmosphere and the interior of the plant or tree is the leaf stomate (Figure 4). These pores on the leaf surface are open during the day to capture and expel CO_2, and in this position they also expose the moist interior of the stomate to the air. Vaporization of moisture (known as transpiration) is therefore an inevitable by-product of photosynthesis. Transpiration is an important process in its own right however, since the water loss induces moisture and nutrient movement through the plant or tree, and the associated uptake of latent heat is a major means of dissipating the energy load on leaves.

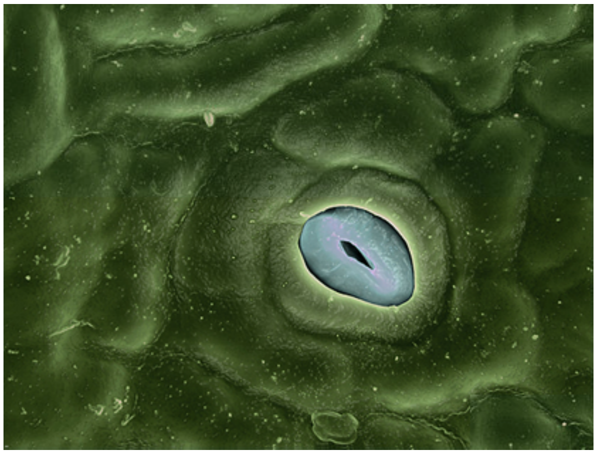

Figure 4: Partially open stomata on Ilex leaf (false color)

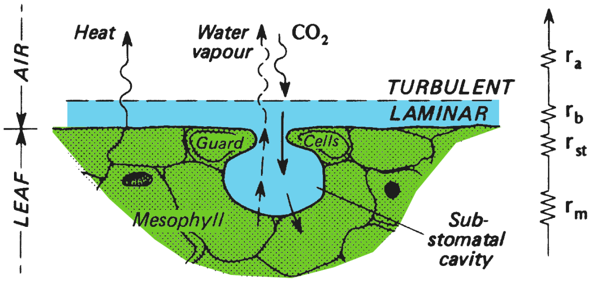

Figure 5: Schematic cross-section through a portion of a leaf illustrating the exchanges of water vapour and CO_2 through a stomate, and of heat from the leaf exterior.

Depending on species stomata are typically 10 to 30 μm in length, and vary in width from zero when closed to 10 μm when fully open. Their density ranges from 50 to 500 per $\mathrm{mm^2}$ of leaf surface, and when open their combined area represents 0.3 - 1% of the total leaf area. In some plants they occur on both leaf surfaces, in others only on the underside. The climatic significance of stomata is their ability to open and shut so that they act as regulatory 'valves' in the transfer of water vapour and CO_2 between plants and the atmosphere. At night the stomata are essentially closed, but with sunlight the *guard cells* controlling the stomatal aperture operate to open the pore. The degree of opening depends on many factors including the light intensity, the ambient temperature and humidity and the CO_2 concentration. Stomatal closure is tied to insufficient light intensity and/or loss of water content (reduction in turgor) by the guard cells. Hence anything producing plant water stress (e.g. excessive transpiration losses, depletion of soil moisture) will constrict or close the 'valves' and hinder the flow of gases. Stomatal activity therefore provides differing degrees of resistance to the exchange of water vapour and CO_2 between the plant and the atmosphere, and makes the plant an active agent in the determination of its climate.

EVALUATION OF EVAPOTRANSPIRATION

Bowen ratio-energy balance method. The energy balance approach to estimating evapotranspiration seeks to apportion the energy available ($Q^* - Q_G$, or $Q^* - \Delta Q_S$) between the sensible and latent heat terms by considering their ratio, β - the *Bowen ratio*. This method has the advantage of not being stability-limited because it only requires similarity between eddy diffusivities of sensible heat K_H and water vapour K_V, and not K_M as the aerodynamic approach (see Lecture 24). Since it can be shown that $K_H/K_V = \Phi H/\Phi V$ and that $\Phi H/\Phi$ for all stability regimes (see Lecture 26) it follows that K_H and K_V indeed are similar, and therefore:

$$\beta = \frac{Q_H}{Q_E} = \frac{C_a \Delta \overline{T}}{L_v \Delta \overline{\rho_v}} \tag{20.8}$$

From the surface energy balance (see Lecture 3):

$$Q^* = Q_H + Q_E + Q_G \tag{20.9}$$

The individual turbulent fluxes are given in terms of β can be written as:

$$Q_E = \frac{Q^* - Q_G}{1+\beta} \tag{20.10}$$

and

$$Q_H = \frac{\beta(Q^* - Q_G)}{1+\beta} \tag{20.11}$$

From Q_E, E can be easily calculated (Eq.20.3) Therefore to evaluate evaportraspiration over an extensive surface all that is required are accurate measurements of Q^* (net pyrradiometer), Q_G (soil heat flux plate) and β from temperature and humidity differences over the same height interval.

Combination Model. The Combination Model approach to estimating evapotranspiration combines aerodynamic *and* energy balance principles into an equation which includes the energy availability, the drying power of the air and the state of turbulence, in a form that only requires relatively easily observed variables.

The approach rests upon a simple but very useful linearization of the temperature vs saturation vapour concentration curve attributed to Howard Penman. Take a small part of the curve, as illustrated in Figure 6; the vapour difference driving evaporation from a wet surface is that between the saturation value at the surface temperature and the actual value in the air, i.e. $(\rho^*_{v0} - \rho_{va})$.

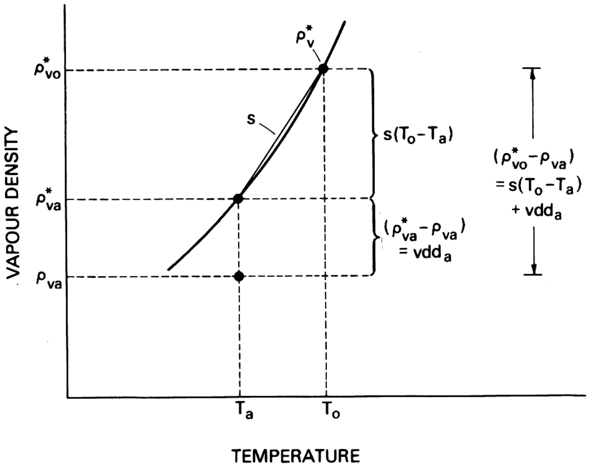

Figure 6: Saturation vapour density vs temperature diagram showing Penman's transformation.

From Figure 6 we see this to be the sum of two parts. The first is the difference between the saturation vapour densities of the surface at temperature T_0 and the air at T_a. This can be expressed simply by approximating the actual slope s of the curve by a straight line, i.e. $s = \partial\rho^*_v/\partial T \approx \Delta\rho^*_v/\Delta T$. The value of s is then applicable to the mid-point temperature $(T_0 + T_a)/2$, although for most cases the error in using the slope at T_a alone is small. With this linearization the required difference $(\rho^*_{v0} - \rho^*_{va})$ is $s(T_0 - T_a)$. The second part is the difference between the actual and the saturation vapour density at the air temperature T_a. This is simply the vapour density deficit of the air; $vdd_a = \rho^*_{v0} - \rho_{va}$. Therefore, we may write:

$$(\rho^*_{v0} - \rho_{va}) = s(T_0 - T_a) + vdd_a \tag{20.12}$$

Saturated surface – Using a simple resistance formulation the evapotranspiration between a saturated surface and the non-saturated air above is

$$E = (\rho^*_{v0} - \rho_{va})/r_{aV} \tag{20.13}$$

where, r_{aV} is the aerodynamic resistance to water vapour transport. Substituting equation 20.12

$$E = \frac{s(T_0 - T_a) + vdd_a}{r_{aV}} \tag{20.14}$$

and

$$Q_E = \frac{L_v\, s(T_0 - T_a) + L_v\, vdd_a}{r_{aV}} \tag{20.15}$$

Since the sensible heat flux density is:

$$Q_H = \frac{C_a\,(T_0 - T_a)}{r_{aH}} \tag{20.16}$$

substituting in Eq. 20.15 gives:

$$Q_E = \frac{L_v\, s\, Q_H\, r_{aH}}{C_a\, r_{aV}} + \frac{L_v\, vdd_a}{r_{aV}} \qquad (20.17)$$

From the surface energy balance $Q_H = Q^* - Q_G - Q_E$, and the psychrometer constant $\gamma = C_a/L_v$, and if we assume similarity of r_{aH} and r_{aV}:

$$Q_E = \frac{s}{\gamma}(Q^* - Q_G - Q_E) + \frac{C_a\, vdd_a/r_{aH}}{\gamma} \qquad (20.18)$$

which after re-arrangement can be written:

$$Q_E = \frac{s}{s+\gamma}(Q^* - Q_G) + \frac{C_a\, vdd_a/r_{aH}}{s+\gamma} \qquad (20.19)$$

This is the Combination Model for the case of a saturated surface. It is similar to the form suggested by Penman in 1948. It consists of two terms: the first, called the 'energy term', depends only upon the absolute temperature and the energy available for turbulent transport; the second is the 'advection' or 'vapour deficit' term which combines the roles of the dryness of the air with the aerodynamic resistance to transport.

The data necessary to implement equation 20 are relatively easily available, and require measurements at only one level. For a simple surface the first term requires only a radiometer, a heat flux plate and a thermometer. Standard humidity measurements will suffice for the vapour density deficit of the air. Re-arranging equation 20.16:

$$r_{aH} = \frac{C_a\,(T_0 - T_a)}{Q_H} \qquad (20.20)$$

then substituting the neutral aerodynamic equation for Q_H (see lecture 24):

$$r_{aH} = [\ln(z_2/z_1)]^2/k^2\Delta u \qquad (20.21)$$

This requires knowledge of wind speeds at two heights. Alternatively if z_0 is known we can set $z_1 = z_0$, and since at z_0 $u = 0$ only u_2 is needed to assess r_{aH}. Because this expression was derived using the aerodynamic equation it already assumes similarity of diffusion coefficients therefore by extension it also gives the values of r_{aM}, r_{aV} and r_{aC}. In strongly stable or unstable cases the calculations should consider ΦM. This, however, would require a measure of stability such as Ri. The Penman equation derived assumes $r_{aV} \approx r_{aH}$. Over some complex surfaces the sources and sinks of water vapour and heat may not be co-located, this could lead to errors because the pathlengths over which the resistances are calculated would be different.

For small objects such as leaves the Penman equation is applicable if r_a is replaced by the boundary layer resistance, r_b. The heat storage term may be small enough to be neglected.

If the air travels a long distance over a wet surface, its lowest layer is likely to be close to being saturated. Then vdd_a becomes very small and equation 20 collapses to:

$$Q_E = \frac{s}{s+\gamma}(Q^* - Q_G) \qquad (20.22)$$

which is referred to as 'equilibrium' evapotranspiration. In practice large-scale weather events and mixing throughout the planetary boundary layer nearly always maintain a vapour deficit in the air, even over extensive oceans. Therefore evaporation over saturated surfaces is usually greater than the 'equilibrium' rate.

Partially saturated surface – Most surfaces are only partly saturated. The Combination Model (equation) can be modified to accommodate this by introducing a surface vapour density deficit; $vdd_0 = (\rho^*_{v0} - \rho_{v0})$:

$$Q_E = \frac{s(Q^* - Q_G) + [C_a(vdd_a - vdd_0)]/r_{aH}}{s+\gamma} \qquad (20.23)$$

which after re-arrangement becomes:

$$(s+\gamma) + \frac{C_a\, vdd_0}{r_{aH}\, Q_E} = \frac{s(Q^* - Q_G) + C_a\, vdd_a/r_{aH}}{Q_E} \qquad (20.24)$$

The appropriate resistance to vapour transfer at the surface is the canopy or surface resistance, $r_c = L_v(vdd_0)/Q_E$. Substituting r_c we have:

$$(s+\gamma) + \frac{C_a\, r_c}{L_v\, r_{aH}} = \frac{s(Q^* - Q_G) + C_a\, vdd_a/r_{aH}}{Q_E} \qquad (20.25)$$

which, writing it in its more familiar form, is the *Penman-Monteith* version of the Combination Model:

$$Q_E = \frac{s}{s+\gamma}\left[\frac{(Q^* - Q_G) + C_a\, vdd_a/r_{aH}}{1 + r_c/r_{aH}}\right] \qquad (20.26)$$

When dealing with an open water surface $r_c \approx 0$ and equation 20.26 collapses to 20. For a full vegetation cover r_c is related to the areal average stomatal resistance.

The difficulty in using the Penman-Monteith model is the evaluation of r_c. For vegetation it is possible to measure the average stomatal resistance of a canopy using a *porometer* but it is a laborious procedure. Considerable effort is being expended to use measurements of Q_E (e.g. from eddy covariance methods) in conjunction with other observations and equation 20.26 to study the behaviour of r_c. The aim is to seek simple parameterization schemes for r_c. It is realized that surface evapotranspiration is controlled by more than the characteristics of the surface and the surface layer. The dynamics of the total planetary boundary layer are involved including the entrainment of dry or moist air from above the mixed layer (e.g. McNaughton and Jarvis, 1983). Low plant covers with relatively large aerodynamic resistances are relatively poorly coupled to the bulk of the boundary layer, hence their evapotranspiration regime tends to be more closely tied to the available energy term, especially Q^*, than to the vapour deficit. Tall vegetation systems are strongly coupled to a deep layer of air and Q_E can be closely related to the variations of vdd_a.

Lysimeter approach. To gain hourly measurements of evaportranspiration E, alternatively the lysimetric approach can be used. A lysimeter is a true water balance device which hydrologically isolates a volume of soil (and its plant cover). In the most accurate examples an undisturbed soil monolith is enclosed in a water- tight container (at least 1 m deep, and 1 to 6 m in diameter) with only its upper side open (Figure 7). This arrangement allows complete specification of the water budget of the soil monolith because p can be measured, Δr is zero, and any deep drainage is either monitored in a sump or kept zero by providing no outlet. Therefore any changes in the mass of the tank must be due to evapotranspiration (including dewfall) or any irrigation applied (i.e. changes in storage, Δ_S must be related to the mass flux of water to or from the atmosphere, E). Mass changes by the lysimeter are monitored either by a mechanical balance system installed under the monolith, or by changes in the hydrostatic balance of a fluid system which transmits pressure changes to a manometer as changes of fluid level. The example in Figure 7 is of the latter, 'floating' design. In the most accurate lysimeters E can be evaluated to within 0.02 mm water equivalent.

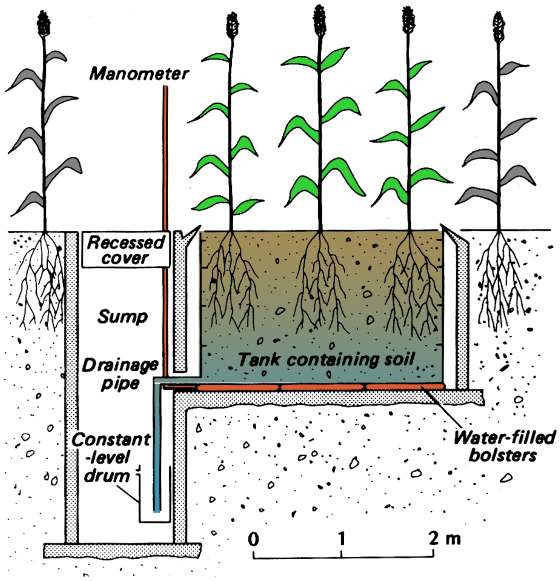

Figure 7: Schematic side view of a lysimeter (modified after Forsgate et al., 1965).