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Water Balance and Evapotranspiration Monitoring in Geotechnical and Geoenvironmental Engineering

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Abstract Among the various components of the water balance, measurement of evapotranspiration has probably been the most difficult component to quantify and measure experimentally. Some attempts for direct measurement of evapotranspiration have included the use of weighing lysimeters. However, quantification of evapotranspiration has been typically conducted using energy balance approaches or indirect water balance methods that rely on quantification of other water balance components. This paper initially presents the fundamental aspects of evapotranspiration as well as of its evaporation and transpiration components. Typical methods used for prediction of evapotranspiration based on meteorological information are also discussed. The current trend of using evapotranspirative cover systems for closure of waste containment facilities located in arid climates has brought renewed needs for quantification of evapotranspiration. Finally, case histories where direct or indirect measurements of evapotranspiration have been conducted are described and analyzed.

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1 Introduction

The interaction between ground surface and the atmosphere has not been frequently addressed in geotechnical practice. Perhaps the applications where such evaluations have been considered the most are in the evaluation of landslides induced by loss of suction due to precipitations (e.g., Alonso et al. 1995; Shimada et al. 1995; Cai and Ugai 1998; Fourie et al. 1998; Rahardjo et al. 1998). Yet, the quantification and measurement of evapotranspiration has recently received renewed interest. This is the case, for example, due to the design of evapotranspirative cover systems for waste containment and mining sites. In cases like this, quantification of the flow boundary condition at the earth-atmosphere interface becomes an integral aspect of the analysis and design of the geotechnical system.

Paradoxically, examination of the atmospheric water balance in different regions through the world has shown that evapotranspiration often exceeds precipitation (Blight 1997). This is the case of arid or semi-arid regions over most of the year as well as of most regions with temperate climate over long portion of the year. Recent assessment on hazards caused by droughts showed that the evapotranspiration is an

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important phenomenon that should be accounted for in natural hazards analysis. In France, the extensive drought from 1989 to 1990 affected shallowly founded buildings of 216 communes in 17 departments (Vandangeon 1992). In the decree of November 1, 2005 (French Official Journal 1.2), more than 870 communes were considered affected by the 2003 drought. In other countries, a number of case studies involving the effect of prolonged periods of evapotranspiration were performed (Driscoll 1983; Biddle 1983; Williams and Pidgeon 1983; Ravina 1983; Holtz 1983; Gao 1995; Allman et al. 1998). It is now recognized that hazards related to droughts have an important economical impact, and deserve additional research.

In addition to the geotechnical engineering problems associated with the changes in mechanical properties of soil induced by infiltration and evapotranspiration, current advances in geoenvironmental engineering have often focused on the effect of infiltration and evapotranspiration on the hydraulic properties of soils. Quantification of evapotranspiration has been particularly relevant for the design of cover systems, which is one of the key engineered components of municipal and hazardous waste landfills as well as mine disposal sites. The cover system should be designed to minimize percolation of rainwater into the waste and prevent leachate generation that may lead to environmental contamination of soil and groundwater. A conventional "resistive barrier" type cover system involves a liner (e.g., a compacted clay layer) constructed with a low saturated hydraulic conductivity (typically 10^{-9} m/s or less) to reduce percolation. Figure 1a illustrates the water balance components in this comparatively simple system, in which percolation control is achieved by maximizing overland flow. However, designing a truly *impermeable* barrier (i.e., one leading to zero percolation) should not be within any engineer's expectations. Instead, the engineer should be able to design a system that *minimizes* percolation to environmentally safe values. Quantification of this minimized, though finite, percolation of liquid into the waste poses significant challenges.

Figure 1b illustrates schematically the water balance components in an evapotranspirative cover system. Evapotranspiration and moisture storage, two components that are not accounted for in the design of resistive barriers, are significant elements in the performance of this system. The uniqueness of this

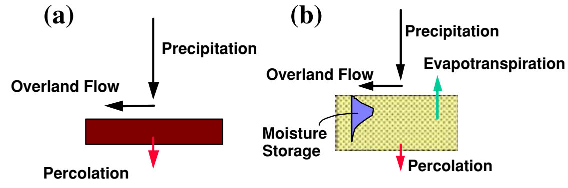

Fig. 1 Water balance components: (**a**) in a resistive barrier; (**b**) in an evapotranspirative cover system

approach is the mechanism by which percolation control is achieved: an evapotranspirative cover acts not as a barrier, but as a sponge or a reservoir that stores moisture during precipitation events, and then releases it back to the atmosphere by evapotranspiration. The adequacy of alternative cover systems for arid locations has been acknowledged by field experimental assessments (e.g., Anderson et al. 1993; Dwyer 1998; Nyhan et al. 1997), and procedures for quantitative evaluation of the variables governing the performance of this system have been compiled in a systematic manner for final cover design (e.g., Zornberg and Caldwell 1998; Zornberg et al. 2003).

Recent developments in unsaturated soil mechanics enable preliminary predictions of evapotranspiration and infiltration of surface water. Water flux on the ground surface is often calculated as a function of the recorded precipitation. On the other hand, evapotranspiration through the upper boundary has been often defined based on the soil suction and temperature at the ground surface. However, evapotranspiration can also be determined based on the energy balance between solar radiation, sensitive heat of air, soil heat flux and heat latent for the need of evaporation. The main advantage of this energy balance approach is that it avoids direct suction measurement. Yet, evapotranspiration has been probably the most difficult water balance component to quantify and measure experimentally. Accordingly, the most common approach has involved quantifying evapotranspiration indirectly by monitoring other components of the water balance. Methods for direct measurement of evapotranspiration, energy balance methods, and water balance methods are discussed in this paper.

2 Solar Radiation

Evapotranspiration is governed by part of the energy coming from the sun. It is well known (see for instance Pidwirny 2006) that the emission rate of the sun is estimated as 63 million W/m^2. However, as this radiation travels away from the sun, the amount that strikes another object depends on its distance from the sun. The portion that reaches the earth's atmosphere is called the solar constant (approximately 1380 W/m^2). Once solar radiation reaches the top of the atmosphere, it is absorbed, scattered, or transmitted through the atmosphere. Since 30% is scattered back to space (earth's albedo), the earth's atmosphere only receives 70% of the incoming (incident) radiation.

Of the 70% radiation that is transmitted through the atmosphere, 19% get absorbed by gases, primarily molecular oxygen and ozone. The remaining 51% is transmitted to the earth's surface. The solar constant can be used to calculate the possible limit to daily evapotranspiration. Specifically, assuming that (i) 51% energy flux (i.e. 704 W/m^2) is used for evaporation of pure water (it is of course not the case as discussed subsequently), (ii) the solar power is parabolically distributed during the day, and (iii) from zero at dawn to zero at sunset, the albeto is 15% (the value can be vary variable according to the soil surface nature, from 5% to 50% in most cases), the maximum solar energy needed to produce evaporation is 17.2 MJ/m^2 for a 12 h day (i.e. $2/3 \times 704 \times 0.85 \times 12 \times 3600 = 17233917$ J/m^2). The latent heat of evaporation of water is about 2.47 MJ/kg and hence the maximum possible evapotranspiration 6.9 mm of water (i.e. $17.234/2.47 = 6.9$ kg water/m^2).

The balance at the earth's surface between incoming and outgoing components of radiant energy is characterized by the net radiation R_n (W/m^2):

$$R_n = (S + D)(1 - a) + \left(L_{\mathrm{down}} - L_{\mathrm{up}}\right) \tag{1}$$

where S (W/m^2) is the direct shortwave radiation corresponding to the shortwave radiation penetrating directly to the surface without being affected by the atmosphere constituents; D (W/m^2) is the diffuse shortwave radiation corresponding to the shortwave radiation scattered or diffused by atmosphere constituents (clouds, dust etc.); a is the albedo corresponding to the proportion of radiation reflected from the ground surface, governed primarily by surface color and incitation angle of the sun; L_{up} (W/m^2) is the terrestrial radiation corresponding to the longwave radiation emitted by the earth's surface; L_{down} (W/m^2) is the atmospheric counter radiation corresponding to the longwave radiation emitted by the atmosphere directed towards the surface. The magnitudes of L_{up} and L_{down} depend on the temperature of the emitting body.

3 Soil Water Balance

The soil water balance can be represented as follows:

$$P - (I_{\mathrm{nt}} + R_{\mathrm{off}}) = ET + R_{\mathrm{wt}} + \Delta S \tag{2}$$

where P (mm/day) is precipitation; I_{nt} (mm/day) is interception; R_{off} (mm/day) is the runoff on ground surface; ET (mm/day) is evapotranspiration; R_{wt} (mm/day) is the water recharged to the water table; ΔS is the change in soil water storage.

Interception corresponds to the storage of water above the ground surface, mostly in vegetation. It is usually negligible but can reduce precipitation intensity in case of dense forest canopy. The term (R_{wt} + S) represents the water infiltration I, therefore:

$$I = (R_{\mathrm{wt}} + S) = P - (I_{\mathrm{nt}} + ET + R_{\mathrm{off}}) \tag{3}$$

In soil water balance, the evapotranspiration term ET is governed by the energy and mass exchanges between soil and atmosphere. It is discussed in more detail in Sect. 4.

4 Evapotranspiration

Evapotranspiration is composed of the direct evaporation, which takes place from the soil surface, and of the transpiration from vegetation. The roots of vegetation capture soil water, part of which evaporates through the stomata (micropores) of the leaves, while the rest is used for photosynthesis. Evapotranspiration depends on two elements: the heat supplied by solar radiation and the water available in the soil. While the quantity of solar energy reaching the ground surface is approximately constant, evapotranspiration is very sensitive to the climate variations and plant characteristics.

4.1 Evaporation

Evaporation involves the change in water state from liquid to water vapor due to an increase of water kinetic energy. During evaporation, hydrogen bonds are broken and water vapor is diffused from regions of higher to lower vapor pressure, i.e., from the ground surface to the surrounding air. Water vapor consists entirely of free water molecules, while liquid water consists of both free and bonded molecules.

Evaporation from soils is an important phenomenon that should be quantified in order to define the surface flux boundary condition in an unsaturated flow analysis. Evaluation methods based on either soil water balance or experimental characterization have been proposed by Philip (1957), Gardner and Hillel (1962), Gardner (1973), Brutsaert (1982), Boast (1986), Evett et al. (1994), Wilson et al. (1994, 1999) and Raghuywanshi and Wallender (1998). Evaporative processes are typically modeled isothermally. This is a simplified analysis because thermally induced flow of water through unsaturated soil may also occur by vapor diffusion. A more realistic evaporation analysis requires consideration of thermal effects. The phenomenon of thermally induced water flow was investigated by Milly (1996) and Kampf and Von der Hude (1995). Heat flux phenomena in soils that lead to thermally induced flow was also investigated by Qualls and Brutsaert (1996a, b). Fischer et al. (1996) modeled soil vapor extraction, and compared experimental data to calculation results. Additional discussion on surface flux boundary conditions was provided by Wilson (1997, 2000).

Evaporation rate is governed by several factors, as follows (see Dingman 1994):

(1) General factors, including: (i) latent heat for evaporation, the major source of which is the solar energy so that the distribution of radiation and evaporation is strongly correlated (maximum evaporation in the tropics and during the warmest part of the day); (ii) sensitive heat of air; (iii) air temperature, which is a measure of heat energy and of the capacity of air to hold water vapor (i.e. the saturation vapor pressure increases with increasing air temperature); (iv) air humidity, an increase of which causes a decrease in the rate of evaporation; (v) wind, which causes eddy (turbulent diffusion) and thereby maintains the vapor pressure gradient between air and the evaporation surface (evaporation increases dramatically with increasing turbulence, which is function of wind speed and surface roughness).

(2) Additional factors controlling the rate of evaporation from water bodies, such as water salinity.

(3) Additional factors controlling the rate of evaporation from soil, including: (i) soil water potential, as the rate of evaporation decreases significantly as soil dries out; (ii) depth of water table, as evaporation rate decreases significantly with increasing depth to the water table to a critical depth below which groundwater does no longer affect the evaporation rate, this critical depth depending on the nature of the soils involved; (iii) soil color, with greater absorption of heat and thus evaporation for dark soils (small albedo); (iv) vegetation, which reduces evaporation by shading soil, reducing wind at the ground surface and increasing vapor pressure by transpiring water pressure.

4.2 Transpiration

Transpiration is the evaporation from the vascular system of plants. Water absorbed by the roots raises by capillary action to stomata cavities in the leaves, from where it evaporates. The vapor pressure gradient between the leaf tissue (and bark, to a lesser extent) and the surrounding air draws water from soil into the roots and up the plant through the xylem. As water evaporates within the leaves tissue, salt can precipitate and further attract water by osmotic effect. However, if the soil is saline, the salt concentration gradient is reversed and water may be even drawn out of plants.

Water uptake by plants and rooting depth of the plant cover are another issue of relevance in the analysis of evapotranspirative covers. Transpiration is also used as boundary conditions in unsaturated flow analyses. Ritchie (1972), Ritchie and Burnett (1971), and Tratch et al. (1995) provide a summary of plant transpiration in terms that engineers are familiar with. The combination of evaporation and transpiration into evapotranspiration has been discussed by Hargreaves (1994) and Pereira et al. (1999), and has been modeled by Chudhury et al. (1986), Levitt et al. (1996), Xu and Qiu (1997), etc. The ecology of plant systems used for transpiration has also been a topic of significant relevance, studied by Anderson et al. (1987) and Anderson (1997). These studies concluded that a diverse group of plant species of different heights and rooting depths are required for a stable plant population.

Transpiration from different plants has been evaluated by Anderson et al. (1987, 1993), and Waugh et al. (1991). Wu and Oster (1997) provide details on several instruments used by agricultural scientists for management of plants and soil water. Because of the difficulty in measuring evaporation and transpiration in the field, a common approach has been to conduct water balance back calculation using lysimeters. Nonetheless, Evett (1994) has used time domain reflectrometry (TDR) to investigate thermal properties in soil, which is related to the amount of potential evaporation. Evett (1993) used TDR and neutron scattering to measure evapotranspiration.

5 Prediction of Evapotranspiration

Numerous predictive methods have been developed to estimate evapotranspiration, including Penman's method, Penman–Monteith's method, and Turc's method (see Guyot 1997 for a comprehensive review). These methods are based on the concept of Potential Evapotranspiration (PET). According to Penman (1948), PET corresponds to the evapotranspiration rate from a large area completely and uniformly covered with growing vegetation which has unlimited supply of water without advection (wind) and heat—storage effects. Since evapotranspiration depends on the type of vegetation, short grass was adopted. Penman (1948) proposed the following equation for PET (kg water/m^2/day, i.e. mm/day) calculation:

$$PET = \frac{1000\Delta R_n/(\rho_w L_v) + \gamma E_a}{\Delta + \gamma} \tag{4}$$

where $\Delta = \frac{4099 P_{vs}}{(T+237.3)^2}$ (Pa/°C) is the slope of the curve of saturated vapor pressure (Pa) versus temperature (°C) at the prevailing temperature; $E_a = 0.165$ $(P_{vs} - P_v)$ $(0.8 + u_2/100)$ (mm/day) where R_n is the net radiation flux (J/(m^{2}day)); ρ_w is water density (kg/m^3); L_v is the latent heat of vaporization of water (J/kg); γ is the psychrometric constant (Pa/°C); P_{vs} is the saturated vapor pressure (mbar); P_v is the actual vapor pressure in air (mbar); u_2 is the wind speed at 2 m elevation (km/day). When wind speed is measured at elevations other than 2 m, the speed at 2 m can be estimated as:

$$u_2 = u_z\left(\frac{4.87}{\ln(67.8z - 5.42)}\right) \tag{5}$$

where u_z is the wind speed at elevation z above the ground surface. Note that in Penman's equation, no vegetation parameters are used even though short grass is referred to.

6 Measurement of Evapotranspiration

6.1 Direct Measurement

For evapotranspirative covers, lysimetry involves the use of buried containers used to collect percolating soil water. Unlike apparatus involving monitoring of suction profiles and indirect determination of flux, lysimetry provides measurement of percolation rate from an soil cover. Among the various types of lysimeters, weighing lysimeters (Fig. 2) allow direct measurement of evapotranspiration as they measure the total weight of soil and stored water (Fayer and Gee 1997; Benson et al. 2001). Changes of soil water storage in a lysimeter can be determined by integrating profiles of water content measured using nests of probes. Consequently, the remaining changes in mass can be attributed to losses by evapotranspiration.

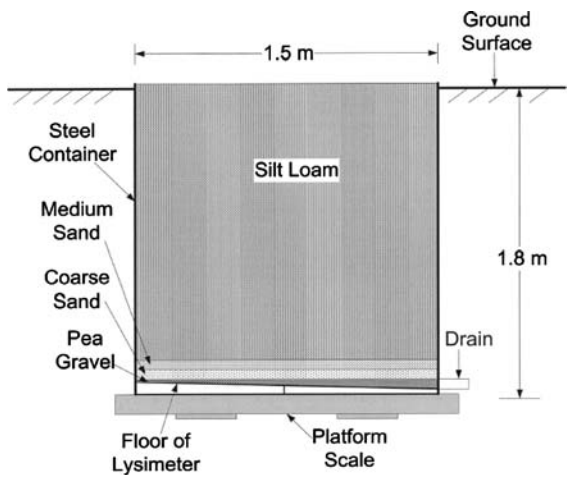

Fig. 2 Weighing lysimeter used in final cover studies at the Hanford reservation in Washington, USA. (from Benson et al. 2001)

Weighing lysimeters are typically limited to small test sections (1–2 m^2) because of the limited capacity of scales (Waugh et al. 1991).

6.2 Energy Balance Approach

This approach involves quantification of exchanges between soil and atmosphere. Specifically, these exchanges involve energy (heat) and mass exchanges, mostly by convection. The energy balance in the earth-atmospheric system can be presented as:

$$R_n = L_e + H + G \tag{6}$$

where L_e is the latent energy transfer (positive for evaporation and negative for condensation); H is the sensitive heat transfer (positive when energy is used to warm the air and negative when the air loses energy due to cooling); G is the ground heat transfer (positive when energy is transferred to the subsurface and negative when energy is transferred to the atmosphere).

There are several methods for evapotranspiration measurement: eddy correlation, flux profiles, residual method and Bowen ratio method Kolle (1996). Consistent with the Bowen ratio method, the sensitive heat flux within a few meters of the surface, H, can be expressed as:

$$H = \rho_a C_p k_H \frac{\partial T}{\partial z} \tag{7}$$

where ρ_a (kg/m^3) = 1.2929 (273.13/T (K)) [(P (mm) − 0.3783 P_v (mm))/760] is air density which depends on vapor pressure; the value of dry air is generally considered: ρ_a (kg/m^3) = 1.2929 (273.13/T (K)); C_p is specific heat of air (also the value for dry air, 1.01 kJ/(kg K), can be generally assumed); T is temperature (K); z is elevation (m); k_H, eddy diffusivity for air. The latent heat flux L_e can be expressed as follows

$$L_e = \frac{L_v \rho \varepsilon k_v}{P} \frac{\partial P_v}{\partial z} \tag{8}$$

where L_v (kJ/kg) = 2501 − 2.361T (°C) is latent heat of vaporization; ε is the ratio of molecular weight of water to molecular weight of air ($\varepsilon = 0.622$); k_v (m^2/s), is eddy diffusivity for vapor; P_v (kPa) is vapor pressure; P (kPa) is atmospheric pressure which depends on elevation, as follows (Wallace and Hobbs 1977):

$$P = 101.325\left(1 - \frac{z}{44307.69231}\right)^{5.25328} \tag{9}$$

In general, k_v and k_H are not known but are assumed to be equal (Blight 1997). The ratio of H to L_e is then used to partition the available energy at the surface into sensitive and latent heat flux. This ratio was first defined by Bowen (1926), and is known as the Bowen ratio β:

$$\beta = \frac{H}{L_e} = \frac{PCp}{L_v \varepsilon} \frac{\partial T}{\partial P_v} = \gamma \frac{\partial T}{\partial P_v} \tag{10}$$

where $\gamma = PCp/(L_v\varepsilon)$ is the psychrometric constant.

Knowing the net radiation flux R_n, the total soil heat flux G and the Bowen ratio β, the latent heat flux can be obtained as:

$$L_e = \frac{R_n - G}{1 + \beta} \tag{11}$$

Bowen ratio can be determined by measuring the temperature T and the vapor pressure P_v at two elevations: water vapor pressure is often measured with a single cooled mirror dew point hygrometer (e.g., Campbell Scientific BR023 1998). Air temperature can be measured using different thermocouples. For example, Campbell Scientific uses two chrome—constantan thermocouples. Soil heat flux can be measured using heat flux plates buried in the soil at a fixed depth. The plates are typically buried at a depth of 8 cm (Campbell Scientific 1998). The average temperature of the soil layer above the plate is measured using 2–4 thermocouples. The heat flux at the surface can be calculated by adding the heat flux measured by the plates to the energy stored in the soil layer. The storage term is calculated by multiplying the soil heat capacity C_s by the change in soil temperature ΔT over the averaging period t:

$$S = \frac{\Delta T C_s d}{t} \tag{12}$$

where d is the plate depth. The soil heat capacity can be calculated by adding the specific heat of the dry soil C_d to that of the soil water C_w:

$$C_s = \rho_d(C_d + wC_w) = \rho_d C_d + \theta \rho_w C_w \tag{13}$$

where ρ_d and ρ_w are soil dry density (kg/m^3) and water density (kg/m^3) respectively; w and θ are gravimetric and volumetric water content respectively. Figure 3 illustrates a soil heat flux system (Campbell Scientific 1998).

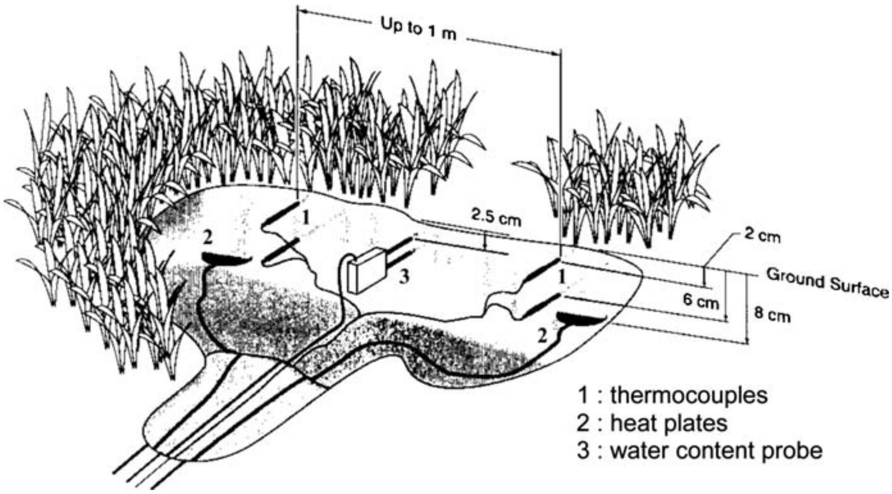

Fig. 3 Soil heat flux measurement (Campbell Scientific system 1998)

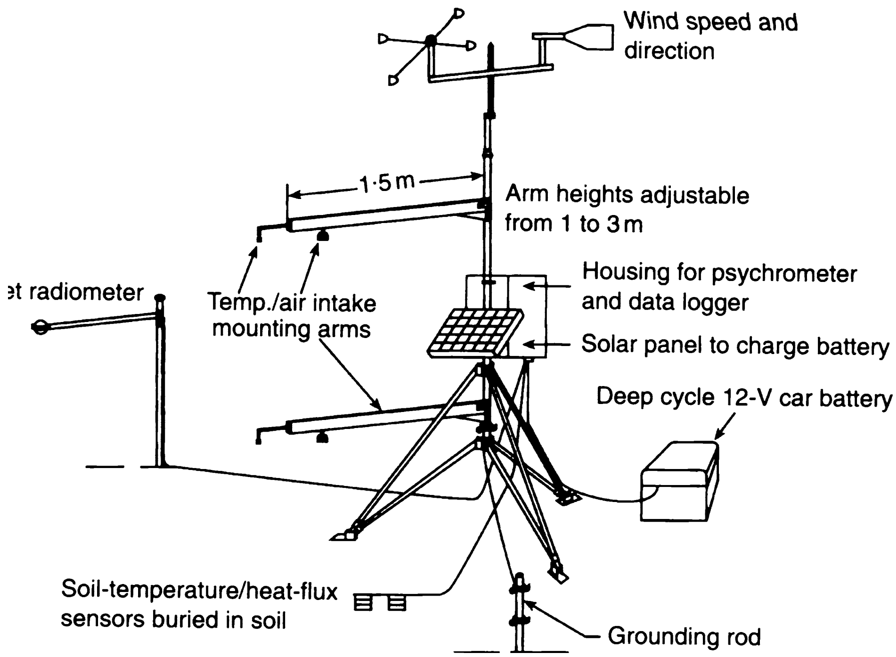

Fig. 4 System of Bowen ratio measurement (Blight 1997)

The measurement of R_n is carried out using a net radiometer. Figure 4 illustrates a Bowen ratio system (Blight 1997), which includes two arms at different elevations where temperature and vapor pressure are monitored. A net radiometer is also installed for R_n measurement.

6.3 Water Balance Approach

Evapotranspiration can also be measured indirectly by quantifying other components in the water balance. Specifically, precipitation, surface water runoff, changes in moisture storage, and basal percolation. As mentioned, lysimeters have been used to monitor basal percolation (e.g., Fig. 5). Note that in a natural soil profile, monitoring of basal percolation is usually not possible because of difficulties associated with the installation of a draining layer. In addition, the possible recharge of the soil from the water table may affect water content measurements. This limits the use of lysimeters in natural soil profiles.

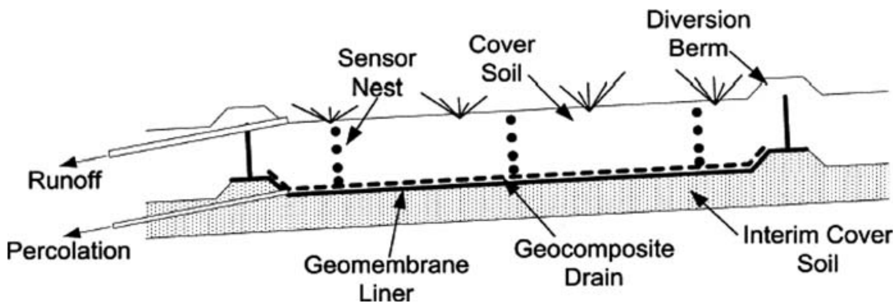

Fig. 5 Schematic of lysimeter used in final cover studies (Khire et al. 1997)

Basal percolation, precipitation, changes in soil moisture storage, and surface water runoff are typically monitored on a daily basis. In addition, solar radiation, wind speed and direction, and percentage cloud cover are also measured. Figure 6 shows typical monitoring layout used in sites where evapotranspiration is defined indirectly using a water balance approach.

Considering the conservation of mass of water into and out of the cover, the evapotranspiration may be obtained as follows:

$$ET = P - G - \Delta S - R_{off} \quad (14)$$

where ET = evapotranspiration; P = precipitation; G = basal percolation; ΔS = change in moisture storage; R_{off} = surface water runoff. Rain and snow can be measured using an all season gauge. Percolation is channeled from the lysimeter by gravity and measured in a sump using a tipping-bucket rain gauge. The moisture content profile is typically measured in the center of the lysimeter using an array of time domain reflectometer sensors, wave content reflectometer (WCR) sensors, or moisture probes spaced evenly with depth. Surface water runoff is typically collected in geomembrane swales around the cover perimeter.

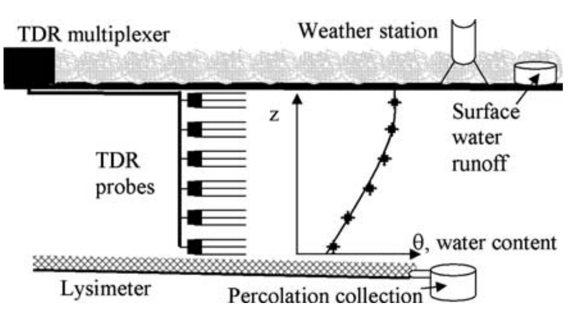

Fig. 6 Monitoring system layout

TDR, nuclear moisture probes and other techniques have been used to measure suction-saturation curves and other unsaturated flow phenomena. Benson et al. (1994) describes the monitoring system for a typical landfill cover. Several field studies were discussed by Gee et al. (1991), Phillips et al. (1991), and Allison et al. (1994). Gee and Hillel (1988) reviewed several methods for estimating percolation through soil covers. Campbell et al. (1991) describes the use of lysimeters to conduct water balances. Additional water balance studies were reported by Nyhan et al. (1990, 1997), Warren et al. (1994, 1996), Hakonson et al. (1994), Khire et al. (1997a, b), Waugh et al. (1991), and Anderson et al. (1998). Water balance approaches have often been used in the US as part of the selection and performance evaluation of alternative cover systems suitable for arid or semi-arid climates, as discussed by Anderson et al. (1998), Stormont (1995), Benson and Khire (1995), Wing and Gee (1989, 1994), Gee and Ward (1997), and Dwyer (1998, 2001).

7 Case Histories Involving Monitoring of Evapotranspiration

This section presents some cases studies aiming at illustrating each of the three methods for quantification of evapotranspiration.

7.1 Direct Measurement: Monticello (US)

The U.S. Department of Energy (DOE) conducted a series of field lysimeter experiments to help design and then monitor the performance of an engineered cover for a uranium mill tailings disposal cell at a Superfund Site in Monticello, Utah (Waugh 2002). The lysimeter test facility evolved as a sequence of

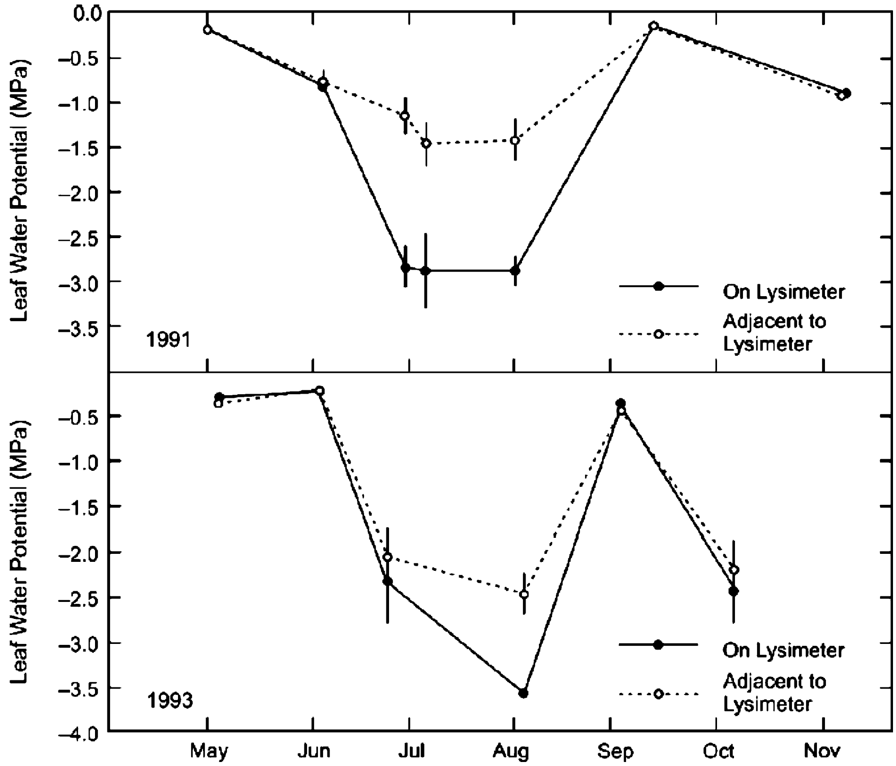

Fig. 7 Predawn leaf water potential values for *P. smithii* growing on and adjacent to the small monolith lysimeters between May and November 1991 and between May and October 1993 (Waugh 2002)

installations, first to test the concept of using an evapotranspirative cover at Monticello, next to evaluate the soil–water balance of the final engineered design, and finally to monitor the hydrologic performance of a large facet of the completed disposal cell cover. Small weighing lysimeters were installing containing intact, 100-cm-deep profiles of undisturbed silt loam soil (monoliths) overlying a pea-gravel capillary barrier and supporting mature native grasses. Leaf water potential and leaf transpiration of plants on and adjacent to the lysimeters were compared to evaluate the effect of the small weighing lysimeter design on plant behavior. Because of favorable monolith lysimeter results, 15 additional small weighing lysimeters were installed to test the effects of varying soil types and soil layer thickness on soil-water balance and water-storage capacity. This study evolved into the construction of large caisson lysimeters to evaluate the water balance of the final cover design for the Monticello disposal cell. The cover layer constructed inside the caissons matched as-built engineering parameters for the actual cover.

Plants growing on and adjacent to the lysimeters (*P. smithii*) were sampled to evaluate the effects of isolating a soil monolith on plant water status. Predawn leaf water potential values were measured monthly during the growing season using a pressure chamber technique (Scholander et al. 1965). Predawn potential values for *P. smithii* growing on and adjacent to the lysimeters were similar early in the growing season, diverged significantly during the mid-summer moisture depletion period, and then reconverged following the late-summer monsoons (Fig. 7).

Some divergence of predawn potential values for *P. smithii* on and adjacent to the lysimeters was observed, indicating that plants were seasonally more stressed inside the lysimeters than in the adjacent plant community. This suggests that small lysimeters moderately underestimates ET. However, for screening tests consisting of multiple treatments and replications, it was concluded that the small lysimeters provided reasonable comparisons of the hydrologic performance of evapotranspirative cover designs.

Overall, the use of weighing lysimeters allowed direct quantification of the ET; on the other hand, the use of small devices may have compromised quantification of flow through macro-fractures as cracks.

The use of large lysimeters (without weighing capabilities) is discussed in the case history presented in Sect. 7.3.

7.2 Energy Balance Approach: Boissy-le-Châtel (France)

In a common meteorology station only data at 2 m elevation are available and thus the Bowen ratio is generally not measured. In this case, it is necessary to use numerical methods to determine L_e indirectly. From a point of view of geotechnical engineering, knowing L_e is essential to subsequently analyze the soil settlement and slope stability problem due to evapotranspiration, because it is possible to determine the variations of soil temperature and water content using an appropriate method. Cui et al. (2005) used the two coupled equations (Eq. 15 and Eq. 16) proposed by Wilson et al. (1994) to calculate β and determine the temperature and water content profiles in the soil:

$$\frac{\partial h_W}{\partial t} = C_W^1 \frac{\partial}{\partial y}\left(k_W \frac{\partial h_W}{\partial y}\right) + C_W^2 \frac{\partial}{\partial y}\left(D_V \frac{\partial P_V}{\partial y}\right) \tag{15}$$

with $C_W^1 = \frac{1}{\rho_W g m_2^W}$ and $C_W^2 = \frac{P+P_V}{P(\rho_W)^2 g m_2^W}$ where h_w = water head; k_w = water permeability depending on the suction; D_v, = vapour diffusivity; P_v = vapour partial pressure; P = atmospheric pressure; m_2^W = slope of the water retention curve expressed in terms of volumetric water content versus suction; y = elevation.

$$C_h = \frac{\partial T}{\partial t} = \frac{\partial}{\partial y}\left(\lambda \frac{\partial T}{\partial y}\right) - L_V \frac{P + P_V}{P} \frac{\partial}{\partial y}\left(D_V \frac{\partial P_V}{\partial y}\right) \tag{16}$$

where Ch = soil heat capacity; T = temperature, t = time; λ = soil thermal conductivity. Note that because of the lack of information, a constant value of G/H ratio was adopted in the calculation, suggesting that the heat fluxes through soil and air remain proportional. An initial β value of 0.01 was considered, enabling the calculation of H and L_e from the energy balance equation (Eq. 6), and thus the determination of initial upper boundary condition at the ground surface. The numerical resolution of the two-coupled equations (Eqs. 15 and 16) provided a profile of temperature T and of partial vapour pressure P_v that allowed the calculation of vapour flux or evaporation at the ground surface. This evaporation was compared with the target value calculated from field data using Penman–Monteith equation and the difference was compared to a maximum acceptable value taken equal to 0.01. In case of a larger difference, the iterative process was renewed based on a different value of β, until the required convergence was reached. The profiles of temperature, volumetric water content were then determined.

Data from the site of Boissy-le-Châtel (France) were evaluated using this method. The site is located about 50 km East of Paris in the South of the Orgeval basin. Main meteorology data of 2003 (air temperature, air relative humidity and solar radiation) are presented in Figs. 8–10, respectively. It is observed that the air temperature is generally above 0°C, with August the hottest month (maximum temperature reaches 40°C). Comparison between Figs. 8 and 10 show that the air temperature changes correlate well with solar radiation. The relative humidity varies between 30 and 100% (Fig. 10), but it does not necessary follow the precipitation pattern. This confirms that relative humidity depends not only on precipitation, but also on air temperature and wind speed.

The potential evapotranspiration PET calculated using Penman–Monteith equation is presented in Fig. 11. Summer is the season where evaporation is most pronounced (July and August). Intense evaporation is also observed by the middle of April (6 mm/day), which corresponds to the particularly low relative humidity observed during that period (30%, Fig. 10).

Figure 12 presents the soil temperature variation obtained from TDR probes. The data shows that

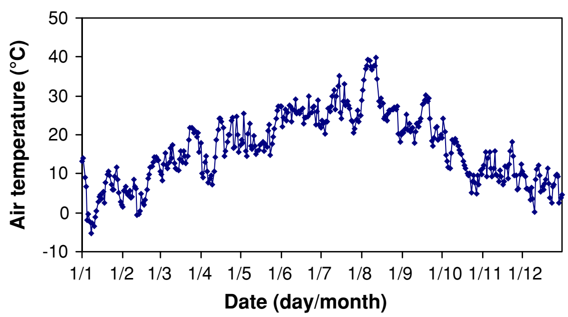

Fig. 8 Air temperature variation during the year 2003 in Boissy-le-Châtel

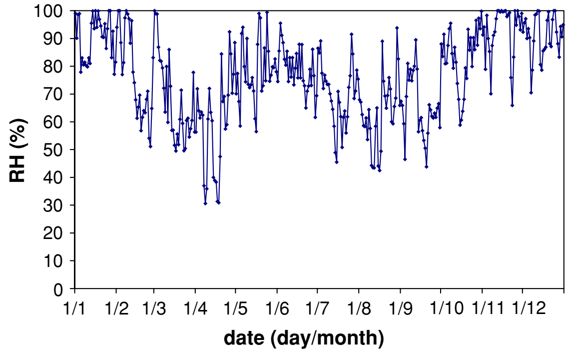

Fig. 9 Air relative humidity variation during the year 2003 in Boissy-le-Châtel

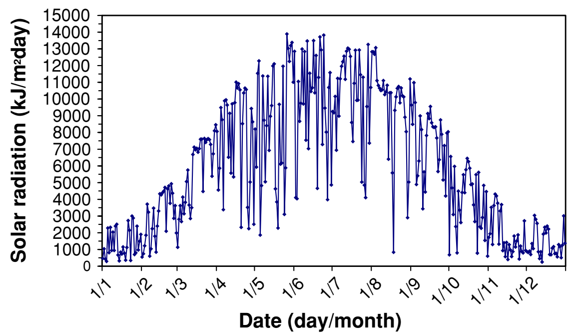

Fig. 10 Solar radiation during the year 2003 in Boissy-le-Châtel

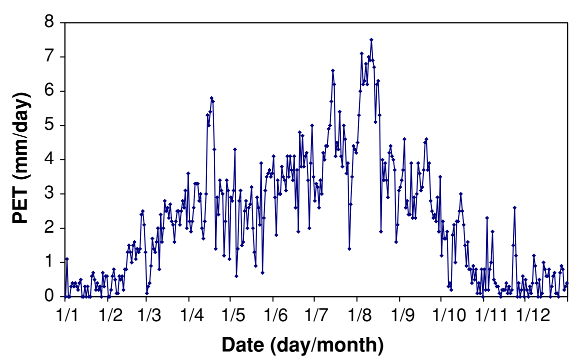

Fig. 11 PET during the year 2003 in Boissy-le-Châtel

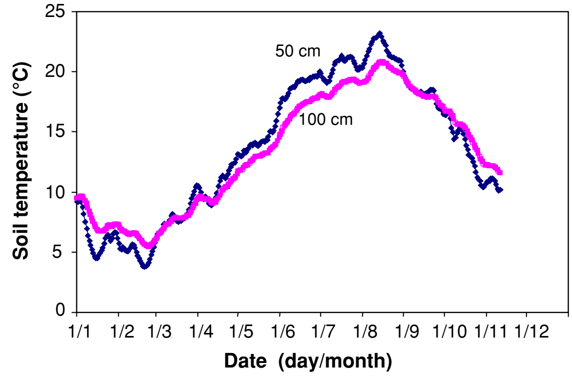

Fig. 12 Soil temperature variation during the year 2003 in Boissy-le-Châtel

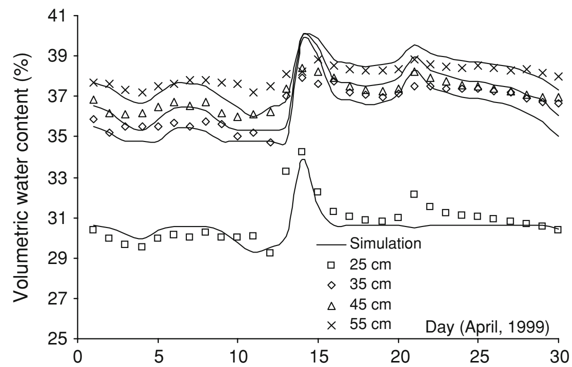

Fig. 13 Comparison between measured and calculated water content at different depth, (Boissy-le-Châtel, Cui et al. 2005)

temperature is higher at deeper soil layer in cold seasons (January–March, October–December), but this trend inverses in the other seasons. Examination of Fig. 11 shows that evaporation is negligible in the cold seasons. Figure 13 shows the comparison between the measured and predicted volumetric water content at four different depths (25, 35, 45 55 cm) for the period of April 1999 (Cui et al. 2005). The good agreement clearly shows the validity of the adopted method. Figure 14 presents the comparison between measured and calculated temperature at 0.5 m depth, with equally satisfactory results.

7.3 Water Balance Approach: Rocky Mountain Arsenal (US)

A series of instrumented test plots were constructed at the Rocky Mountain Arsenal, located near Denver, Colorado, USA, in Summer 1998 (Kiel et al. 2002; Zornberg and McCartney 2003). The climate in Denver is semiarid, with an average annual precipitation of 396 mm and an average pan evaporation of 1,394 mm (quantified from 1948 to 1998). The wettest months of the year (April–October) are also the months with the highest pan evaporation; which are optimal conditions for an evapotranspirative

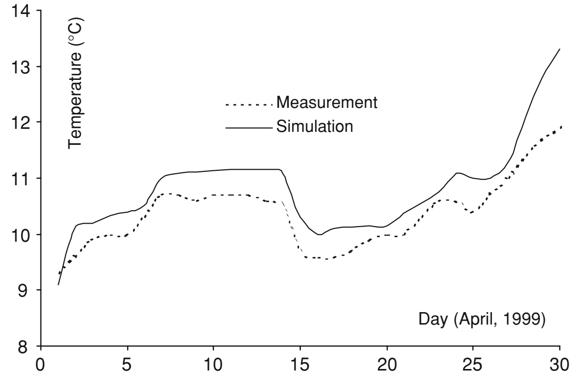

Fig. 14 Comparison between measured and calculated soil temperature at 0.5 m depth, (Boissy-le-Châtel, Cui et al. 2005)

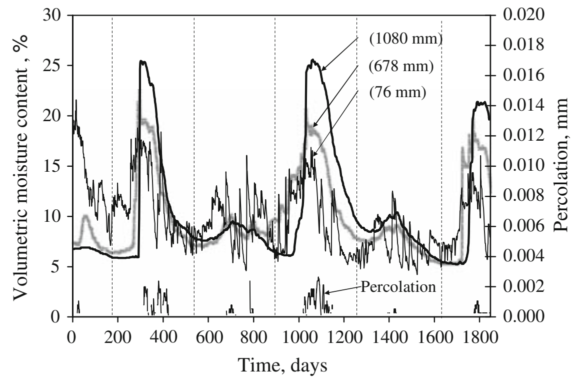

Fig. 15 Percolation and volumetric moisture content at three depths (76 mm, 678 mm, and 1080 mm)

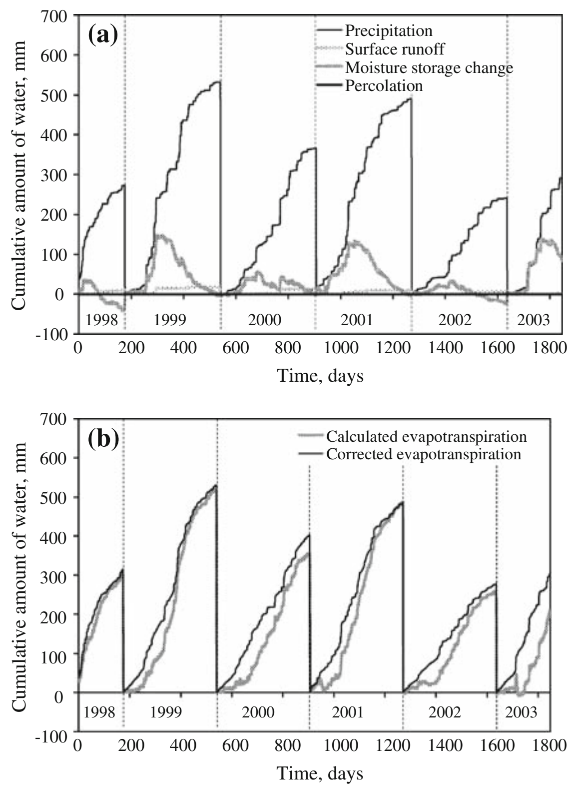

Fig. 16 Water balance variables: (**a**) Measured values; (**b**) Calculated values

cover. The test cover analyzed in this study was constructed by placing a 1,168 mm layer of low plasticity clay soil atop a large pan lysimeter (9.1 m by 15.2 m). The soil was placed at 70% relative compaction with respect to standard proctor maximum dry density (1,960 kg/m^3). The lysimeter consists of a geocomposite for water collection (consisting of a geonet for in-plane drainage sandwiched between two geotextiles) underlain by a geomembrane. The lysimeter was placed on a 3% grade, which allows gravity drainage through the geocomposite. The soil used was a low plasticity clay (CL), with an average fines content of 43%, and an average plasticity index (PI) of 15.4. The cover and surrounding buffer zone were vegetated with local grasses and shrubs, such as Cheatgrass.

Measurements obtained for water balance components in the evapotranspirative cover at the Rocky Mountain Arsenal were used to define the evapotranspiration component. The indirectly measured evapotranspiration was subsequently compared against predictions obtained using energy balance methods and numerical simulations. Monitoring commenced on July 10, 1998 (day 1), and continued until July 31, 2003. Figure 15 shows the variation in moisture content with time at three depths in the test cover along with the percolation collected from the lysimeter. The vertical dashed lines in the figures denote January 1st of each monitoring year.

This figure indicates that the time periods when percolation was collected in the lysimeter correspond with the periods of increased moisture within the cover. The surface moisture content fluctuates on a daily basis, while the basal moisture content changes in response to significant wetting events. The moisture content was integrated over the cover depth to calculate the cover moisture storage. Figure 16a shows the cumulative values for the measured water balance. Above average amounts of precipitation occurred in 1999 and 2001, which corresponds to the periods of increased moisture content observed in

Fig. 15. The cover moisture storage increases in the early portion of each year in response to higher precipitation in the spring, while it decreases in response to high evapotranspiration in the summer and fall. Runoff was minimal, but it was observed to follow the pattern of precipitation and was greatest in the spring during heavy storms. Little runoff was collected from melting snow. The percolation was a comparatively small component of the water balance, typically less than 0.02% of the precipitation.

Figure 16b shows the cumulative ET calculated on a daily basis using Eq. 14. The program REF-ET was used to calculate the potential evapotranspiration (PET) for the years 1999 to 2002 (Allen 2001). This program solves for the PET using the Penman–Montieth equation. The potential evapotranspiration from REF-ET must be partitioned into the potential transpiration T_p and the potential evapotranspiration PET. This was achieved by using the Ritchie model (Ritchie and Burnett 1971) to correlate the variation in the leaf area index *LAI* with the partitioned evapotranspiration PET. LAI corresponds to the ratio of the leaves area of plants to the area occupied by the plants. Transpiration by root uptake is modeled using a sink term in the Richards' equation at each node (Simunek et al. 1998). The Feddes model was used to calculate the actual root uptake based on the available moisture at each node and the capacity of the plants (Feddes et al. 1978). The model requires a distribution of root length density with depth, and an estimation of the range of water contents at which plants will transpire.

Figure 17a shows the calculated change in moisture content at three depths. The results shown in this figure indicate that the numerical results obtained by solving Richards' equation yield similar results to those observed in Fig. 15. However, the wetting front does not reach the base of the cover (1,080 mm) until 2003. This may be due to preferential flow in the field, or to difficulties in modeling the boundary condition representative of a lysimeter. Figure 17b shows a comparison between the simulated surface evaporation and transpiration values. This figure indicates that the surface evaporation contributes approximately 1.5 times more to the removal of water from the cover than plant uptake. The depth of influence of evaporation depends on the moisture content of the near ground surface soil. Roots remove moisture from the full cover profile, but the amount of removal depends on water availability and the season of year. Evaporation occurs throughout the year, while transpiration occurs mostly during the vegetation growing season.

Figure 17c shows a comparison between the simulated and the measured evapotranspiration. The two quantities compare quite well. The measured ET typically is slightly greater than the calculated ET. Over the four-year monitoring period, ET removed 96% of the precipitation (1,565 mm out of 1,626 mm). Negligible runoff was collected. Although Fig. 15

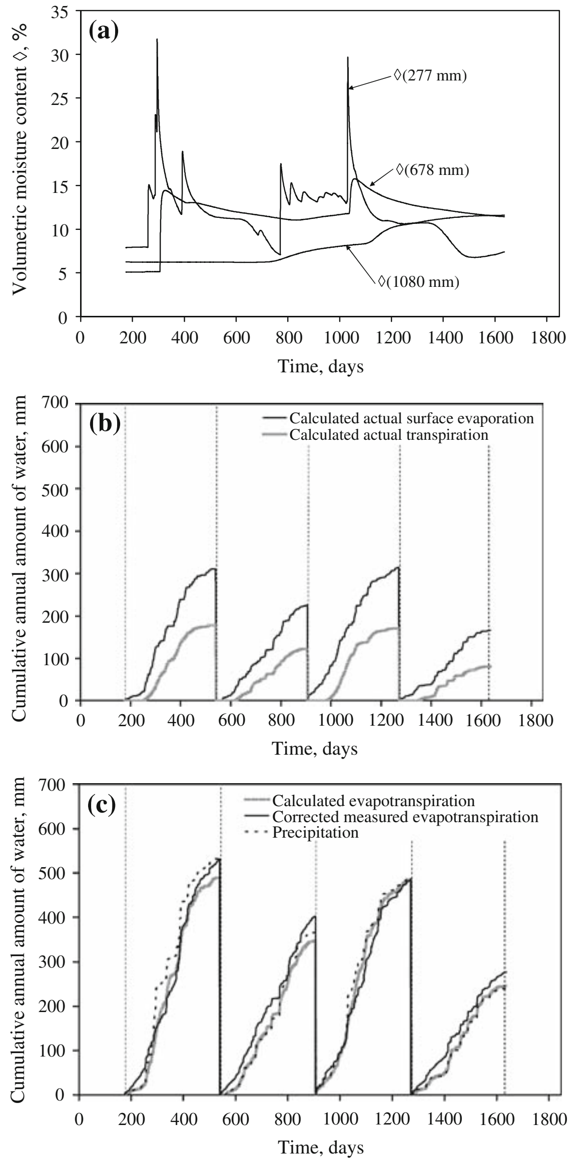

Fig. 17 Hydrus-1D results: (**a**) Moisture content at three depths (277 mm, 678 mm and 1,080 mm); (**b**) Surface evaporation and root flux (transpiration); (**c**) Comparison between calculated and measured evapotranspiration

indicates an increase in moisture content on several occasions, ET led to relatively low moisture contents throughout the soil profile at the end of the simulation. Also, the percolation throughout the four-year simulation period was less than 0.1 mm (0.02% of the precipitation), indicating that the ET adequate enough to lead to satisfactory cover performance.

8 Final Remarks

Among the various components of the water balance, the evapotranspiration component has probably been the most difficult component to quantify and measure experimentally. Some attempts for direct measurement of evapotranspiration have included the use of weighing lysimeters. However, quantification of evapotranspiration has been typically conducted using energy balance approaches or indirect water balance methods that rely on quantification of all other water balance components. This report initially presented the fundamental aspects of evapotranspiration as well as of its evaporation and transpiration components. Typical methods used for prediction of evapotranspiration based on meteorological information were also discussed. The current trend of using evapotranspirative cover systems for closure of waste containment facilities located in arid climates has brought renewed needs for quantification of evapotranspiration. Accordingly, a brief overview of evapotranspirative cover systems was presented in this paper. Finally, case histories in which direct or indirect measurements of evapotranspiration have been conducted are described and analyzed.

Overall, significant improvements have been recently made regarding monitoring of evapotranspiration using direct methods (weighing lysimeter), energy balance methods, and water balance approaches. However, significant additional advances should be made towards integrating the unsaturated soil mechanics concepts with other areas such as meteorology, agronomy, and biology in order to further advance our ability to predict evapotranspiration.

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