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Evaporation estimates in arid environments: an evaluation of some methods for the Patagonian steppe

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ABSTRACT

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Different ways of estimating soil evaporation losses were compared in an arid region of Patagonia. Microlysimeters were used as a reference estimate and tested against approaches based on (1) surface energy balance and (2) meteorological data.

Cumulative losses during a 12-day drying cycle, calculated from microlysimeter data and from the model that uses meteorological data, were very similar. The energy balance method yielded estimates higher than the model based on meteorological data and overestimated the evaporation rate, especially at low rates of evaporation. This feature could be related to the plausibility of an assumption made by the method.

INTRODUCTION

The separation of plant transpiration and soil evaporation fluxes is essential to understand the water economy of both natural and agricultural ecosystems. These fluxes are parallel ways of water loss from the soil, but are subjected to different controls. The proportion of water lost by evaporation increases with climate dryness and in arid zones can account for up to 50% or more of total precipitation (Hide, 1954, 1958). For sahelian grasslands, evaporation accounts for > 80% of the water lost by evapotranspiration (Stroosnijder and Koné, 1982).

Often, transpiration is estimated from the difference between measured (or calculated) evapotranspiration and evaporation. Several methods are available to estimate both potential and actual evapotranspiration, but there is no standard procedure to estimate evaporation (Stroosnijder, 1987).

The objective of this paper is to compare some methods for estimating evaporation losses in an arid region of Patagonia. Microlysimeters (Boast and

Robertson, 1982; Boast, 1986) were used as a reference estimate and tested against results obtained with the surface energy balance method (Ben Asher et al., 1983, 1984) and meteorological data (Ritchie, 1972). We chose microlysimeters as a reference method because gravimetry is the most direct way to estimate soil evaporation.

MATERIALS AND METHODS

Study site

The study area was a shrub steppe with < 50% of plant cover, located in the southwest of Chubut Province (Argentina) (45°30′S, 70°10′W). Mean annual rainfall is 168 mm, concentrated in the autumn and winter months (Soriano and Sala, 1983). Soils, petrocalcic Calciorthids, show a coarse texture (sandy) and a gravimetric gravel content (> 2 mm) > 47%. At 45–60 cm, the profile presents a cemented, calcareous layer (Paruelo et al., 1988).

Field measurement of evaporation

Water losses by evaporation were measured during a 12-day drying cycle during summer. Four plots of 10×10 m, in which vegetation was removed by hand, were watered with the equivalent of 30 mm at the beginning of the measurements (9 January).

Gravimetric measurements

In the centre of the plots, microlysimeters 8 cm in diameter and 10 cm deep (Boast and Robertson, 1982; Boast, 1986) were placed at ground level. They were filled with soil, watered and allowed to drain for 48 h (avoiding evaporation) to bring soil to field capacity. They were weighed daily using a 1-g precision balance to estimate water losses. Field capacity values of soil contained in the microlysimeters did not differ statistically from that of the first 20 cm of the profile. The depth of the microlysimeter was chosen taking into account that soil psychrometer data did not show significant differences among water potential at depths of 7.5, 15, 30 and 60 cm, after a 17-day drying cycle (R.A. Golluscio, unpublished data, 1989).

Energy balance method

In the energy balance method (Ben Asher et al., 1983), evaporation is calculated using the difference in surface temperature between the evaporating soil and a reference dry soil. The basis equation is

$$E=S(T_d-T_e) \qquad (1)$$

where E is the daily evaporation (J m^{-2} day^{-1}), T_d is the surface temperature of dry soil (K), T_e is the surface temperature of evaporating soil (K) and S is the following expression

$$S=8.7(\rho C_p r^{-1}+4\epsilon\sigma T^3) \tag{2}$$

where ρ is the air density (kg m^{-3}), c_p is the specific heat of air at constant pressure (J kg^{-1} K^{-1}), r is the aerodynamic resistance to heat transport (h m^{-1}) ($r=0.035\,u^{-0.96}$, where u is the wind speed in m s^{-1}), ϵ is the emissivity of the bare soil surface (0.95), σ is the Stefan–Boltzmann constant (2.04 10^{-4} J h^{-1} m^{-2} K^{-4}), T is the average of the surface temperature of drying and dry soil (K), and 8.7 is a unit conversion factor (hour to day) that incorporates a component which accounts for the periodic nature of evaporation. Evaporation values will be expressed in mm day^{-1} taking into account that 2.4×10^6 J m^{-2} is equivalent to 1 mm of evaporation.

We recorded daily the surface temperature of the drying soil and the dry soil, placed in trays at ground level. Previous exploratory data indicated that the surface temperature of dry soil in trays did not differ significantly from that of dry bare soil. Measurements were carried out near 14:00 h. We used an infrared thermometer (Barnes), taking five temperature readings of the drying soil and five of the dry soil per plot. Values were averaged to calculate the evaporation per plot. For the calculation of r (eqn. (2)), we used the 24-h average wind speed measured with an anemometer at 40 cm height.

Evaporation calculation from the Ritchie (1972) model

The model presented by Ritchie (1972) calculates evaporation from mean air temperature, radiation, relative sunshine and soil hydraulic conductivity at −0.01 MPa. Two phases could be defined in the evaporation processes, the constant and the falling rate phases (Phillip, 1957). In Phase I (wet soil), actual evaporation matches potential evaporation. As the surface dries out, available energy ceases to be the limiting factor for the process and the soil begins to control the water flux. In this second state (Phase II), cumulative evaporation is proportional to the square root of time (Black et al., 1969)

$$\Sigma E=\alpha t^{1/2} \tag{3}$$

Both the amount of water evaporated (mm) during Phase I and the slope of the relationship (3), α, are proportional to hydraulic conductivity at −0.01 MPa (Ritchie, 1972).

The meteorological data necessary for the calculation of evaporation with the Ritchie model were obtained from an agrometeorological station located 5 km from the experimental plots. Radiation values were taken from the Smithsonian Meteorological Tables (1958) and corrected by cloudiness after Feddes et al. (1978). Unsaturated hydraulic conductivity at −0.01 MPa was taken from Paruelo (1990). Potential evaporation was estimated using the Priestley and Taylor (1972) formula.

RESULTS AND DISCUSSION

Cumulative evaporation calculated for the 12-day drying cycle using microlysimeters was 14.3 mm (standard deviation (SD)=2.31, $n=4$). Estimates obtained with Ritchie's model (16.7 mm) showed small differences compared with that made from microlysimeter data. The energy balance method produced cumulative evaporation estimates about 10 mm higher than the other values (26.8 mm; SD=3.05, $n=3$) (Fig. 1). Daily values estimated gravimetrically and using both the surface energy balance method of Ben Asher et al. (1983, 1984) and the method based on meteorological data of Ritchie (1972) showed a high correlation ($r^2=0.93$ and 0.85, respectively, $n=12$, $P<0.01$).

Latent heat flux was positively correlated with T_{max} ($r^2=0.75$, $P<0.01$), as Ben Asher et al. (1983) postulated. Large cumulative values of energy balance estimates resulted from daily overestimation of evaporation rates during the drying cycle. The relative overestimation (ROv) was higher at low daily evaporation rates (DER) (ROv$=12.41-2.97\times$DER, $r=0.60$, $n=12$, $P<0.10$).

Overestimation at low evaporation rates could be related to one of the assumptions of the method: the sum of the integrated difference in short-wave radiation and the integrated difference in soil heat fluxes, between transient dry and dry surfaces, can be neglected because it is much smaller than the latent heat flux density. Nevertheless, the integral over time of the sum of

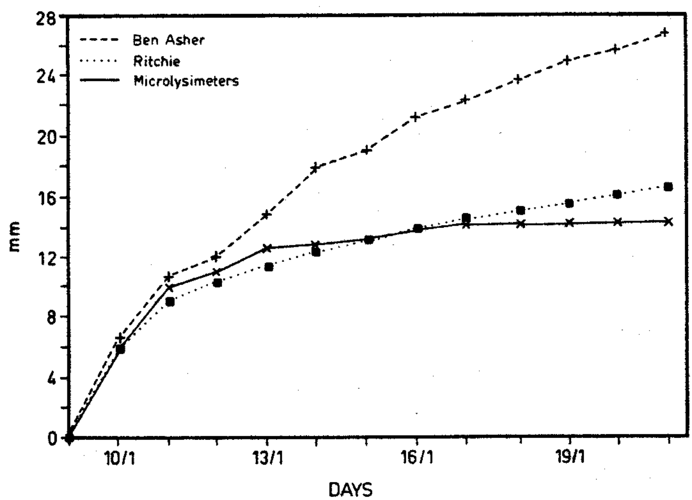

Fig. 1. Cumulative evaporation during the drying cycle, estimated in three different ways.

those differences could be negative (Ben Asher et al., 1983) and, at low evaporation values, of the same order of magnitude as the latent heat flux term. This leads to an overestimation of evaporation rates.

The selection of one of the analysed methods will depend on the objectives of the work and on time and equipment availability. Microlysimeter methodology needs a minimum of equipment and allows the monitoring of the daily evaporation course, but it cannot be used for more than one drying cycle without reinstallation.

The energy balance method would mainly produce some degree of overestimation of evaporation, according to the particular environmental conditions. Nevertheless, a site-specific calibration with gravimetry would allow its use when other methods were not feasible. It can be utilized to estimate the relative evaporation losses at very detailed scales (i.e. relative evaporation in different microsites) or in bare patches with simultaneous evaporation and water absorption by roots. Energy balance methods may also be a valuable tool to estimate evaporation from remotely sensed data (Idso et al., 1975).

The closeness between the results of Ritchie's model and those estimated from microlysimeters, if representative, ensures a reliable use of the former in the conditions of the Patagonian steppe. Its adaptation to a particular environment is related to the change in simple parameters associated with hydraulic conductivity at -0.01 MPa. This versatility, together with the few input data required, makes Ritchie's model especially useful to estimate evaporation at an intermediate spatial and temporal scale, i.e. computing evaporation in regional water balance models and/or during long time periods.

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