Ecosystem water and energy balances

JELENKA SAVKOVIĆ STEVANOVIĆ
Faculty of Technology and Metallurgy Belgrade University, Karnegijeva 4,
11000 Belgrade, SERBIA

Abstract: In this paper water and energy balance in ecosystem was studied. A number of models for plant environment interactions for the utilization of water and energy by plant, have been developed. Soil water balance, precipitation and interception, Horton overland flow and deep drainage were developed. Energy balance equation of ecosystem was derived. Environments are highly complex systems whose evolution is determined by complicated networks of positive and negative feedback loops.

Keywords: Ecosystem, environment, water, energy, balance.

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1. Introduction

A number of models for plant environment interactions and particularly for the utilization of water and energy by plant, have been developed since the 1950s . Several of these attempt to understand plant growth and water use as related to specific physiological and environmental parameters [1], [2]. These models, however, can not be applied to situations for which few data are available unless a number simplifying assumptions are used. On the regional and geographical levels, other models of a predominantly qualitative character have been suggested [3],[4]. As a consequence of this dichotomous development, attempts have been made to unify these two approaches with a view to simplifying the comprehensive models, so as to make them applicable to regional use to areas with limited data, without introducing misleading over simplifications.

In this paper water balance and energy balance of ecosystem were studied.

2. Soil-water balance

Examples of soil-water balance studies that appear to partly fill in this middle ground are those of Fuchs and Stanihll [5], Slatyer [6], Benecke [7].

The soil-water balance equation is generally written in the form:

$$\begin{aligned} & P - R_S - (GWR + TF) - ET + \\ & \Delta(SM + SW + GW) = 0 \end{aligned} \quad (1)$$

where P denotes precipitation, R_S the surface runoff, GWR the ground water runoff, TF the through flow, ET evapotranspiration, SM soil moisture, SW the perched water (above compact horizon), and GW ground water. For the sake of greater convenience, ($(GWR + TF)$ is termed deep drainage, and $\Delta(SM + SW + GW)$ is the change in soil water storage (initial minus final) during the period and for the depth of measurement. For a number of cases, deep drainage may be adequately defined as the amount of water passing beyond the root zone or, for experimental purposes, as the amount passing below the lowest point of measurement. All symbols have dimension of length, for example millimeters. Equation (1) can be written more specifically as

$$\int_{t_1}^{t_2} [(P - R_S) - (ET) - v_z] dt = \int_{t_1}^{t_2} \int_0^z \frac{\partial (SM)}{\partial t} dz dt \quad (2)$$

where ($t_2 - t_1$) is the time interval over which the measurements are made, z is the depth to the lowest point of measurements (cm), v_z is the net downward flux of water at depth $z(cm/s)$, SM soil moisture, i.e., the volumetric soil water content (cm^3 water per cm^3 of soil). P, R_S, and ET are in units of mm/s or $gcm^{-2}s^{-1}$. In the absence of a water table near the surface, v_Z is generally positive.

Compared to above ground measurements of estimates of the water vapor flux, the soil water balance approach has the advantages of ease of data processing and integration, since the soil – water reservoir (SM) automatically integrates extraction rates between observations. The disadvantages are associated largely with a somewhat lower level of measurement accuracy and the difficulty of adequately assessing evapotranspiration during periods of rainy weather. Therefore, its applicability is, to a considerable degree, restricted to regions of relatively high potential evaporation rates and sufficiently well defined alternations of rainy and dry weather.

3. Precipitation and interception

The measurement of precipitation N at a site or in a region is generally considered simpler and more straightforward than that of the other terms in the water balance equation. Marked differences in the pattern of precipitation actually reaching the ground normally develop in many plant communities because of the gross interception of precipitation by the vegetation. Subsequently, precipitation is partly transferred to the soil by channeling down the main stream "stemflow" and partly by dripping from branches, twigs, and foliage "canopy leaching", or it may be lost by evaporation from the wet surface. This latter proportion constitutes the term "net interception" . Further differences in the amount of precipitation reaching the ground between plants "throughfall" are due particularly to the disturbed wind structure and are most noticeable in the case of snow.

In view of the complicated physical nature of net interception, attempts to assess it by means of measurement and by indirect estimations are quite numerous. In the framework of a comprehensive agroclimatological model, Braden developed the following interception estimate:

$$N_i(R) = a \bullet LAI(1 - \frac{1}{1 + N_0 b / aLAI}) \quad (3)$$

where $N_i(R)$ is the net interception, $a \bullet LAI$ the saturation parameter, dependent on leaf area index (LAI) and a special-specific maximum interception a, N_0 the above –canopy precipitation, and b the density of vegetation cover. The Braden approach has the double advantage of mathematical simplicity and physical foundation, and its validity has been widely tested.

With rain, drip or stemflow is usually observed after an area rain total about $2\,mm$ has been received. However, with freezing rain or snow under conditions favoring retention on the leaves, twigs, branches, and stems (i.e. low wind, temperatures a few degrees below freezing), several times this amount may be accumulated. Stemflow is enhanced by a smooth bark and by branches and leaves that are inclined upwards. Thus, in deciduous (beech) and evergreen forests the amounts vary considerably, beech providing much more stemflow and much less interception loss, while oak would have an intermediate position.

4. Horton overland flow

Surficial runoff in the sense of Horton overland flow occurs whenever the rate of effective precipitation (i.e. precipitation net interception) exceeds the rate of infiltration (F^*) and the resultant accumulation of surface water exceeds the pondage capacity (S^*) at the point of measurement. The most important regulator is F^*. It is useful in applied hydrology and in relation to pollutant transport to characterize the dynamics of infiltration by a small number of parameters. Philip developed a simple physical model of infiltration, which is, however, closely related to more precise diffusion descriptions of infiltration [8].

$$i = St^{05} + At \quad (4)$$

where i cumulative infiltration at time t, and the constants S ("sorptivity") and A have a physical meaning related to the diffusion analysis of infiltration. The first term on the right hand side of equation (4) describes the contribution to infiltration due to capillarity; the second term represents mainly the contribution due to gravity . The differential form of equation (4) is equation (5), where v is the rate of infiltration , cm/s:

$$v = 1/2St^{-1/2} + A \quad (5)$$

Some values of the infiltration rate. mm/h For particular soils in specified conditions are given below:

Sand, loess, silt	11-7
Sandy loam	7-4
Clayey loam, soils poor in organic matter	4-1
Clays, alkaline soils (solonets)	<1

These values are much lower than those determined by means of a field rain simulator. The experimental conditions were such that a constant surface runoff rate was produced by a constant amount of rainfall when the soil had reached its saturation point. Since evaporation is negligible during the short duration of the experiment in comparison to the rainfall applied (100-250 mm/h) and the increment in soil and water storage is assumed to be zero, the difference between precipitation and surface runoff may be equated to infiltration. For two clay soils tested, the minimum infiltration capacity amounted to 58 and 60 mm/h; for sandy soils 79 mm/h was measured, while the infiltration rates of loamy soils varied between 63 and 76 mm/h. These results are, on the one hand, indicative of a marked enhancement of infiltration due to organic matter and, in particular, desiccation cracks and root voids; on the other hand, they point to high trough flow rates close to the surface (e.g. piping). As a consequence, no differences in runoff characteristics could generally be found on soils in coniferous, broadleaf, or mixed forests. Variation in the runoff rates was high when the soils were initially dry, but low when the soils were wet. A comparison of these minimum infiltration rates with the maximum net precipitation rates recorded within the last hundred years leads to the conclusion that surface runoff is an exceedingly rare phenomenon in temperate forests due to preferential seepage.

Horton overland flow appears to be a common process in semiarid and arid regions, where precipitation intensities are high and the infiltration capacity of the sparsely vegetated soil is low. It is further caused or intensified by the development of a crust on the soils because the surface layer becomes compacted and the pores blocked as a result of the redistribution of soils particles following raindrop impact. Crust formation due to lateral iron translocation is a particularly widespread phenomenon in ferric luvisols of the semi humid tropics, where it largely contributes to enhance pediplanation processes.

In temperate environments with normally modest precipitation rates and well-structured soils, Horton overland flow is the exception rather than the rule, except under certain conditions of cultivation and when the ground is frozen. In temperate environments, all the pores may become filled with water after a period of prolonged rainfall, thus saturating the soil. At this point the water table has risen to the surface and the effective infiltration capacity is consequently reduced to zero. Subsequent rainfall runs off directly across the surface of the slope as saturated overland flow. This situation is likely to come about toward the base of a slope or in microtopographical depresions on a slope where both local infiltrations and throughflow received from higher up the slope contribute to soil moisture. Smith and Parlange describe simple relationships that enable saturating or ponding times to be estimated from values of saturated hydraulic conductivity, K_s and sorptivity, s. In Morocco Imeson [9][11] found that the amount of rain required to pond the soil p_r could be estimated reasonably well with one of these equations, namely.

$$\int_0^{tp_r} p_r dt = \frac{A}{K_s} \ln \frac{r_p}{r_p - K_s} \tag{6}$$

where A is o.5s^2 and r_p is the rainfall intensity.

Because of the number of boundary conditions operative in runoff, the latter varies considerably with the amount, intensity, and duration of precipitation, as well as with slope configuration and soil fabric, which determine the degree and extent to which pondage can occur [9], [10]. In natural situations, the slope is rarely constant and, while runoff tends to reduce soil water recharge at the top of the slope and increase it at the bottom, minor changes of slope generally modify the slope-runoff interrelation.

5. Deep drainage

The deep drainage term in balance equation (1) comprises throughflow and groundwater flow and can be equated to a vertical flow that, in turn, may be calculated from hydraulic conductivity and soil water potential data. The normal equation for vertical flow of water ν_z is

$$\nu_z = K = K\frac{\delta h}{\delta z} \tag{7}$$

where K is the hydraulic conductivity, cm/s and $\delta h/\delta z$ is the rate of change of soil water suction, h, cm with depth, z, cm. Soil water suction is derived from the soil water potential $\psi, dyn/cm^2$ by the relationship $h = -\psi/\rho_w g$, where ρ_w is the density of water and g the vertical acceleration due to gravity. Unless h is very small, $\delta h/\delta z$ is usually much greater than unity so the K term in equation (7) is often negligible [6]. Under these circumstances, the deep drainage term of the water balance equation in given by

$$GWE + TF = \int_{t_1}^{t_2} \nu_z dt \tag{8}$$

where $(t_2 - t_1)$ is the time between observations.

In other situations, a net upward flux of soil water into the root zones can occur from wetter underlying soil horizons or, in particular, from water table closer to the surface. For comprehensive reviews of method available for the determination of deep drainage with a particular emphasis on ground water recharge.

Under certain soil conditions, diffuse water movement through the interganular pore spaces and voids may be supplemented by concentrated turbulent throghflow in networks of pipes. These results from large voids that exist in many soils and are enlarged by soil fauna (e.g. mice, rats, hamsters moles, weasels, ground squirrels) and the growth and decay of roots. Frequently, soil pipe soil pipes develop at the interface between organic soil and the underlying mineral soil. Discharge in completely filled pipes varies depending on pressure and gravity, and in party and gravity potentials, and in partly filled pipes in response to the gradient of the water surface. Usually, therefore pipe flow velocity is much more rapid than that matrix flow. Table 1 quotes estimates of flow velocities. From these, pipe flow may be seen to attain considerable

Table 1 . Flow velocities along different routes in a catchment [11]

Type of flow	Flow route	Velocity, m/h
Surface	Channel flow	300-10000
	Overland flow	50-500
Soil flow	pipe flow	50-500
	matrix throughflow	0.005-0.3
Groundwater flow	Limestone (jointed)	10-500
	sandstone	0.001-10
	shale	$10^{-8}-1$

importance in chemical transport, although the distances covered are normally small in comparison to channel flow. Piping is perhaps more strongly associated with semiarid areas than with humid regions. Drainage and slope development in badlands all over the world are frequently dominated by piping.

Measurement of changes in soil water storage is conducted most accurately by the use of weighing or hydraulic lysimeters, provided they are properly designated and sited [12]. Lysimeters can not be used, however, when the nature of the species compositions, the spatial structure of the vegetation cover, the depth and ramification of the root system, or other factors make it impossible to simulate the natural environment inside the lysimeter itself. A in such case s, determinations of changes in soil water storage at different points in the plant community provide the only technique for evaluating $\Delta(SM+SW+GW)$. Soil water sensing equipment may still fall short of operator requirements, although marked advances have been made since the 1980s. Probably the most commonly used techniques, at the present time, are those of neutron moderation and tensiometers or tensiographs.

6. Energy balance of ecosystem

The disposition of radiant energy at the surface of the earth is of prime importance for understanding soil-water balances and the related chemical transport and transformation processes. A formalized transcription of the energetic relationships yields the following form of the energy balance equation:

$$R_n + H_S + H_a + r_{ET} = 0 \qquad (9)$$

In this equation, which summarizes the solar energy cascade with its numerous regulation and transformation components, any form of energy that is flowing toward the surface is considered positive and any form that is moving away from it is considered negative. R_n is the net radiation flux, H_S is the heat flow in soil at the surface, H_a is the sensible heat transfer through air at the surface, and r_{ET} is the transfer rate of latent heat due to evapotranspiration. Conventionally, all terms are expressed in joules per minute and square centimeters, $4.19cm^{-2}\min^{-1}$ equals ca $1mm$ of water depth evaporated per hour.

The earth's surface receives short- wave and long- wave counter radiation from the atmosphere, while itself emitting long-wave radiation. Whereas photochemical reactions are controlled mostly by short- wave radiation , the diurnal variability of surface temperature and the related energy fluxes are very much under the influence of long-wave radiation. Thus, the net radiation flux R_n may be derived from the short and long-wave radiation balances:

$$R_n = ES(1-\alpha) + R_c - \varepsilon\sigma T^4 \qquad (10)$$

where the solar radiation ES is the sum of direct beam radiation ES_S and diffuse radiation ES_D, each term comprising short-wave and solar long-wave components R_c is the long-wave counter radiation counter of the lower atmosphere $\varepsilon\sigma T^4$ the long-wave radiation from the earth's surface (Stefan-Bolzmann radiation), α the reflectivity.

The rate at which heat is transferred down wards into the soil and subsurface substrate H_s is directly related to the nature and efficiency of the distribution mechanisms. In solids, heat is redistributed by conduction, and the flow rate depends on thermal conductivity. Units are watts per meter per degree, $Wm^{-1^\circ}C^{-1}$ or joules per centimeter per second per degree Celsius, $Jcm^{-1}s^{-1^\circ}C^{-1}$. Under steady state conditions, the flow of heat $\partial Q/\partial t$ is related to the temperature gradient $\partial T/\partial x$ and the thermal conductivity λ by

$$H_S = \frac{\partial Q}{\partial t} = \lambda\frac{\partial T}{\partial x} \qquad (11)$$

A major problem in soils is that steady state conditions are rarely achieved and thermal conductivity is a complicated function of granulometry, mineralogical composition, compaction, and water content. The conventional determination of thermal conductivity is consequently fairly difficult. An alternative parameter, thermal diffusivity, is used, which is given by $\lambda / c\rho$, where c denotes the specific heat and ρ the density, units are square meters per second. For homogenous medium, thermal diffusivity defines the rate at which temperature changes
$\partial T / \partial t$ take place:

$$\frac{\partial T}{\partial t} = \frac{\lambda}{c\rho}\frac{\partial^2 T}{\partial x^2} \tag{12}$$

The magnitude of the heat flow into the air at the surface H_a can be obtained by a difference when the other components of the energy balance in equation (10) are measured, or it can be determined directly. Among the latter approaches, two methods, the aerodynamic method Sverdrup–Albrecht method merit particular attention.

The *aerodynamic methods* is based on very precise determinations of the vertical temperature and wind profiles, and involves a horizontal homogeneity of the surface over considerable distances in the luff of the measuring station.

The *Sverdrup-Albrecht method* determines either H_a and ET as components of the energy balance in equation (13):

$$H_a = \frac{H_a}{ET}(R_n - H_s)(1 + \frac{H_a}{ET})^{-1} \tag{13}$$

The ratio H_a / RT (Bowen ratio) can be estimated from measurements of the vertical gradients of temperature $\partial T / \partial z$ and vapor pressure $\Delta e / \Delta z$ above the surface.

The latent heat flux can be estimated as a result of vaporization or condensation of water. Consequently, five general methods are used to evaluate the water vapor flux caused by evaportranspiration.

$$ET = (R_n - H_s)(1 + \frac{H_a}{ET}) \tag{14}$$

Thornthwaite and Holzman [3] suggested

$$ET = \frac{\rho k(U_2 - U_1)(q_1 - q_2)}{\{\ln(z_1 / z_2)\}^2} \tag{15}$$

where q average water vapor, ρ, g / cm^3 the air density, U, cm / s the average wind speed, z the average elevation above surface, cm and k von Karman's constant, 0.4.

A continuous record of water vapor convention by means of eddy-correlation techniques as described by Dyer [13] yields estimates of ET by

$$ET = \frac{1}{t}\int_0^t \rho wq dt \tag{16}$$

where w the momentary vertical wind speed, and q the instantaneous value of water vapor concentration. The problem of measuring ET as the time average of the vertical flux of water vapor thus becomes one of designing instrumentation capable of measuring the turbulent air motion and structure of q. This involves equipment whose response time is sufficiently short to take account of all frequencies in the turbulent spectrum contributing significantly to the flux. This requirement depends on both the height of measurement and the ability of the air, which means, in general terms, that the sensing elements should respond adequately to signals with a period of $1\,s$.

If all terms of energy budget are known except the flux due to evaporation (or evaportanspiration), the latter can be obtained by a difference. Finally the latent heat flux can be found by direct measurement, i.e., by means of weighing lysimeters on the local scale or as the difference term of the water budget of a drainage system on the regional scale.

Notation

c -denotes the specific heat, $Jg^{-1}step.^{-1}$
ET -evapotranspiration
ES -solar radiation
ES_D -diffuse radiation
ES_S -direct beam radiation
GW - ground water
GWR - groundwater runoff
H_a - the sensible heat transfer through air at the surface
H_S - the heat flow in soil at the surface
P - denotes precipitation
R_S - the surface runoff
r_{ET} -the transfer rate of latent heat due to evapotranspiration.
SM - soil moisture
SW - the perched water (above compact horizon)
TF - the through flow
R_n - the net radiation flux
q - average water vapor

U - the average wind speed, cm/s
w - momentary vertical wind speed, cm/s ,
z - average elevation above surface, cm

Greek symbols
α - reflectivity
λ - thermal conductivity, $Jcm^{-2}step.^{-1}$
ρ - density, g/cm^3

7. Conclusion

In this paper water balance equation and energy balance equation in ecosystem were studied. Soil water balance, precipitation and interception, Horton overland flow and deep drainage were developed. The measurement of precipitation at a site or in a region is generally considered simpler and more straightforward than that of the other terms in the water balance equation. Surficial runoff in the sense of Horton overland flow occurs whenever the rate of effective precipitation exceeds the rate of infiltration. The deep drainage term in balance equation comprises through flow and groundwater flow.

The disposition of radiant energy at the surface of the earth is of prime importance for understanding soil-water balances and the related chemical transport and transformation processes.

Reference

[1] J.L. Monteith, *Symp. Soc. Exp. Biol.* **29**, 205-234,1965.
[2] C. T. de Wit, *Simulation of Assimilation, Respiration and Transpiration of Crops*, PUDOC, Wageningen, 1978.
[3] C. W. Tornthwaite, *Geogr. Rev*., **38**, 85-94, 1948.
[4] L. Turc, *Ann. Agrar*., **5**, 1954,491-596.
[5] M. Fuchs, G. Stanhill, I*sr. J. Agric. Res*. **13**, 63-78,1963.
[6]R. O. Statlier, *The use of soil water balance relationships in agroclimatology* (Natural Resources Res. VII) UNESCO, Paris,73-87, 1968.
[7] P. Benecke, R. R. van der Ploeg, *Forstarchiv*, **49**, 1-7, 26-32, 1978.
[8] J. R. Philip in E. S. Hills (ed.) : *Water Resources, Use and Management,* Melbourne University Press, Melboune,, 1964, pp. 257-275.
[9] A. C. Imeson, *CATENA Suppl*., **4**, 79-89, 1983.
[10] J. Drescher R. Horn, M. de Boodt (eds.) Impact of water and external forces, on soil structure, *CATENA Suppl*., **11**, 1988.
[11]I. D. White, N. D. Mottershead, S. J. Harrison *Environmental Systems,* Allen &Unwin, London, 1984.
[12] M. Shroeder, *Dtsch. Gewasserkd Mitt*., **20**,8-13, 1976.
[13] A. I. Dyer, *Q. J. R. Meteorol. Soc*., **8**6, 401-412, 1961.