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Research Article

Model for ecosystem water circulation

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Abstract

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

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

Introduction

The water cycle refers to the continuous cycle of water as it evaporates from the surface of the earth, rises into the atmosphere, cools and condenses into rain or snow in clouds, and falls again to the surface as precipitation. This water then collects in rivers and lakes, soil, etc. and flows back into the oceans. From there, it evaporates again to reach the atmosphere.

About 71 percent of the earth's surface is water-covered, and the oceans hold about 96.5 percent of all earth's water. Water also exists in the air as water vapor, in rivers and lakes, in icecaps and glaciers, in the ground as soil moisture and in aquifers, and even in you and your dog. Water is never sitting still.

How much water is on the earth? The breakdown of where all that water resides is estimated as follows: Oceans (saline) 1,338,000,000 cubic kilometers. Ice caps and glaciers (fresh) 24,064,000 cubic kilometers.

Just a minuscule amount (1%) is in freshwater lakes, streams, and in the atmosphere. Glaciers store approximately 3/4 of earth's freshwater. This makes glaciers the largest reservoir of freshwater on earth. Finally, groundwater is the second-largest reservoir of freshwater on earth.

It suggests that most of earth's water was on the surface at that time, during the Archean Eon between 2.5 and 4 billion years ago, with much less in the mantle. The planet's surface may have been virtually completely covered by water, with no land masses at all.

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 attempts to understand plant growth and water use as related to specific physiological and environmental parameters [1], [2]. These models, however, cannot 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 [5].

In this paper water cycle 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 [6], Slatyer [7], Benecke [8].

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

$$P_r - R_S - (GWR + TF) - ET + \Delta(SM + SW + GW) = 0 \qquad (1)$$

where P_r 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.

Precipitation P_r can be expressed as:

$$\frac{\partial \rho}{\partial t} + \frac{\partial(\rho v_x)}{\partial x} + \frac{\partial(\rho v_y)}{\partial y} + \frac{\partial(\rho v_z)}{\partial z} + \frac{\partial(\rho v_i)}{\partial \xi} - D_f\left(\frac{\partial^2 \rho}{\partial x^2} + \frac{\partial^2 \rho}{\partial y^2} + \frac{\partial^2 \rho}{\partial z^2}\right) + \rho g_z + (r_i) = 0 \qquad (2)$$

The surface runoff R_S can be described in the x direction if assuming that in y, z changes are constant:

$$\rho[(\frac{\partial v_x}{\partial t} + v_X\frac{\partial v_x}{\partial x} + v_y\frac{\partial v_x}{\partial y} + v_z\frac{\partial v_x}{\partial z} + \sum_i^n \frac{\partial}{\partial \xi}(v_i v_x)] = -\frac{\partial P}{\partial x} + \mu_d\left(\frac{\partial^2 v_x}{\partial x^2} + \frac{\partial^2 v_x}{\partial y^2} + \frac{\partial^2 v_x}{\partial z^2}\right) + \rho g_X \qquad (3)$$

The ground water runoff GWR and through flow TF can be expressed, assuming changes in the x, y, z direction, its deep drainage:

$$\rho[(\frac{\partial v_x}{\partial t} + v_X\frac{\partial v_x}{\partial x} + v_y\frac{\partial v_x}{\partial y} + v_z\frac{\partial v_x}{\partial z}] = -\frac{\partial P}{\partial x} + \mu_d\left(\frac{\partial^2 v_x}{\partial x^2} + \frac{\partial^2 v_x}{\partial y^2} + \frac{\partial^2 v_x}{\partial z^2}\right) + \rho g_X \qquad (4)$$

$$\rho[(\frac{\partial v_y}{\partial t} + v_x\frac{\partial v_y}{\partial x} + v_y\frac{\partial v_y}{\partial y} + v_z\frac{\partial v_y}{\partial z}] = -\frac{\partial P}{\partial y} + \mu_d\left(\frac{\partial^2 v_y}{\partial x^2} + \frac{\partial^2 v_y}{\partial y^2} + \frac{\partial^2 v_y}{\partial z^2}\right) + \rho g_y \qquad (5)$$

$$\rho[(\frac{\partial v_z}{\partial t}+v_x\frac{\partial v_z}{\partial x}+v_y\frac{\partial v_z}{\partial y}+v_z\frac{\partial v_z}{\partial z}]= -\frac{\partial P}{\partial z}+\mu_d(\frac{\partial^2 v_{z_y}}{\partial x^2}+\frac{\partial^2 v_z}{\partial y^2}+\frac{\partial^2 v_z}{\partial z^2})+\rho g_z \quad (6)$$

Evapotranspiration ET is expressed as:

$$\frac{\partial \rho}{\partial t}+v_x\frac{\partial \rho}{\partial x}+v_y\frac{\partial \rho}{\partial y}+v_z\frac{\partial \rho}{\partial z}+\frac{\partial (v_i\rho)}{\partial \xi}- D_f(\frac{\partial^2 \rho}{\partial x^2}+\frac{\partial^2 \rho}{\partial y^2}+\frac{\partial^2 \rho}{\partial z^2})-\rho g_z+(r_i)+ k_L(\rho_L-\rho_G)=0 \quad (7)$$

Soil moisture SM and ground water GW can be expressed:

$$\frac{\partial \rho}{\partial t}-D_f(\frac{\partial^2 \rho}{\partial x^2}+\frac{\partial^2 \rho}{\partial y^2}+\frac{\partial^2 \rho}{\partial z^2})+\rho g_z \quad (8)$$

The perched water (above compact horizon) SW is described as:

$$\frac{\partial \rho}{\partial t}+v_x\frac{\partial \rho}{\partial x}+v_y\frac{\partial \rho}{\partial y}+v_z\frac{\partial \rho}{\partial z}+\frac{\partial (v_i\rho)}{\partial \xi}- D_f(\frac{\partial^2 \rho}{\partial x^2}+\frac{\partial^2 \rho}{\partial y^2}+\frac{\partial^2 \rho}{\partial z^2})+\rho g_z=0 \quad (9)$$

where ρ density, v geometrical velocity, D_f effective diffusion coefficient, g gravity acceleration, r_i reactions, x, y and z space coordinate, ξ attribute of interest, and t is time. P is pressure, μ_d viscous diffusivity, k_L interphase transfer coefficient.

Equation (1) can be written more specifically as:

$$\int_{t_1}^{t_2}[(P_r-R_S)-(ET)-v_z]dt=\int_{t_1}^{t_2}\int_0^z\frac{\partial (SM)}{\partial t}dzdt \quad (10)$$

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, 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.

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.

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:

$$N_i(R) = a \bullet LAI(1 - \frac{1}{1 + N_0 b / aLAI}) \qquad (11)$$

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 [9].

$$i = St^{05} + At \qquad (12)$$

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 (12) describes the contribution to infiltration due to capillarity; the second term represents mainly the contribution due to gravity. The differential form of equation (12) is equation (13), where v is the rate of infiltration, cm/s:

$$v = 1/2St^{-1/2} + A \qquad (13)$$

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

Sand, loess, silt	s11-7
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 depressions 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 [10]-[13] found that the amount of rain required to pond the soil p_r could be estimated reasonably well with one of these equations, namely.

$$\int_0^{tp_r} p_r dt = \frac{A}{K_s} \ln \frac{r_p}{r_p - K_s} \qquad (14)$$

where A is 0.5s^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 [10], [11]. 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 v_z is

$$v_{z'} = K = K\frac{\delta h}{\delta z'} \qquad (15)$$

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 ψ, 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 (15) is often negligible [7]. Under these circumstances, the deep drainage term of the water balance equation is given by

$$GWE + TF = \int_{t_1}^{t_2} v_z dt \qquad (16)$$

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 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 [12]

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 [14]. Lysimeters cannot 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.

Notation

D_f - effective diffusion coefficient, cm/s

g - gravity acceleration, cm/s^2

GW - ground water

k_L - interphase transfer coefficient, cm/s

P - pressure, Pa

GWR - groundwater runoff

P_r - denotes precipitation

R_S - the surface runoff

r_i - chemical reaction

SM - soil moisture
SW - the perched water (above compact horizon)
TF - the through flow
t - time, s
v geometrical velocity, cm/s
x, y and z space coordinate
z' - average elevation above surface, cm

Greek symbols
α - reflectivity
ξ - attribute of interest
μ_d - viscous diffusivity
ρ - density, g/cm^3

Conclusion

In this paper water balance equations in ecosystem were developed. Soil water balance equations, precipitation and surface runoff were derived. 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.

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, 491-596, 1954.
5. J.Savkovic Stevanovic, Ecosystem water and energy balances, International Journal of Applied Physics, **7**, 48-53, 2022.
6. M. Fuchs, G. Stanhill, Isr. J. Agric. Res. 13, 63-78,1963.
7. R. O. Statlier, The use of soil water balance relationships in agroclimatology (Natural Resources Res. VII) UNESCO, Paris,73-87, 1968.
8. P. Benecke, R. R. van der Ploeg, Forstarchiv, 49, 1-7, 26-32, 1978.
9. J. R. Philip in E. S. Hills (ed.): Water Resources, Use and Management, Melbourne University Press, Melbourne, pp.257-275,1964.
10. A. C. Imeson, CATENA Suppl., 4, 79-89, 1983.
11. J. Drescher R. Horn, M. de Boodt (eds.): Impact of water and external forces, on soil structure, CATENA Suppl., **11**, 1988.
12. I. D. White, N. D. Mottershead, S. J. Harrison Environmental Systems, Allen &Unwin, London, 1984.
13. M. Shroeder, Dtsch. Gewasserkd Mitt., 20,8-13, 1976.
14. A. I. Dyer, Q. J. R. Meteorol. Soc., 86, 401-412, 1961.

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