WATER EXCHANGE BETWEEN LAND AND ATMOSPHERE

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Contents

Summary

Since the hydrological atmosphere-land (AL) system is an integral interacting system, there must be joint analysis of atmospheric and land-based branches of the hydrological cycle. This article discusses issues of accuracy of the calculation of atmospheric and land water balance components and shows that with a sufficiently dense network of aerological stations the atmospheric water balance (AWB) equation works out most accurately for areas larger than 10^6 km^2 and with an averaging period of at least a few days days, i.e. on a scale that filters out much of the synoptic variability.

An equation is proposed for the relation between atmospheric and land water balances, reflecting interrelation between all the parameters that determine moistening (M). In particular, it connects together the processes of atmospheric moisture transport and the processes of water motion in soil and deeper under the ground. It works only with sufficiently large space and time averaging scales when evaporation is determined mainly by climatic factors, and the net effect of local (landscape) factors may be neglected.

The basics of the semi-empirical water circulation theory are discussed. This is a constituent part of AWB defining in more detail the exchange processes of moisture of external and internal origin. Estimates of the parameters of moisture circulation over the continents are presented, along with discussion of the factors of formation of land surface M from the point of view of joint analysis of moisture exchange in the AL system. Characteristics of M for various territories of the Earth are presented.

1. Interrelation of moisture exchange processes in the atmosphere–land system

The hydrological atmosphere-land (AL) system is a major component of the hydrological cycle (HC) and an indispensable condition for existence of human civilization. Investigation of processes in the AL system began more than a century ago, but for many decades study of atmospheric humidity and on-land hydrological processes developed independently of one another. It was only after 1950, when the well known work by Benthon, Blackbourne and Snead was published, devoted to water balance analysis of the Mississippi River using atmospheric moisture flows, that the study of moisture exchange in the AL system began as an integral interacting system. Then, in the works by V. Starr, A. Oort, J. Peixoto, E. Rasmusson, O. Drozdov, L.Kuztnestova, V. Malinin, and others, regular features of moisture exchange in the AL system and specific features of synoptic, seasonal, and annual oscillations of moisture exchange characteristics were established. Methods for calculation, parametrization and long-term forecast of difficult-to-determine water balance components and so on were then developed.

First of all we should consider the basic equations of the atmospheric and land components of HC. For the purpose, let's choose an arbitrary volume in the system AL, the upper boundary of which is the surface at a height of z, where air humidity is close to zero, and the lower boundary is the surface where water exchange with the soil layer below may be ignored. The water balance equation for the land surface for any arbitrary time interval must include atmospheric precipitation P, surface and underground inflows of water Q_{sl} and $Q_{u,}$ evaporation from land surface E, surface and underground outflows Q_{sl} and Q_{u2}, giving rise to changes in total surface and underground water outflows that cause changes in total storage of surface and ground water ΔS. As a result, we have the following water balance equation:

$$\Delta S = P + Q_{s1} - Q_{u1} + E - Q_{s2} - Q_{u2}. \qquad (1)$$

In hydrological calculation practice the given equation may be used for any natural objects. Depending on the purpose of investigations, availability of initial data, type of water object and its sizes, equation (1) may be either simplified or complicated. For example, for a river basin where surface and subsurface drainage divides coincide ($Q_{s1} = Q_{u1} = 0$), water balance equation will acquire the following form

$$\Delta S + Q = P - E, \qquad (2)$$

where $Q = Q_{s2} + Q_{u2}$ – river flow in the outlet of the basin. Random errors in determination of water balance components essentially depend on space and time averaging scales. For example, for month-long intervals and for large basins, random errors usually amount to: 5% for Q, 10% for P, 30% for E, and 100% for ΔS. With increase in the area and a sufficiently dense network of stations, the errors in water balance components usually decrease.

Atmospheric water balance (AWB) is composed of changes in water vapor content ΔW_v and water content ΔW_w (in clouds), difference between horizontal inflow and outflow of water vapor $F_{v1} - F_{v2}$ and water $F_{w1} - F_{w2,}$ total evaporation from land surface E and precipitation P. Horizontal transport of water in the solid phase is considered to be negligibly small. The AWB equation for any arbitrary time interval may be written as follows

$$\Delta W_v + \Delta W_w = F_{v1} + F_{w1} - F_{v2} - F_{w2} + E - P. \qquad (3)$$

Water content and transport are the terms that are most difficult-to-determine in equation (3). As to water storage of the clouds, it is measured sufficiently reliably by means of satellites. As a result of multiple measurements it was established that the ratio W_w/W_v is small at all latitudes and, as a rule does not exceed 1%. Estimation of liquid-drop water transport in clouds is, however, much more complicated. To date there are no reliable measurements of this component of AWB. Since atmospheric water storage is small, it is difficult to expect that water transport would make up a considerable share of water vapor transport. Model calculations have shown that when averaging over a sufficiently long time interval the difference between inflow and outflow of moisture even over not very large flat territories becomes small, i.e. $F_{w1} - F_{w2} = 0$. However, for regions with clearly pronounced orography, the resulting inflow of moisture in liquid phase cannot be neglected. Apart from mountainous regions, orographic effects take place under transition of moisture flow from ocean to land. In this case favorable conditions are also created for convective activity which leads to formation of clouds and fall out of additional amounts of precipitation. To neglect resulting inflow of liquid moisture would be the same as adopting a hypothesis of instantaneous fall out onto the underlying surface of all the water vapor condensed in an atmospheric column. This hypothesis is among the basic ones in calculation of precipitation using numerical models of general circulation in the atmosphere and ocean.

Giving due account for these assumptions and normalizing equation (3) by area, let's put it down in the integral form

$$\partial W/\partial t + \mathrm{div}\, \mathbf{F}_v = E - P, \qquad (4)$$

where $\mathbf{F}_v$ is vertically integrated horizontal flow of water vapor. It is in this form that expression (4) has become widespread as the AWB equation, although in practical terms it is an equation of water vapor balance. However, this difference in terminology is not a fundamental one, especially when studying large-scale hydrological processes. A distinctive feature of the AWB equation is that its left and right sides are determined with the use of fundamentally different observation systems. The left side (atmospheric components) is calculated using data from aerological sounding of the atmosphere, while its right side (hydrological components) may be determined with data from land-based stations.

As regards changes in atmospheric moisture content (MC), i.e. $\partial W/\partial t$, the contribution of this component is considerably smaller as compared to div F_v, but it essentially depends on the averaging period. With decrease in τ, the contribution of $\partial W/\partial t$ to AWB equation increases; for small τ under sharp transformation of air masses its value becomes comparable with div $\mathbf{F}_v$. The role of $\partial W/\partial t$ becomes appreciable when evaporation approximately corresponds to precipitation, i.e. div $\mathbf{F}_v \to 0$. It is natural that maximum contribution to AWB equation is made by div $\mathbf{F}_v$ value, and at the same time it is this value that brings lots of difficulties to calculations. The accuracy of div $\mathbf{F}_v$ calculations depends on a great many factors including instrumental errors of wind speed and air humidity, the method for numerical realization of div $\mathbf{F}_v$, and systematic errors arising when approximating the daily moisture flow by instantaneous observations. In their turn, these errors depend on many other factors, including the number and representativity of aerological stations, homogeneity of observation systems, the number of sounding periods, terrain relief, etc. In general, the error of calculations of div $\mathbf{F}_v$ due to net effect of all types of errors decreases with increase in the area A and averaging period τ. However, opinions of many different scientists on minimum values of A and τ, ensuring "sufficient" accuracy of calculations of div $\mathbf{F}_v$ differ considerably. For example, summarizing equations (2) and (4) for mean long-standing conditions it is easy to have

$$\text{div } \mathbf{F}_v = -Q. \tag{5}$$

Taking into consideration the high accuracy of determination of mean long-term flow of large rivers, this equation may serve as a criterion for assessment of the accuracy of div $\mathbf{F}_v$ as a function of the area A. Therefore, using estimates of div F_v for a number of big river basins of Russia and North America it is not difficult to plot its relative error η^1 as a function of area (see Figure 1). It is seen from Figure 1 that relative error η^1 (in %) decreases with increase in the area following a non-linear law, and may be approximated by the following expression

$$\eta' = \begin{cases} 154\text{A}^{-1,45}, & \text{A} \le 14.8 \\ 3, & \text{A} \ge 14.8 \end{cases} \tag{6}$$

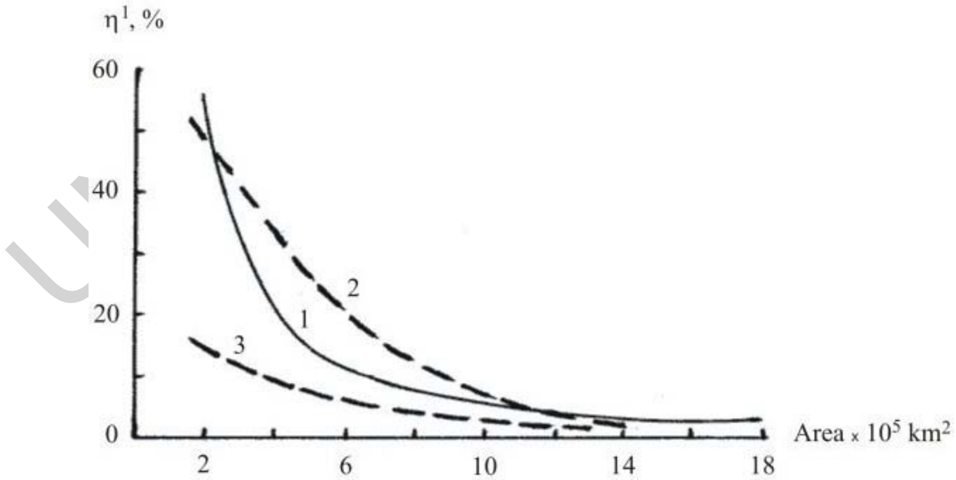

Figure 1. Dependence of relative computation error of div F_v on area, under long-term annual conditions. 1 – from the author's data; 2 and 3 – from the data of E. Rasmusson.

Curves 2 and 3 in Figure 1 were obtained from model calculations made by E. Rasmusson under certain assumptions. For example, curve 2 corresponds to normal annual flow equal to 50 mm/year, and curve 3, to 200 mm/year. These curves may be considered as upper (insufficient M) and the lower (excessive M) limits of dependence

(6) obtained from experimental data. Divergences between "theoretical" and empirical curves are small for A > 10^6 km^2—the value of η^1 makes up less than 5%. This means that the accuracy of determination of div$\mathbf{F}_v$ and Q for long-term period are comparable with each other when A ~ 10^6 km^2.
To estimate the accuracy of calculation of div$\mathbf{F}_v$ as a function of averaging period, all the components of the AWB equation were calculated for the Volga river basin up to the outlet of the city of Samara (A = 1.2 ×10^6 km^2) for the winter period of 1966 from instantaneous data. As a result 90 daily fields were obtained for each of the terms of equation (4). These were then successively summed up to 9 days, inclusive. After that correlation coefficients r for the left and right sides of equation (4) were calculated, and for precipitation and div$\mathbf{F}_v$ for all averaging intervals τ. Correlation curves were then plotted reflecting the closeness of the relationship of these characteristics (see Figure 2). These curves are approximated by the following expressions

$r = -0.94\exp(-0.63\tau) + 0.94$,
$r = -0.75\exp(-0.58\tau) + 0.75$.

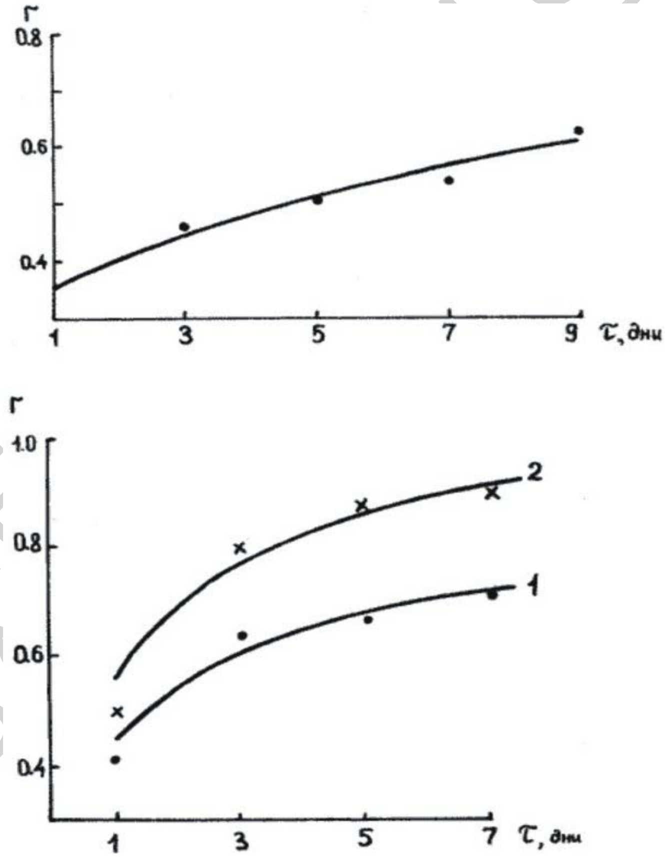

Figure 2. Correlation dependencies between P and div$\mathbf{F}_v$ (1) and between E – P and ∂W /∂t + div$\mathbf{F}_V$ (2) for different averaging intervals in a summer (a) and winter (b) periods for the Volga basin

From here it follows that when τ increases to seven days, r rapidly increases, but under further increase of τ, the value of r changes only slightly. For example, when τ increases from 7 to 20 days, r increases only by a few hundredths. Similar results were obtained when comparing values of P and div $\mathbf{F}_v$ for a summer period. But in this case the minimum averaging period proved to be a bit longer than seven days; this is connected with greater influence of evaporation in moisture exchange. Consequently, it can be assumed that a 7-day averaging period is minimal and sufficient, for measured and calculated AWB equation components to be consistent with each other. Thus, under a sufficiently dense network of aerological stations, the AWB equation works most accurately for areas of more than 10^6 km^2 and an averaging period exceeding several days, i.e. filtering out synoptical variability.

The AWB equation opens considerable opportunities for estimation of some WB components that are difficult to determine by traditional hydrological methods. First of all it is true for total evaporation from large areas that may be presented in the following way

$$E = E_{1a} + E_{sw} + E_g + E_{un} + E_{sn} + E_t + E_d + E_{ur}, \quad (7)$$

where E_{1a} – evaporation from water surface, including that of lake, water storage basin and channel network; E_{sw} – evaporation from swamps; E_g , E_{un} –evaporation from soil and underground water; E_{sn} – evaporation from snow and glaciers; E_t –transpiration of vegetation; E_d – evaporation of the portion of atmospheric precipitation retained by vegetation; E_{ur}, - evaporation from urban territories. In order to estimate E by traditional hydro-meteorological methods, usually the value of evaporation in certain points is determined, and then averaging over the area is performed. Due to the variety of natural landscapes and terrain relief, the task of determining evaporation for large regions becomes extremely difficult. Moreover, almost all modern methods only allow the calculation of a certain part of total evaporation. More adequate methods for estimation of total evaporation are equations of water balance for river basins and the atmosphere. The first of them makes it possible to calculate the climatic normal E from flow and precipitation data. Space and time limitations for AWB equations are considered above.

Summing up equations (2) and (4), we obtain a water balance equation for the system atmosphere–lithosphere with the following form

$$\Delta S + \Delta W = - \mathrm{div}\mathbf{F}_v - Q \quad (8)$$

Value ΔS contains almost the same components as total evaporation in (7).

Variations in ground water storage are especially difficult to estimate. As shown above, random errors in determination of ΔS by traditional methods may exceed 100%. Therefore, using aerological and hydrological data one may determine the value of ΔS for large regions with accuracy sufficient for practical purposes.

And, finally, equation (3), apart from control over the accuracy of div$\mathbf{F}_v$, may be used for assessment of river water flow divergence (local flow) for arbitrarily chosen territory (for example, administrative area, republic, state etc.). In this case estimation of the local flow becomes difficult if there are no hydro-meteorological stations at the boundaries of the territory,

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Biographical Sketch

Valery Nicolaevich Malinin was born in 1948. Having graduated from Sea Academy after adm. Makarov, he succeeded to the speciality of oceanologist. He worked in the Arctic and Antarctic Scientific Research Institute, and the State Hydrological Institute. From 1981 to the present he has been working in the Russian State Hydrometeorological University, where he progressed from teacher to professor. In 1978 he took a Ph.D (Geography) degree and in 1994 a D.Sci (Geography) degree. He has been a professor since 1996. He is the author of more 100 printed works, including six monographs and five textbooks, including:

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- The Problem of Forecasting the Level of the Caspian Sea. (1994), RSHU Publ., 342 p. (in Russian),
- Sea Ice and Climate (co-author) (2000), Gydrometeoizdat, 91 p. (in Russian), and
- Water Balance of the Atmosphere as a Hydrological Problem (co-author) (1988). LSU Publ, 200 p. (in Russian).
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