MOUNTAIN RIVER FLOOD RUNOFF THEORY,
MEANS OF ITS INVESTIGATION AND CALCULATION

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ABSTRACT

Results of extensive experimental studies of mountain slope runoff formation carried out by a number of institutes in the USSR are treated. A theory of rain water mountain runoff and a substantiation of regional design formulae elaboration methods are presented.

In this case a great variety of mountain runoff genetic types, the basic being the surface and contact ones, is indicated. The latter takes place in friable surface sediments on an underlying relative confining layer—a slightly fissured rock or loam-skeleton layer.

The contact layer has a number of subtypes;

1. Draining — in extensive natural drains washed by water on landslide mount slopes with abundance of fine earth;
2. Water blanket — in rubble and weathering residue layers;
3. Fissure — in largely fissured rocks;
4. Runoff — in stone placer deposits;
5. Lode runoff, etc.

The theory of contact and the above mentioned types of surface runoff is developed on the basis of integration of differential runoff equations in partial derivatives with a subsequent approximation of complex mathematical formulae. As a result, theoretical formulae determining a slope flood runoff component, a maximum runoff module and a surface and contact flow discharge hydrograph equation are obtained.

Parameters of different types of slope flow time overrunning and parameters of losses (surface infiltration in case of surface runoff and storage; infiltration into the underlaying rock and soil storage in case of contact runoff) are the main components of theoretical equations.

The performed investigations made it possible to substantiate the regional formula calculation method for layer and slope inflow discharge hydrograph computation.

RÉSUMÉ

Des résultats d'études expérimentales étendues de la formation de l'écoulement sur les pentes montagneuses, réalisées par nombre d'instituts en URSS, sont discutés. Une théorie de l'écoulement des pluies en montagne et une discussion des formules régionales sont présentées.

Une grande variété d'écoulements de montagne de divers types sur la base de la surface et du contact sont indiqués. Le contact se fait dans une couche sédimentaire friable reposant sur une roche légèrement fissurée ou sur une couche à squelette limoneux.

La couche de contact présente divers types différents.

1. Drainante — dans des drains naturels étendus lavés par l'eau sur les pentes avec de la terre fine en abondance.
2. Dans des couches résiduaires en décomposition.
3. Fissurée — avec des roches largement fissurées.
4. Écoulement — dans des dépôts pierreux, etc.

La théorie du contact avec les types mentionnés ci-dessus est développée à l'aide de l'intégration d'équations différentielles aux dérivées partielles exprimant l'écoulement avec une précision suffisante. En conclusion, des formules théoriques déterminant les composantes de la crue sur les pentes, un module d'écoulement maximum et une équation de l'hydrogramme du débit en surface et au contact sont obtenus.

Des paramètres pour les différents types d'écoulements de pente et des paramètres de pertes (infiltration de surface dans le cas d'écoulement et d'emmagasinement de surface; infiltration dans la roche sous-jacente et emmagasinement dans le sol en cas d'écoulement de contact) sont les principales composantes des équations théoriques.

Les recherches réalisées permettent de discuter les formules de calcul régionales pour la recherche de l'hydrogramme de débit.
Runoff formation processes, especially those from mountain slopes are distinguished for their exceptional variety. Strictly speaking runoff agents are found in their unique combinations on any specific watershed, thus assigning individual characteristics to any of the runoff hydrographs. A mathematical theory of runoff processes may establish only their most general characteristics meeting a certain abstract idealized model of the phenomenon. To pass from theoretical to design formulae it is necessary to study the runoff agents, their distribution curves and indices, and subsequently on the basis of the experimental data, to estimate the variables of the theoretical equations. To solve this task one has to combine observational and experimental data analyses. Only such a procedure permits one to fill the gap in the rather scanty information about each type of these experiments.

Active experiments and observations of rainfall runoff in mountains have been used widely in recent years. These added greatly to the development of a mountain slope runoff theory. Discussed below are the peculiarities of rainfall runoff formation in mountainous areas, main conclusions of the theory and methods of computation of the flood design formulae.

From the hydrological point of view the cover mass of mountain slopes may be divided into three layers, the lower one being a relative confining layer comprising a slightly fissured rock or a skeleton clay layer which emerged on the rock in the course of friable rock kaolinization. Sometimes permafrost or a ground-water surface plays the part of a confining layer. The intermediate stratum is the drainage layer pierced through by thick macropores washed in loose talus ground by atmospheric precipitation water. Washing out of coarse water-conducting cavities is accompanied by a colmatage of the more finely porous surrounding plots of ground, resulting in a heterogeneous structure of the drainage layer. Sometimes for lack of aleurite or a great aggregation of rock debris the mentioned layer is represented by a continuously washed stone bed, crushed stone, or landwaste. Sometimes it is of a fibrous nature, i.e. the washed cavities are surrounded by silted ground; often it is a fine-grained rock mass pierced by rather infrequent coarse drains (spaced up to 1 metre and sometimes even more). The upper layer is a mantel surface layer. Its properties vary greatly. Now it is of a great thickness, now it disappears altogether (bare placers, regions of exposed landwaste, and so on). Depending upon its texture, slope steepness, forestation and agricultural utilization of slopes, the permeability may vary within a fairly wide range. Nevertheless, it is always smaller than that in the underlying draining layer.

In broad-leaved pine and thick mixed forests with rough soil texture overcrowded with plant roots and macropores, rainwater is almost immediately falls through down to the confining layer and in contact with the latter it is removed into the draining layer. This is the way the near-surface runoff emerges. It can also be termed as contact runoff. One distinguishes bed, vein, drainage, and fissured contact runoff. All the listed runoff forms are characterized by common hydraulic features, viz, a turbulent regime and lack of relationship between the depth of flow and its mean velocity, i.e. the time constancy of the latter. As a matter of fact, along with coarse water-conducting cavities (drains) the drainage layer has a lot of small pores. Their diameter is rather small and water flow through them is of a laminar nature. The lengths of the latter are negligently short and they are feeding the drains of the turbulent regime.

It is through these drains that the stored slope water is being discharged. Similar to the fine-grained rock mass of the draining bed, the mantel soil layer also discharged into the channel network not directly but through the underlying drainage bed. The channels can also be directly supplied with runoff only when the soil layer is flooded. At the beginning of runoff, so long as the large drains are filled a little, the velocity grows from zero to its normal value which is repeated in reverse order at the end of the recession when the most fine-grained beds of the surrounding soil discharge into drains. It is however experimentally established that the periods with an alternating
velocity are short-termed, particularly during the stage of rise. Thus the velocity of
the contact layer may be estimated according to the formula:

\[ v = a_0 \sqrt{\beta} \]  

(1)

where \( \beta \) is the slope (a relative value), and \( a_0 \) - the velocity parameter, depending upon
the draining layer structure (bed, drain, vien, etc.). Nevertheless, as has been shown
by exploration, the mean value of the \( \alpha \)-parameter (considering the width of the slope)
varies for different kinds of the drainage layer on a narrow range (4). The velocities
of the constant layer may be rather considerable often amounting up to 10 per cent
of surface velocities or even more. Contact runoff is apparently one of the genetic
categories of the flood runoff. In comparison with the surface floods on small drainage
areas the contact ones are highly spread out; but on large drainage areas where the
main factor of runoff transformation is not the slope but the channel travel the surface
and contact runoff floods present merely an identical picture.

Low flow of mountain rivers is formed at the expense of infiltration into the near-
surface confining layer. Percolating through deep fissures, displacement and fault
planes, water reaches the horizons which are feeding the drainage network.

Deepseated infiltration may be either uniformly distributed over the area (for instance,
in case of a skeleton clay confining layer) or it can be of a focussed nature as is the case
on monolith rock with chiefly obtussed crevassing. The deep-seated infiltration inten-
sity is practically constant in time during the wet periods of the year, when the capaci-
ties of small pores and fissures at the surface of the confining layer are filled up.

In the course of an abundant rainfall the whole permeable mass of the soil is
easily filled, especially during the wet periods of the year. At this time not only the fine
pores are filled up, but the contact water which remained from the previous rainfall
may also be stored in coarse drains. Then, a backwater surface runoff emerges. As the
relative confining layer has a more or less wavy outline, one can observe a transversal
concentration of contact water in depressions (talwegs) of the underlying layer. Here
the permeable layer is partially silted, sometimes it is eroded, or even fully washed out.
That causes a rapid emergence of surface run-off over slope hollows, which is fed with
local water formations and contact inflow rapidly saturating the rock. Under
the described conditions, high and sometimes moderate floods are composed of both
contact and backwater surface runoff waters. In this case, depending upon the degree
of slope undulation contact and surface streams may be either independent over the
entire length of the slope or mixed, i.e. first contact and then superficial (in hollows).

If there is a mantel ground layer of a comparatively low permeability, a suspended
surface runoff emerges on mountain slopes, which is the result of surplus of rainfall
intensity over the intensity of infiltration. In dense mixed broad-leaved forests the mantel
clay layer is cut through with macropores (from the overrotten roots, at tree trunks,
etc.). Therefore, synchronously with the progress of suspended surface runoff, owing
to collapse imbibition into macropores, a contact runoff is also formed, which is usually
more abundant. The soil-ground capacities having been inundated, the suspended
runoff is converted into a backwater one. On not very steep open sites of slopes, here
and there where the forest is thin and is being used for pasturing, the macropores are
entirely or almost completely colmatated, and the runoff is of purely surface nature,
as it is the case on plane territories. On steep grassland slopes, where macropores are
supported by displacement of debris by rock-slides and also in fir-wood forests there
exist transient conditions, viz. the surface runoff is more frequent but its ratio to the
contact one is defined by the ground layer texture, density of macropores free of silt,
and progress of colmatated microstream network.

The theory of runoff from mountain slopes may be elaborated on the basis of
several models of the phenomenon which are different in the accuracy of reproduction
of its main peculiarities. It is easier to solve a plane task, based on data analysis of a
metre-long slope-band with constant gradient length, depth of the permeable layer and its filtration parameters. In this case the run-off from natural slopes is estimated by integration of its equations for an elementary plane with allowance for distribution curves of the depth of permeable layer and the length of contact travel of the flow. A spatial task may be solved as well. This, however, leads to very complex equations, the actual application of which due to the deficient investigation of transverse concentration parameters does not give any extra advantages. It is expedient to assume the treatment of runoff under collapse imbibition (a plane task) as a principal of the theory since the solutions so derived may also become the basis for deduction of formulae applicable for slopes with a heavy mantel layer. The task may be solved with a good precision if the differences in the properties of the draining and mantel layers are taken into consideration. While elaborating the contact runoff theory one can nevertheless average the parent parameters (accumulation coefficient and coefficient of drainage water yield over the whole permeable layer depth since both variants of the study give rather close results).

For conditions involved in a given simplified model of the phenomenon it is easy to get the following differential equation of runoff:

1. For contact component:

\[ v \delta_a \frac{\partial y_c}{\partial x} + \delta \frac{\partial y_c}{\partial t} = q_{xt} \]  

Here \( y_c \) is the depth of the contact stream at the moment \( t \) at the distance \( x \) from the source; \( v \) is the velocity of the contact stream defined by the formula (1), \( \delta_a \) is the coefficient of drainage water yield equal to the ratio of the drain areas or the ratio of washed-out part of the drainage bed to the section of the whole permeable soil-ground mass; \( \delta \) is the accumulation coefficient equal to the ratio of the total free porosity of the layer to its volume; \( q_{xt} \) is the lateral inflow per running metre of slope band, \( q_{xt} = 0 \) being the equation for water formation (rain formation) stage \( q_{xt} = -k \) for the recession stage, and \( q_{xt} = 0 \) for the period of surface flowing with backwater runoff. The intensity of contact water formation equal to the difference between rain intensity \( h_0 \) and infiltration intensity into the underlying layer \( k \) is denoted by \( h_t \).

2. For surface runoff component:

\[ \frac{3}{2} c y^{0.5} \frac{\partial y}{\partial x} + \frac{\partial y}{\partial t} = q_{xt} \]  

Here \( y \) is the average conventional depth of surface sprays (attributed to the entire width of the slope), \( q_{xt} = h_t \), or \( k \) are the same as for contact runoff, and \( c \) is the parameter of the test formula of surface velocity

\[ v = a_n \sqrt{3y} = c \sqrt{y} \]  

(i.e. \( c = a_n \sqrt{3} \))

The values of \( a_n \) are cited in [4].

Equation (3) is true both for backwater and suspended runoff, these cases being distinguished according to the initial conditions and the nature of infiltration intensity which is constant (or varies slowly) with a backwater runoff. In case of a suspended runoff the infiltration is reduced rapidly in time. The principal conclusions obtained upon integration of the foregoing equations are reported below.

The most important elementary slope runoff components (i.e. runoff from a unity slope band) are the time of contact travel \( t_c \), the time of surface travel \( t_s \), the duration of contact \( (\tau_c) \) and surface \( (\tau) \) water recession, and also the coefficient of soil inun-
The \( \sigma \) coefficient is the ratio of the momentary contact water formation depth \( S_d \) during the time of travel \( t_c \) to free porosity \( \delta H \), i.e. \( \sigma = S_d / \delta H \) where \( S_d = (h_0 - k) t_c = \gamma c \Sigma T \) (\( \gamma c \) is the precipitation depth formed during the time of contact travel \( \Sigma T \) (the total duration of rainfall, which occurred during the time \( t_c \) and \( h_0 \) is the average precipitation intensity). For other components the following formulae have been derived:

\[
t_c = \frac{\delta l_c}{\delta d v} \quad (4)
\]

\[
t_s = \left[ \frac{l - x_{to}}{c \sqrt{h_M}} \right]^{\frac{1}{2}} \quad (5)
\]

\[
\tau_c = \frac{t_c}{1 + \eta} \quad (6)
\]

and

\[
\tau = t_s \frac{1}{(1 + \eta)^{\frac{1}{2}} \eta^{\frac{1}{2}}} \quad (7)
\]

Here

\[
x_{to} = \delta_d v \frac{H}{h_M} \quad \text{and} \quad \eta = \frac{k}{h_M} \cdot h_M
\]

is the average intensity for the estimated time. The value \( l_c \) is the length of contact travel which is equal for the band of the slope to its full length \( l \). If transverse concentration is the case, the length \( l_c \) is less than \( l \) (frequently far less). In the latter case the travel over the slopes becomes mixed; the full time of travel is defined by the equation:

\[
t_f = \frac{\delta l_c}{\delta d v} + \left[ \frac{l - l_c}{c \sqrt{h_M} \cdot n} \right]^{\frac{1}{2}} \quad (8)
\]

where \( n \) is the coefficient of surface water concentration.

If \( \sigma < 1 \), than an elementary flood is formed solely by contact waters. In the most frequent case \( t_c > T \) (where \( T \) is the duration of an effective rainfall) the contact runoff is incomplete, i.e. its discharge is formed by an inflow only from the lower part of the slope. Unit hydrographs of incomplete runoff (from a single rainfall) are characterized by a rise branch, duration \( T_r \), and two recession stages: a slow linear recession during the time \( t_c - T_r \), and an intensified recession; the duration of which is \( T + \tau_c - t_c \).

The equation of water rise will be:

\[
q_t = \frac{\delta_d v}{\delta l} S_t \quad (9)
\]

where \( q_t \) is the rate of contact runoff and \( S_t \) is the waterformation layer at the moment \( t \). The maximum rate of the incomplete runoff is formed from the water formation layer \( S_T \) (for the whole period of rainfall). The precipitation intensity being uniform.
the rising curve acquires a linear form. In general, it reproduces the accumulation curve of precipitation storage and is usually of a mild S-form. The first stage of recession is also determined by an equation (9) with substitution of \( S_t \) by \( S_T - kt \) where \( S_T \) is the water formation layer at the end of the rainfall and \( t \) is read off from the beginning of the recession. Theoretically the second stage of recession has a linear form as well, but in practice due to a hindered water yield of the fine-grained soil mass before the end off the recession, the decrease of runoff becomes somewhat decelerated. On the whole the fall of an incomplete contact runoff hydrograph from a single rainfall may be approximated by a triangle.

In case \( \sigma > 1 \), at the moment \( t_0 = \delta H/h_{0M} \), the capacity of the soil layer is filled, and a backwater runoff emerges. At this moment the rate of contact runoff achieves the maximum possible magnitude \( q_0\text{max} = (\delta a e/l) H \) and remains on that level during the whole period of surface runoff \( T_{sur} = T + \tau - t_0 \). The hydrograph acquires a trapezoidal form with the upper base \( T_{sur} \).

On very short slopes, usually when \( \delta < 1 \), a direct contact runoff having a linear stage of rise (9) may emerge, which is replaced by a stage of a direct runoff with a duration \( T - t_0 \).

During this stage the rate of runoff is determined by the same formula, but as to \( S_t \) it is estimated as a water formation layer for the preceding time of travel. When the intensity of rainfall is constant the direct runoff stage is characterised by constant discharge. Generally its hydrograph is of a smooth convex outline reproducing the rainfall graph in a smoothed form.

As the time of contact water flow is measured by at least tens of hours, superimposing of inflow graphs of several rainfalls is a common event. The maximum discharge is therefore frequently estimated by total formation from several rainfalls. Equation (9) is the design formula when

\[
S_t = \sum \delta-a - kT_r - \beta k \tau_p \approx \sum \delta-a - kT_d \tag{10}
\]

Here \( T_r \) is the general duration of rainfalls in the interval of the time of travel, \( \tau_p \) is the sum of durations of periods without rainfall and the magnitude \( \tau \), \( T_d \) is the design time, \( \Sigma \delta-a \) is the precipitation depth during the time of travel \( t_0 \), estimated by the precipitation storage formula:

\[
\sum \delta-a = P \epsilon \cdot t_c \tag{11}
\]

In this formula \( P \) is the precipitation depth of a given frequency for a unit of time (for instance, diurnal maximum for diurnal interval), \( \epsilon \) is an index which is usually individual for the three ranges of time: interstorm (up to 1-2 hours) interdiurnal and a multiday ones.

For the sake of numerical forecasting the general hydrograph of an elementary contact runoff is obtained by means of summing up to unity graphs. A frequency hydrograph may be constructed on the basis of the precipitation storage formula (11) subdivided into stages of rise and recession of the inflow during the rainy period according to the average ratio of these stages (in a number of regions inflow graphs are assumed to be symmetrical). The rise can approximately be assumed to be linear, which permits to schematize the whole contact hydrograph according to a triangular form.

The rate of backwater surface runoff in the water formation and maximum stages is determined by an equation:

\[
q_r = \frac{cS_t^\frac{1}{2}}{l} \tag{12}
\]
Here it is for the stage of incomplete runoff, i.e. if \( t - t_0 < t_S \) the value \( S_t \) is defined as a water formation depth accumulated during the whole elapsed time of backwater \( t - t_0 \); the maximum rate of runoff corresponds to the excess of the general water formation depth over the free capacity \( \delta H \). In the stage of direct runoff, i.e. at \( t - t_0 > t_S \), the value \( S_t \) is assumed to be a water formation depth covering the preceding time of travel in that case, \( q = \delta H \) where \( \delta H \) is the water formation intensity (average within the time of travel). Thus, the hydrograph of the surface component in the stage of an incomplete runoff is determined by ordinates of the precipitation storage curve in the power 3/2, and with a direct runoff it reproduces a transformed rainfall hydrograph which is all the more smoothed the greater is the \( t_S \). At a direct runoff the surface recession curve has a rather complex equation, which can be presented in relative co-ordinates by a family of curves marked by a value of \( \eta \) or by an average characteristic curve. With an incomplete runoff prior to elapse of time \( t_S \), a stage of slow recession is observed. On the whole a hydrograph of a surface component is characterized by a variety of forms, being composed of a number of curves. It is expedient to construct design hydrographs on the basis of a characteristic rainfall graph derived from the well-known precipitation reduction formula

\[
\overline{h}_T = \frac{\Delta}{T^b} \tag{13}
\]

where the rainfall intensity \( h_T \) is the longest out of average intensities within the time \( T \), \( \Delta \) is the force of rainfall determined with the help of isoline maps suitable for different design frequencies, and \( b = 1 - e \).

The reduction formula is conjugated with the relationship

\[
h_T = \frac{(1 - b) \Delta}{T^b} \tag{14}
\]

where \( h_T \) is the intensity exceeded during the time \( T \).

Utilizing this relationship it is easy to derive the march of rainfall and the accumulation curves of its storage in relative co-ordinates expressed both analytically and in the form of characteristic graphs marked by values of \( b \) and an assymetry rainfall co-efficient. The latter value is a ratio of rainfall rise to its fall; its modulus or design values being constant for homogeneous geographical regions. The constancy of \( z \) and \( b \) over great territories permits to obtain regional typical curves of march and curves of precipitation storage (such curves are constructed for most of the territories of the USSR). Utilizing these typical curves one can construct hydrograph families of surface slope inflow in relative coordinates, marking them by values

\[
\beta = \frac{t_0}{T - T_0} \quad \text{and} \quad \gamma = \frac{t_S}{T - t_0}.
\]

For rainfalls of unit length the principal parameter of the hydrographs is \( t_0 \). The values of \( t_0 \) for different watersheds varies slowly over the territory of mountainous countries, while on individual small watersheds deflections from average conditions are possible. Outlines of elementary surface runoff hydrographs at different \( \beta \) and \( \gamma \) are different, nevertheless they can be schematized for different ranges of the indicated parameters. Thus for \( \gamma \) up to 0.20 a schematization in accordance with a simple parabolic triangle is adequate.

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A general hydrograph of an elementary slope inflow is derived by simple adding of graph ordinates of contact and surface runoff. On the bands of a mixed runoff the general hydrograph has its peculiarities, being more smooth and concentrated. It can be constructed according to the foregoing relationships using the time of travel $t_{tr(8)}$. The maximum discharge in this case is defined as the highest intensity of water formation within the time $t_{tr}$.

Integration of the elementary runoff equation in time and a balance analysis allow to obtain an equation of the mountain slope runoff depth:

$$y = P - kT - AH - a_c \varphi$$

(15)

where

$P$ - the precipitation depth;
$AH$ - the soil retention;
$\varphi$ - the relief index equal to

$$\frac{l}{\sqrt{3}}$$

and

$$a_c = \frac{k\delta}{2\delta_d a_0}.$$

The ratio between backwater surface and contact runoff influences only upon the form of the hydrograph, not upon the runoff depth. For the suspended part of the surface runoff

$$y_{sur} = S_{sur} - a_{sur} \varphi ^\lambda$$

(16)

where $a_{sur}$ is a slope parameter linked with its infiltration capacity.

A hydrograph of inflow from natural mountain slopes is obtained by means of summing up runoff hydrographs from elementary slope bands. The form of the net hydrograph depends upon the manner of distribution of the two values over the territory, viz. the depth of the permeable layer $H$ and the length of the band of contact travel $l_c$. The distribution parameters of these values are fairly stable. It is easier to solve the task for plane slopes, i.e. at $l_c = l$ under the simplest law of depth distribution $H$ over the slope width. As it has been shown by the analysis the maximum rate formula has five types, the range of their application being determined by the values

$$\frac{t_s}{T}, \frac{t_s}{T - t_{0, \max}} \text{ and } \frac{S_T}{\delta H_{\max}} \text{ (Here } H_{\max} = 2 H_m).$$

Thus in extreme cases we obtain:

1. At $T < t_s$ and $S_T < \delta H_{\max}$

$$q_{\max} = \frac{0.4 c S_T^\delta + \delta_d S_T}{\delta H_{\max} l} + \frac{v \delta_d S_T^2}{2 \delta^2 H_{\max} l}$$

(17)

2. At $S_T > \delta H_{\max}$ and $t_s < T - t_0$

$$q = \frac{H_{\max} \delta_d v + h_d}{2 l}$$

(18)

The form of hydrographs varies accordingly. The latter are in general similar to those obtained for an elementary band with an average depth $H_m$ but they are somewhat more spread out, getting a comparatively smooth transition from one stage to another.
An approximate calculation method is possible which is based on singling out contact and surface flood runoff components and constructing a net hydrograph as an ordinate sum of schematized graphs of contact and surface runoff (most commonly triangular surface and trapezoidal contact hydrographs).

At $S_T < \delta H_{\text{max}}$ for the contact runoff is determined by the formula

$$a_c = \xi \frac{S_T^2}{2H_{\text{max}} \delta (S_T - a_c \varphi)}$$  \hspace{1cm} (19)

At $S_T > \delta H_{\text{max}}$ the ratio will be:

$$a_c = \xi \frac{\delta H_{\text{m}} - a_c \varphi}{S_T - a_c \varphi}$$  \hspace{1cm} (20)

Here $\xi$ is the coefficient of transverse concentration; for a plane slope $\xi$ is equal to 1 while generally it is less than a unity.

Concentration of contact waters in the direction of surface runoff hollows is observed on rough slopes; owing to which the time of contact travel $l_c$ varies on the range from 0 to $l$.

In this case three variants of analysis are possible:

1. Summing up of elementary mixed runoff if the value $l_c$ is predetermined by one or another law of slope distribution;
2. Summing up of contact runoff only (surface travel over the length $l - l_c$ is considered in this case as an element of channel travel);
3. Calculation according the modulus value of $l_c$.

The hydrograph formulae are derived for linear and parabolic distribution of $l_c$; they are in general similar to the foregoing ones (for $l_c = \text{const}$), but they are somewhat more complex and are not cited in this report. A composite mixed runoff hydrograph has a complex shape and smooth outlines. In most cases it can be roughly presented by a triangle with a linear loop of slow recession attached to it on the right. Observational analysis showed that a design based upon the modulus value of $l_c$ is also suitable.

It is expedient to determine the parameters of the principal mountain slope runoff equation (15) (i.e. $k$, $A_H$ and $a_c$ values) by combining the analysis of test data with that of observations. The principal means of test work on mountainous terrain is the distant sprinkling of large slope plots and natural drainage systems. On choosing the latter one must either avoid systems having a lateral diffluence or, when this is the case, the diffluence should be measurable. To determine the parameters of the surface components use is made of sprinkling small plots isolated up to the confining layer. To account for contact water allowance, filling of pits with water (the lateral diffluence being considered) and sprinkling of isolated long strips is also used. Tests carried out with the help of infiltrometers mounted on different levels of the underlying relative confining layer give useful results. Travel of surface waters is determined according to the bent points of the hydrograph; measurement of flow velocities is carried out by means of hydrochemical colouring, etc. This makes the problem of transition to a mean velocity of a runoff wave not clear enough. On the basis of the specified and similar methods one can define the local values of parameters of losses and travel inherent in different types of soils, vegetation, geology, and so on. Taken from different individual watersheds these parameters present a data analysis of observations carried out on discharge sites. Thus the graphic analysis of the precipitation runoff relation (1,4) permits to determine all the elements of losses and to link them with the antecedent
soil moistening and rainfall features. Linking the values taken from single watersheds with local values permits to subdivide them according to the types of slopes, and sometimes to map them, conforming to the soil-geological map (as it was done for the East Carpathians and South Primorye regions).

The variety of forms of theoretical hydrographs has been mentioned above. On natural slopes, however, extreme deviations are compensated, which due to an indirect relationship between rates and distribution of a number of initial parameters (such as $H_0$, $l_c$ as well as $a_c$, $k$) results in a certain stability of slope inflow graph outlines on different slopes (or course within a certain range of effective rainfall duration). The slope inflow hydrograph can be reproduced according to the hydrometrical data, its ordinates being determined from the channel discharge formulae (1, 2) in reverse order. Such estimates are quite satisfactory because present-day investigations of a channel process allow to determine the channel travel velocities with an adequate accuracy (3). A direct extrapolation of a number of channel line gauge hydrographs over channelless watersheds is also possible. Carrying out parallel computations of synthetical hydrographs in accordance with theoretical formulae and employing test data for several distribution types or model magnitudes of values variable over the water-shed, one can establish both a typical hydrograph (differentiated over the length of rainfall) and design values of varying magnitudes. Next regional generalizations may be effected.

The most important problem of flood estimation is the establishment of that combination of precipitation recurrences and antecedent moistening which forms a runoff of a required (estimated) frequency. A widely used runoff calculation method involving precipitation data of the same frequency is not justifiable as in this case the specific determination of losses is hindered. Taking one’s cue from maximum moistening one gets overestimated results. As to the popular probability flood calculation carried out according to modulus (average) moistening, the latter will give underestimated results due to classifying of storms into rainfall periods. The two following calculation methods may be advised:

1. Calculation of yearly maxima of runoff depth and water formation depth according to the difference of the peak precipitation and losses, corresponding the actually observed antecedent moistening. Thus for the suspended surface runoff zone a surface water formation with utilization of imbibition formulae expressed through the moistening indices (1,2,5,6) is established. The obtained water formation rows are statistically treated. Then the water formation depth and the parameters of design frequencies are mapped. For mountainous areas the runoff depth is calculated similarly in conformity with the prevalent slope type and triniminal and the recurrent magnitudes of the stated mapped values are then determined. For other types of slopes transition coefficients are estimated. An important condition for the estimation development is the combination of analytical and synthetical approaches, i.e. calculations based on formulae of losses and inverse estimation of runoff on the basis of maximum discharge formulae obtained in turn from recorded maxima (2).

2. Determination of a conventional runoff coefficient, i.e. of the ratio of runoff depth or estimated frequency to precipitation of the same frequency. This allows the maps or formulae of the initial characteristics of precipitation to be the basis of the investigation. The first method is more accurate but it is labour-consuming and is possible only when extensive one may conclude that the design formulae of floods must have a regional nature. This was shown by the developments for a number of regions of the Ukraine and the Far East. The problem of flood estimation has a number of aspects which for lack of space are not set forth here.
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