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ION ZĂVOIANU

# MORPHOMETRY OF DRAINAGE BASINS

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# MORPHOMETRY OF DRAINAGE BASINS

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# MORPHOMETRY OF DRAINAGE BASINS

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# Contents

# Preface

*Water, "the blood of the Earth", a substance vital for the very existence of man, has played an important role in society's progress. Today, and even more so in the future, such progress would be inconceivable without hydropower plants and storage lakes, without drinking-water systems, or without the use of water for industry and agriculture, requiring systems for water transfer from one region to another, irrigation networks, shipping canals, etc. However, the network of gauging stations which supply the data required for such projects was established only recently and is of low density, so that, on a world scale, few such stations have reported data over a period sufficiently long to be considered statistically significant. The measurement and establishment of natural drainage regimes have been accompanied by a desire to manage water resources in order to meet the steadily growing needs of the economy and of civilization in general. In many instances, man's intervention in the utilization and spatial redistribution of water resources in a region has resulted in changes of the natural drainage regime before sufficient data have been collected to ascertain the region's natural potential. Furthermore, there are many regions which do not have a network of gauging stations, or have very poor networks, where it is necessary to rely on detailed determinations made on the basis of empirical formulae.*

*Unintended changes in drainage regimes and the inadequacy of gauging-station networks make it necessary to find other ways of evaluating the natural water-resources potential of a region, using both existing measurements and series of generalized regional relationships based on morphometrical parameters. Although land forms have been measured sporadically for practical purposes since very early in mankind's history, it was only in the 19th century that extensive measurements were made. The general progress of science in the 19th and 20th centuries produced multiple and diverse measuring methods, with as many and diverse applications, relating to the fields of hydrology and geomorphology. Thus, a need arose to unify all these methods into a single discipline with its own principles, with a coherent terminology and with a methodology for collecting morphometrical data, which would allow comparison and interpretation of results obtained in different countries.*

*The purpose of the present work is to describe the drainage basin as a system unit resulting from the interaction between runoff and topography, which is a lengthy process of evolution that occurs according to* *well-defined laws. It purports not to quantify the agents which created the present forms, but only to analyze the latter in order to establish the laws according to which they develop and to define a series of interrelationships between morphometrical parameters and river discharge.*

*Although quite a number of specialized works were consulted, it was not the author's intention to present an exhaustive description of all the models found in them. Only the best and most frequently used have been selected, analyzed and verified by application to a sufficiently large area with diverse geographical conditions. This has led to a number of original contributions expressed mathematically, which have been verified rigorously so that, in the author's opinion, they can also be applied to other regions.*

*A number of aspects, such as the analysis of drainage networks and stream order, have been dealt with extensively in the literature, and for this reason are treated more briefly. Instead, greater emphasis is placed on questions of practical interest or on less well-known aspects, concerning which original ideas are advanced. The results obtained from the examination of a fairly large number of cases have permitted the formulation of new rules and relationships concerning the basin perimeter, the form factor, the drainage and stream densities and the mean channel slope, the introduction of new relationships between channel slope and basin slope and between mean slope and discharge, and determination of the discharge and total duration of flood events as functions of drainage-basin area and mean altitude. In presenting the mathematical formulae, the simplest form has been sought so that the results may be accessible to the largest number of readers, even to those lacking specialist mathematical knowledge. However, even in this form the formulae can easily be used to elaborate programs for automatic data-processing. From this point of view, it is the author's opinion that the work contains many questions which can be enlarged upon.*

*Considering its small volume, this work discusses numerous questions of consequence for future research into the utilization of water resources and for river engineering projects. It is intended as a landmark in this field and as a starting point for further research aimed towards investigating the interrelationships between morphometrical parameters and hydrological features, or towards estimating discharge from its relationships to generalized areal and topographical factors. It is intended for a wide circle of specialists working in design or research departments who must deal with the utilization of water resources, for geographers, teaching staff and students of higher education, and for all those interested in the morphometry of drainage basins either for practical or theoretical purposes.*

*I would like to thank all my colleagues at the Bucharest Institute of Geography who, through their suggestions, contributed to improving this book, Editura Academiei for support, and finally my family for understanding my need to spend more time at my desk and less with them.*

*ION ZĂVOIANU*

## List of Main Symbols Employed

| | |
|---|---|
| $C_v$ | coefficient of variation ($C_v = [\Sigma(x - \bar{x})^2/n]^{1/2}/\bar{x}$) |
| $S_k$ | skewness coefficient ($S_k = (x - \bar{x})^3/ns^3$, where $s$ is standard deviation) |
| $N_u$ | number of stream segments of highest order $u$ |
| $N_x$ | number of stream segments of order $x$ |
| $N_{1,\ldots,u}$ | summed numbers of stream segments of orders $1,\ldots,u$: $N_1 = \sum_{i=1}^{n} N_{i(1)}$, etc. |
| $R_c$ | confluence ratio |
| $N$ | total number of stream segments in a drainage basin: $N = \sum_{i=1}^{n} N_{i(1)} + \sum_{i=1}^{n} N_{i(2)} + \ldots + \sum_{i=1}^{n} N_{i(u-1)} + N_u$ |
| $A'_{1,\ldots,u}$ | summed areas required for the appearance of drainage basins of orders $1,\ldots,u$ : $A'_1 = \sum_{i=1}^{n} A'_{i(1)}$, etc. |
| $a'_{1,\ldots,u}$ | average areas required for the appearance of drainage basins of orders $1,\ldots,u$ : $a'_1 = \sum_{i=1}^{n} A'_{i(1)} \Big/ \sum_{i=1}^{n} N_{i(1)}$, etc. |
| $r_{a'}$ | mean ratio of average areas required for the appearance of drainage basins of successively higher orders |
| $A_{1,\ldots,u}$ | summed areas of drainage basins of orders $1,\ldots,u$: $A_1 = \sum_{i=1}^{n} A_{i(1)}$, etc. |
| $R_A$ | mean ratio of summed areas of drainage basins of successively higher orders |
| $a_{1,\ldots,u}$ | average areas of drainage basins of orders $1,\ldots,u$: $a_1 = \sum_{i=1}^{n} A_{i(1)} \Big/ \sum_{i=1}^{n} N_{i(1)}$, etc. |
| $r_a$ | mean ratio of average areas of drainage basins of successively higher orders ($r_a = R_c/R_A$) |
| $P_{1,\ldots,u}$ | summed perimeters of drainage basins of orders $1,\ldots,u$ : $P_1 = \sum_{i=1}^{n} P_{i(1)}$, etc. |

| | |
|---|---|
| $R_P$ | mean ratio of summed perimeters of drainage basins of successively higher orders |
| $p_{1,\ldots,u}$ | average perimeters of drainage basins of orders $1,\ldots,u$: $p_1 = \sum_{i=1}^{n} P_{i(1)} \Big/ \sum_{i=1}^{n} N_{i(1)}$, etc. |
| $r_p$ | mean ratio of average perimeters of drainage basins of successively higher orders ($r_p = R_c/R_P$) |
| $\bar{L}$ | average basin length |
| $\bar{B}$ | average basin width |
| $RC$ | circularity ratio for drainage basins |
| $R_e$ | elongation ratio for drainage basins |
| $R_f$ | form factor for drainage basins |
| $\overline{RF}_{1,\ldots,u}$ | average form factors for drainage basins of orders $1,\ldots,u$: $\overline{RF}_1 = \sum_{i=1}^{n} \overline{RF}_{i(1)} \Big/ \sum_{i=1}^{n} N_{i(1)}$, etc. |
| $T_s$ | coefficient of topographical sinuosity |
| $H_s$ | coefficient of hydraulic sinuosity |
| $K_s$ | sinuosity coefficient |
| $HSI$ | hydraulic sinuosity index |
| $TSI$ | topographical sinuosity index |
| $K_b$ | braiding coefficient |
| $L_{1,\ldots,u}$ | summed lengths of stream segments of orders $1,\ldots,u$: $L_1 = \sum_{i=1}^{n} L_{i(1)}$, etc. |
| $R_L$ | mean ratio of summed lengths of stream segments of successively higher orders |
| $\Sigma L$ | total length of entire stream network in a drainage basin: $\Sigma L = \sum_{i=1}^{n} L_{i(1)} + \sum_{i=1}^{n} L_{i(2)} + \ldots + \sum_{i=1}^{n} L_{i(u-1)} + L_u$ |
| $l_{1,\ldots,u}$ | average lengths of stream segments of orders $1,\ldots,u$: $l_1 = \sum_{i=1}^{n} L_{i(1)} \Big/ \sum_{i=1}^{n} N_{i(1)}$, etc. |
| $r_l$ | mean ratio of average lengths of stream segments of successively higher orders ($r_l = R_c/R_L$) |
| $l_0$ | length of overland flow |
| $D_s$ | density of stream segments |
| $D_d$ | drainage density |
| $F_{1,\ldots,u}$ | summed falls of stream segments of orders $1,\ldots,u$: $F_1 = \sum_{i=1}^{n} F_{i(1)}$, etc. |
| $R_F$ | mean ratio of summed falls of stream segments of successively higher orders |
| $\Sigma F$ | fall of all stream segments in a drainage basin: $\Sigma F = \sum_{i=1}^{n} F_{i(1)} + \sum_{i=1}^{n} F_{i(2)} + \ldots + \sum_{i=1}^{n} F_{i(u-1)} + F_u$ |

$f_{1,\ldots,u}$ average falls of stream segments of orders $1,\ldots,u$: $f_1 = \sum_{i=1}^{n} F_{i(1)} \Big/ \sum_{i=1}^{n} N_{i(1)}$, etc.

$r_f$ mean ratio of average falls of stream segments of successively higher orders ($r_f = R_c/R_F$)

$H_m$ mean altitude

$H_{max}$ maximum altitude

$H_{min}$ minimum altitude

$H_{1,\ldots,u}$ summed mean altitudes of drainage basins of orders $1,\ldots,u : H_1 = \sum_{i=1}^{n} H_{i(1)}$, etc.

$R_H$ mean ratio of summed mean altitudes of drainage basins of successively higher orders

$\Sigma H$ total mean altitude of all subbasins in a drainage basin: $\Sigma H = \sum_{i=1}^{n} H_{i(1)} + \sum_{i=1}^{n} H_{i(2)} + \ldots + \sum_{i=1}^{n} H_{i(u-1)} + H_u$

$\bar{H}_{1,\ldots,u}$ average mean altitudes of drainage basins of orders $1,\ldots,u : \bar{H}_1 = \sum_{i=1}^{n} H_{i(1)} \Big/ \sum_{i=1}^{n} N_{i(1)}$, etc.

$r_{\bar{H}}$ mean ratio of average mean altitudes of drainage basins of successively higher orders ($r_{\bar{H}} = R_c/R_H$)

$R_{1,\ldots,u}$ summed maximum heights of drainage basins of orders $1,\ldots,u : R_1 = \sum_{i=1}^{n} R_{i(1)}$, etc.

$R_R$ mean ratio of summed maximum heights of drainage basins of successively higher orders

$r_{1,\ldots,u}$ average maximum heights of drainage basins of orders $1,\ldots,u : r_1 = \sum_{i=1}^{n} R_{i(1)} \Big/ \sum_{i=1}^{n} N_{i(1)}$, etc.

$r_r$ mean ratio of average maximum heights of drainage basins of successively higher orders ($r_r = R_c/R_R$)

$h_{1,\ldots,u}$ summed mean heights of drainage basins of orders $1,\ldots,u : h_1 = \sum_{i=1}^{n} h_{i(1)}$, etc.

$R_h$ mean ratio of summed mean heights of drainage basins of successively higher orders

$\bar{h}_{1,\ldots,u}$ average mean heights of drainage basins of orders $1,\ldots,u : \bar{h}_1 = \sum_{i=1}^{n} h_{i(1)} \Big/ \sum_{i=1}^{n} N_{i(1)}$, etc.

$r_{\bar{h}}$ mean ratio of average mean heights of drainage basins of successively higher orders ($r_{\bar{h}} = R_c/R_h$)

$S_{b_{1,\ldots,u}}$ summed mean slopes of drainage basins of orders $1,\ldots,u : S_{b_1} = \sum_{i=1}^{n} S_{b_{i(1)}}$, etc.

$R_{S_b}$ — mean ratio of summed slopes of drainage basins of successively higher orders

$\bar{s}_{b_1,\ldots,u}$ — average mean slopes of drainage basins of orders $1, \ldots, u : \bar{s}_{b_1} = \sum_{i=1}^{n} S_{b_{i(1)}} \Big/ \sum_{i=1}^{n} N_{i(1)}$, etc.

$r_{\bar{s}_b}$ — mean ratio of average mean slopes of drainage basins of successively higher orders ($r_{\bar{s}_b} = R_c / R_{S_b}$)

$s_{1,\ldots,u}$ — average mean slopes of stream segments of orders $1, \ldots, u : s_1 = \sum_{i=1}^{n} F_{i(1)} \Big/ \sum_{i=1}^{n} L_{i(1)}$, etc., or $s_1 = f_1/l_1$, etc.

$r_s$ — mean ratio of average mean slopes of stream segments of successively higher orders

$\bar{s}_n$ — mean slope of entire river network in a drainage basin

$Q_{1,\ldots,u}$ — summed mean discharges of stream segments of orders $1, \ldots, u : Q_1 = \sum_{i=1}^{n} Q_{i(1)}$, etc.

$R_Q$ — mean ratio of summed mean discharges of stream segments of successively higher orders

$\bar{Q}_{1,\ldots,u}$ — average mean discharges of stream segments of orders $1, \ldots, u : \bar{Q}_1 = \sum_{i=1}^{n} Q_{i(1)} \Big/ \sum_{i=1}^{n} N_{i(1)}$, etc.

$r_{\bar{Q}}$ — mean ratio of average mean discharges of stream segments of successively higher orders ($r_{\bar{Q}} = R_c / R_Q$)

# Brief History

Measurement of the elements of the Earth's surface is a concern that can be traced back to mankind's early history, when such measurements were made primarily for immediate practical purposes. With the Egyptians, measurements appeared and developed from the necessity to reconstitute the boundaries separating estates which were wiped away by each flood of the Nile. These measurements, together with those made for the erection of the impressive Egyptian temples and pyramids, formed the embryo of geometry as eventually developed by the Ancient Greek scholars. The initial significance of the practice and the way in which the science of land measurement developed into a branch of mathematics are generally known. Since the etymology of the word geometry (*geometrein*, to measure the land) bears witness to this origin, it is too late to attempt to appropriate it for the geographical sciences, whose content, it might be thought, has become remote from the original concerns.

Nonetheless, geography today cannot do without data concerning the shapes and sizes of land forms. Moreover, in quantitative geography the employment of precise data concerning land forms is a prerequisite for fundamental research into the laws governing relief development, for elaborate mathematical models, and for practical applications such as forecasting discharge and the regional modelling of hydrological features.

Thus, the measurement of forms, i.e., morphometry, appears to be of increasing importance. It is equally fundamental to all sciences studying the forms of the natural world. Botany, zoology, geology, mineralogy, anthropology, etc., cannot undertake any thorough studies without resorting to morphometrical data. Hydromorphometry, for example, deals with the determination of variables characterizing channel size and shape.

In order to avoid misunderstanding with regard to morphometry, the term *geomorphometry* as used by Morisawa (1962) has been adopted as denoting the science dealing with measurements of the form of the Earth's crust. Of course, no geomorphological or hydrological study can ignore the quantitative values supplied by geomorphometry.

Concomitant with society's development and the progress of science, methods for investigating land forms have also improved. During the Renaissance, Leonardo da Vinci in the 16th century and

Galileo Galilei in the 17th established the qualitative observation that the longitudinal profiles of rivers take the form of a concave curve. During the 17th and 18th centuries, Lombard engineers stated correctly that river slope depends on discharge, on suspended load and on the shape of the channel.

The expansion of trade and the increasing number of voyages and geographical expeditions in the 16th century led naturally enough to a desire to record as correctly as possible the shapes and sizes of newly discovered lands, resulting in the achievement of the first world map by Mercator. Gradually, accumulated observations allowed scholars to establish the forms of the continents, from which Alfred Wegener ultimately deduced the hypothesis of continental drift. First Immanuel Kant and then Alexander von Humboldt noted that the Atlantic ocean has an S-shape. Research showed that the Antarctic is a landmass, in contrast to the Arctic ocean, and so on.

The growing amount of information rendered ever more necessary the generalization and classification of the forms of both land and water. To facilitate this task, these forms have frequently been compared to various regular geometrical figures, the outcome being a number of relationships between corresponding elements of length, surface or volume. For example, the length between two points on a river course or on the shore of a lake, etc., can be related to the straight-line distance between them, in terms of a nondimensional scaling factor.

A first step towards comparability was made in 1826 by Carl Ritter, who related the areas of the continents to the squares of their perimeters or to the areas of the smallest circumscribed circles, thereby obtaining a circularity index. Measurement of the relief of the continents of the ocean floors lagged behind the recording of surface areas and lengths. Nevertheless, in 1816 Alexander von Humboldt, on the basis of heights of points determined in the Pyrenees, succeeded in computing the average altitudes of peaks and passes. In 1842, employing a series of parallel profiles, he sought to calculate the average heights of the continents, with the exception of Africa and Australia for which the data were insufficient.

The principle of the hypsometric curve was established in 1854 by Carl Koritska, who showed that relief should be divided into horizontal sections which can be assimilated to frustums of cones with the same base and height. In this case, the mean altitude is given by the ratio of the volume of these figures to the mean basin surface area. In 1883 this method was used by Albert de Lapparent to establish a distribution curve of altitude and depth in polar coordinates for the whole globe. In 1888 John Murray applied the method to ocean floors, and in 1894 Albrecht Penck corrected it to the form used subsequently for continents, islands, mountain ranges, etc.

The appearance of topographical maps also lent impetus to geomorphometrical research. Thus, in 1873 the Austrian C. von Sonclar determined twelve characteristic values for each of several mountain ranges, the most noteworthy being the mean relative altitude of ridge lines. His example was followed in 1886 by L. Neumann who, while carrying out a detailed study of the Schwarzwald mountains, calculated among other parameters seven characteristic values for each of a large number of valleys. From a comparison between the Schwarzwald mountains, the Thuringerwald mountains and the Northern Calcareous and Eastern Alps, he noted that levelling in the old Hercynian massifs was in a much more advanced stage (Baulig, 1959).

In 1890, when R. Manning proposed the formula that bears his name, S. Finsterwalder and K. Peucker arrived independently at a formula for the computation of mean slope, using the equidistance between two contour lines, the area delimited by them and their lengths.

The last decades of the 19th century saw the first use of morphometrical elements in hydrology. In 1884, while A. I. Voeikov showed that rivers can be regarded as a product of climate, Kestlin (quoted by Ogievsky, 1952) employed a formula for the determination of discharge from small drainage basins, in which the major variables were the surface area of a basin and a coefficient depending on its length. The first map with isolines of mean specific discharge, for which computations of basin surface areas were absolutely necessary, was drawn by Newell in 1892 (quoted in "Mean specific discharge of Romania's rivers", Anon, 1954).

The achievements of the Russian school of hydrology in the early 20th century were remarkable. Numerous elements, such as the mean surface area and altitude of drainage basins, the length of the main river, etc., were employed to determine water flows and other hydrological characteristics on the basis of empirical formulae. In this respect, the hydrological studies carried out by D.I. Kocherin (quoted by Ogievsky, 1952) are particularly noteworthy. In 1927, Kocherin plotted for the European territory of the Soviet Union the first map showing isolines of mean specific discharge, calculated on the basis of data supplied for 34 drainage basins, and produced the first formula for establishing maximum specific discharge.

Basin surface area was already being used as a variable at the end of the last century by Kestlin. Then, in 1930, D. L. Sokolovski, S.N. Kritski and M. F. Menkel used it to establish an empirical formula for determining the variation coefficient of annual runoff, in the absence of direct observations. N. P. Chebotarev and M. E. Shevelev in 1937, and later N. D. Antonov in 1941, used this formula to determine the variation coefficient of maximum runoff. In 1938, A. V. Ogievsky suggested a formula, which he amended in 1945, for determining discharge taking into account basin surface area and river length. Sokolovsky (1945) used basin surface area in an empirical formula for calculating the maximum discharge originating from rainfall for small rivers.

With the development of geographical research, morphometrical data have been used increasingly to compare and confirm scientific conclusions, and many such data are found in most treatises on physical geography. Thus, in 1926, Emm. de Martonne, who worked for more than a year determining the mean altitude of France and its major geographical regions, viewed morphometrical computations as presenting great advantages in precise evaluations. Between 1918 and 1938, in work on the French Alps, other French researchers used known methods of computing morphometrical indices, to which they added a carving coefficient for mountain relief.

Continuing the list of researchers who made important contributions to the development of morphometrical analysis, the name of P. S. Jovanović, who applied genetic analysis in a study of the longitudinal profiles of three rivers in Yugoslavia (Jovanović, 1940), should not be omitted. The attempt to single out the quantitative influences of various factors on the longitudinal profile was original, but the data from which Jovanović started and the fact that he ignored suspended load from the very beginning explain why his results were unrealistic. The ideas resulting from the above studies were quite valuable, but they were developed too early in relation to the general progress of science and thus could not solve the questions being asked. As Baulig (1959) pointed out, in formulating the laws governing a phenomenon a necessary condition is that the factors determining the phenomenon and their interdependences should all be known. However, in the case of the geographical sciences, there still were many questions to be answered in 1940.

In 1942, Ch. P. Péguy applied the principles of morphometry in a study of the Haute Durance and Ubaye regions, determining for the first time a clinometric curve to complement the hypsometric curve. In addition, Péguy constructed a hydrodynamic curve showing the erosion power of a water course in a long profile as a function of flow rate and channel slope, defining, among other hydrological characteristics, a volumetric runoff coefficient (Péguy, 1951).

In the last few decades, and especially in recent years, a great number of scientists have dealt either directly or indirectly with morphometrical questions, so that their mere enumeration would require considerable space. The contributions of each have gradually accumulated, growing into an individual discipline with specific methods and principles.

The genetic explanation of hydrological phenomena has benefitted considerably by the introduction of the classification system for channel networks proposed by Horton (1945) and the establishment of laws of development for river networks. His thorough mathematical training and theoretical mastery enabled Horton to arrive at a number of valid mathematical relationships.

In an important study of the topographical characteristics of drainage basins completed in 1941, but published only some years later because of the Second World War, Langbein (1947) presented calculation methods for the surface area and length of basins, for river density, slope and channel form, and for hypsometric curves.

Use of the classification system proposed by Horton spread quite rapidly, and the subject of geomorphometry has been enriched substantially by the work of numerous American researchers, among whom may be mentioned Strahler (1952a, b, 1953, 1956, 1957, 1958, 1964, 1966), Schumm (1954, 1956, 1977, 1979), Maxwell (1955, 1960), Leopold and Miller (1956), Hack (1957), Melton (1957, 1958a, b, 1960), Morisawa (1957, 1958, 1959, 1962, 1967, 1968), Bowden and Wallis (1964, 1966), Leopold et al. (1964), Shreve (1966, 1967, 1969), Woldenberg (1966, 1969, 1971, 1980), Scheidegger (1965, 1967, 1968a, b, 1970), Ranalli and Scheidegger (1968a, b) and Smart (1968a, b, 1969).

Special attention should be given to the work of Strahler who, after having complemented (Strahler, 1952a) the classification system for river networks elaborated by Horton (1945), carried out a number of studies which provided theoretical and mathematical foundations for many concepts and laws relying on mathematical, physical and statistical properties, and on general system theory. In order to illustrate the diversity of his morphometrical studies, it suffices to mention that in one study, Strahler (1958) singled out 36 indices to define the geometry of forms of river erosion, and 16 to define the kinematic and dynamic features of catchments.

A classification system very close to that proposed by Horton (1945) was elaborated in 1948 by B. P. Panov (quoted by Morariu et al., 1962), in which the first class of river networks included the smallest streams.

For hilly and plain regions, it has been established that the basin surface area is the main element in estimating water flow; for mountainous areas, the mean basin altitude is the basic parameter. A first step along these lines was made by V. L. Schultz (quoted by Ogievsky, 1952), who determined the variation coefficient of annual runoff in relation to mean drainage-basin altitude. In the last three decades an increasing number of studies have used morphometrical elements in geographical research or as a means for making regional generalizations in hydrology. Of these, it is worth mentioning a study carried out by Makkaveyev (1955) concerning river beds and erosion in various basins, which is in accordance with knowledge since accumulated in the rest of the world on this topic. An original contribution concerning the relationship between the morphometrical and hydrological characteristics of river networks was made by Rzhanitsyn (1960). Starting from the classification system developed by Horton and complemented by Panov and Strahler, Rzhanitsyn discovered new laws relating the classification of rivers to characteristic phases of their hydrological regime.

The French school has distinguished itself with a synthesis of the history of morphometrical research. Thus, Baulig (1959) showed inter alia that morphometry can be of considerable service on condition that it considers simple and well-defined problems. In this respect, he proposed that small areas, homogeneous in terms of structure, morphological processes and stage of development, be chosen for study. This remark is valid provided that theoretical substantiation of the relevant morphometrical laws is well developed. If not, the study of small areas may not always reflect more general laws, since the cases analyzed may represent exceptions or particular instances of insufficiently known general laws. In order to test a model, or to substantiate and verify morphological principles and laws, research involving larger areas is also necessary : the complexity of geographical reality and the variety of environmental conditions provide a good means of verification.

The work carried out between 1961 and 1965 by F. Hirsch at the Strasbourg Centre for Applied Geography is also noteworthy. In addition to applying and verifying Horton's laws for a number of rivers in France, Hirsch effected a critical analysis of certain morphometrical relationships, noting a direct link between water discharge and basin surface area. The next step was an attempt to forecast water flow on the basis of morphometrical characteristics (Tricart and Hirsch, 1960 ; Hirsch, 1962, 1963, 1964).

Merlin (1965, 1966) published studies concerning the methods used and the results obtained in a morphometrical analysis of several North African massifs, using hypsographic curves and mean altitudes. He also revived the method of decimal profiles used by Jovanović, on the basis of which he calculated and analyzed critically a series of mean durability indices computed for various geological formations.

Valuable work has been produced in Great Britain concerning both morphometrical relationships and the application to geography of mathematical modelling, statistical methods and general system theory : the work of Chorley (1957, 1958, 1962, 1967, 1969), Chorley and Haggett (1967), Chorley and Kennedy (1971), Ferguson (1975), Pethick (1975), Gardiner (1975, 1977, 1978), Doornkamp (1968), Gregory (1966, a, b, 1971, 1976, 1981) and Gregory and Walling (1968, 1973) may be mentioned.

In Brazil, special attention has been given to morphometrical studies by Christofoletti, who alone (1969, 1970, 1973, 1978, 1979, 1981), and in collaboration (Christofoletti and Perez Filho, 1975 ; Christofoletti and Filizola, 1978 ; Christofoletti and Oka-Fiori, 1980) made a substantial contribution to the application and verification of morphometrical relationships for the northeastern part of the Brazilian plateau. In a work of synthesis (Christofoletti, 1970), he expounded the methodology for determining the indices used and presented morphometrical data for ten representative basins situated in a region in which crystalline and volcanic rocks prevail. The exposition, analysis and interpretation of the results were followed by calculation of a matrix of correlation coefficients between the morphometrical indices employed. By its approach, by its verification of known morphometrical indices and laws, and by its interpretation of results in relation to the lithology and geomorphological evolution of the region considered, this work marked a major step forward in morphometrical research.

In the last two decades, an increasing number of researchers in other parts of the globe have shown interest in the verification and application of morphometrical methods. Noteworthy amongst these are Milton (1966, 1967) and Abrahams (1970, 1972, 1980) in Australia, Cotton (1964), Selby (1968) and Eyles (1974) in New Zealand, Sakaguchi (1969), Kayane and Shimano (1974) and Takayama (1977) in Japan, L. Singh (1970), Pofali (1979, 1981) and S. Singh and Sharma (1979) in India, Frey (1965) in Switzerland, Verhasselt (1961) in Belgium, Seyhan (1975, 1976, 1977a, b) in The Netherlands and Drexler (1979) in the Federal Republic of Germany.

In Romania, morphometrical elements have been used both in geographical research and in hydrology for the generalization of several hydrological parameters. In geography, noteworthy results have been obtained by the Cluj-Napoca School headed by Morariu in determining the density of the river network in Romania (Morariu et al., 1956) and the maximum energy (Morariu et al., 1957) and mean fragmentation (Morariu and Savu, 1959) of Romania's relief. The density of river networks, as a factor influencing runoff, has been the focus of a large amount of research (e.g., Morariu and Savu, 1959; Ujvári, 1956, 1959, 1972; Neamţu, 1961, 1962; Podani, 1961).

An ever-increasing number of studies have relied on morphometrical data to characterize relief (Martiniuc, 1955; Micalevich-Velcea, 1961; Băcăuanu, 1968; Hârjoabă, 1968; Coteţ, 1973; Ichim, 1979). In a detailed study, Grumăzescu (1973) used extensively, instead of morphometrical elements, the notion of geometrical features of relief. A special merit of this work was the use, for the first time in Romania, of a development and flattening index for interfluves to characterize land forms. More recently, Grigore (1979) paid special attention to the mapping of morphometrical features with the aim of achieving as graphic and accurate an image of land forms as possible.

Morphometrical elements have been used successfully in limnology, where morphometrical data concerning lakes represent basic elements both in the classification of genetic types of lake basins (Gâştescu, 1960, 1963, 1971) and in the characterization of certain types of lake or lake complexes (Pişota, 1971; Breier, 1976).

In hydrology, the morphometry of drainage basins and river networks has always proved of value in estimations of water discharge and maximum and minimum specific runoff, and of their coefficients of spatial variation, especially as concerns conditions involving an obvious vertical zonal pattern of runoff. One of the results worthy of mention is the drawing of the first map of mean specific runoff in Romania by the General Water Office as early as 1954. The main elements for this were basin surface area and mean basin altitude.

In addition to the influence of zonal climatic factors on runoff, the studies mentioned so far all highlighted the influence of an azonal factor, basin surface area, which plays a compensatory and regulatory role. Accordingly, as runoff became better understood, its coefficient of variation was shown to depend on drainage-basin surface area (Diaconu, 1961) and mean altitude (Dumitrescu, 1964). The vertical zonal pattern demonstrated for almost all the elements of the water balance accounts for the mean altitude of drainage basins having remained a basic element in regional characterizations of runoff parameters (Diaconu, 1966).

Good results have been obtained by correlating the total duration of floods with various morphometrical parameters of drainage basins and water courses, such as mean river length and slope or surface area and mean basin slope (Diaconu et al., 1961 ; Mustaţă, 1964). Relationships have also been established for the dependence of maximum specific runoff on mean basin altitude and surface area (Mociorniţa, 1961, 1964), and of maximum discharge on basin surface area (Platagea and Platagea, 1958, 1965).

Research has been undertaken in Romania to determine the most adequate relationships between the form of drainage basins and the characteristics of their river networks, on the basis of statistical processing of the number of rivers and the lengths and surface areas of the corresponding basins. This has thrown light on a number of well-defined relationships between river length and basin surface area, and between the density and length of river networks and basin surface area (Diaconu, 1971). These relationships, established for Romania's territory, confirm the validity of relationships determined for very large areas situated in a variety of other physiographical conditions.

The classification system developed by Horton was first applied in Romania to the characterization of the river network between the Ialomiţa and Trotuş rivers, for which, besides other morphometrical features, a dependence of the mean multiannual flow (calculated from data covering a period of several years) on the river size in various sectors has been established (Platagea and Popa, 1963).

The above results have encouraged further research in various geographical areas, using the classification systems devised by Horton and Strahler both to verify existing laws and to establish new relationships (Zăvoianu, 1967, 1968, 1969, 1970, 1972 a,b, 1974, 1975, 1978, 1980 ; Ichim, 1979 ; Grecu, 1980, 1981 ; Sandu, 1980).

Chapter I

# The Drainage Basin as a System Unit

The evolution of any landscape on our planet, and hence of any drainage basin, is the result of interactions between the flows of matter and energy entering and moving within its limits and the resistance of the topographical surface. Under normal conditions, precipitation is the major source of matter and solar radiation the major source of energy. The resistance of the topographical surface is determined by its altitude, the resistance to erosion of the constituent rocks, the percentage of plant cover, the presence of a layer of soil, etc. The interrelationships between these factors and their distributions in time and space govern to a great extent the evolution and present state of drainage-basin topography. Th e qualitative aspect of these relationships is fairly well known and has been analyzed in detail in the geographical literature, in accordance with the discipline's mandate to deal with interrelationships between the components of the environment.

However, quantitative analysis is much more difficult, owing to (i) insufficient observations and (ii) the lack of a unified method: such a method has become a necessity with the current explosion of information, for which data systematization, processing and interpretation are vital requirements. Science has provided answers to these problems : system theory may be applied to overcome problems of the first type, and statistical methods to resolve the remainder. Thus, geography has enlarged its stock of methods in accordance with the need to study the interrelationships between geographical phenomena, and can now make use of quantitative studies and new principles to analyze and synthesize geographical information.

## Theory of the General System

System theory, the basis of which was laid down early by von Bertalanffy (1951), resulted from an analysis of a large number of laws of nature which proved to be almost identical to the laws of theoretical physics. This suggests the existence of structural similarity in very many areas, as a result of correspondence between the general principles governing the behaviour of entities that at first sight appear to differ greatly.

In the light of the principles of general system theory, the organization of the real world appears as a hierarchy of groups of interacting elements, each of which, irrespective of its taxonomic position, functions as a whole according to well-defined laws. Thus, the possibility exists that the same laws and conceptual models can be applied to various fields of research, enabling the researcher better to carry out quantitative and qualitative analyses. Because of these attributes, system theory is very helpful in the nonphysical fields of science, contributing to the unity of principles from the general to the individual sciences, and hence to the unity of science (von Bertalanffy, 1951).

In the broadest sense, a system is "...a set of objects together with relationships between the objects and their attributes" (Hall and Fagen, 1956, p. 18). Such a definition naturally permits the application of the concept to the most diverse areas, since the term "objects" can be interpreted to cover an unlimited number of parts of a system, which may well be the morphometrical variables of a drainage basin or a river bed and a series of mathematical variables, laws, processes, etc. A number of obvious interrelationships will exist between the components of a system or among their specific attributes, which make the aggregate of these components function as a whole.

Application of system theory to the study of the real world poses a number of difficulties, stemming primarily from the great complexity of nature. For geography, the solution of this question has required greater depth in interdisciplinary research, knowledge of the basic law governing the circulation and transfer of mass and energy in nature, knowledge of physics, statistics, system theory, etc. By such means the geographer can more easily divide the environment into subsystems, systems and supersystems, or move to various levels of generalization in order to detect the relationships between variables and their significance. The working methods depend on the extent of knowledge of the internal structure of a system's components and of the causal relationships between the established variables, on the amount of existing data, on the aims of research, etc.

The creative application of system theory to geography has become a steady concern of an increasing number of scientists (Melton, 1957; Morisawa, 1959; Leopold et al., 1964; Chorley, 1967; Haggett and Chorley, 1969; Chorley and Kennedy, 1971; Strahler and Strahler, 1973).

As stated above, system analysis is of great help in the study of the real world, which is viewed accordingly as a hierarchy of organized, interrelated sets of systems which condition one another (Chorley and Kennedy, 1971). A more careful analysis of the principles of system theory in relation to the stages of development of geography proves that the interrelationships between phenomena and the permanent links between cause and effect continue to be among the concerns of geographical research, causality being one of the basic laws of this discipline.

According to the way in which they exchange matter and energy with their surroundings, systems can be divided into two categories, *isolated* and *nonisolated* systems. The former, achievable only under laboratory conditions, have no exchanges of matter and energy with their surroundings, while for the latter category the reverse is true. In turn, *nonisolated* systems can be grouped into *closed* systems, which "...prevent the import and export of mass, but not of energy", and *open* systems, which "... are characterized by an exchange of both mass and energy with their surroundings" (Chorley and Kennedy, 1971, p. 2). In the latter case, the relationships between the components tend to result in self-regulation of the system, so that once the steady state has been achieved, equilibrium exists between the system's inputs and outputs of matter and energy. According to the internal complexity of the systems being considered, geographical studies are using increasingly refined morphological systems, e.g., cascading, process response and control systems (Chorley and Kennedy, 1971).

## Mass and Energy Exchanges with the Surroundings

With regard to the above statements, a drainage basin can be described as an open system which permanently exchanges matter and energy with its surroundings. The fact that such an areal unit, which may be assigned to one of various taxonomic orders, can be delineated precisely by its drainage divide offers a great advantage in quantitative studies, since the researcher is able to estimate accurately the inputs and outputs of matter and energy. Considering a drainage basin as a whole, with all its component parts and their attributes, the basin area is thus subjected to continuous flows of matter and energy. Precipitation accounts for the major input of matter. To this should be added inputs due to man's action, underground inputs from other basins, and those due to wind (Fig. 1). These inputs of matter are accompanied by quantities of energy commensurate with their masses; in addition, the surface of any basin receives energy from the sun through insolation.

The amounts of matter and energy received act upon the variables defining the characteristics of a basin. Some of these quantities are stored as a result of physico-chemical and biochemical processes, while others leave the system by various routes. Thus, unstored water can evaporate, run off, or move into other basins through man's action, via underground paths or following the action of wind. Part of the solar energy received is reflected by the basin area and is lost into the atmosphere (Fig. 1). This permanent exchange with the surroundings is a prerequisite for the existence of any drainage basin, but equally important is the fact that the nature of this exchange bears considerably upon the processes and interrelationships existing between the variables that define the present state of the basin area.

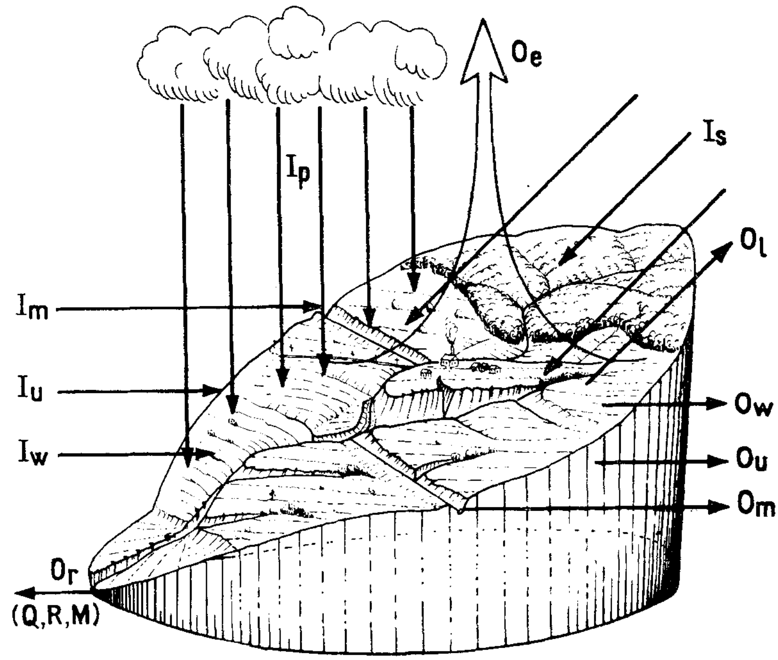



Fig. 1. Scheme of exchanges of matter and energy between a drainage basin and its surroundings : $I_p$, inputs of matter due to precipitation ; $I_m$, inputs due to man's action ; $I_u$, underground inputs ; $I_w$, inputs due to the wind ; $I_s$, inputs of energy due to solar radiation ; $O_r$, outputs of matter and energy through stream flow (water discharge $Q$, sediment yield $R$, and dissolved-load concentration $M$) ; $O_e$, outputs due to evapotranspiration ; $O_m$, outputs due to man's action ; $O_u$, underground outputs ; $O_w$, outputs due to the wind ; $O_l$, outputs of energy due to reflection and radiation.

## Inputs of Matter and Energy

*Precipitation.* The most important form of matter entering a drainage basin is provided by precipitation. Like any mass in motion, precipitation possesses a certain amount of energy which is mostly consumed within processes taking place at the surface of the basin. The volume of precipitation falling over unit area depends upon the latitude and altitude of the basin, which also influence the form of precipitation. For instance, a basin situated at a greater altitude or latitude will receive a larger proportion of its precipitation in the solid state. For Romania, the average annual precipitation is estimated as approximately 670 mm.

The amount, form and distribution over time of precipitation are of great importance in relation to morphoclimatic zones, governing both the processes taking place at the surface of a basin and the amplitude of the response reactions and adjustments of the drainage basin's morphometrical characteristics.

Analysis of this flow of matter proves that, at least for Romania, it has a daily regime, with three maxima, one in the early morning, one at noon, which is the most important, and another in the evening. In relation to the seasons, the noon maximum has its highest value in the summer, and the next highest in spring and autumn. With regard to daily maximum intensity, the rainfall occurring around 2 p.m. has the highest intensity and shortest duration. The distribution of precipitation within the year varies from one place to another, but in all cases a maximum is evident at the end of spring and the beginning of summer, which lies between two periods with lower values, at the beginnings of spring and autumn.

In addition to the above recurring features, special attention must be given to torrential rainfall, which in most instances has disastrous effects due to the huge amounts of energy involved : remarkable quantities of soil may be displaced from the basin area or catastrophic floods may be caused, such as those recorded along most inland rivers in Romania in 1970, 1972 and 1975.

The amount of precipitation received by a basin is estimated on the basis of daily or weekly data provided by nonrecording or recording rain gauges. For a particular drainage basin, precipitation is determined with the aid of Thiessen polygons or isohyets drawn by interpolation. In order to determine the amount of precipitation in litres per s and km$^2$, $P(\mathrm{l\ s^{-1} km^{-2}})$, a value that is more easily comparable with the mean specific discharge, the following relation is used :

$$P(\mathrm{l\,s^{-1}\,km^{-2}}) = P\,(\mathrm{mm\ \ y^{-1}})10^6/32{,}556{,}926$$

For a more detailed analysis concerning both the installation of rain gauges and the processing of data, account should be taken of the following conditions (Serra, 1951) :

the effect of ground slope on precipitation is obvious, in that an inclined surface receives a larger or smaller amount than a horizontal one, depending on whether or not it is exposed to rain ;

the greater the altitude, the greater the amount of precipitation : in Romania, this relationship is used successfully in drawing isohyets (there are, nevertheless, cases in which precipitation decreases with altitude, e.g. for the eastern coast of Madagascar, which is affected by the alternate action of trade winds and monsoons) ;

precipitation decreases towards the interior of a continent, commensurate with the distance from the coast, due to the fact that precipitation is discharged progressively as cloud systems advance towards the interior.

Research has also cast light on the relationship between the size of an area affected by torrential rainfall and flood frequency, to the effect that the larger the area of a drainage basin, the less extensive the effects of torrential rain, and vice versa.

*Inputs due to man's action.* The rapid development of industry and agriculture, parallelled by the population explosion, has resulted in higher consumption of water. In very many instances, in order to cope with these problems the natural limits of drainage basins are no longer observed and considerable amounts of water are transferred from one basin into another. The supply of industrial and drinking water to great urban centres demands the transport of harnessed streams through pipes tens or hundreds of kilometers long. Huge farming areas situated in regions poor in water resources are criss-crossed by irrigation canals through which water from neighbouring basins is carried. Lowlands in which phreatic water is near the surface are drained and reclaimed for agriculture. Shipping requirements have often led to the building of canals, in which water movement is controlled in relation to man's interests.

Given the omnipresence of human activities, it is fairly difficult to carry out quantitative system studies without taking into account the resulting movements of matter from one basin to another. Temporal variations of the flow of matter should also be studied taking human activities into account, since, for example, strings of storage lakes have been created along many rivers in order to produce a different temporal distribution of discharge, related to man's needs. Such storage lakes play a very important role in attenuating floods and, implicitly, affect the intensity of river-bed processes and thus the evolution of a river's longitudinal profile.

All these situations must be known thoroughly when studying a drainage basin using a systems approach, since any quantitative change in a variable causes a chain reaction in a fairly large number of other components of the system. Therefore, detailed knowledge is required of the origins and sources of matter in the system and of the ways in which this matter circulates.

*Underground inputs.* As is known, a diversity of geographical landscapes with varied tectonic structure and lithology may characterize drainage basins. This explains why, in very many instances, surface and underground water divides do not always correspond; hence there is no homogeneity in this respect. There are instances when springs appear on fault lines, carrying to one basin water that has fallen and infiltrated into another. Sometimes, an underground flow of water from other basins may appear on the flank of a syncline or anticline, either through a stream developed expressly in relation to this flow, or via the bed of a drainage channel fulfilling a runoff function when the channel depth is such that its bed intersects the phreatic water. In karst areas, there are instances where large flows of water move from one basin into another regardless of the morphologically obvious drainage boundary.

*Inputs due to wind.* In addition to inputs of matter in the form of precipitation, matter carried by the wind may also penetrate into a drainage basin. This may be liquid precipitation carried over a drainage divide by strong winds ; snow, which can also be moved easily by strong winds from one basin into another, especially in plain areas; or dust and sand carried by the wind into a drainage basin and sometimes capable of altering the surface topography considerably. This is more obvious in areas where the wind is an important modelling agent. Sand dunes can change the drainage divide and hence the area of a basin. By virtue of its high permeability, sand influences runoff processes and may disturb the circulation of water within a basin.

In the above situations, no account has been taken of the climates of large sub-tropical deserts or of regions covered permanently by ice, where the very concept of a drainage basin is not applicable.

*Solar radiation.* Besides the energy possessed by matter in motion, the major source of energy for a drainage basin is solar radiation, which at the upper limit of the atmosphere is of the order of 2 cal $cm^{-1}min^{-1}$, a value known as the solar constant. More than one-half of this energy returns into outer space following certain processes that occur in the atmosphere and at the Earth's surface; approximately 9 % is returned by diffuse reflection and 25 % by diffusion from clouds and the ground; approximately 19 % is absorbed by the atmosphere and clouds. Hence, leaving these losses (totalling 53 %) aside, the remaining 47 % reaches the Earth's surface either directly through the atmosphere (41 %) or through diffuse sky radiation (6 %) (Strahler and Strahler, 1973).

Therefore, a major source of energy at the surface of a drainage basin is solar radiation or insolation, flows of both direct (in the form of parallel rays) and diffuse solar radiation being intercepted by the topographical surface. Obviously, the amount of radiation received by unit surface area varies in relation to a number of factors which affect it quantitatively. One such factor is the activity of the sun, and another the distance of the Earth from the sun at a given moment. The height of the sun above the horizon and the physical state of the atmosphere are further factors, the amount of radiation reaching the ground decreasing with increasing humidity and cloudiness.

Taking into account that the received intensity $I'$ of direct solar radiation is directly proportional to the sine of the angle $\alpha$ at which the source intensity $I$ falls on a given surface, its value can be calculated from the relation

$$I' = I \sin\alpha \tag{1}$$

Insolation is thus directly related to latitude and varies both annually and daily. For a particular drainage basin it can be calculated according to the methodology developed by Lee (1964).

This flow of energy with its variations over time and space is the major factor determining the great diversity of environmental conditions. The energy received forms the basis of the complex chemical processes of photosynthesis, whereby the supply of energy required by other biotic systems in the environment is ensured. The ways in which this energy is conveyed, stored and transformed in the real world are so numerous that it is not possible to give an exhaustive account in this presentation.

In addition to direct radiation, the area of a drainage basin receives an amount of diffuse radiation from the atmosphere, to which can be added that reflected back to the surface by clouds. It is known that a proportion of the parallel rays passing through the atmosphere are diffused by gas molecules, water vapour and solid particles, and for this reason take a longer time to reach the Earth's surface. As a result of the thermal energy thereby imparted to the atmosphere, the latter acts as a further source of long-wave radiation, of which a certain proportion reaches the Earth's surface and is to be included among the sources of energy arriving at the latter.

It is absolutely necessary to take into account other forms of energy in addition to insolation, such as mechanical energy (potential or kinetic), which can be studied only in relation to the matter bearing it. A large proportion of the water falling as precipitation within the area of a drainage basin will have absorbed solar energy, storing it in the form of latent heat of evaporation which is carried in its ascent to form clouds, to be released again during condensation, thus providing the energy necessary for the wind and storms that accompany most rain. Within a cloud, the potential energy of a raindrop is commensurate with its mass and the falling distance. The moment fall begins, the potential energy turns into kinetic energy, part of which is absorbed by the surrounding air through friction, the remainder being released on impact with the Earth's surface.

## Outputs of Matter and Energy

The quantities of matter and energy contained within a drainage basin can circulate in a large number of ways. Matter and energy can be stored in various processes, or may interact with other components of the drainage basin; they may leave by surface or underground outflows, by evapotranspiration or as the outcome of man's action. Without going into details of the processes of transformation and circulation of matter and energy within a basin, this Section deals with their outputs.

*Runoff*. This is the amount of water which enters a stream network via overland flow and leaves the respective basin through its mouth. It is estimated as volume per unit time ($m^3s^{-1}$) or as average specific discharge ($l\,s^{-1}km^{-2}$) with dimensions of speed ($L\,T^{-1}$).

On contact with the soil, the kinetic energy of falling raindrops is consumed in the actions of splitting, scattering and transporting soil aggregates, or is absorbed by vegetation if the raindrops are intercepted by plant cover. The subsequent paths followed by the water depend on the characteristics of the basin area in relation to the amount of precipitation and to the amount of energy received from the sun.

By virtue of capillary action and gravity, water tends to infiltrate through the pores of soil layers or through rocks, but in many cases the infiltration capacity is easily exceeded and, in the case of a sloping area, runoff will then occur. In this process, the surface conditions at the slope determine the way in which the elementary processes unfold. For a slope which is well protected by plant cover, a large amount of the kinetic energy of the falling water will be consumed through friction, and consequently the transportation capacity will be decreased. In contrast, a slope protected insufficiently, or with noncompacted soil, will allow the moving layer of water to transport huge suspended saltation and contact loads. Transport of these materials takes place as a function of the force of the current, which depends on the overland slope as well as on the discharge, so that the coarsest grains are deposited first when the slope decreases, as a result of the decreasing transport capacity of the current. Only the finer materials are carried further and enter the channel network. Nevertheless, fairly large amounts reach the organized channel network in this way, and may thereby eventually be carried out of the drainage basin. Sometimes, especially in karst areas, colouration of the water is a sign of huge amounts of dissolved substances being transported in solution.

Once in the channel network, the water discharge, together with its mineral and organic contents, both solid and in solution, is transported to the mouth of the basin, where it enters another river or flows into a lake or the sea, thus eroding the drainage basin it has left. Analysis of the amount of alluvial matter transported by rivers into the world's oceans reveals that the largest amounts (600 t $km^{-2}y^{-1}$) are transported by rivers in Asia, while the smallest amounts (27 t $km^{-2}y^{-1}$) are carried by those in Africa (Holeman, 1968). From Romania, some 44.5 million tonnes of solid material are carried away each year, which represents an average specific discharge of 1.88 t $ha^{-1}y^{-1}$ (Diaconu, 1971 b). The conditions specific to Romania's relief favour a fairly direct relationship between the amount of runoff from drainage basins and their average altitude.

For a given drainage basin, the *theoretical runoff energy* $E_r$ can be computed as the sum of the products between the average specific discharge $q$(l $s^{-1}km^{-2}$) over a given period, e.g. a year, and the areas $a_i$($km^2$) and heights $h_i$ (above a reference level) of the corresponding surface regions :

$$E_r = K \sum_{i=1}^{n} q\, a_i\, h_i \;(\mathrm{GWh\ y^{-1}}) \tag{2}$$

Sea level may be taken as the reference level, but the most correct reference is the altitude of the mouth of the respective basin. The constant $K$ in eqn. (2) takes the value $86 \times 10^{-6}$. Another value used in practice is the *theoretical linear energy* $E_l$, which is the sum of the products between the amounts of discharge at different levels and the differences $\Delta H_i$ between the altitudes of the points of inflow and outflow in the sectors taken into account:

$$E_1 = K_1 \sum_{i=1}^{n} [(Q_{in} + Q_{out})/2]\,\Delta H_i \text{ (GWh y}^{-1}\text{)} \tag{3}$$

Therefore, the average annual discharges ($m^3 s^{-1}$) at the points of inflow and outflow and the latter's differences in altitude must be known. The constant $K_1$ in eqn. (3) takes the value $86 \times 10^{-3}$.

*Evapotranspiration.* This phenomenon concerns water, from both precipitation and underground sources, which is returned to the atmosphere in the form of vapour, incorporating a large amount of energy. At the Earth's surface, the ratio between precipitation and evapotranspiration varies as a function of latitude, being greater than unity in equatorial-forest zones and in areas more than 40° north or south of the equator. In savannah and subtropical-desert zones, the ratio falls below unity. In Romania, 529 mm of the total yearly precipitation returns to the atmosphere through evapotranspiration (Diaconu, 1971b).

The amount of water leaving a basin area through evapotranspiration depends on a number of general and local factors. One quantitative general factor determining this phenomenon is the amount of insolation the drainage basin receives, which depends on its location on the globe. Local factors include plant cover, and the nature and type of superficial deposits. A well-vegetated basin not only retains a greater amount of water which evaporates easily, but large amounts of water are also released into the atmosphere through transpiration. In contrast, soil poorly protected by plant cover allows greater amount of runoff, which carries a proportional quantity of mineral matter away from the basin area.

Generally, few data exist concerning actual evapotranspiration rates, and it is possible only to estimate the value of this variable, either directly or indirectly. Direct measurements may be made by means of evaporimeters, but the number of evaporimetric stations is fairly small. For indirect determinations, empirical formulae are used which take into account a number of climatic elements on which the phenomenon depends. For instance, the most general formula ($E = P - Q$) may be employed for estimating the average annual value of evapotranspiration when sufficient rain gauges are available for adequate assessments of precipitation $P$ and water discharge $Q$.

*Outputs due to man's action.* The construction of domestic and industrial water supplies, or of canals linking various waterways of the same system or of adjoining basins, transferring variable amounts of water, cannot but influence the flow and transfer of matter within a drainage basin. Therefore, detailed knowledge of these losses and their range of variation is essential in order to analyze a drainage basin when such factors are present.

*Energy outputs.* When considering energy outputs, account must be taken of the fact that any amount of matter leaving a basin carries away a proportional amount of energy. If only a change in state of water is considered, such as evaporation, each gram of water consumes 539 cal, most of which is provided by solar radiation. Hence, huge amounts of energy may leave a basin through latent heat of evaporation. Among the losses of energy should be included the solar energy which, although incident on the basin area, is not retained by the latter but is reflected due to particular surface features. The percentage of the incident radiation energy reflected by a surface is known as the albedo, and in the case of snow and ice it can be as high as 45—80%.

Some of the diffuse long-wave radiation received at the surface is also reflected back into the atmosphere. Furthermore, the heated surface of the Earth emits long-wave infrared radiation which continues to be emitted during the night when no solar energy is received.

Over several years, the average levels of underground inputs and those due to the wind are by and large equal to the corresponding outputs. Hence the number of variables affecting the flows of matter and energy within a basin can be reduced, thus allowing quantitative estimates to be made under appropriate conditions.

*Underground outputs.* Underground outputs occur when surface water divides do not coincide with those underground, and consequently a number of underground water horizons supplied within a basin flow out into other basins. This can happen because of the occurrence of faults or of a structure which is not confined to a single drainage basin, as often occurs when the latter is small, and sometimes even as a result of neotectonic movements. The magnitude of underground losses depends on the transmissivity coefficient and on the discharge capacity of the respective horizon. As regards underground losses, detailed studies must be made in particular of karst areas which do not observe the surface limits of drainage basins, in order not to introduce erroneoul values.

*Outputs due to the wind.* During strong storms, drainage basins may receive an additional although usually insignificant amount of precipitation carried by the wind, depending on the wind direction. In plain areas, snow may be swept along by strong snowstorms and can travel far beyond the limits of the drainage basin in which it fel originally.

In very many instances, the wind carries impressive amounts of mud and sand from the beds of rivers to other basins shaping these materials into dunes which can reach tens of kilometers in both width and length. Examples of this phenomenon are frequent, and although most such dunes are fixed at present, they may have been responsible for changes from the initial configuration of the corresponding basins.

## Variables Defining the Characteristics of a Basin

The present surface of any drainage basin is the result of a long process of evolution, in the course of which dynamic equilibrium has been achieved between the general flows of matter and energy and the variables which define the behaviour of the basin towards these flows. Generally, there are two groups of factors with differing tendencies: on the one hand, there are agents which, through the flows of matter and energy introduced into the system, act as forces that tend to lower the basin surface continuously; on the other hand, there are factors which resist this process of erosion, lending unity to the whole basin and undergoing a continuous process of adaptation to achieve dynamic equilibrium. The main elements contributing to the definition of a basin's characteristics are rock type, relief, soil type and depth, and plant cover (Fig. 2). All of the above factors appear to be qualitatively distinct variables with well-defined interrelationships, but they are hard to separate and estimate quantitatively.

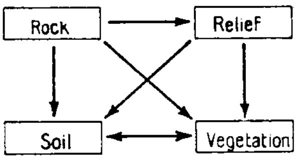


Fig. 2. Variables defining the characteristics of a drainage basin, and their interrelationships.

*Rock* is a very important element, both as a support for other elements and in forming the morphometrical features of drainage basins. The relief of a basin depends to a great extent on the degree of resistance of the rocks making up its surface. The type of rock also influences the soil layer and vegetation (Fig. 2).

On a geological scale, in the process of paleogeographical evolution rock type has played a remarkable part in the formation of the current configuration of drainage basins. With regard to the behaviour of rocks under the action of external agents, we are less interested in their genetic classification than in their resistance to erosion. Therefore, a simple classification of rocks as *unconsolidated*, with a low resistance to mechanical stress and with low or nonexistent cohesion, or *indurated*, with a greater resistance to mechanical stress, is sufficient. The presence of unconsolidated rock, which can be eroded and transported easily, favours large suspended loads. The fairly high porosity (between 25 and 65 %) of this type of rock, which confers high permeability on its deposits, facilitates the infiltration of water from precipitation, resulting in lower surface runoff. Hence, deposits of this type often contain important reserves of underground water, which in most instances are of good quality and suitable for drinking-water supplies. In the case of unconsolidated rock, the more rugged the basin relief, the faster the latter's evolution and the greater the tendency for the basin area to be eroded under the action of external agents. The porosity of indurated rock varies between 5 and 25 %, depending upon the nature of the material. In the case of metamorphic and volcanic rock, it may decrease to 3 % or even below 1 % (Gregory and Walling, 1973). The low permeability of indurated rock explains why the infiltration capacity is rapidly exceeded and runoff is more abundant during torrential rain, while the high resistance to erosion favours a low suspended load.

Although the role of rock type in the morphometry of drainage basins is acknowledged unanimously, no coherent method has so far been developed for estimating quantitatively the resistance of rock to the action of eroding agents. It is known that the response of rocks to such action depends on their physical and mechanical properties. The main physical properties of rocks are porosity, specific weight, compactness, permeability and solubility, while the main mechanical properties are cohesiveness, resistance to mechanical stress and perforation, and especially hardness and resistance to wear, and to compression, traction, shearing and bending.

It is very difficult to determine an index incorporating all these properties because of the diversity and spatial distribution of the rock types in a given region. Nevertheless, a number of attempts have been made to classify rocks from the technico-geological viewpoint taking these practical difficulties into account, although most of them have had limited success in relation to the practical goal pursued.

A quantitative rock classification was suggested by Protodiakonov and Boblikov (1957), who divided rocks into 15 classes. According to this classsification the *coefficient of hardness* $h$ is given for unconsolidated rock by the tangent of the angle of internal friction ($h = \tan \alpha$), and for indurated rock by the average resistance to compression $R_c$. In the latter case, the coefficient of hardness is obtained on the basis of the relationship

$$h = R_c/100 \tag{4}$$

In Romania, a classification of rocks according to their resistance to perforation was made by the Geology Department of the Ministry of Mines, Oil and Geology, rocks being divided into six groups with twelve categories from the softest to the hardest. A first observation was that the rocks with the greatest resistance to erosion are generally older, highly indurated rocks which produce a relief with steep slopes. As the resistance to erosion diminishes, the relief formed by the various rock types becomes correspondingly less extreme, with gentler slopes. The resistance of poorly consolidated or even unconsolidated rock to boring and perforation decreases while the resistance to erosion increases with lower values of slope and relief energy.

Zdražil (1965) took into account the influence of rock type on erosion processes in estimating soil losses. He distinguished four groups of rocks on the basis of a coefficient (the coefficient of geological influence) characterizing the effect of rock type, which depends on permeability and other rock characteristics. The group with the highest coefficients (1.30 — 1.50) includes rocks with low permeability, but which, through weathering processes, produce fine particles. Then follow rocks with low permeability which produce clay—sand material and whose coefficients range from 1.10 to 1.30. The third group, with coefficients between 0.90 and 1.10, includes the more permeable rocks with little resistance to weathering and which disintegrate to produce sand—clay materials. The last group, with the lowest coefficients (0.70 — 0.90), contains the most permeable rocks which weather to produce coarse-grained sandy or stony material. Starting from this classification, Stehlik (1970) calculated a coefficient of geological influence for each geological formation in Czechoslovakia. These values were then used to elaborate a methodology for mapping soil erosion.

Quantitative estimation of the resistance of various rock types to erosion is possible only after a detailed analysis of the lithology of a basin and correct classification of the rocks according to their resistance. By mapping the areas with their coefficients, the areas corresponding to each type of rock within a basin can be determined by planimetry. The *average coefficient of geological resistance* $R_g$ is then given by the area-weighted average of the partial coefficients :

$$R_g = (R_{g1}a_1 + R_{g2}a_2 + \dots + R_{gn}a_n)/A = \sum_{i=1}^{n} R_{gi}a_i/A \qquad (5)$$

where $R_{g1,\dots,n}$ represent the geological resistances of the formations within the area of the basin according to the classification chosen, and $a_{1,\dots,n}$ the areas corresponding to each formation; $A$ is the area of the whole basin. This coefficient, determined for a large number of basins, has given good results when correlated with a number of other morphometrical elements of drainage basins and channel networks.

*Relief* is the totality of land forms in a given drainage basin. The current aspects and characteristics are the result of a long process of evolution and interaction between internal and external factors. Internal factors are controlled by the extents of tectonic and eustatic movements, which determine the altitude and potential energy of a drainage basin. This energy varies within the basin in relation to its height and relief, and over time : it would have been maximum during the period of most intense orogenic activity, decreasing slowly with erosion and levelling by external agents. The action and intensity of external agents depend on the global position and altitude, which determine climatic conditions and hence the rates of flow of matter and energy. From an analysis of the orogenic, neotectonic and eustatic movements which have affected a basin, the cyclicity of the increases and decreases of the internal energy of the system can be inferred.

The altitude and latitude of a drainage basin are very important with regard to the inputs and outputs of matter and energy. A basin in a mountainous region, for instance, receives a larger amount of precipitation and a smaller quantity of solar radiation in comparison with one situated in a plain area. The high potential energy at the surface of such a basin allows a fast flow of water down the slopes and partial storage at their bases and in valley deposits. Hence, the surface runoff is larger while the amount of water leaving by evapotranspiration is much lower. The relief of a basin also exerts an influence on the plant cover and soil layer, a close interrelationship existing between these elements.

*Soil* is another component which depends on the type of rock, on vegetation, on the spatial position of a basin, and hence on climatic factors (Fig. 2). Being the outcome of a lengthy evolution, the characteristics of a soil can provide information about the conditions existing at the time of its formation. A good example in this respect is the spatial distribution of forest reddish-brown soil.

However, formation of a soil layer presupposes the existence of a dynamic equilibrium between the soil-forming factors. If this equilibrium is altered, the new situation will be reflected by a vertically zoned soil profile, as in the case of underlying soils.

A well-formed and conserved soil layer is of particular importance in the processes resulting from the circulation of matter and energy in a drainage basin. The hydrophysical properties of soils govern the circulation of water, the major source of matter. These properties are of decisive importance in the processes of runoff, water infiltration into the soil, and soil washing and erosion. Thus, soils with high permeability result in a lower surface runoff but enhance underground water reserves, thereby modifying the temporal distributions and the ratio of surface to underground runoffs, the main routes of circulation of matter.

In defining the hydrophysical properties of soils, texture is one of the most important characteristics. It determines to a great extent the permeability and the capacity to retain or lose water, as well as the cohesion and resistance to the action of external agents. Soil texture is estimated in relation to the percentages of sand, loam and clay found in the mineral mass. According to these percentages, soils may be basically sandy, loamy or clayey, but the various ratios between these components determine a whole series of intermediate groups with well-defined properties. Sandy and slightly loamy soils, for instance, contain up to 80% of sand, which results in high permeability to water and air but lowers their retention capacity. In the case of soil with a slight or moderate clay content, these properties are balanced, the soil being moderately permeable to water and air and retaining absorbed water well (Chiriţă, 1974). For clay—loam soils the percentage of clay increases to 35 — 45%, which results in lower permeability and higher capacity to retain water. If the clay content exceeds 50% of the mineral mass, the hydrophysical characteristics of the soil are altered considerably. Thus, the presence of a large proportion of particles of diameter less than 0.002 mm leads to an increase in capillarity and a decrease in the number of large pores, which results in lower permeability. Water infiltrates only with great difficulty into such soils, and frequently gathers in pools at the surface when the slope is gentle. In the case of sloping areas, surface runoff is abundant and the soil is subject to strong surface washing and erosion, which in most instances results finally in destruction of the soil profile. The low permeability of these soils proves considerably detrimental during torrential rainfall, when most of the precipitation runs off and leads to excessive flooding.

The particle size distribution and texture of soils play a very important role in infiltration and surface-runoff processes. Thus, it has been noted that an increase in the percentage of sand in a soil lowers surface runoff. There is a close relationship between infiltration and runoff in that the factors favouring runoff hinder infiltration, and vice versa (Bertrand et al., 1964).

Moisture content is also an important factor in processes of soil erosion. If, for instance, torrential rain falls on soils where moisture capacity has been saturated by previous rainfall, runoff will appear immediately. Such rapid surface runoff will possess a high erosion and transport capacity, favouring both the deepening of old rills and gullies and the formation of new ones.

*Vegetation* is another element of the basin surface which depends on the soil, rocks, relief and location of a basin. A close interdependence exists in particular between soil formation and vegetation (Fig. 2). Through its influence on climatic conditions and hydrological processes, vegetation affects the circulation, spatial distribution and annual regime of energy and matter within a basin. The canopy and litter of a forest retain a large amount of water and hence diminish runoff, while at the same time increasing the quantity of water that infiltrates into the soil, thereby augmenting underground reserves. Thus, the distribution over time of the amount of water falling as precipitation is altered, and the ratio of underground to surface supply is changed. In basins for which afforestation coefficients are low, runoff is greater, and vice versa.

In winter, a forest may accumulate snow, and hence increase its water reserves. In spring, such accumulated snow thaws much more slowly than in open country, and thus surface runoff and infiltration are prolonged.

Grass-type vegetation plays a protective role for the soil, depending on the plant type and density. Grasses such as *Graminae* have extensive root systems which consolidate the higher levels of the soil, modifying their structure and enhancing their hygroscopicity. A dense plant cover considerably lowers the kinetic energy of rainfall reaching the soil, while through the rugosity it creates, overland flow is arrested to a great extent and the force of the rain is diminished.

Vegetation also plays a considerable role in the water balance of a region, through evapotranspiration. Indeed, this depends to a great extent on the type of vegetation and its needs for water. At the latitude of France, the plant cover is estimated to cede to the atmosphere by transpiration approximately 2000 — 3000 tonnes of water per hectare per year. In order to produce 1 t of wheat, for instance, some 250 — 550 t of rain water are consumed, which is tantamount to a layer 25 — 55 mm in depth (Duvigneaud and Tanghe, 1962). One hectare of wheat in the growth season may use some 3750 t of water, equivalent to 375 mm of precipitation.

Forests are also major consumers of water. One hectare of spruce forest, with a foliage weight of 31 000 kg, transpires approximately 43 000 l $d^{-1}$. One hectare of beech forest loses into the atmosphere through transpiration 20 000 — 50 000 l $d^{-1}$, while one hectare of birch forest, with a foliage weight of 4940 kg, transpires over 47 000 l $d^{-1}$ (Duvigneaud and Tanghe, 1962). If it is remembered that these amounts originate from precipitation, the great importance of vegetation in the transfer of mass and energy by various routes within a drainage basin is apparent.

In conclusion, all the variables mentioned work as a whole, resulting in a certain resistance of the basin surface to the action of external agents. The interrelationships between these variables are analyzed frequently in geographical studies. It is known, for instance, that destruction of the plant cover produces a number of reactions of the other variables, either directly or by altering the balanced distributions in time and space of matter and energy. Thus, the amount of water stored may decrease, resulting in greater surface runoff. In turn, the latter acts with greater energy on the soil layer, which it gradually carries away. Reaching the rock, it works upon this too and finally, by virtue of the tendency to achieve a new steady state, lowers the basin area. These relationships as well as their causal patterns have been described well from a qualitative viewpoint. Unfortunately, quantitative assessments and hence the elaboration of mathematical models are still goals to be attained in this field. Geographers have insufficient knowledge of how to estimate quantitatively the role of plant cover as a component of the resistance of a drainage-basin area. The same is true for the resistance of the soil layer, and for geological resistance. Although it is obvious that all these variables act as a whole, with well-defined attributes and interrelationships, no unanimously accepted method has so far been developed for quantitative estimation.

Nevertheless, it is common knowledge that the above-discussed interrelationships and exchanges of matter and energy with the surroundings have moulded in the course of time the shapes and structures of drainage basins, and these, thanks to modern recording techniques, can be studied in the minutest detail.

Chapter II

# Formation of Channel Networks

The action of the various modelling agents on the Earth's crust results in a fairly wide range of processes, all of them tending ultimately towards a state of equilibrium (steady state) between the force of the agents and the resistance of the topographical surface. In these processes, a particularly important role is played by the river network through which the water from a basin flows towards its mouth and, in relation to the energy of the flow, carries a certain volume of material either in suspension or in solution. However, before an elementary course of river with a temporary or permanent runoff function is individualized, the water flow will have acted to model the topographical surface. According to the location of the erosion processes leading to the formation, maintenance and development of a channel network, we distinguish slope erosion and bed erosion.

In its broadest acceptance, the term erosion denotes the processes whereby particles of matter or soil are transported from their initial location by a fluid agent (Strahler, 1956b). Because of the introduction of the notion of a fluid agent in this definition, it is obvious that erosion does not include movements of the landmass caused by the action of gravity, since this implies no fluid agent. The fluid media working in nature upon the Earth's crust are water, air and ice.

According to R. T. Knapp (quoted by Strahler, 1956b), "...erosion is a mechanical process, whose vital components are the forces which cause erosion, those which resist it, and the resulting motion of the eroded material". It is then evident that erosion will occur only for values of the ratio of these forces above unity, that is, when the forces tending to produce movement or shearing of materials (acting via a fluid agent) are greater than those acting to resist them : the greater the difference between the two groups of forces, the greater the extent of erosion.

The shaping and evolution of a river network constitute a complex process including all the successive stages from elementary forms of slope erosion to complex stream forms. We deal first with pluviodenudation, or sheet erosion during overland flow; then, with the development of linear erosion through rills and gullies, we can consider the first elements of a drainage system through which mass transfer proceeds in accordance with the laws of hydrodynamics.

To obtain a clearer picture of the processes which lead to the formation of a river network, it is necessary to study the kinetic energy of rainfall and the path followed by water resulting from precipitation concomitantly with the resulting processes, from the stage of contact with the topographical surface to the development of the channel network.

## Kinetic Energy of Rainfall

In order to characterize the effects of rainfall on the soil, account must be taken both of the rainfall characteristics (quantity, duration, intensity) and of the properties of the soil surface (permeability, granulometric composition, structure, slope, vegetation cover). It is also necessary to note that not all rainfall works aggressively on the soil, but only torrential rain with high kinetic energy. Rain is generally considered to be torrential when the amount of precipitation is greater than 10 mm and its intensity exceeds 0.4 — 0.5 mm min$^{-1}$. Yarnell (1935) developed a formula to express rainfall intensity, which, converted into metric units, is as follows :

$$i_t \geqslant 0.254 + 5.08\, t^{-1} \tag{6}$$

where $i_t$ represents the mean intensity (mm min$^{-1}$) for the duration $t$ (min) of the torrential nucleus.

Rain aggressiveness $I_w$ can also be determined as the product between the kinetic energy $E_k$ and the average intensity $i$ (mm h$^{-1}$) during 30 min (Wischmeier and Smith, 1958) :

$$I_w = E_k i \tag{7}$$

It is possible to estimate the energy of torrential rain starting from the kinetic energy of a raindrop, which depends on its mass $m$ and velocity $v$, using the equation

$$E_k = m\, v^2/2 \tag{8}$$

The mass of raindrops is estimated taking into account that their diameter cannot exceed 9 mm in a calm atmosphere or 5 mm in a stormy one. During rainfall the diameter also varies in relation to rainfall intensity. The velocity at which raindrops fall also depends on their diameter, showing a more marked increase up to a diameter of 3 mm after which, beyond a velocity of some 9 m s$^{-1}$, the rate of increase of the limiting velocity becomes progressively lower (Fig. 3) : air resistance considerably reduces the limiting velocity of fall of large raindrops so that it can never exceed 10 m s$^{-1}$. There is no relation of direct proportionality between the diameter of raindrops and the intensity of rainfall, since if the latter has a greater value, more raindrops will collide with one another and break up. Since not all the relationships existing between the parameters of torrential rainfall are sufficiently well quantified to serve as a basis for determining the kinetic energy

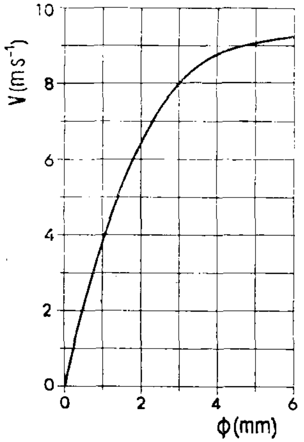



Fig. 3. Relationship between limiting velocity of fall $V$ (m s$^{-1}$) of raindrops in a calm atmosphere and their diameter ∅ (mm) (after Laws, 1941; Gunn and Kinzer, 1949).

of rainfall, the most accurate method is to measure the kinetic energy during rainfall directly (Hudson, 1961). For direct computation of the kinetic energy $E_k$ of rainfall, the following formula was proposed by Wischmeier and Smith (1958):

$$E_k = 916 + 331 \log i_h \tag{9}$$

where $i_h$ is the average intensity per hour of segments of uniform intensity on the pluviogram. In metric units, the above formula becomes

$$E_k = 10^5(3.4 + 2.5 \log i_h) \text{ (kg m ha}^{-1}\text{mm}^{-1}) \tag{10}$$

In practice, the kinetic energy of rainfall is given by the sum of the partial products of the amounts of precipitation falling over certain time intervals with the same intensity, and the average intensities in the respective intervals.

In Romania, detailed studies have revealed a number of indicators that can be computed more simply and can be well correlated with erosion (Stănescu et al., 1969). These indicators are

$$I_1 = Pi_{15} \tag{11}$$

$$I_2 = Pi_{30} \tag{12}$$

$$I_3 = i_{15}\Sigma p_k i_k \tag{13}$$

$$I_4 = i_{30}\Sigma p_k i_k \tag{14}$$

where $P$ is the total amount of precipitation (mm), $i_{15}$ and $i_{30}$ are the average intensities (mm min$^{-1}$) for respectively 15 and 30 min of the torrential nucleus of rainfall, $p_k$ is the amount of precipitation (mm) on a uniform segment of the pluviogram, and $i_k$ the average intensity (mm min$^{-1}$) of the rain within the respective segment. The best of these indicators for characterizing pluvial erosion is that incorporating intensity for 15 min. Moţoc et al. (1975) recommended that only rainfall whose torrential nucleus has an intensity of at least 0.6 mm min$^{-1}$ for 15 min should be considered aggressive.

The amount of water reaching the surface during a rainstorm may subsequently take several paths, the share of each depending on the total volume of precipitation and on local conditions.

## Interception

Interception is the process whereby part of the precipitation falling in a region is retained by aerial components of the vegetation cover or by the soil surface. Research has shown that interception is least (2 mm) for flat soil with little vegetation, and greatest (15 — 20 mm) for old, dense forest with a well-developed canopy (Hâncu et al., 1971).

Horton (1945) proposed the following formula for the estimation of interception $I_n$:

$$I_n = b + cP^n \tag{15}$$

where $P$ denotes precipitation (mm) and $b$, $c$ and $n$ are parameters taking various values from one type of vegetation to another.

Of all types of vegetation, forest plays the most important role in interception. In order to assess forest interception, thorough research has been carried out in France, 15 km east of Nancy (N 48°44′ E 06°14′), at an altitude of 250 m (Aussenac, 1968). For two years Aussenac studied the throughfall, stemflow and amount intercepted by the canopy in forests of various species of resinous and deciduous trees (see Table 1).

The results obtained prove that a fairly large proportion of the total annual precipitation is retained by the canopy and returned to the atmosphere by direct evaporation. The role of forests in the interception of precipitation depends very much on both the duration and intensity of rainfall and on the total amount of water falling. Accordingly, rainfall involving less than 0.6 mm precipitation is intercepted fully by a spruce-fir forest, while a deciduous forest can intercept completely only the rainfall less than 0.3 mm. In the case of more abundant rainfall, it is also important to take into account saturation of the canopy (Table 1): in fact, the heavier a rainfall, the smaller the amount of water intercepted by the canopy, and vice versa. From this point of view, it was noted in the above studies that the most important role in interception is played by spruce and fir forests, followed by pine and deciduous forests (Fig. 4).

TABLE 1

**Precipitation intercepted by various types of forest (after Aussenac, 1968)**

| Type of forest | Age (y) | Average height (m) | Density of trees (No. $ha^{-1}$) | Intercepted precipitation (% of annual sum) | Rainfall intercepted totally (mm) | Precipitation necessary for saturation | |
|---|---|---|---|---|---|---|---|
| | | | | | | Foliage (mm) | Canopy (mm) |
| Pine (*Pinus silvestris*) | 28 | 13 | 1520 | 30 | 0.4 | 3.0 | 3.5 |
| Spruce fir (*Picea abies*) | 30 | 12.5 | 2160 | 34 | 0.6 | 3.1 | 4.0 |
| Vancouver fir (*Abies grandis*) | 35 | 23 | 620 | 42 | 0.5 | 3.8 | 4.5 |
| Beach and hornbeam (*Fagus silvatica, Carpinus betulus*) | 30 | 12.5 | 1300 | 17 | 0.3 | 1.9 | 2.0 |

Apart from interception, vegetation cover plays a very important protective role by breaking raindrops and absorbing their kinetic energy, thereby diminishing considerably the energy of rainfalls and hence their aggressiveness. J. Tricart (quoted by Tufescu, 1966), was

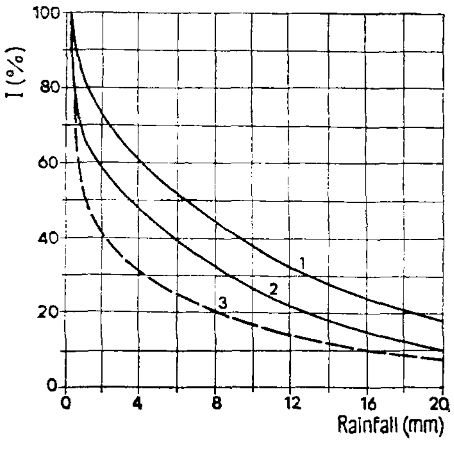



Fig. 4. Forest interception $I$ (%) as a function of rainfall (mm) : 1, Vancouver fir and spruce fir; 2, pine; 3, deciduous trees (beech and hornbeam) (after Aussenac, 1968).

able to distinguish several situations in relation to the percentage of vegetation cover. Thus, pluviodenudation is greatest on bare soil, while on grassland it is alleviated almost entirely as a result of the plant cover breaking raindrops and absorbing their kinetic energy. In a forest, even if the soil is bare the energy of the raindrops reaching it will be small, and hence the effects are insignificant. For agricultural areas, the effect of rainfall depends very much on the type of crop, and on the plant size and density. In the case of maize, for instance, almost one-half of the rainfall reaches the soil, while in the case of a soybean field, the corresponding proportion is two-thirds.

If the soil is not protected, pluviodenudation may be extensive, depending on the characteristics of the soil. Ellison (1952) has shown that a rain shower of 100 mm can move over 300 t of soil per hectare. Tracing the processes occurring in the course of rainfall, it can be stated first that the initial raindrops will penetrate into the soil in relation to the latter's compactness. Their kinetic energy is consumed in lifting soil particles into the atmosphere : this explains the smell of dust after the first raindrops have fallen. Moistening of the soil continues through absorption of the infiltrating water, securing greater cohesion of the soil. During torrential rainfall, soil elements 2 mm in size may be carried by the fall of raindrops to a distance of 40 cm, and those of 4 mm to a distance of 20 cm. These large values are explained by the fact that at the moment they reach the soil, raindrops have a kinetic energy approximately 1000 times greater than the energy of an equivalent mass of water running off at the soil surface (Ellison, 1952).

## Infiltration

Any rainfall is accompanied by infiltration of water into the soil under the influences of gravity, capillarity and absorption. The amount of infiltrated water (mm) depends on the physico-mechanical properties of the soil, the nature of the plant cover, the duration and intensity of rainfall, and the slope. The multitude of local factors on which infiltration depends, and the lack of an expression to render this dependence correctly, explain why most existing formulae are valid only for the conditions for which they were established. For other conditions, they are of use as a reference indicating the general structure of the relationships, but greater accuracy demands local determination of the parameters they incorporate and which are not defined in empirical terms (Hâncu et al., 1971).

Infiltration is more frequently measured using the formula established by Horton (1945) :

$$= f_c + (f_0 - f_c)\, e^{-K_f t} \qquad (16)$$

$f$where $f$ is the infiltration capacity, $f_0$ an initial value, $f_c$ a constant, e is the base of Naperian logarithms, $t$ the period elapsed (h) from the beginning of rain, and $K_f$ is a proportionality factor to be determined experimentally. The infiltration capacity $f$ thus does not remain constant throughout the rainfall. Starting from the initial value $f_0$, infiltration decreases sharply at the beginning to reach a constant value $f_c$ after 0.5 to 3 – 4 h. Infiltration capacity is very important from the geomorphological viewpoint, since a high value presupposes a reduction of runoff and erosion during torrential rainfall, while a low value leads to an increase in the depth of the water layer on gentle slopes and to strong runoff and erosion on steep slopes.

Determinations effected in Romania at the Valea Călugărească, Drăgăşani and Tîrgu Jiu experimental stations have shown that in the case of soils with a fairly high clay content infiltration is related to the moisture deficit $D$ according to the equation

$$I_n = a\, D^r + b \tag{17}$$

where $I_n$ is the infiltration beneath the surface after $n$ min, and $a$, $b$ and $r$ are locally determined parameters (Stănescu, 1964). For agricultural soil, infiltration depends more on the average weighted porosity of the layer over a depth of 30 cm :

$$I_n = m\, p_a + n \tag{18}$$

where $p_a$ is the porosity and $m$ and $n$ are parameters (Stănescu, 1964).

In practice, a number of graphs are used to estimate this indicator. Thus, D. L. Armand's graph (quoted by Hâncu et al., 1971) gives the stabilized infiltration intensity (mm min$^{-1}$) in relation to the intensity of precipitation (mm min$^{-1}$), that of Frewert (1955) allows the estimation of infiltration (mm) as a function of the duration of rainfall (min) for various land uses, while N. N. Chegodaev—E.V. Boldakov's graph (quoted by Hâncu et al., 1971) gives infiltration (mm) in relation to rainfall duration for various soil categories (Fig. 5).

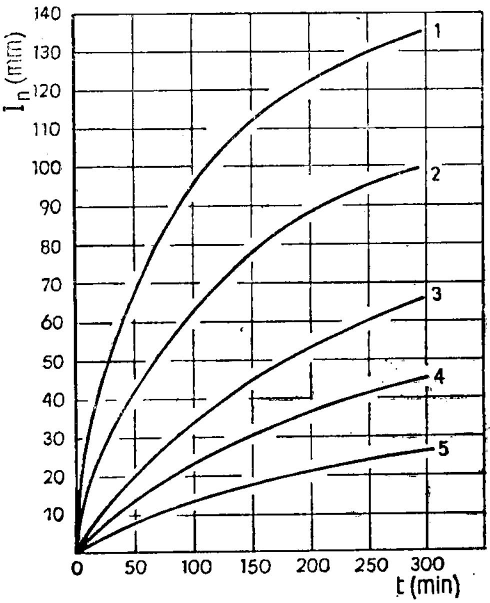
To the extent to which the amount of precipitation falling exceeds the infiltration capacity of the soil, a film of water appears at the surface which, once reaching a certain thickness, prevents the displacement of further soil particles, by absorbing the kinetic energy of the raindrops. By using artificial rainfall with raindrops of identical size falling on calibrated sands and varying the height of fall, Ekern (1950) established a relationship between the amount $E$ of material displaced and the kinetic energy $E_k$ of rain falling for 5 min with an intensity of 25 mm h$^{-1}$ :

$$E = -0.515 + 0.1\, E_k \tag{19}$$

This proves that displacement of soil particles occurs only if the kinetic energy exceeds a certain value. Hudson (1961) showed that the erosive effect of rainfall at an intensity below 25 mm h$^{-1}$ is weak, but that generally the amount of soil displaced in the course of one year is a function of the cumulated kinetic energy. Certainly, the effective energy is greatly diminished by the formation of a water film at the soil surface, and by the presence of vegetation which, depending on the percentage cover, may intercept up to 50 % of the rainfall.

It is thus obvious that once slope runoff has begun it must also carry along material displaced by shear stresses which are expressions of the force of gravity. Indeed, the collision force of raindrops is also a

Fig. 5. Infiltration $I_n$ (mm as a function of time $t$ (min) for various categories of soil: 1, sand; 2, sandy; 3, sand—clay (chernozem with good structure); 4, clay—sand (podzols, degraded chernozem); 5, clay or mainly clay.




result of the action of gravity. Strahler (1952a) also included a number of molecular forces among the forces tending to weaken the soil: water absorption by colloids, hydrolysis of silicates, increases of capillarity on contact between particles, dilatation and contraction, increases of ice and salt crystals and other phenomena may all reduce soil or rock cohesion.

The forces resisting erosion are related to internal friction, to soil cohesion and even to the infiltration capacity of the soil. The latter element implies flow of water into the soil through the pore system, and consequently disappearance or diminution of overland flow.

In addition to considering the external factors which protect the soil, an analysis should also be made of those related to the soil's physico-mechanical properties. Taken together, these result in a certain capacity of the soil to break into particles which may or may not be transported. This property is termed detachability. It is closely related to the degree of cohesion of a soil, being enhanced by mechanical disaggregation or freezing, and diminished in the case of compacted soils or those cemented by organic material.

In conclusion, the process of soil erosion and transport starts the moment raindrops reach the soil, their kinetic energy being absorbed in displacing and translocating soil particles.

## Overland Flow

In the initial moments of rainfall, almost the entire amount of precipitation is retained either by the plant cover or by the soil when the latter's infiltration capacity is higher than the rain intensity. However, to the extent to which the absorption capacity of the soil is exceeded, an excess of water will remain which will flow downwards if the surface is sloping, filling first of all the microdepressions of the soil surface. At this stage the layer of water, although very thin, can nonetheless carry downstream the smallest soil particles, which it deposits in the microdepressions. These, concomitant with filling, are thus subject to a process of silting, which reduces their volume and tends at this stage to result in levelling of the soil. When the rain intensity remains at values higher than the infiltration capacity, the microdepressions become overfilled and streamlets of water start to run off which may cover large slope areas. In the case of a small slope, the water layer may be as thick as 5 — 10 mm. In its movement it will detach soil particles and carry them along downstream. Erosion by water streamlets gives birth to a number of rills which continuously change their routes due to micropiracy. The material detached by surface runoff is usually deposited at the base of the slope, or carried further into the river network, depending on its weight. Of course, an important role in this process is played by the soil properties, and by its structure and texture in particular. Another important factor is vegetation cover, which enhances the rugosity of the soil and markedly lowers the velocity of slope runoff, hence increasing the quantity of infiltrated water. As may be seen, all these effects tend to reduce slope erosion, because a great part of the kinetic energy of the water is spent in overcoming the resistance due to them. Sometimes, piled-up soil fragments and plant remnants form small barriers behind obstacles to their movement, which retain water and lower the velocity of flow. However, when such barriers collapse, the resulting streamlets of water will have even greater erosion and transport effects. Obviously, slope runoff has very many parameters, which have been described in detail by Horton (1945).

Once raindrops have destroyed and detached soil aggregates, overland flow serves as a means of transporting the displaced elements. In cases in which the overland flow possesses sufficient kinetic energy, it too can participate effectively in dislodging soil particles. However, as has been seen, the occurrence of overland flow is linked to the formation of a water layer which can no longer be absorbed by the soil, under conditions of rain of given intensity. Experiments have proved that water infiltration into soil depends on the latter's properties, each type of soil having a limited infiltration capacity. Generally, attempts to find the moment when overland flow begins have been made by relating rain duration to its intensity. The rain intensity $I_r$ (mm h$^{-1}$) which triggers erosion is calculated on the basis of the Yarnell (1935) and Riou (1963) formula

$$I_r = 1500\,(0.20 + 0.01\,t)/t \qquad (20)$$

where $t$ is the rain duration. Subsequent research has demonstrated, however, that the relationship is not applicable for short rainfalls, because the total rainfall is too small to produce overland flow. Thus, it was proposed that only rain intensities greater than 20 mm $h^{-1}$ be considered excessive. However, this assessment cannot be generalized if no account is taken of the properties of each type of soil, which may accelerate or arrest the start of overland flow. Of these properties, the primary factor is the amount of moisture already present in the soil, since this reduces the effective porosity : as a consequence, the greater the interval between two rainfalls, the higher the absorption capacity and the greater the amount of precipitation needed to produce overland flow. Likewise, the texture and structural stability of a soil also influence infiltration, in relation to whether the soil is compacted or formed from loose aggregates. The continuity of the soil profile must also be known, since the presence of an appreciable amount of clay arrests infiltration considerably, which also happens in the case of wet soil which has frozen.

It can thus be stated that the appearance of overland flow is directly related to the intensity, duration and kinetic energy of rainfall and to the soil's structure and structural stability, as well as to its moisture content and profile continuity. Hence, a large number of factors must be considered in any quantitative evaluation of this phenomenon.

Once it has occurred, overland flow works upon the soil to detach particles and aggregates as a function of stream energy. In order to determine the transport capacity $Q_s$ of streams, that is the saturation content of suspended alluvial matter, E. M. Laursen (1959) (quoted by Moţoc et al., 1975) proposed the relationship

$$Q_s = K\, S^{1.67}\, Q^{1.67} \tag{21}$$

where $K$ is a soil factor, $S$ the slope and $Q$ the liquid flow. The material detached is transported in suspension, by saltation and by rolling, depending on its size and the velocity of the stream. For instance, the limiting velocity required to move a soil particle by rolling will depend on its density and volume, and on the shape and surface it presents to the stream (Feodoroff, 1965). Since this limiting velocity is very hard to measure, it has been suggested that it should be replaced by the critical traction force $F_c$ given by the formula

$$F_c = \gamma h S \tag{22}$$

where $\gamma$ is the fluid density, $h$ the thickness of the water layer, and $S$ the slope. Using this formula, the limiting diameter of particles which can be carried by the critical traction force has also been determined.

All the above processes result finally in carving of the soil surface and in the appearance of rills and gullies along which the water flow carries large amounts of soil towards the base of the slope or into the channel network. In most instances, these rills disappear after rainfall, either naturally or as a result of agricultural work. Only when the streamlets are sufficiently strong to deepen their courses, either in soil or in rock, will they remain until the subsequent torrential rain. Repeated rainfall with erosive effects plays an important role in the orientation and organization of overland flow, creating embryonic drainage paths which may develop into rills, gullies and other forms of organized drainage. Certainly, man's action can also result in the appearance of rills and their development into gullies, either through the use of inadequate methods of soil tilling, or through the creation of paths or bridleways down steep slopes.

*Rills and gullies*. As has been noted, excess water on a hillside always tends, under the influence of gravity, to move down the line of steepest slope. If several streamlets coalesce on the same path, the volume of localized runoff increases and with it also the kinetic energy of the moving mass. This may be sufficient to dislodge soil particles, and sometimes even coarser materials which enhance the errosive power of the streamlet. Thus, by the concentration of several streamlets on a common path, surface erosion develops into linear erosion. The outcome of this process is the formation of an incipient water course. Research has led to a number of formulae establishing the transport capacity of a water stream, taking into account the flow, depth of current and slope (Moţoc et al., 1975). In the apparently simple process of stream development, which is nonetheless difficult to express in quantitative terms, alluvial matter in the channel is disintegrated and subsequent transport is thereby facilitated.

The depth and size of incipient water courses also depend on the physiographical conditions of the slope. If their depth exceeds 20—30 cm, it is likely that they will remain until the subsequent rainfall, when they will be deepened still further and will become permanent. Such a path, representing the smallest conduit for the runoff of excess rainfall, is known as a *rill*. Its depth in the topographical surface will vary in relation to the nature of the surface in which it is carved (Tufescu, 1966). As a rule, the length of such formations is linked to the distance over the hillside for which approximately the same slope is maintained. Downstream, once the slope changes, rills may turn into more evolved forms, creating a continuous route whereby materials brought from upstream are evacuated. In other instances, the slope may change such that the rill is obliterated by the surface of its alluvial fan.

The above features, which have the capacity to orient and organize runoff, evolve by deepening with the passage of time. If the rill has reached the base rock, we are already dealing with a *gully*. It should be noted from the beginning that these classifications in relation to depth are arbitrary and subjective, the elements characterizing them being insufficient from the quantitative viewpoint. However, it is obvious that such forms develop, although for very short times, in accordance with the laws governing the general processes of river erosion. Thus, an element of very great importance is the slope of the longitudinal profile, which has a marked influence on the erosion and transport power of the water flow, and thereby governs the rate of development of the water course. As a rule, all these elementary forms resulting from linear erosion fulfil the runoff function temporarily, while most of the time their banks and sides evolve under the influence of gravity, especially in the case of nonconsolidated rocks. This explains why before torrential rainfall, the courses of these formations become covered with a layer of material which has fallen from the banks and which is carried downstream to larger channels on the occasion of the first flood.

If, during deepening, one of the above forms intersects a groundwater table, the runoff function for which it has been created becomes seasonal or permanent. In this case, the laws of hydrodynamics act for longer periods or permanently, and thus the new stream fashions the morphometrical elements of the bed in relation to the available kinetic energy of the flow.

As is known, any mass in motion possesses a certain capacity to effect mechanical work. In the present case, we are dealing with a certain hydraulic energy, kinetic and potential, which is equal to the product of the weight of water and its distance of fall. It is thus obvious that the energy available for erosion and transport is directly proportional to the amount of runoff and distance of fall specific to a channel network. Most of this energy is consumed in overcoming resistance due to bed rugosity and to internal friction, and in transporting material in suspension, by saltation and by rolling along the channel. Erosion will occur when the energy of a water stream is greater than that required to overcome friction and to transport material in suspension. However, as erosion proceeds, the slope at a given point decreases: concomitantly, the flow velocity will decrease and, if the discharge is constant, also the energy. When the energy consumed in overcoming friction and in transporting suspended material exceeds the available energy of the stream, the latter will deposit a proportion of the material previously carried in order to attain energy equilibrium. Once this has happened, we are dealing with a channel network, i.e., with well-formed channels along which water runs according to the laws of hydrodynamics.

In conclusion, it should be pointed out that the movement of precipitation from its contact with the soil until it reaches well-formed channels takes place in several stages. First comes direct action of raindrops on the soil; then, surface runoff, which prevails in cases of intense or prolonged rainfall, or snow thawing. On sloping ground, this two-dimensional surface flow is soon replaced, at a short distance from the water divide or upon a small change of slope, by a linear form in which several streamlets unite, their combined energy being sufficient to carve into the topographical surface. At this stage, the first elements of a new stream have already appeared. Repetition of the phenomenon and the deepening produced by erosion result in a greater distance between the initial point and that at which the stream flows into a larger collector. During torrential rainfall, the discharge increases and the slope steepens as a result of deepening: concomitantly, the erosion and transport forces are enhanced. Thus the prerequisites are created for the appearance of an organized network of rills and gullies which are the incipient elements of a new drainage basin.

The above forms, even if they fulfil only temporary runoff functions related to precipitation, should be taken into consideration as soon as an elementary, morphologically outlined water course has been shaped which is capable of maintaining itself and evolving for a longer period. On deepening to reach the water table, such channels may acquire a seasonal and then a permanent runoff function. However, hierarchization of water courses according to the time during which they exercise their function is unidimensional and hence inadequate. It does not take into account all the morphometrical features characterizing a drainage basin. As will be seen, this problem has been solved by the system of classification developed by Horton and Strahler.

Chapter III

# Systems of Stream Classification

The classification of channel networks has always been among the concerns of researchers. As basic criteria, both the plane outline of a network and various dimensional elements (basin area, length of streams, depth, direction, water discharge, position in relation to main river, etc.) have been considered.

A variety of physiographical and geological conditions (rock type, tectonics, structure, etc.) are decisive in defining the plane configuration of a channel network, but these alone are insufficient to provide a genetic basis to be used in the classification of river networks.

An important step towards a more accurate classification of channel networks was made as early as 1800, when the British geologist John Playfair (quoted by Horton, 1945) showed that: "Every river appears to consist of a main trunk, fed from a variety of branches, each running in a valley proportioned to its size, and all of them together forming a system of vallies, communicating with one another and having such a nice adjustment of their declivities that none of them join the principal valley either on too high or too low a level". On this basis, which was quite progressive for its time, when river valleys were still considered to be the outcome of cataclysms, the idea can be derived that valleys are proportionate in size to the rivers which carved them and which flow along them, which has proved an idea of considerable importance in the subsequent development of morphometry.

Of the attempts to typify the plane aspect of river networks, that of V. A. Troitsky (quoted by Morariu et al., 1962) deserves mention. Troitsky characterized several types of river system in the USSR, including the radial, centripetal, tree, right angle, pinnate, right-pinnate, parallel, grill and tree-crown patterns.

V. G. Bondarchuk (quoted by Morariu et al., 1962) singled out three of the multitude of network types as being the most important, which are generally determined by the magnitude of the confluence angle. Thus, in denticulate systems the tributaries flow into the major collector at an acute angle, in pinnate systems the angle increases to $65° - 90°$; while radial systems are specific to volcanic cones and insular relief.

The first attempt to classify rivers starting from the positions of streams in relation to the main collector was made by Gravelius

(1914). According to this system, the largest river is considered to be of first order from source to mouth. The tributaries flowing directly into it are of second order, all streams flowing into a second-order tributary are of third order, and so on down to the smallest streams. With the passage of time this system has been recognized and employed in most European countries. Analysis of the system indicates that it is unidimensional since, apart from the positions of the tributaries in relation to the main stream, it reveals nothing of the characteristics of rivers of one order or another.

According to the Gravelius system, the rivers Ialomiţa, Mostiştea and Blahniţa, as tributaries of the Danube, are of the same order, although they differ in size, basin area, water discharge, suspended load, etc., and it is obvious that they cannot be compared, although they are of the same order. Applying this system down to the smallest streams, the latter will appear to be of various orders by virtue of their positions in relation to the main stream, although they may be highly similar in many respects and can be viewed as the basic cells of any river system. Given all this, the system mentioned does not facilitate genetic hierarchization of river networks, and remains only a classification of water courses in terms of their position with respect to the main stream (Fig. 6).

In 1945, the American hydrotechnical engineer R. E. Horton inverted this classification system, attributing the first order to the smallest fingertip tributaries appearing as an outcome of concentrated runoff. According to this system, a second-order stream is one which receives at least one or several tributaries of the first order and only of this order. When a second-order stream joins another of the same order, a third-order stream results, and so on. It should be noted that using this system, a water course preserves its order and its name from source to mouth (Fig. 6). In order to identify which is the parent stream and which the tributary at a confluence undisturbed by tectonic movement, Horton recommended the observance of two rules. The first rule states that the tributary is the stream which makes the greatest angle with the headward direction of flow downstream from the confluence. To establish this stream, the direction of the parent river course downstream is extrapolated upstream and the angles made by the two branches with the extrapolation are measured. In this case, the stream of lower order, that is, the tributary, will be the one which makes the greater angle. The second rule states that in cases in which the two angles determined are equal, the tributary is the stream of smaller length. This rule is also applicable in cases in which streams have developed under the influence oı tectonic movements which have disturbed the normal development of the river network. With the elaboration of Horton's classification system, the basis was laid for a method to study comparatively basins of the same order, and hence of the same size.

Three years later, the Soviet researcher B. Panov (quoted by Morariu et al., 1962) again attributed the first order to the smallest fingertip tributaries. Two first-order streams give a second-order stream two second-order streams give one of the third order, and so on up to the main stream, which will be of the highest order (Fig. 6). This system differs from the one described above in that the order of the

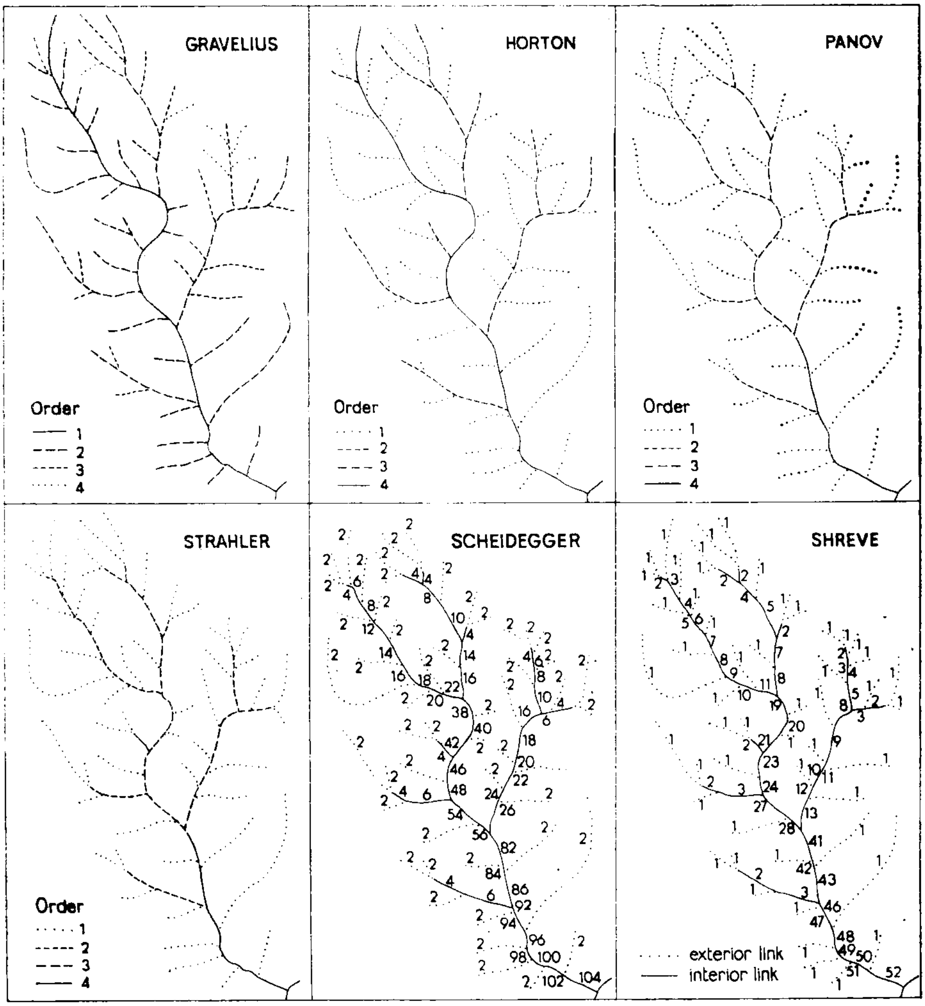



Fig. 6. Various classification systems for river networks.

main stream is not maintained from the source to the mouth, but changes at the confluence of two streams of lower order. Hence the main stream passes successively through all orders, to the extent to which it is augmented due to the confluence of streams of lower or identical order. Panov also found a relationship for determining the class *cl* to which a tributary belongs, on the basis of regional parameters

$a$ and $b$ which determine runoff (climate, relief, vegetation, slope, etc.), and the basin area $A$ :

$$cl = b + a \log A \tag{23}$$

The task of classification was continued by Strahler (1952b), Leopold and Miller (1956), Schumm (1956), Rzhanitsyn (1960) and Hirsch (1962), among others. In their classifications of river networks, all these researchers started from the simplest and smallest elements and progressed to the largest and most complex, taking as the basic element the smallest fingertip tributary. Thus, a river segment of second order appears as a result of the junction of two first-order streams. The new course has characteristics entirely different from those of the two branches forming it (Fig. 6). A third-order river appears through the junction of two second-order stream segments. This third-order river may also receive tributaries of lower orders without changing its own order; for the qualitative change to a higher order it must unite with another stream segment of the same order, and so on.

Analysis of the above classification systems shows that in a drainage basin the number of first-order streams according to the Panov—Strahler system is equal to the total number of rivers in the Horton system. Studying river networks on the basis of these classification systems, it appears that rivers which are of the same order are generally very similar in terms of drainage-basin area, mean length of channel network, mean slopes, water discharge, etc., provided that they have developed under similar physiographical conditions.

The Horton—Strahler classification system has two drawbacks. First, the order intervals are not known, and second, these intervals are such that, for instance, two basins classified as fourth-order may differ greatly as regards their characteristics for the reason that they contain different numbers of streams of lower order, and hence vary in size. In this situation, the analysis of basins of small area is not an easy task, since it is difficult to choose basins of the same size for comparison.

Scheidegger (1965) showed that in evaluating the order of an entire basin, the Horton—Strahler classification system ignores tributaries of order lower than that of the river into which they flow. If, for instance, a segment of fourth order unites with one of third order, the segment resulting downstream from the confluence is also of fourth order, which means that the classification of the new segment throws no light on the contribution made by the lower-order tributary. Given this drawback, Scheidegger proposed another classification system which takes into account all tributaries of lower order (Fig. 6). Thus, the second order is attributed to exterior (terminal) segments, while the size of interior segments is obtained by summation. Consequently, when a 16th-order segment unites with one of 22nd order, the resulting segment will be of 38th order, the order of each segment depending on the contributions of the previous segments (Fig. 6).

Shreve (1966) analyzed the then existing classifications and remarked that both in Horton's system and in the improved version due to Strahler, all basins, with the exception of that of the highest order, are considered complete because they flow into streams of higher order. The main stream in turn may be complete or incomplete, although in most instances it is considered complete. Based on these considerations, Shreve proposed a new system which takes into account all tributaries and provides a realistic picture of the size of a basin in terms of the number of stream segments it contains. He showed that in a channel network there are two types of segment : exterior segments, which end with a spring and which, as in the Strahler system, are considered to be of first order, and interior segments, which at their upstream end are connected to another two segments. In such a network, the size of a stream segment can be defined on the basis of the following two rules : (1) each exterior link is of first order (Fig. 6) ; (2) if two interior links of orders $n_1$ and $n_2$ unite, then the resulting link will be of order $n_1 + n_2$ (Fig. 6). In analyzing this system, it should be noted that each exterior link determines an interior one, and in this case the size of the respective basin is given by the number of first-order streams in the Strahler system. Hence, knowing the number of first-order streams it is possible to obtain the size of the respective basin. Given these conditions, the probability of accidentally omitting an exterior or interior link from a network is 1/2.

However, comparative studies seem to be hindered if the above approach is adopted, since it is difficult to find basins of the same size which can be compared. Being the best verified system, the Horton—Panov—Strahler classification will serve as the basic system in the present work, since, in spite of its drawbacks, it is a good means of evaluating the order of a given stream and offers the possibility of comparative studies with statistical processing of data by value classes. Besides, knowing the number of first-order streams, transition is easily made to the classification systems developed by Shreve and Scheidegger.

## Number of Stream Segments

Morphometrical analysis of any river network demands first of all the adoption of a classification system. Then, each stream segment and drainage basin may be assigned an order according to the principles of the system and to the extent to which the network has developed.

In the present work, the classification system developed by Horton (1945) and complemented by B. Panov (1948) (quoted by Morariu et al., 1962) and Strahler (1952 b) has been adopted because, having a genetic basis, it allows comparative analysis of drainage basins. It has been verified for very many cases involving a wide range of physiographical conditions that, using this classification system, rivers belonging to the same order and located in similar physiographical conditions have approximately the same catchment areas, water discharges, mean slopes, stream lengths, etc. In this system, the first order is assigned to the smallest elongated depressions which have the capacity to organize runoff, have an elementary channel and receive no other tributary, while the whole amount of water is drained by a single main stream that is of the highest order (Fig. 6).

Since the most difficult problem is to establish and compute the number and lengths of first-order streams, Bauer (1980) recommended use of the following methods to this end :

*The blue-line method.* This consists in setting up a hierarchy and considering only the streams represented on maps. Of course, in this case the first-order network will be omitted even on 1 : 25 000 maps. Then, the drawing of blue lines on these maps depends on the time of the year when the aerial photographs on which map drawing is based were taken : if they were taken in the rainy seasons, more streams will appear on the map ; the contrary will happen for dry periods. Furthermore, hilly, plateau and piedmont regions, which generally comprise very permeable rocks, will have a less dense network, since certain streams will frequently become dry even in basins hundreds or even thousands of square kilometers in area, whereas in mountainous regions, where the rock is impermeable and the moisture level is higher, even streams with small basins never become dry and will appear on maps. Since one purpose of studying the laws described later is to allow evaluation of the capacity of a drainage basin to react to rainfall of medium or low frequency, it is necessary to consider the entire network of channels capable of orienting and favouring maximum runoff.

*The contour crenulation method.* This method, proposed by Strahler (1953), is based on the interpretation of topographical maps, the insertion of "streams" being indicated by the configuration of contour lines. In this case, Bauer (1980) considered that the presence of a first-order stream can be assumed only when a contour crenulation forms an angle no greater than 120° and appears in at least two consecutive contour lines. This detail is important as a rigid criterion for establishing first-order streams, which are usually evaluated fairly subjectively.

*The aerial-photograph interpretation method.* This consists in deciphering aerial photographs by adequate means, but it also requires topographical maps for comparison and verification.

*The field-inspection method.* This method is adequate but time-consuming, because the areas to be investigated are quite large, and is recommended only for small basins where details are to be studied.

Generally, such studies resort to more than one method, depending on the goal pursued and the material available. The basic method is nevertheless the contour crenulation method which, for greater accuracy, may be complemented by aerial-photograph interpretation or by field inspection.

The Horton—Strahler classification system and the morphometrical relationships deriving from it have been checked for the Ialomiţa basin situated on the southern side of the Southern Carpathians in Romania (Fig. 7). The data were extracted from 1 : 25 000 maps, and values were obtained for 550 representative points in basins of various orders (from the first to the eighth order in the classification system adopted). The basin investigated is approximately 10 000 km$^2$ in area and has a highly diversified relief, extending from an altitude of 10 m at the confluence with the Danube to 2505 m at the Omu peak in the Bucegi mountains (Fig. 8). The mountain relief consists of crystalline rock, Jurasic limestone and Cretaceous grit—conglomeratic flysch, and constitutes 14.6% of the total area. The Subcarpathian region, comprising Paleogene and Neogene rock, with active tectonics reflected by high fragmentation and very active modelling processes, constitutes 22.8% of the area. The flatland relief formed on Quaternary deposits, with a small relief energy and poor fragmentation, represents 62.6% of the total drainage-basin area (Fig. 8).

*Law of stream numbers*. Once an entire network has been classified, the stream segments are counted by order, to facilitate analysis. This yields a series of data with decreasing values corresponding to increasing order. By plotting such values on semilogarithmic paper, Horton (1945, p. 291) established the following law : "*the numbers of streams of different orders in a given drainage basin tend closely to approximate an inverse geometric series in which the first term is unity and the ratio is the bifurcation ratio*". Using the properties of geometric progressions, Horton then gave an expression for the number $N_o$ of streams of a given order $o$ knowing the bifurcation ratio $r_b$ and the order $u$ of the main stream :

$$N_o = r_b^{u-o} \tag{24}$$

The total number $N$ of streams in a given basin can thus be determined using the formula

$$N = r_b^u - 1/(r_b - 1) \tag{25}$$

This relationship has been verified in numerous instances and its validity proved in various physiographical conditions (e.g. Strahler, 1952b ; Maxwell, 1955 ; Schumm, 1956).

In order to determine which of the properties of the law of stream numbers are due to geomorphological factors, both accidental and natural networks have been studied (Leopold and Langbein, 1962). In accidental networks, the lines supposed to be streams may flow without restriction in any direction (they may thus intersect or coil), while in natural networks limits are imposed by the force of gravity, which makes water run only along the lines of steepest slope. This is a general theoretical law which may be applied to any ramified system.

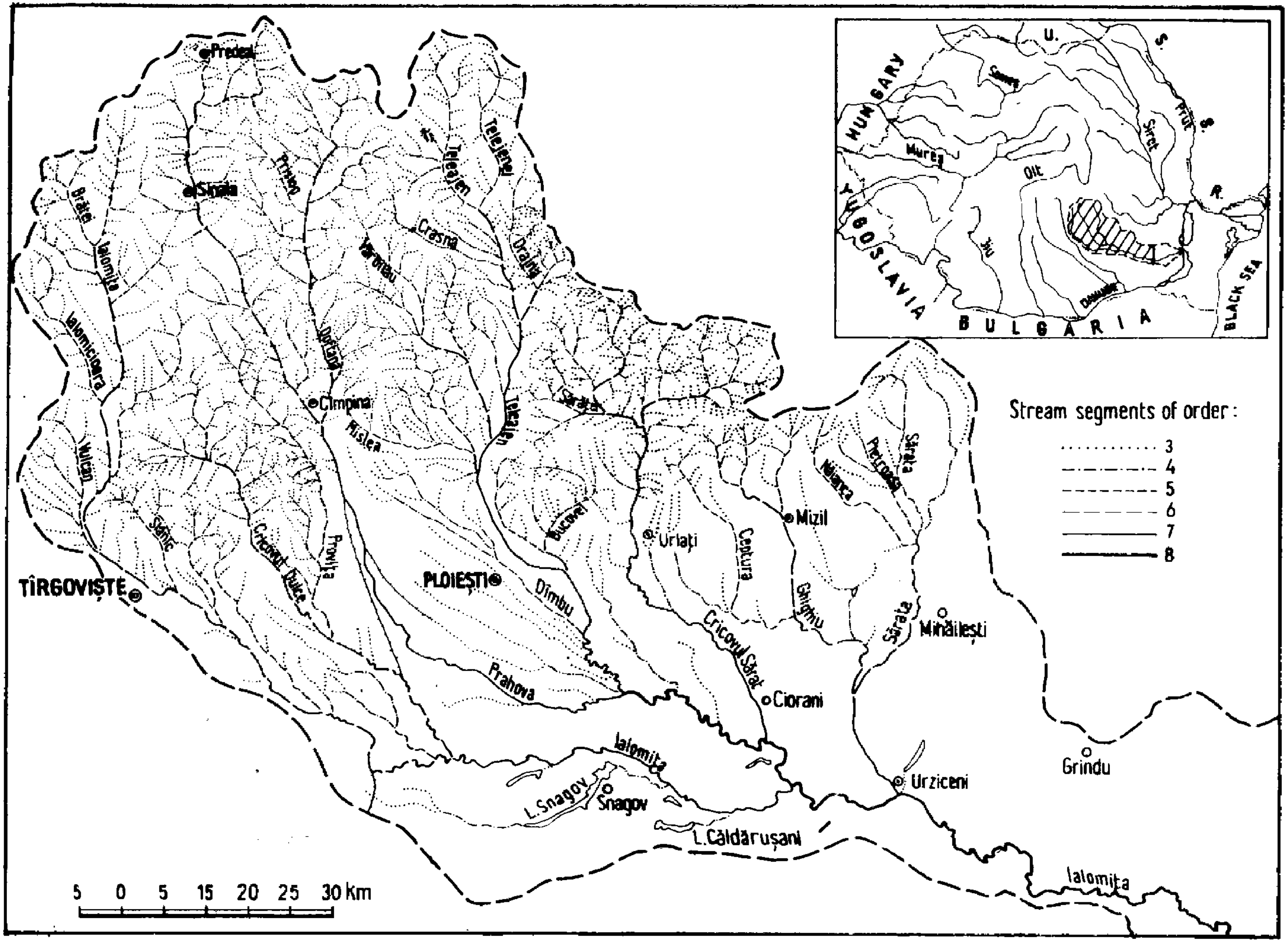



Fig. 7. Ordering of river network in the Ialomița drainage basin according to the Horton–Strahler clasification system

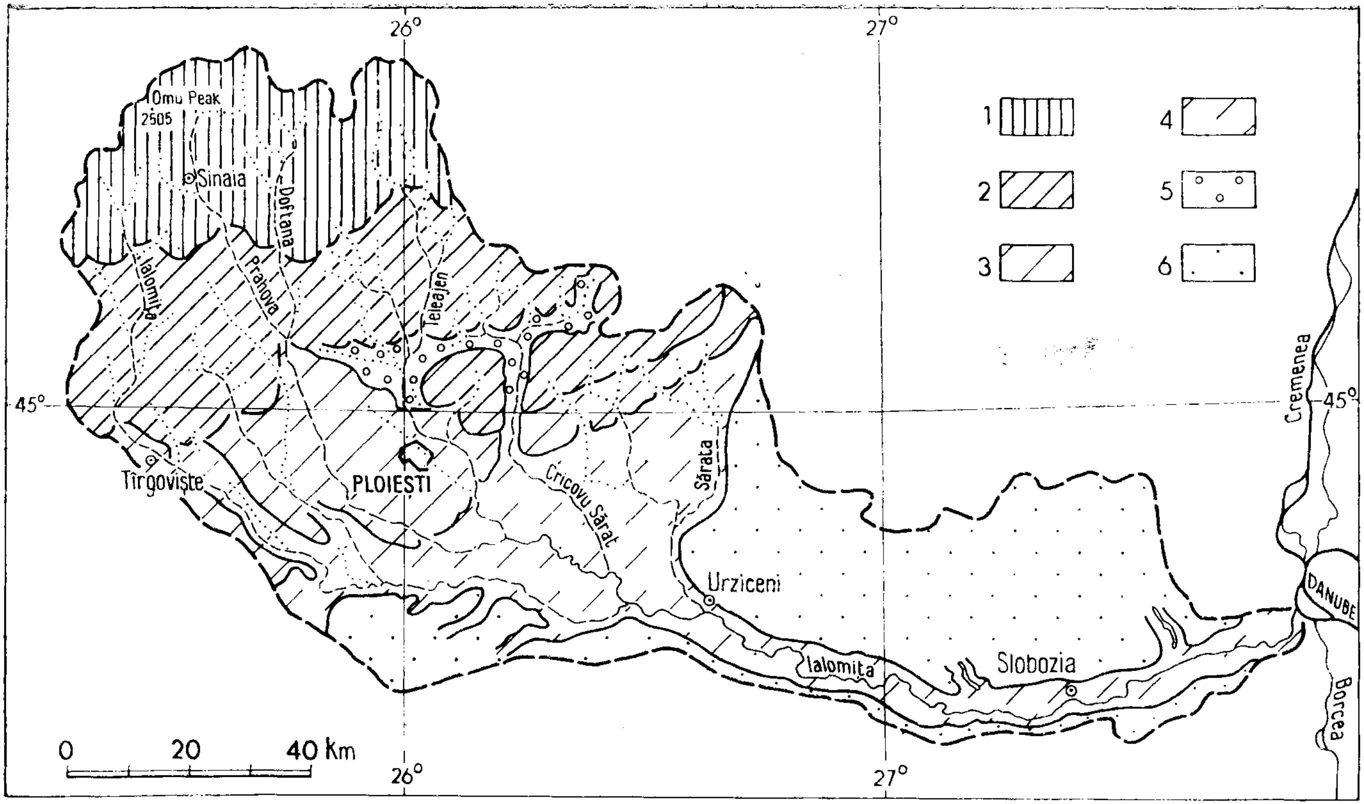



Fig. 8. Sketch of relief units in the Ialomța basin: 1, massifs and mountain summits of the Carpathian Orogene; 2, hills developed on strongly folded Neogene sedimentary formations with marked tectonic mobility; 3, piedmont plain of Quaternary accumulation; 4, divagation (subsidence) Holociene alluvial plain with flood-plain appearance; 5, depressions or subsidence areas with proluvio—alluvial—plain appearance; 6, nonfragmented tabular plain covered with loess-like deposits (according to the *Atlas of the Socialist Republic of Romania*, Sketch III-1).

The most difficult problem is to determine correctly the number of streams of first order, especially in large basins, for which the congruity of cartographic documents with the terrain must be ascertained. Field research and comparisons of large-scale maps with the terrain and aerial photographs have shown that hierarchization may instead be started either from the second order onwards or using the inverse order method (Horton, 1945 ; Hirsch, 1962).

After determining the number of streams of each order, the *confluence ratio* * $R_c$ may be established in various ways (Haggett and Chorley, 1969) by calculating :

individual values based on the numbers of stream segments of successive pairs of orders :

$$R_{c1} = N_1/N_2,\ R_{c2} = N_2/N_3,\ R_{c3} = N_3/N_4,\ \text{etc.}$$

the arithmetic mean of the $n_i$ values of the confluence ratio obtained for the whole river network ;

the weighted mean, obtained by weighting the numbers of stream segments of each order considered in the computation (Strahler, 1952b) : this method yields good results and has been adopted in most cases ;

the geometric mean, represented by the slope of the line drawn through the experimental points ;

the antilogarithm of the best-fit regression coefficient of the straight line drawn on the basis of the experimental points.

Horton (1945) showed that the bifurcation ratio has low values in flat regions and high values in intensely fragmented areas. It is therefore subordinate to geomorphological factors, which is fully expected taking into account that the present configuration and morphometrical features of a river network reflect the effects of climatic change, orogenic movements, stratigraphic conditions and erosion over a period extending from the geological past.

The representation on semilogarithmic paper of the number of stream segments in relation to order always gives a certain amount of scatter around the regression line, since a probability function is involved which diminishes with a decrease in the number of values on which it is based (Milton, 1966). Therefore, since it is known that the number of segments decreases in proportion to order, the higher the latter, the greater the scatter of the points. The existence of an experimental relationship between order and number of segments represents a probability optimum to which a drainage system tends under given conditions (Leopold and Langbein, 1962).

In the series of numbers of stream segments, the first term represents the number of first-order stream segments and the last, the number

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* We prefer the term confluence ratio (Hirsch, 1962) to the bifurcation ratio (Horton, 1945), because we are studying river networks from small to large segments, and not in decreasing order. See also the entry for 'bifurcation ratio' in *A Dictionary of Geography* (F. J. Monkhouse, London, 1972) : "the term is somewhat of a misnomer, since b. would imply division, not uniting".

of stream segments of the highest order. Horton considered this latter term as being equal to unity, and hence the stream to be complete, as remarked by Shreve : it is thus no longer possible to take it into account in subsequent computations.

More detailed analysis shows that in the above formulation, the law of stream numbers, although verified in many instances, presents certain drawbacks. Take for instance the example of the fifth-order Crane Creek basin described by Horton (1945, p. 288). Note that the last two orders have a single segment each. Using the bifurcation ratio $r_b = 2.22$ determined by Horton and starting compulsorily from the highest order (the fifth), a straight line is obtained which, although dividing the points equally, cannot be considered to have the most correct position (Fig. 9A(a)). In another very detailed morphometrical study dealing with the drainage network on the Brazilian plateau the number of streams of successive order were determined for several basins (Christofoletti, 1970). Taking as an example the fourth-order Vargens basin, Christofoletti determined the bifurcation ratio using the weighted mean. If the law of stream numbers, according to which the last term of the progression must be equal to unity, is used in the case of the basin mentioned, starting from unity a straight line is obtained which has nothing in common with the value of the lower orders (Fig. 9B(a)).

Applying the law of stream numbers to several sections of the Ialomiţa river, from upstream to downstream, streams of lower orders accumulate to the extent to which further confluences are added (Fig. 10A, B, C). Thus, after the junction of the fifth-order rivers Ialomiţa and Brătei, a qualitative leap occurs to a sixth-order river and this class is maintained up to the confluence with the Cricovu Dulce, which is also of sixth order (Fig. 10C). In between the two extreme points, the Ialomiţa basin increases substantially in terms of area, length, water discharge, number of streams of lower orders, etc. Depending on local conditions, the transition to a higher order may occur before the accumulation of all streams of lower orders, as is the case for the Ialomiţa upstream from its confluence with the Brătei (Fig. 10B). Generally, in elongated basins, the accumulation of lower-order streams may proceed so far as to exceed the expected numbers of streams for these orders (Fig. 10C).

If the law is accepted as valid, and it has been verified in very many cases, the implication is that for any drainage basin the numbers of streams of successive orders tend to a geometric progression. Under certain physiographical conditions this may be achieved, but it represents only cases of dynamic equilibrium, i.e., when the drainage system is adapted perfectly to a certain state of the transfers of mass and energy within it. However, should a certain imbalance occur in the state of a variable or in the flow of matter, the drainage network will react so as to adjust to the new state. For instance, extensive deforestation in a drainage basin will enhance the intensity of surface runoff and favour the appearance of new phenomena under torrential rainfall, which may result in an increase of the order of the given basin.

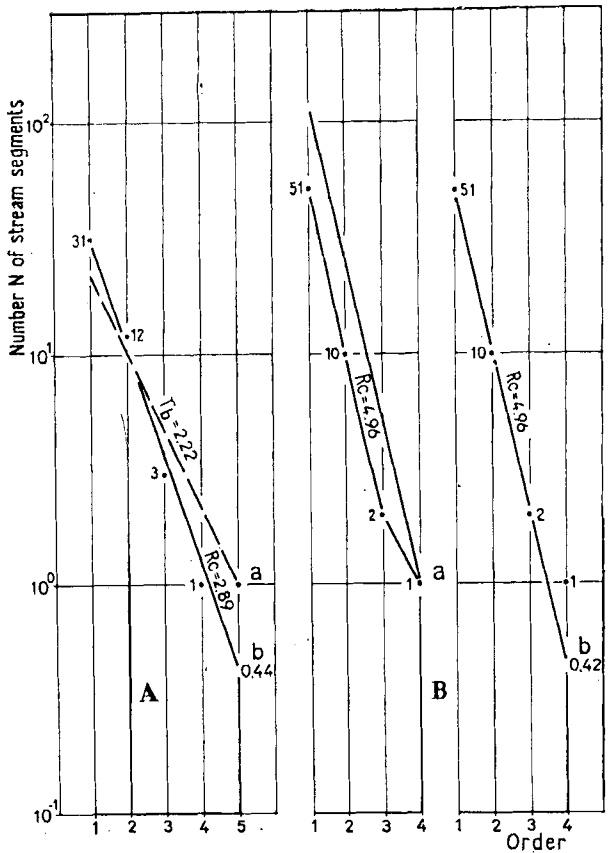



Fig. 9. Regression of numbers of stream segments on order for (A) Crane Creek in Wedsport, NY, and (B) Vargens on the Brazilian plateau (according to Horton, 1945; Cristofoletti, 1970) (a, with assumption that basin is complete; b, without this assumption).

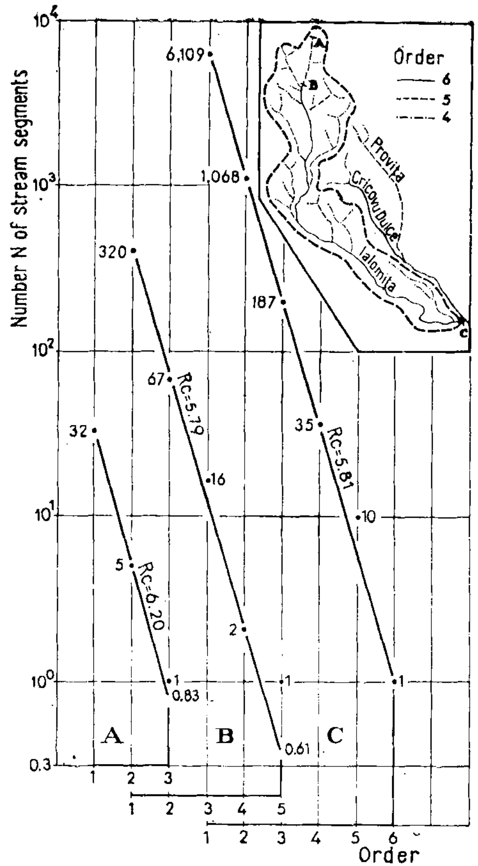



Fig. 10. Regression of numbers of stream segments on order for the drainage basins of the Ialomiţa (A) upstream from its confluence with the Sugările stream (B) upstream from its confluence with the Brătei stream, and (C) upstream from its confluence with the Cricovu Dulce.

For full concordance to exist between the relationship established by Horton and the above considerations, the law of stream numbers can be worded as follows : *the numbers of stream segments of successively higher orders in a given drainage basin tend to form a decreasing geometric progression in which the first term $N_1$ is the number of first-order streams and the ratio is the confluence ratio $R_c$*. The points determining the position of the straight line are then analyzed in relation to the proportions of streams of various orders. It is obvious that the greatest proportion is constituted by the lower-order streams, and the smallest by the higher-order ones. The confluence ratio can be determined using one of the above mentioned methods and the values corresponding to each order, starting from the points with the largest proportions and thus the smallest deviations from the law. The equation of the straight line which best delimits the points may be also determined by means of the chosen-point method (Bloh, 1971).

It is obvious that if the above procedure is followed, a value either above or below unity may be obtained for the stream of highest order. In this case, the value obtained indicates the extent to which the segment concerned is concordant with the order assigned to it as concerns the numbers of lower-order streams. Applying this method to the Crane Creek basin, a confluence ratio of 2.89 is determined from the weighted mean, and a value of 0.44 is found for the fifth-order basin. This means that under the given physiographical conditions, and with the given confluence ratio, the fifth-order stream accumulates only 44 % of the lower-order streams necessary for its completion (Fig. 9A(b)).

In the case of the Vargens basin, using the confluence ratio determined by Christofoletti but starting from the first-order streams, a straight line is obtained which delimits the points well. The value of the fourth-order term indicates that this basin is only 42 % developed in relation to the given physiographical conditions (Fig. 9B(b)). If the degree of ramification of the drainage network increases while the confluence ratio retains a constant value, the system will become complete when the fourth-order term is equal to unity, corresponding to a much larger number of lower-order streams. In this way, an answer is provided to the above criticisms of the Horton—Strahler system, namely, that its class intervals are not known and that two basins may be considered of the same order although they have differing characteristics.

Starting from the first term, which can be obtained by counting or computation, note that the number of streams of a given order $x$ is equal to the ratio between the number of streams of the immediately lower order and the confluence ratio $R_c$:

$$N_x = N_{x-1}/R_c \tag{26}$$

Following this reasoning, a value is finally obtained for the term of highest order which represents the last term of the geometric progression and at the same time provides an indication of the extent to which the main stream is developed as concerns the number of lower-order streams accumulated.

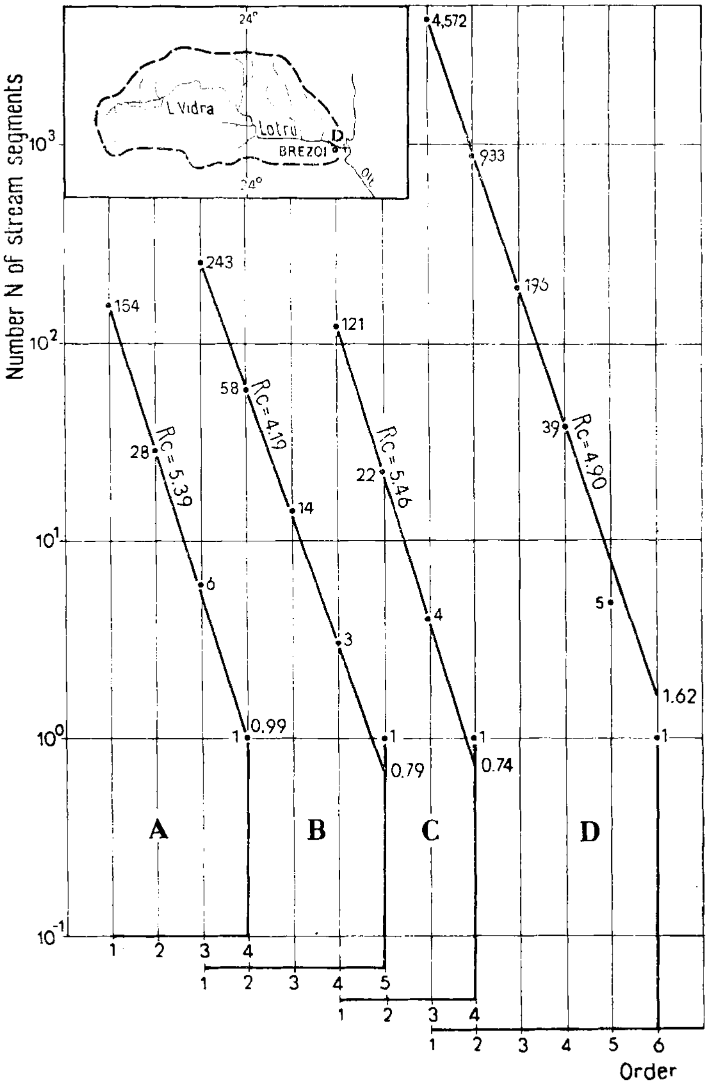



Fig. 11. Regression of numbers of stream segments on order for the drainage basins of (A) the Sultanu at its confluence with the Cociani, (B) the Brătei upstream from its confluence with the Ialomiţa, (C) the Raciu upstream from its confluence with the Ialomiţa, and (D) the Lotru upstream from its confluence with the Olt.

For instance, in analyzing the basin of the Sultanu stream, a tributary of the Cricovu Dulce (Fig. 11A), the calculated value of the last term is approximately unity, and therefore for the calculated confluence ratio sufficient streams of lower orders are accumulated to consider this fourth-order basin to be completed. The relevant data are $N_1 = 154$, $N_2 = 28$, $N_3 = 6$, $N_4 = 1$. Thus

$$R_{c1} = 5.50,\ R_{c2} = 4.67, = R_{c3} = 6.00$$

$$N_1 + N_2 = 182, N_2 + N_3 = 34, N_3 + N_4 = 7$$

$$R_{c1}(N_1 + N_2) = 1001,\ R_{c2}(N_2 + N_3) = 159,\ R_{c3}(N_3 + N_4) = 42$$

$$\Sigma(N_i + N_{i+1}) = 223$$

$$\Sigma R_{ci}(N_i + N_{i+1}) = 1202$$

and therefore

$$R_c = 1202/223 = 5.39$$

Knowing $N_1 = 154$ and $R_c = 5.39$ it is possible to determine the numbers of streams of each order, and hence the value of the term of highest order, using the formula

$$N_u = N_1/R_c^{u-1} \tag{27}$$

or, in the present case,

$$N_4 = N_1/R_c^3 = 154/5.39^3 = 0.99$$

The value obtained for the fourth-order term shows that it is 99% complete in terms of accumulation of lower-order streams: the value may be considered equal to unity. Unfortunately, such cases are fairly rare in nature.

To facilitate the computation, it is also possible to consider the progression as being increasing; then, a given term $N_x$ is determined using the formula

$$N_x = N_u\ R_c^{u-x} \tag{28}$$

where $u$ is the order of the term for the main stream.

In the case of the Brătei upstream from its confluence with the Ialomiţa (Fig. 11B), the last term of the progression has a value of 0.79, which shows that under the physiographical conditions of this basin and for the existing confluence ratio, the number of streams of lower orders should be much larger for the basin to be considered completed for the order assigned to it. In contrast, in the case of the elongated Lotru basin upstream from the confluence with the Olt, lower-order streams have gradually accumulated in a larger number than necessary for a sixth-order basin with a confluence ratio of 4.90 (Fig. 11D).

*Total number of stream segments.* Knowing the numbers of stream segments of each order in a given drainage basin, the total number of segments may be determined. This can be done either directly, knowing that

$$N = N_1 + N_2 + N_3 + \ldots + N_u$$

or using the properties of geometric progressions. The sum of the terms of a decreasing geometric progression is determined according to the relationship

$$N = N_u(1 - R_c^u)/(1 - R_c) \tag{29}$$

The relationship given by Horton (1945, p. 286) for the sum of the numbers of streams can be applied only in the case of an increasing geometric progression in which the first term is equal to unity.

The formula for the sum (eqn.(29)) may be verified in the case of the fourth-order basin of the Sultanu upstream from its confluence with the Coceani. Equation (29) yields

$$N = 0.99\ (1 - 5.39^4)/(1 - 5.39) = 190$$

which compares with the 189 segments obtained by direct counting.

The law of stream numbers has been proved valid for a large number of basins in mountainous, hilly and plain regions, despite a wide diversity of rock type, tectonics and structure. When drawing the straight line in semilogarithmic coordinates account should first be taken of the lower-order streams, which are the most numerous and best indicate the position of the straight line. The results for a large number of the cases analyzed demonstrate the absolute necessity of computing the value for the stream of highest order and not considering it to be always equal to unity.

If, for instance, in the case of the fifth-order basin of the Ialomiţa at its confluence with the Brătei (Fig. 10B), or the sixth-order basin of the Teleajen at its confluence with the Drajna (Fig. 12B), the computations were started from a value of unity for the highest order, the straight lines obtained would provide no indication of the actual situations in these basins. If the ratio $R_c$ of the progression were determined on the basis of the weighted mean starting from unity for the highest order, the straight line obtained would be parallel to the real one and would reflect values much higher than those existing in reality for the lower orders. Starting again from unity, but using the average for the other points, once more quite erroneous values would be obtained in comparison with the real situation.

We therefore consider that the best solution for drawing the straight line is to take into account only the streams of lower orders (Figs. 10A, B and 11B, D), and not that of highest order, the value for

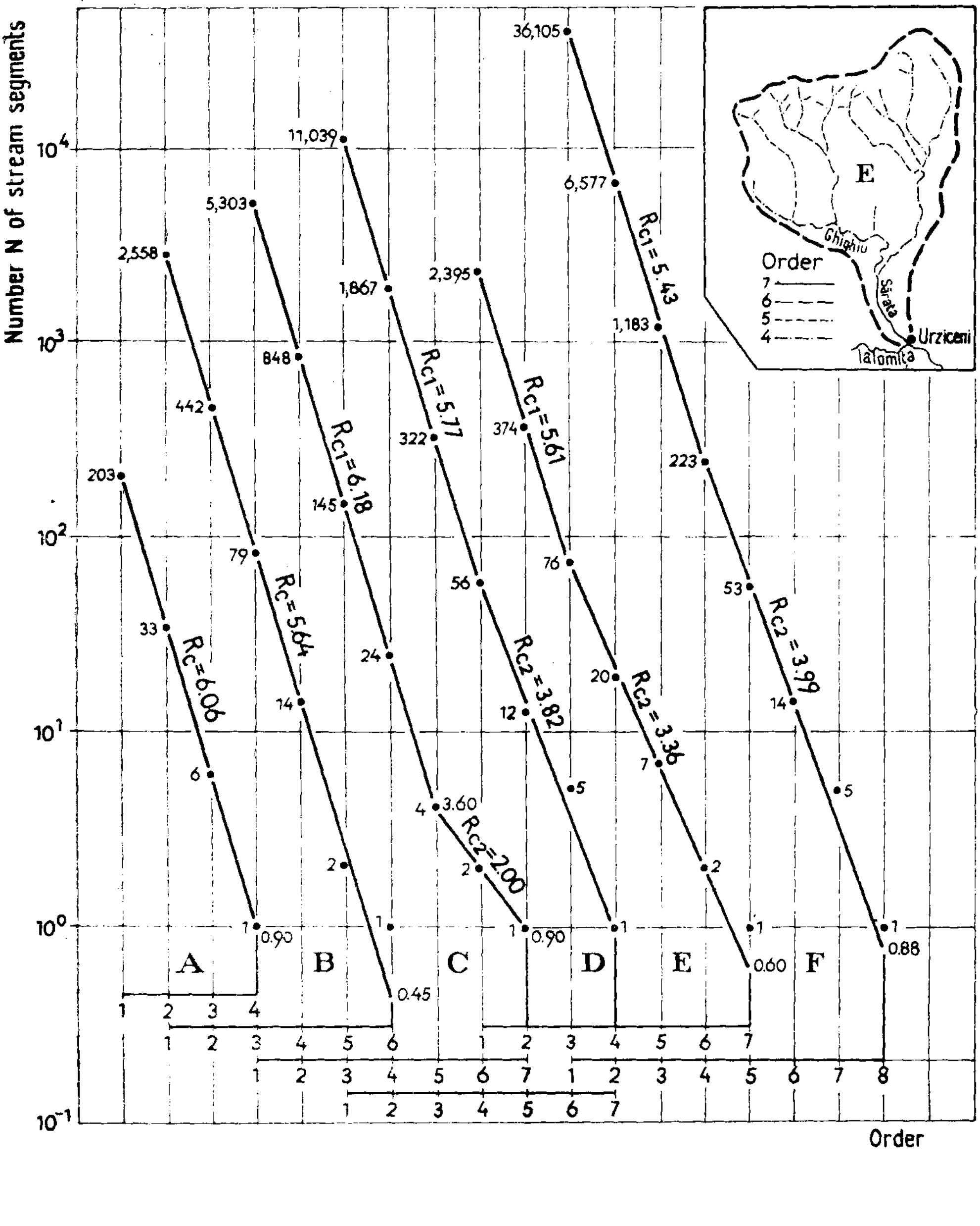



Fig. 12. Regression of numbers of stream segments on order for the drainage basins of the Teleajen (A) upstream from its confluence with the Gropşoarele, (B) upstream from its confluence with the Drajna, (C) at its confluence with the Vărbilău, and (D) at its confluence with the Prahova ; (E) for the drainage basin of the Sărata at its confluence with the Ialomiţa ; and (F) for the drainage basin of the Ialomiţa at its confluence with the Danube.

which has the lowest statistical significance. The latter should be determined only after the straight line for all the other streams in the drainage basin concerned has been well established, since it depends on the

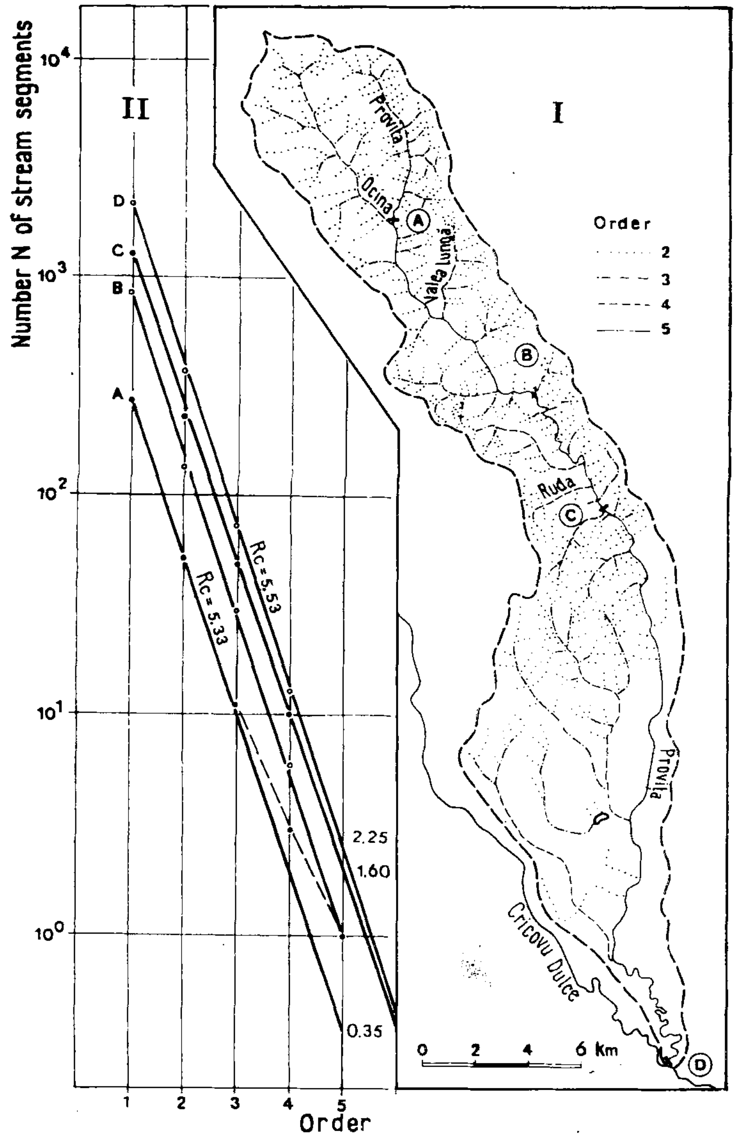



Fig. 13. The Provița drainage basin : (I) hierarchy of river network ; (II) regression of number of stream segments on order for various sections.

numbers of streams of lower orders accumulated in relation to the specific conditions within the basin : i.e., its rock type, structure, tectonics, shape, etc.

A basin carved in friable rock will have a high confluence ratio, which means that to be considered complete for a certain order it will need more streams of lower orders in comparison with a basin of the same size in a plain area, which will have a much lower confluence ratio.

Basin shape is also important in the hierarchization of river networks and in the law of stream numbers. As seen above, a stream cannot be considered to have the same order from spring to mouth, irrespective of its number of tributaries. Consider the Provița stream, which has a very long basin and a low form factor ($RF = 0.329$). Developing in the Subcarpathian area, characterized by friable rock and a high degree of relief fragmentation, this stream passes quite quickly from the fourth to the fifth order already in its upper basin (Fig. 13(I)). The law of stream numbers for the basin area accumulated down to the confluence with the Ocina stream indicates that the fifth-order stream is only 35% completed. Then, down to the confluence with the Cricovu Dulce, there is no other stream of fifth order which would allow transition to a higher order. The elongated form of the basin allows only the accumulation of lower-order streams. In this situation, the numbers of streams of the first to fourth orders increase sufficiently to satisfy the completion of a fifth-order stream (Fig. 13(II)B). The shape of the basin does not change downstream, and so the accumulation of lower-order streams continues until upstream from the confluence with the Murele valley, 29% of the streams necessary for a completed sixth-order basin have been accumulated (Fig. 13(II)C). At the confluence with the Cricovu Dulce there are sufficient streams of lower orders for 2.25 fifth-order basins, or 41% of the segments necessary for a sixth-order basin, although this transition has not occurred. In fact, this is one of the few cases in which the curve has downward concavity at its lower end. In most instances, this concavity is upward (Schumm, 1956), which proves that streams generally join prior to accumulation of the number of segments necessary for the completion of the order of the largest segment. Therefore, the transition to a higher order may occur at quite different values of completion, depending on specific local conditions.

From the above examples it may be seen that the straight line determining the number of streams in relation to their order is well established for most drainage basins. However, the latter are very heterogeneous as concerns their geological and geomorphological features. Most of the larger rivers in Romania develop in at least two major relief units, each with its specific conditions. The passage from one relief unit to another, and even from one type of rock to another, will alter the degree of ramification of a river network and, with it, the confluence ratio which shows how many lower-order streams are necessary to complete a stream of a higher order under the given conditions. In the case of the Ialomița basin, for instance, the passage from the Subcarpathian region into the piedmont area and subsidence plain brings about a substantial decrease of the confluence ratio for the higher-order streams in the lower parts of their curves (Fig. 12F).

This happens because numerous lower-order segments accumulate in the highly fragmented Subcarpathian and mountainous regions in which the upper and middle basins of these streams develop, resulting in higher values of the confluence ratio as compared with the piedmont area and plain proper, where the number of such segments is reduced.

A very good example of change in the number of streams on passage from the Subcarpathian to the plain region is provided by the Sărata basin (Fig. 12E). The channels of streams forming this basin strongly fragment the southern side of the Istriţa hills and then drain the plain area of Mizil. In this case, the stream segments of lower order are very numerous on the hill slopes. The passage to the plain area corresponds to organization of the network into streams of higher orders, which are now very reduced in number. This may be related to the decrease of the relief energy, which can no longer lead to marked fragmentation. In the hilly area of this basin, the confluence ratio reaches an average value of 5.61, which drops to 3.36 in the plain area. In this situation, the straight line for the higher orders is established more correctly using the second confluence ratio. When determining the total number of rivers in a basin according to eqn. (29), the procedure when there are two straight lines is to calculate

$$\sum_{i=1}^{u} N_i = \sum_{i=1}^{k} N_i + \sum_{i=k+1}^{u} N_i = N_k(1 - R_{c1}^{k})/(1 - R_{c1}) +$$

$$+ N_u\,(1 - R_{c2}^{u-k})/(1 - R_{c2}) \tag{30}$$

In the case of the Ialomiţa at its confluence with the Danube (Fig. 12F), the following result is obtained:

$$\sum_{i=1}^{8} N_i = \sum_{i=1}^{4} N_i + \sum_{i=5}^{8} N_i = N_4(1 - R_{c1}^{4})/(1 - R_{c1}) + N_8(1 - R_{c2}^{4})/(1 -$$

$$- R_{c2}) = 223(1 - 5.43^4)/(1 - 5.43) + 0.88(1-3.99^4)/(1-3.99) = 43\,787$$

*Assessment of order.* One of the major objections brought against the Horton–Strahler classification system is that no values can be determined for the intervals between two orders, and hence no quantitative assessments can be made. This problem can be solved if, as shown by Morisawa (1962), the law of stream numbers is considered as being a relationship of the form

$$\log N = B + Cx \tag{31}$$

in which $x$ is the stream-segment order. For instance, considering the fourth-order basin of the Valea Raciu upstream from its confluence with the Ialomiţa, the geometric progression 121 : 22 : 4 : 1 is obtained.

To determine the coefficients $B$ and $C$ in eqn. (31), the method of chosen points may be used. In this case, we take the values corresponding to orders 1 and 3 and form two equations which on solving the system yield the result

$$\log N = 2.82315 - 0.74036\, x$$

Substituting $N = 1$ yields

$$x = 2.82315/0.74036 = 3.81$$

which represents the real order of the stream in relation to the total number of lower-order stream segments accumulated (Fig. 14A). Thus, owing to the accumulation of lower-order stream segments, the real order of a stream may be equal or higher than the order assigned to it (Fig. 14B).

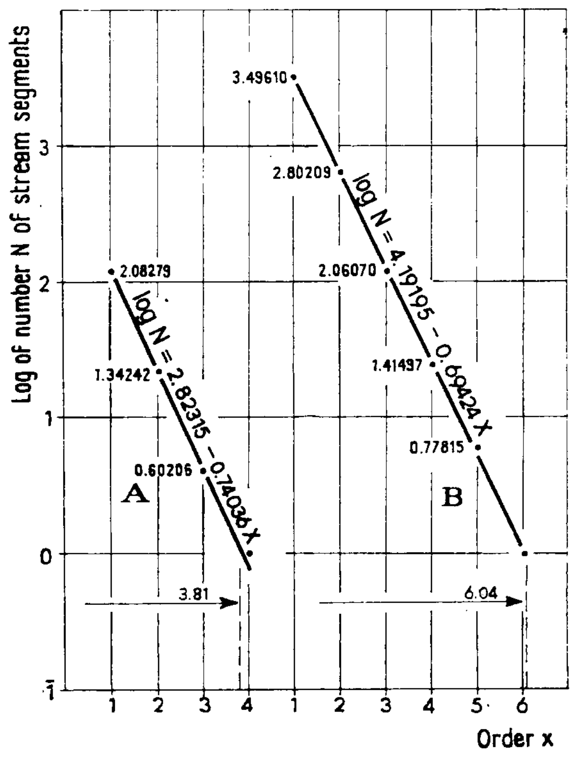



Fig. 14. Regression of numbers of stream segments on order, and equations of exponential form using logarithmic values, for the drainage basins of (A) the Raciu at its confluence with the Ialomiţa, and (B) the Doftana at its confluence with the Prahova.

The same result is obtained starting from eqn. (27) by taking logarithms :

$$\log N_u = \log N_1 - (u - 1) \log R_c$$

from which

$$u = (\log N_1 - \log N_u)/(\log R_c) + 1 \quad (32)$$

Knowing the value of $R_c = 5.46$, for $N_u = 1$ the real order of the main stream is obtained as

$$u = (2.08279/0.73719) + 1 = 3.82$$

It is true that in the classification of streams of any order it is necessary to assign a certain order to the main stream, but its true value is ascertainable only from the geometric progression determined by the data for all streams of lower orders. Otherwise, subsequent computations should use the value determined on the basis of eqn. (27). Therefore, the real order is obtained from eqn.(31) or by calculating the ratio of the logarithm of the number of first-order streams and the logarithm of the confluence ratio, and adding unity.

*Frequency distribution of confluence ratio.* The confluence ratios determined for a large number of basins of various orders bear a certain relation to basin order. Thus, for low-order basins the confluence ratio generally has a high value, while for streams of higher order it decreases. For a large sample of third-order basins, for instance, the calculated confluence ratio ranges from 2.5 to 11.

For a sample of 230 fourth-order basins, the confluence ratio ranges from 2.95 to 10.8, with a mean of 5.74. In this case, the scatter of the values around the mean in relation to the standard deviation $\sigma$ indicates a normal distribution, 60% of the values falling in the interval $\bar{x} \pm 0.6745\ \sigma$, and 75% in the interval $\bar{x} \pm \sigma$, or between 6.94 and 4.54. The coefficient of variation $C_v$ of 0.21 indicates a mean variance of values, and the skewness $S_k$ of 0.94 shows positive asymmetry (Fig. 15A).

The confluence ratios calculated for a sample of 53 fifth-order basins (considered as representative) vary between 2.88 and 6.98, indicating a statistically significant decrease of the average value and thus a shift of the frequency-distribution curve to lower values. The scatter of the values around the mean (4.92) is much smaller ($C_v = 0.17$), and the skewness ($S_k = 0.06$) indicates that the distribution is normal (Fig. 15B). If a normal-distribution curve calculated with the same parameters as those of the sample is superimposed upon the real distribution, only small deviations are observed.

For 14 sixth-order basins, although this number is too small for statistical significance, a mean value of 4.79 was obtained, smaller than in the case of the orders analyzed previously.

On calculating the probabilities of occurrence of various confluence-ratio values for fourth- and fifth-order basins, it is clearly apparent that much smaller confluence ratios correspond to a given probability in the case of the fourth order (Fig. 15A), the difference being more marked for the maximum values. For instance, while for a fourth-order basin a confluence ratio of 9.32 can be reached with a probability of 1 %, for a fifth-order basin a confluence ratio only of 6.94 can be reached for the same probability (Fig. 15B).

The great variability of the confluence ratio, especially for small drainage basins, reflects the existence of a wide diversity of local physio-

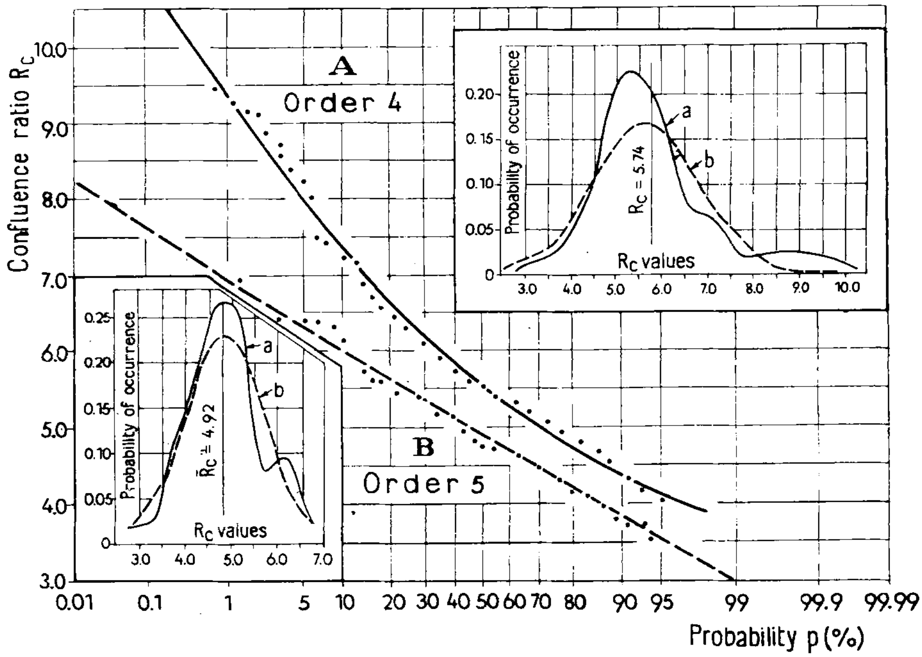



Fig. 15. Probability distributions of the confluence ratio for (A) fourth- and (B) fifth-order basins : a, real frequency distributions ; b, normal-distribution curve calculated with the same parameters as those of the sample.

graphical conditions, including the influence of tectonic movements, of structure and especially of rock type. In basins of markedly elongated shape, there is a great difference between the numbers of lower-order and higher-order streams, which leads to an increase of the confluence ratio. Relatively high values of the latter indicate the existence of basins in the process of evolution, and of a young relief permanently subjected to fragmentation. A steady state of drainage is very hard to reach in highly fragmented areas such as the Carpathian and Subcarpathian regions in Romania, with still active tectonics and a markedly heterogeneous lithology. The fact that most basins of various orders are not fully developed for the order assigned to them indicates that they are in continuous evolution, involving very active erosion and fragmentation processes. The evolution of such drainage basins in terms of numbers of streams of various orders nonetheless tends towards a steady state.

*Relationship between total number of stream segments and number of first-order segments.* Of the total number of stream segments in a drainage basin, the greatest proportion comprises first-order segments. The total number $N$ of segments is closely related to the number $N_1$ of first order. In the case of the Ialomiţa basin, the relationship is

$$N = 1.204\ N_1 \quad \text{or} \quad N_1 = 0.831\ N \tag{33}$$

This relation is valid also for the average values for basins of various orders, and even for basins which have developed under differing tectonic and rock conditions. Therefore, if the number of first-order segments in a given basin is known, the total number of segments can be easily determined.

The fact that the greatest proportion of streams are first order indicates the link between the classification system adopted here and those of Scheidegger (1965) and Shreve (1966), which lay great stress on the first-order segments within a studied drainage basin. This facilitates the transition from the present system to those mentioned, which is required in order to estimate basin size with sufficient accuracy. A direct relationship is found between basin size and the number of first-order stream segments, which is generally the greater, the higher the order of a basin. This means that an increase in the number of such segments should result in an increment in the order of the main stream if the latter is to reflect basin size.

Chapter IV

# Drainage-Basin Area

A major geometrical characteristic of any drainage basin is its area, delimited by the water divide linking the highest points at the boundary with neighbouring basins and extending down to the final section of the main stream. As a rule, any stream has both a surface drainage basin and an underground one, the water divides of which coincide fairly seldom. Owing to the difficulty of delimiting underground drainage basins, only surface drainage basins are here taken into account in analyzing runoff in relation to basin size. In the case of large basins, the resulting differences in discharge are small and do not exceed admissible errors. Greater errors appear in the case of detailed studies of small basins, especially of those which have developed in karst regions with high underground runoff.

The area of a drainage basin is the morphometrical element which determines, in relation to local physiographical conditions, the magnitudes of the exchanges of matter and energy with the surroundings. The excess water (the main form of mobile matter) which a basin receives, and which is not lost through evaporation or physico-chemical and biological processes, moves by surface and subsurface runoff together with the mineral and organic masses circulating through the channel network to the mouth of the main stream. The way in which the basin area is divided for drainage purposes depends on a whole series of factors. Among these are the relief ratio, the nature of the substituent rocks, and the structure of deposits, which impart a certain resistance to infiltration. Then, the basin area may be protected by soil or vegetation, and will be affected by various environmental agents, particularly rainfall. Interaction among all these elements over the course of time is responsible for the present geomorphological aspects of any basin, and in many cases these aspects may still be in the process of evolving.

## Correction of the Map Projection of an Area

The area of a drainage basin delimited on a map by a perimeter corresponding to the drainage divide can be measured using a planimeter, by the grid-square method, or on the basis of coordinates, and the results may be expressed in various units. A value thus calculated represents the horizontal projection of a real topographical area which makes a certain angle with the horizontal plane as a function of the mean gradient of the multitude of planes it comprises. This angle will range from

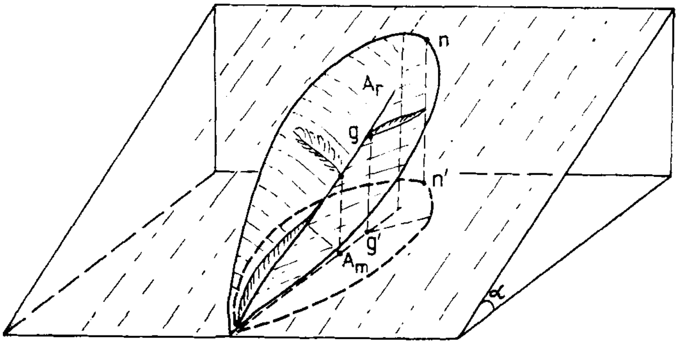



Fig. 16. Real area $A_r$ and its horizontal projection $A_m$ for a drainage basin.

0° for horizontal areas in plain regions, to 90° for vertical walls and steep slopes in mountainous regions. Hence, the greater the gradient, the greater the difference between the horizontal projection and the real area of a valleyside (Fig. 16). For instance, a steep wall with a gradient of 70 – 90° may appear to be very small in a horizontal projection, while in reality its area is much bigger. In such instances, there will be a large difference between the real area and that determined from a map projection. In the case of the basins of large area and small mean slope, the error may be ignored. However, there are cases when greater accuracy is required in establishing hydrological parameters. In detailed studies and especially for experimental basins where precision measurements are effected, ignoring the basin slope will be a source of error. Thus, in the case of mean specific runoff, determined as the ratio between measured discharge and the map projection of an area, values greater than the real ones will be obtained. As shown by Serra (1951), in order to determine the precipitation input with greater accuracy it is necessary to take into account both the orientations of valleysides in relation to the prevailing wind, and the land slope, in order to evaluate the true area in relation to its map projection.

In order to eradicate these shortcomings, when dealing with small areas, and especially with regions of high relief, it is desirable to utilize the real area $A_r$, which may sometimes be considerably larger than its map projection $A_m$. The real area of a drainage basin may be calculated knowing the mean slope $\alpha$, using the formula

$$A_r = A_m / \cos \alpha \tag{34}$$

Thus, a basin in a mountainous area, the Rîul Mare, which collects its water from the Retezat mountains (Southern Carpathians, Romania) and has a map area of 347 km² and a mean slope of 418‰, has a real area given by

$$A_r = A_m/\cos\alpha = 347/0.923 = 376 \text{ km}^2$$

a difference of 29 km² (8.4%) from the map area. The mean specific runoff (Pădeşel gauging station) on the basis of the map projection is 33.4 $l\ s^{-1}\ km^{-2}$, while for the true area it is only 30.8 $l\ s^{-1}\ km^{-2}$. For the Someşul Mare (Rodna gauging station), for which the mean slope of the drainage basin is 426‰, the area given by the map projection is 290 km², while the real area is 315.2 km². In considering basins in plain regions, where the mean slope generally has low values, the map projection can be used without any correction. Thus, in the case of the Obedin, a right-side tributary of the Jiu, whose basin area is 475 km² and has a mean slope of 58‰, the difference between the true and projected areas is only 1 km².

## Arrangement of Area Elements in Relation to Drainage Axis

Several methods may be used to provide a clearer picture of the arrangement of a basin area in relation to the main drainage axis.

*Circular diagram of arrangement of area elements on valleysides and in subbasins.* Using the area values obtained by planimeter for an entire basin and for its main subbasins, it is possible to draw a circular diagram which renders the whole basin area, the areas of the two valleysides and of the main subbasins, and their positions in relation to the drainage axis (Fig. 17). The circle radius is taken arbitrarily, as conve-

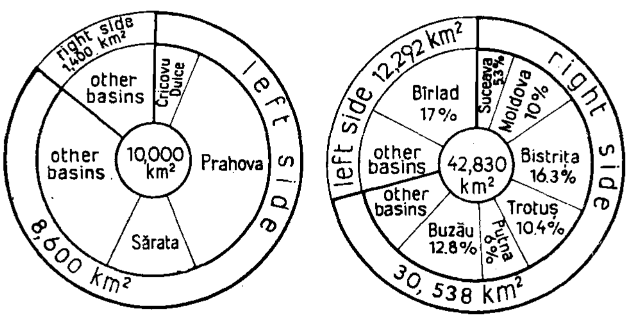



Fig. 17. Circular diagram of arrangement of area elements on valleysides and in subbasins for (A) the Ialomiţa and (B) the Siret basins.

nient. The areas of the valleysides and of the subbasins may be given as absolute values (Fig. 17A) or as percentages of the whole basin area

(Fig. 17B). To evaluate the sectors of the circle corresponding to each of the subbasins, the following formula can be used :

$$A_b^o = 360°(A_b/A) \tag{35}$$

where $A_b^o$ is the angle (deg.) of the circle sector corresponding to a subbasin, $A_b$ the area (km$^2$) of the subbasin, and $A$ the area (km$^2$) of the whole basin (Morariu et al., 1962).

*The asymmetry index of a drainage basin.* This index ($a$) is a measure of the arrangement of the area elements in relation to the drainage axis. It is computed from the area $A$ of the whole basin and the areas $A_R$ and $A_L$ on the right- and left-hand sides, respectively. These values are introduced into the formula (Chebotarev, 1953)

$$a = (A_L - A_R)/A \tag{36}$$

to give the asymmetry coefficient, whose value may range from $+1$ to $-1$. When the value of $A_L - A_R$ approaches zero, the basin is perfectly symmetrical; negative values correspond to basins developed on the right-hand side, as is the case for the Siret (Fig. 17B), while positive values correspond to basins with a larger area on the left-hand side, as is the case for the Ialomiţa (Fig. 17A).

Sometimes, the following formula is used:

$$a = 2(A_L - A_R)/A \tag{36 a}$$

in which the asymmetry index ranges from $+2$ to $-2$, the zero value and the signs having the same significance. As is obvious, in this case the values are doubled. In the case of the Ialomiţa basin, the area on the right-hand side of the main stream is 1400 km$^2$, while that on the left-hand side is 8600 km$^2$, which correspond to indices of 0.72 (former formula) and 1.44 (latter formula).

*Diagram of basin area in relation to length of main stream.* This diagram can be constructed on the basis of the arrangement of the area elements in the various sectors of the main stream marked by confluences and along the more important tributaries. This is also possible using the Horton—Strahler classification system for drainage basins, and helps in following the arrangement of the area elements belonging to basins of various orders and their growth with increasing basin order (Fig. 18).

*The afforestation coefficient.* In studying the role of forest in determining hydrological parameters, a quantitative measure of the degree of afforestation is needed. Such a measure is provided by the afforestation coefficient, which is simply the ratio of the forested ($A_f$) and total ($A_b$) areas of a basin, determined planimetrically :

$$I_f = A_f/A_b \tag{37}$$

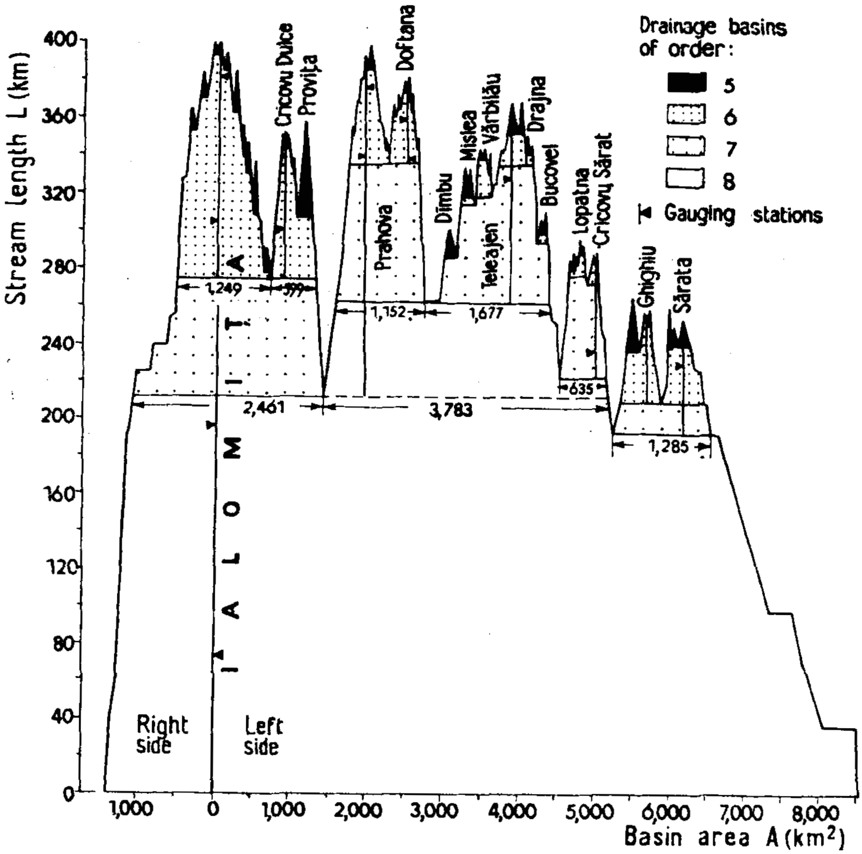



Fig. 18. Increase in basin area of the Ialomița in relation to main-stream length.

which gives a dimensionless coefficient ranging from zero (when there is no forest) to + 1 (when the basin is completely forested). The values can be also expressed as percentages, employing the relationship

$$I_f = 100(A_f/A_b) \tag{37a}$$

As a genetic runoff factor, the afforestation coefficient of a drainage basin cannot be ignored. The higher this coefficient, the lower the mean multiannual runoff for smaller basins (below 200 km²) and the higher the mean multiannual runoff for larger basins, as noted for the Tîrnave basin (Stănescu et al., 1967). A greater afforestation coefficient also implies smaller maximum runoff and lower flood volume and suspended load, as a consequence of the resulting diminution of erosion. The water layer that runs off is of greater depth in deforested basins than in forested ones. In wintertime, the mean daily runoff has higher values in deforested basins because of the more rapid thawing of snow layers, and lower values in forested ones (Miţă et al., 1970).

Similar computations can be made to determine the *lacustrine coefficient* (the extent to which a basin is covered by lakes), the *swampiness coefficient*, when there are swamps, and even a coefficient for a certain type of soil if the latter influences runoff processes.

## Areas Required for the Appearance of Drainage Basins of Various Orders

Establishment of the area necessary for the organization of surface runoff, and hence the threshold from which denudation processes begin under the impact of linear erosion, is very important both in field activities and in hydrological and geomorphological studies. Such questions are of urgent importance in cases of excessive utilization of sloping ground, when areas affected by erosion may grow rapidly. Man's action may thus set in motion a complex chain of phenomena. Numerous zones may appear in which the natural equilibrium between the forces opposing erosion and those causing it is destroyed in favour of the latter, resulting in increasingly large areas being rendered useless for agriculture. Taking into account that a considerable proportion of the Earth's arable land is affected by erosion, and that the major cause is river networks, knowledge of the laws governing the appearance and evolution of the latter must be seen as of paramount importance.

As early as 1945, Horton showed that a water course can appear only when there is a threshold area sufficiently large for the water volume accumulated to carve such a form. This idea was taken up by Strahler (1964), Leopold et al. (1964), Tricart (1962, 1965) and Schumm (1977), and studied in detail by Lambert (1975), who complemented Horton's statement by showing that the discharge must also have a frequency sufficient to maintain the water course once carved.

Considering a first-order basin in the Horton—Strahler classification system, at its origin there will be an area of variable size lacking an elementary water course with definite morphological features (Fig. 19A(a)). The fact that unorganized runoff prevails in this area leads to the conclusion that the appearance of elementary channels along which water may run requires a certain area over which water must accumulate with sufficient energy to carve a channel and sufficiently often to maintain it. This minimum area, which Lambert (1975) termed the *threshold area of surface runoff*, is a result of the interaction of local geographical factors. Thus, plant cover with its high interception power, small slopes of the land, or the presence of soil with a thick and well-formed profile, and hence great permeability, all act to hinder the concentration of water, and thus a greater drainage area is required to accumulate a discharge with sufficient energy to form a water course. In contrast, a lower vegetation density or a complete absence of vegetation, steep slopes of the land, impermeability of the soil and subsoil,

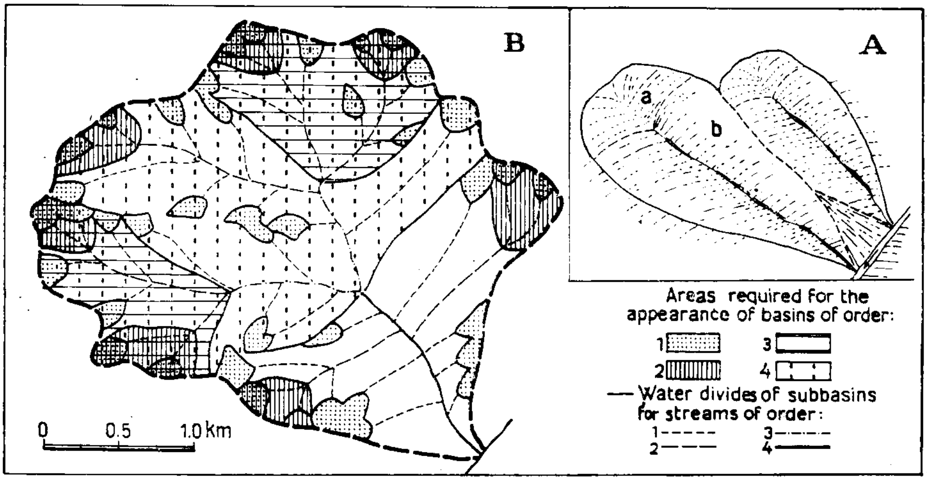



Fig. 19. Areas required for appearance of basins of various orders : (A) scheme for first-order basins (a, minimum area required for concentration of runoff and appearance of a channel, and hence of first-order basins ; b, area required for maintenance and development of channel) ; (B) scheme for a fourth-order basin.

the presence of soil with a thin profile and heavy texture, and higher frequency, duration and intensity of rainfall favour concentration of water along the slopes, and thus a smaller area is necessary for concentrated flow.

The threshold area of surface runoff can be determined by field measurements, from aerial photographs or from large-scale maps. Since this is more difficult to achieve in the case of large basins, it is possible instead to employ various indices and relationships for the factors mentioned above. Taking into account that the opposition between the forces of erosion and resistance continues under almost identical conditions during very long periods of time, the present topographical surface of a basin will obviously reflect these effects. Therefore, by examining the morphometrical features of an area, it should be possible to derive from a sufficiently large number of measurements a series of indices and laws which will be of use not only in understanding the present situation but also in forecasting its evolution.

As is known, in the Horton—Strahler system, a second-order basin appears only after the junction of two first-order basins. The junction of two second-order basins represents both the instant of creation and the threshold area for appearance of a third-order basin, and so on. However, between the appearance of such a basin and its union with another of the same order, a number of quantitative accumulations take place which also result in an increase of the basin area. This leads to the conclusion that a stream of any order will require a minimum basin area in order to appear which must be sufficiently large for it to persist and develop (Fig. 19A). To obtain the threshold area of surface runoff we start from the minimum areas necessary for the formation of basins of various orders, knowing that a basin of a certain order $x$ appears only through the junction of two basins of immediately lower order $x - 1$:

$$A'_x = 2A'_{x-1} \tag{38}$$

Thus, the minimum area required for the appearance of a fourth-order basin is established from the sum of the areas of the two third-order basins through whose junction it appeared (Fig. 19B). The same method may be used for all the streams of lower order existing in a basin. The resulting series of sums of minimum areas are then divided by the cor-

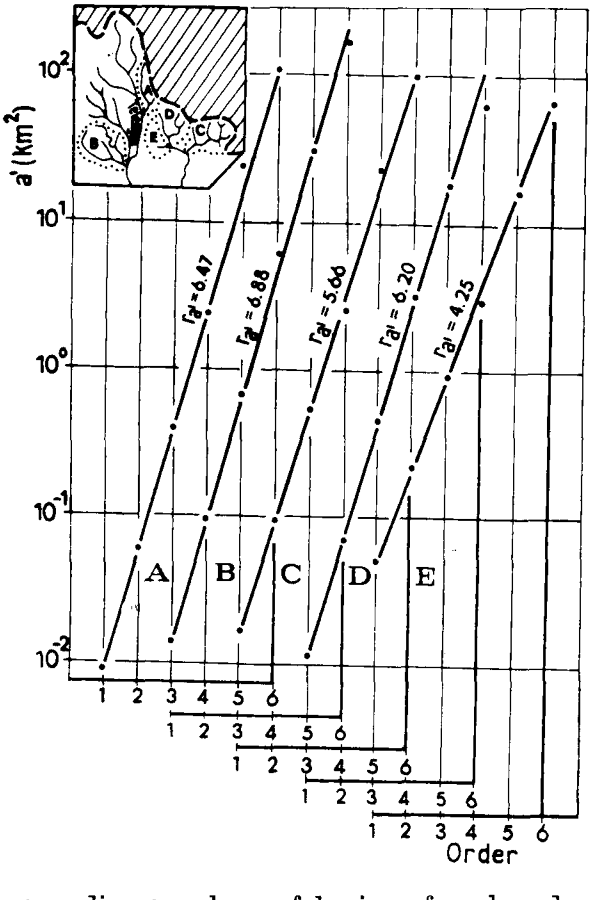



Fig. 20. Relationship between areas required for appearance of basins of various orders, and order of basin, for (A) the Drajna at its confluence with the Teleajen, (B) the Mislea at its confluence with the Teleajen, (C) the Cricovu Sărat at its confluence with the Lopatna, (D) the Lopatna at its confluence with the Sărățel, and (E) the Sărățel at its confluence with the Lopatna.

responding numbers of basins of each order analyzed. If the average values obtained are represented on semilogarithmic paper in relation to stream order (Fig. 20), a straight-line relationship reflecting direct proportionality is obtained. Verification of this relationship for a large number of cases has led to the following general law : *the areas required for the appearance of drainage basins of successively higher orders tend to form an increasing geometric progression in which the first term is the area $a'_1$ required for the appearance of first-order basins and the ratio is the ratio $r_{a'}$ of successive areas.* Denoting the series of average areas obtained as above by $a'_1, a'_2, a'_3, \ldots, a'_n$, the general term of the geometric series can then be computed using the relationship

$$a'_u = a'_1 r_{a'}^{u-1} \tag{39}$$

The ratio of this progression may be calculated using the weighted mean of the values obtained for successive pairs of orders, the arithmetic mean of the partial ratios, or the slope of the straight line passing through the average of all points.

Equation (39), represented by a straight line on semilogarithmic paper, can also be solved as an equation of the form

$$\log a'_u = c' + d'x \tag{40}$$

where $c'$ and $d'$ are constants and $x$ is stream order.

Thus, the area required for the appearance of first-order streams, which is simply the threshold area of surface runoff and which is most difficult to establish cartographically, can be determined very easily on the basis of the above law. It is given by the first term of the geometric progression, and can be established graphically or from the relationship

$$a'_1 = a'_u / r_{a'}^{u-1} \tag{41}$$

On the basis of this relationship, it is possible to establish for a given system the minimum area required for the appearance of a first-order stream, and hence the area necessary for the appearance of concentrated flow capable of creating an elementary water course which is able to carry away from the area considered both excess water and suspended matter.

From the situation analyzed it has been found that the minimum areas required for the appearance of streams of any order, and hence also those of first order, correspond to hilly areas. Thus, considering the basin of the Drajna river upstream from its confluence with the Teleajen, which has developed on strongly folded Neogene sedimentary formations with marked tectonic mobility, it may be noted that at the origin of the elementary network an area of only 0.9 ha is required for the concentrated flow to be able to carve a distinct bed in the topographical surface (Fig. 20A). In the case of the Lopatna river upstream from its confluence with the Sărăţel, the threshold area of surface runoff is 1.17 ha (Fig. 20D), while in the case of the Sărăţel upstream from its confluence with the Lopatna, with a relief carved in a Subcarpathian depression, it amounts to 5 — 6 ha (Fig. 20E). This difference is easily explained by the fact that one of the most important tributaries of the Sărăţel (the Bălţeşti stream) drains the Podeni depression where the relief energy is low and the alluvial deposits have high permeability, which requires a greater area for the excess water to be able to carve a course under the given physiographical conditions.

Detailed studies were conducted by Lambert (1975) in the basin of Hers-Mort, a right-side tributary of the Garonne downstream from Toulouse. For this basin, which has an area of 768 $km^2$ and has developed on Tertiary molasse formations with layers of clay and sand, sand, sandstone and limestone, with altitudes ranging from 60 to 400 m and homogeneous geomorphological conditions, the measurements revealed a threshold area of surface runoff of 12.5 ha. Taking into account the greater diversity of rock conditions, relief and tectonic mobility in Romania's Neogene sedimentary formations, and the climatic regime, with a high frequency of precipitation in shower form, the smaller threshold area is quite explicable.

Analysis of the data obtained indicates that the threshold area of surface runoff is closely related to an aggregate of geographical factors, among which important roles are played by rock conditions, tectonics, structure, plant cover, etc. In mountainous areas with a relief developed on pre-Alpine crystalline massifs or on folded Mesozoic and Paleogene flysch formations, the threshold area of surface runoff is larger than in regions developed on Neogene sedimentary formations. In plain regions, where the relief energy and slope have very low values and where highly permeable alluvial deposits prevail, it is natural that the threshold area of surface runoff should be much larger. The fact that the lowest values are generally found in deeply fragmented hilly regions is accounted for by the great diversity of rocks with various resistivities to erosion (which explains the high degree of fragmentation). On friable rock such as the Cîndeşti gravel, only very small areas are required for the appearance of stream flow. There is thus a close connection between lithology on the one hand, and relief fragmentation and drainage density on the other.

*Relationship between threshold area of surface runoff and rock type*. Considering the threshold area of surface runoff in relation to rock type only, it is clear that the value of the former will be closely related to a rock's ability to resist mechanical stress. Thus, on the Precambrian chloritic schists in the Leaota mountain, which have a high resistance to erosion, a threshold area of 6 ha is necessary for the carving of the first channel unit. This value decreases to the extent to which the rock resistance to erosion also drops, reaching approximately 1 ha in areas of Cîndeşti gravel or of friable formations subject to extensive erosion, as in the Drajna and Lopatna basins which have developed on Neogene sedimentary formations. Thus, as the rock resistance becomes lower, with concomitant decreases of slope, and its permeability higher, with increased retention of water originating in precipitation, the threshold area of surface runoff increases. Of the aggregate of physiographical factors affecting the threshold area, lithology thus seems to be the most significant. It is true that a number of factors, such as soil, vegetation, etc., may protect the topographical surface and increase its resistance to erosion, but these effects operate only over relatively small depths. Once gully erosion has started and the topsoil or deposits covering a slope have been washed away, the evolution of the newly formed channel takes place in relation to the slope and resistance to erosion of the underlying rock. This resistance also seems to be the main factor determining the gradual advance of the channel, and hence the threshold area of surface runoff.

Knowledge of the threshold area is of practical importance, since on its basis, measures can be taken to protect a zone or increase its resistance to erosion. For example, uncontrolled deforestation, especially if carried out in regions with highly friable rock, such as the Cîndeşti gravel, may lead in a very short time to strong ramification of the existing channel network. Erosion in these regions then becomes much more rapid, since rills and gullies tend to appear in geometric progression. Thus, these phenomena must be understood so that appropriate measures can be taken immediately in regions subject to erosion.

*Relationship between the area required for the formation of third-order basins, and rock type.* As shown above, the areas required for the appearance of basins of various orders are greatly influenced by overall physiographical factors, primarily rock type, tectonics and geological structure. In the Subcarpathian region, lithological heterogeneity increases in basins of order higher than the fourth, and decreases with decreasing order, so that first-order basins develop mostly under homogeneous conditions.

In order to study the relationship between the areas required for the formation of basins of various orders, and rock type, a classification was made of geological formations in the Ialomiţa basin according to mechanical strength, on the basis of norms established by the Geological Committee of Romania. Then, some 550 third-order basins were analyzed which showed lithological homogeneity, without taking into account structure and tectonics. Of course, these also play important roles, but the decisive factor is rock type.

In examining the relationship between the various categories of rock and the average area required for the formation of a third-order basin, a difference was noted between the influences of consolidated and nonconsolidated rock. Thus, the Precambrian chloritic schists in the Leaota mountains, which have high resistance to erosion, although their high relief energy is striking, appear at first sight to be highly fragmented. In fact, of the rocks classed as consolidated, these are characterized by the largest average area required for the formation of third-order basins (Fig. 21). There follow, in terms both of resistance and of average area required for the formation of third-order basins, sandstone, marl, marl—sandstone and marl—limestone, schist, marl—clay and calcareous Lower Cretaceous sandstone. The average area required for the appearance of third-order basins continues to decrease with decreasing rock consolidation, so that in the case of the Helvetian deposits of sandstone, marl and conglomerates, its value is almost half that for the Precambrian schists in the Leaota mountains. The series of consolidated rocks ends with the Cîndeşti layers formed from clay,

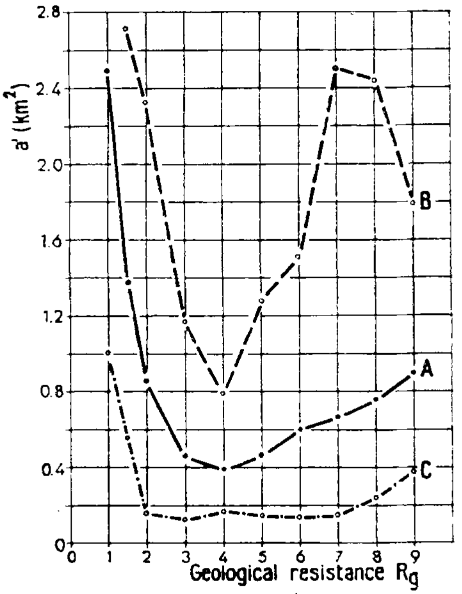



Fig. 21. Areas required for appearance of third-order basins within the Ialomiţa basin, as a function of geological resistance $R_g$: A, mean areas; B, maximum areas; C, minimum areas.

sand and gravel of the Lower Pleistocene, which are sometimes sufficiently well consolidated to resist on slopes up to 90°. Thus the area required for the appearance of third-order basins decreases continuously from the Precambrian schists to the Cîndeşti gravel, reaching its lowest value in the case of the latter (Fig. 21).

In the case of nonconsolidated rock with low resistance to erosion, the average area required for the formation of third-order basins increases to the extents to which the hardness decreases and the permeability becomes greater. Thus, for the complexes of clay, marl, sand and gravel found from the Pontian to the Romanian, the minimum third-order area increases, reaching a maximum for gravel of the Upper Pleistocene or loess deposits (Fig. 21). The relationship thus established shows that the smallest areas required for the appearance of third-order basins are found not in the mountainous area but in the Subcarpathian regions consisting of poorly consolidated and highly friable rock, such as the layers of clay, sand and gravel at Cîndeşti, or Mio-Pliocene deposits. The largest areas, on the other hand, occur in plain regions covered by loess deposits.

Although the extreme highest and lowest values of the area required for the appearance of a third-order basin in a given region may be determined accidentally by structure and tectonics, the influence of the latter factors is restricted, since the extreme values show a tendency similar to that of the mean values across the various regions (Fig. 21B, C).

Taking into account that the response of a given area to external factors (precipitation, erosion, man's action) is commensurate with the potential resistance it possesses, it is obvious that rock type will be an essential factor.

As has been seen, rock type accordingly plays a decisive role in determining the areas required for the formation of basins of various orders. Although the other geographical factors mentioned cannot be ignored, over longer periods of time their role is smaller and they do not significantly affect the outcome.

*Relationship between number of streams and drainage area.* A direct relationship exists between the total number of streams in a drainage basin and the latter's area, although, given the important role played by rock type, a given relationship will be valid only for lithologically similar regions.

For Romania, with a territory of 237 500 km², Diaconu (1971) determined the numbers $N^A$ of rivers with basin areas larger than certain limiting areas $A$, and from these data he calculated, for each category, the specific number of rivers $N_S^A$ for a unit area of 10 000 km² (Table 2).

TABLE 2

**Numbers $N^A$ of rivers with basin areas larger than limiting areas $A$, and specific numbers of rivers $N_S^A$ (after Diaconu, 1971)**

| | A (km²) | | | | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | >20 | >50 | >100 | >200 | >300 | >400 | >500 | >700 | >1000 | >1500 | >2500 | >5000 | >10 000 |
| $N^A$ | 2545 | 1124 | 586 | 302 | 210 | 165 | 131 | 85 | 67 | 44 | 26 | 15 | 7 |
| $N_S^A$ | 107 | 47.0 | 24.7 | 12.7 | 8.85 | 6.95 | 5.50 | 3.60 | 2.80 | 1.85 | 1.10 | 0.630 | 0.295 |

The values obtained both on the national scale and for individual drainage basins allow comparative studies. It is found that the logarithms of the specific values ($N_S^A$) correlate well with the logarithms of the areas $A$ (Fig. 22).

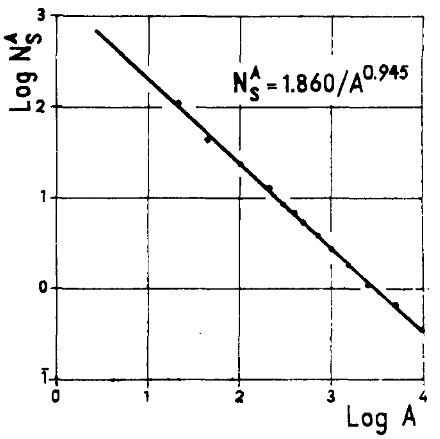



Fig. 22. Relationship between logarithm of river specific number $N_S^A$ conditioned by a given area, and logarithm of basin area $A$ for the overall network of rivers in Romania (after Diaconu, 1969).

## Frequency Distribution of Drainage Areas

In the Horton—Strahler classification system, the order of the main stream is also the order of the basin feeding it. The drainage basin areas measured by planimeter may be grouped by size and can be then processed statistically. The areas of drainage basins of various orders are determined by specific conditions pertaining to the relief units on which they have developed, and in particular by the resistance of the substituent rocks, to which should be added the influence of paleogeographical evolution, tectonics, structure and even human activities. In order to assess the roles of these various factors in determining drainage-basin area, a sample of 1202 third-order basins in the Ialomiţa system with an average area of 3.42 km$^2$ has been analysed. Of these, 337 basins (average area 2.45 km$^2$) are in the mountainous region, with a relief developed on Precambrian amphibolites and Cretaceous sedimentary formations. Passing to the hilly regions, where the highly folded Neogene formations have a lower resistance to erosion, a decrease of the average area to 1.67 km$^2$ (771 basins) is found. In the piedmont and plain regions, the average area of a third-order basin is 21.2 km$^2$ (97 basins). This latter value is a result of the small slope of the relief, which determines the drainage conditions in plain areas.

Significant differences between the average areas by relief unit are also found for fourth-order basins. Thus, for 72 basins in the mountainous region, the average area is 12.7 km$^2$; in the hilly regions this value decreases to 9.84 km$^2$, with 138 drainage basins; and in plain regions it increases to 47.3 km$^2$ for 16 drainage basins.

Statistical processing of the data for a sample of 210 fourth-order drainage basins in the mountainous and hilly regions, and construction of a histogram of the frequency distribution by class shows marked positive skewness consequent upon accentuated fragmentation of the relief, i.e., smaller drainage basins are expected in the hilly areas (Fig. 23A).

In order to determine the probabilities of occurrence of extreme values, it is possible to calculate either empirical curves or appropriate theoretical probability curves. The calculation of empirical probabilities requires the construction of a table containing the individual values arranged in order of decreasing magnitude, from which the probability of occurrence of each value may be calculated using the formula

$$p(\%) = 100(m - 0.3)/(n + 0.4) \tag{42}$$

where $m$ is the number of the term in the series and $n$ the total number of cases. By graphical representation of the calculated values on probability paper, the empirical probability curve may be drawn through the average of the points. Such a curve is precise only for the interval for which direct data are available. There are instances when the probability of occurrence must also be determined at the extremities of the empirical curve, for which no data may exist even if the sample analyzed

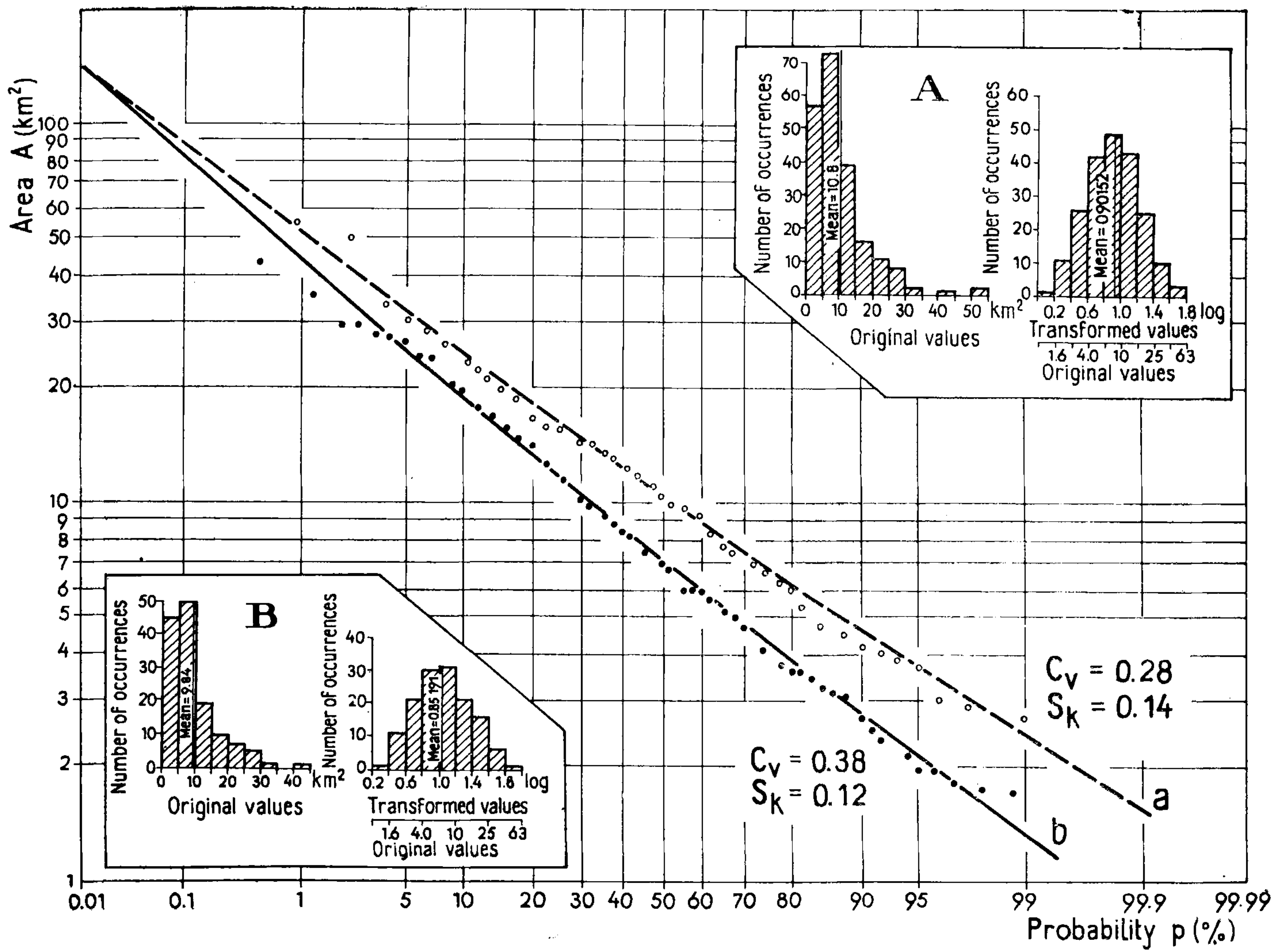



**Fig. 23.** Probability distribution of drainage areas for samples of (a) 78 fourth-order basins in mountainous areas and (b) 138 basins in hilly areas of the Ialomița basin, and distribution of original values compared to that of transformed values for (A) the population of both samples, and (B) the hilly regions alone.

is large. This problem may be solved either by simply extrapolating the empirical curve, which implies a fairly high degree of approximation, or by calculating the theoretical probability curve. This requires computation of the mean of the series, the variance, the standard deviation and the coefficients of variation ($C_v$) and skewness ($S_k$). After determination of these parameters, the deviations of the ordinates of the binomial curve are taken from a table in relation to the average $\Phi$ for $C_v = 1$ and for the calculated $S_k$. Then, to reduce the deviations, $\Phi C_v$ is calculated using the coefficient of variation determined for the series of data considered, i.e. $\Phi C_v = K_i - 1$ or $K_i = \Phi\ C_v + 1$, where $K_i$ is the ratio of the mean of the series to the *i*th value. The result obtained is multiplied by the average for the sample to give the points which define the theoretical probability curve (Diaconu and Lăzărescu, 1965).

The occurrence of skewness coefficients of high value (2 – 4) for the series analyzed indicates that the probability curve is highly concave, and hence that the extreme values may be extrapolated or computed only with difficulty. In such instances, it is recommended that the curve be corrected or the data transformed in such a way as to approximate as closely as possible a normal distribution of values around the average. A graphical method for correcting the empirical probability curve consists in representing the data calculated according to eqn. (42) on logarithmic probability paper (Fig. 23). If a normal logarithmic distribution is involved, the plot will be a straight line, allowing easier and safer extrapolation to obtain the extreme values.

Transformation of the data is also recommended in calculating theoretical probability curves in such cases (S. Gregory, 1973). In fact, Schumm (1956), in studying the density of first- and second-order basins, showed that the data could be suitably corrected by using the logarithms of area. The use of transformed data in calculating theoretical probability curves for a decreasing series requires logarithmic treatment of the values employed in establishing the statistical parameters, including the coefficients of variation and skewness. Considering the density of areas for the preceding sample of 210 drainage basins in mountainous and hilly regions, it can be seen how a highly asymmetrical distribution is turned by this method into a normal one. The calculation method for establishing the ordinate of the theoretical probability curve is identical to that employed in the computations using the untransformed data; however, having determined the ordinates, it is necessary to return again to the original values (Table 3).

Representation of the data obtained on logarithmic probability paper allows accurate calculation of the probability of occurrence of any value on the curve (Fig. 23). The very low values of the skewness coefficient and the straight lines obtained by this method prove that a normal logarithmic distribution represents the data well. The positions of the curves show that whereas for the largest fourth-order basins the difference between the areas in mountainous and hilly regions is very slight and the two curves nearly coincide, for the lower portions of the

TABLE 3

**Determination of the ordinates of the theoretical binomial curve at various probabilities for a sample of 138 fourth-order drainage basins in the hilly area of the Ialomiţa basin, with $C_v = 0.379$ and $S_k = 0.116$ obtained by processing transformed data**

| | Probability | | | | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|
| | 0.01 | 0.1 | 1 | 5 | 20 | 50 | 70 | 90 | 99 | 99.9 |
| $\Phi$ | 3.984 | 3.260 | 2.414 | 1.676 | 0.838 | 0.022 | 0.534 | 1.268 | 2.236 | 2.922 |
| $\Phi C_v$ | 1.510 | 1.236 | 0.915 | 0.635 | 0.318 | 0.008 | 0.202 | 0.481 | 0.848 | 1.108 |
| $\Phi C_v+1$ | 2.510 | 2.236 | 1.915 | 1.635 | 1.318 | 1.008 | 0.789 | 0.591 | 0.152 | 0.108 |
| $(\Phi C_v+1)\bar{x}$ | 2.138 | 1.905 | 1.631 | 1.383 | 1.123 | 0.859 | 0.679 | 0.443 | 0.130 | 0.099 |
| Final calcd. values | 137.6 | 80.3 | 42.8 | 24.7 | 13.3 | 7.23 | 4.78 | 2.77 | 1.35 | 1.23 |

curves the difference increases because of the tendency of drainage-basin areas to decrease in hilly regions in comparison with mountainous ones.

## Morphometrical Model of Basin Area

The distributions of area for drainage basins of various orders are not random but follow certain laws which can be ascertained. From this point of view, each drainage basin can thus be individualized in relation to the parameters characterizing it. In order to establish a morphometrical model of drainage areas it is necessary to consider the law of stream numbers, as discussed previously, and the laws of summed and of average areas. The parameters of these laws are important in characterizing the state of a drainage system as concerns the delineation of drainage areas.

*Law of summed areas.* After delimiting the drainage basins of various orders within a given drainage system, their perimeters are drawn on planimetric paper and the areas of basins of each order are then summed. By effecting this operation, a series of data are obtained for the summed areas of basins of successive orders. These values are denoted by $A_1$ (the sum of the areas of all first-order basins), $A_2$ (the sum for the basins of second order), and so on, up to the area $A_u$ of the entire basin (of highest order).

On plotting these values on semilogarithmic graph paper in relation to basin order, it is found that they define a straight line (Fig. 24). This functional dependence, verified in a large number of cases, confirms the existence of a law governing the variation of the summed areas for successive orders : *the summed areas of drainage basins of successively higher orders in a given system tend to form a geometric progression in which the first term is the summed area of $A_1$ first-order basins and the ratio is the ratio $R_A$ of successive summed areas.* The resulting geometric progression can be written as $A_1, A_2, A_3, \ldots, A_{u-1}, A_u$. The general term of this series is

$$A_u = A_1/R_A^{u-1} \tag{43}$$

or

$$\log A_x = D - Ex \tag{44}$$

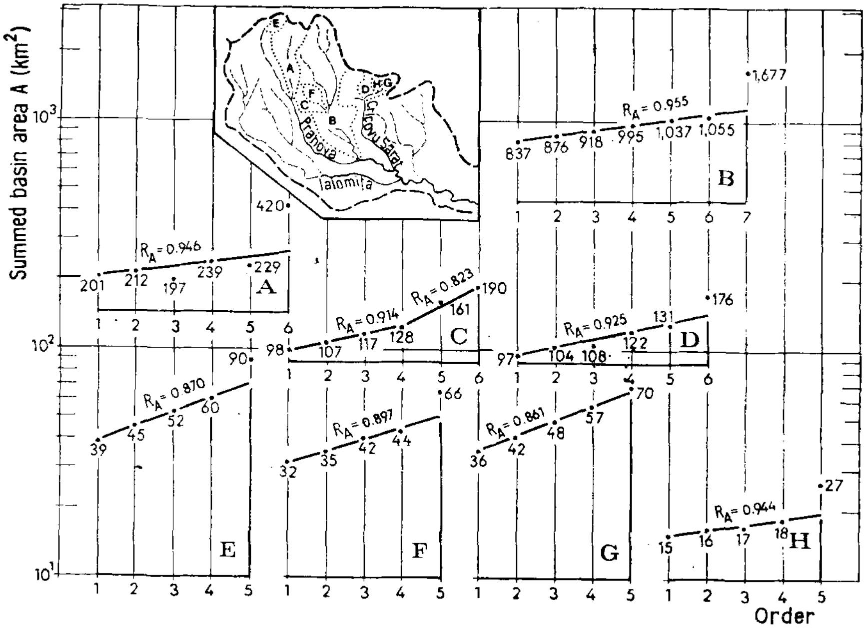


Fig. 24. Regression of summed basin areas on order for the basins of (A) the Doftana at its confluence with the Prahova, (B) the Teleajen at its confluence with the Prahova, (C) the Mislea at its junction with the Teleajen, (D) the Cricovu Sărat at its confluence with the Lopatna, (E) the Azuga at its confluence with the Prahova, (F) the Cosmina at its confluence with the Mislea, (G) the Cricovu Sărat at its confluence with the Salcia, and (H) the Salcia at its confluence with the Cricovu Sărat.

where $D$ and $E$ are constants and $x$ is basin order. To determine a term $x$ when the value of the general term is known, the following relationship can be used:

$$A_x = A_u\, R_A^{u-x} \tag{45}$$

However, this relationship is seldom resorted to because, as may be noted from the graphical representation (Fig. 24), the value for the area of the highest-order basin generally deviates from the rule.

The series ratio, which also gives the slope of the line, is equal to the ratio of successive summed areas and may be calculated on the basis of the weighted mean or the arithmetic mean, or by determining the slope of the straight line which best fits the measured points. Using the chosen-point method, it is possible to consider two points at an interval of two orders (the second and fourth, or the third and fifth), and knowing that $A_3 = A_5 R_A^{5-3}$, or $R_A^2 = A_3/A_5$, $R_A$ is obtained as $\sqrt{A_3/A_5}$. Considering the second and fourth orders for the drainage basin of the Salcia at its confluence with the Cricovu Sărat (Fig. 24H), for which $A_2 = 16.5$ km$^2$ and $A_4 = 18.4$ km$^2$, $R_A = \sqrt{16.5/18.4} = 0.944$. If terms situated at an interval of three, four or more orders are considered, the order of the radical will be equal to the difference between the two orders. For instance, the measured areas of drainage basins of successive orders in the case of the Azuga system at the confluence with the Prahova yield the following series of sums, corresponding to the first to fifth orders : 39.59 : 45.5 : 52.3 : 60.2 : 89.6. From a graphical representation of these values (Fig. 24E), it can be seen that the first four terms form a geometric progression whose ratio is $R_A = 0.870$.

The results for a large number of fifth-order basins show increasing geometric progressions with ratios below unity, the mean value for all fifth-order basins in the Ialomiţa river system being 0.93. However, there are also situations in which the values obtained indicate decreasing or increasing geometric progressions, or even series for which $R_A$ is equal to unity. Decreasing progressions are found more frequently in the case of markedly elongated basins. In the case of higher-order basins, there are instances when the values obtained define two distinct lines, one valid for lower-order basins and the other for higher-order ones, as in the case of the Mislea basin at its confluence with the Teleajen (Fig. 24C).

When the value of the first term cannot be measured because the operation is difficult, and the last term deviates considerably from the line, so that its value cannot be taken into consideration, the ratio is determined starting from the values falling best on the line. For instance, the summed areas of drainage basins of successive orders in the Cricovu Sărat system at the confluence with the Lopatna form the following series :

| | Order | | | | | |
|---|---|---|---|---|---|---|
| | 1 | 2 | 3 | 4 | 5 | 6 |
| $A_{meas.}$ ($R_A = 0.925$) | – | 104.2 | 108.3 | 121.8 | 131 | 176 |
| $A_{calcd.}$ | 96.4 | 104.2 | 112.6 | 121.8 | 131.6 | 142.3 |

Graphical representation (Fig. 24D) shows that it is possible to compute the ratio of the summed areas starting from the values obtained for the second and fourth orders, which fall on the line, and then the values corresponding to the other orders.

The law of summed areas for drainage basins of successive orders is directly related to relief fragmentation. High fragmentation results in a lower average area for each order, in greater summed areas and in smaller interbasin spacing. This reflects the appearance in such basins of greater numbers of new streams of various orders, which implies redistribution of the drainage network. The new generation will take over part of the area fed by the previous generations of gullies and will result in a decrease in the interbasin spacings corresponding to each order.

*Law of average areas.* Knowing the series of values given by the summation of the basin areas corresponding to each order, and the

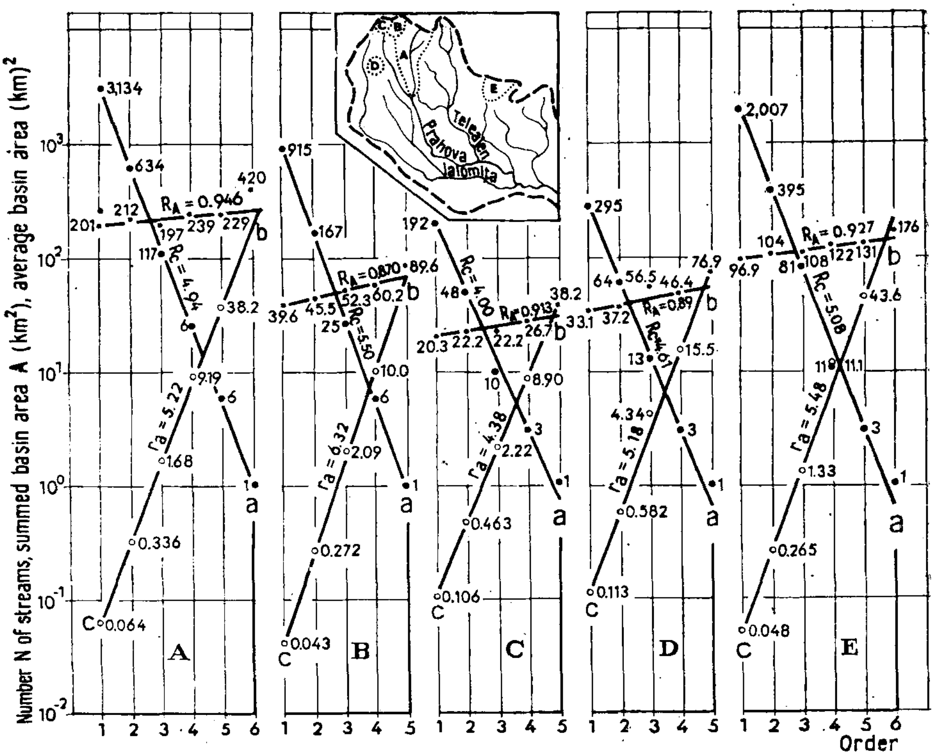


Fig. 25. Morphometrical models of basin area for the basins of (A) the Doftana upstream from its confluence with the Prahova, (B) the Azuga upstream from its confluence with the Prahova, (C) the Prahova upstream from its confluence with the Azuga, (D) the eastern Ialomicioara at its junction with the Ialomiţa, and (E) the Cricovu Sărat at its confluence with the Lopatna (a, regression of numbers of streams; b, regression of summed basin areas; c, regression of average basin areas).

numbers of basins of each order in the drainage system analyzed, it is possible to determine the third progression mentioned above, that of the average areas. Representation of the first two series on semilogarithmic paper gives two straight lines (Fig. 25(a, b)). The ratios of the corresponding terms in the two series will then yield the average areas of drainage basins of successive orders, which again form an increasing geometric progression (Fig. 25(c)).

On the basis of the expressions for the general terms of the first two series, eqns. (27) and (43), the general term in the third progression is obtained from

$$a_u = A_u/N_u$$

By substituting the corresponding expressions, and noting that $A_1/N_1 = a_1$ and $R_c/R_A = r_a$, the general term of the new series is obtained as

$$a_u = a_1 r_a^{u-1} \tag{46}$$

From graphical representation of the values on semilogarithmic paper the relationship between the two variables can also be written as

$$\log a = d + ex \tag{47}$$

where $d$ and $e$ are constants and $x$ is basin order. As in the case of the stream numbers, the two constants are determined from a system of two equations.

In the case of the fifth-order drainage basin of the Prahova at its confluence with the Azuga (Fig. 25C), only small differences are found between the measured values and those calculated on the basis of the

TABLE 4

**Measured and calculated data for the morphometrical model of the drainage areas of the Prahova at its confluence with the Azuga**

| | Order | | | | |
|---|---|---|---|---|---|
| | 1 | 2 | 3 | 4 | 5 |
| $N_{meas.}$ | 192 | 48 | 10 | 3 | 1.0 |
| $N_{calc.}$ ($R_c = 4.00$) | 192 | 48 | 12 | 3 | 0.75 |
| $A_{meas.}$ | | 22.2 | 22.2 | 26.7 | 38.2 |
| $A_{calc.}$ ($R_A = 0.913$) | 20.27 | 22.2 | 24.3 | 26.6 | 29.15 |
| $a_{meas.}$ | | 0.463 | 2.22 | 8.90 | 38.2 |
| $a_{calc.}$ ($r_a = 4.38$) | 0.106 | 0.463 | 2.03 | 8.88 | 38.9 |

preceding formula (Table 4). Therefore, the ratio of the new series is equal to the quotient between the ratios of the two previous series.

The new law shows that *the average areas of drainage basins of successively higher orders tend to form an increasing geometric progression in which the first term is the average area* $a_1$ *of first-order basins and the ratio is the ratio* $r_a$ *of successive average areas.* This law was sketched as early as 1945 by Horton, who showed that the average areas of progressively greater basins increase in a geometric progression. Thus, Horton had already found a very important law according to which the drainage area of river basins develops. Schumm (1956) resumed the matter and gave a definitive formulation of this law, which was eventually confirmed by Leopold and Miller (1956) for drainage basins with temporary runoff in New Mexico, by Morisawa (1962) for the Appalachian plateau, by Christofoletti (1970) for a number of basins on the Brazilian plateau, by Drexler (1979) for basins in the Karwendel mountains (Tyrol), and by a whole series of other researchers. Since a fairly large number of situations have been analyzed, the law of average areas may be considered well verified in comparison with that established for summed areas (Fig. 26).

The ratio of successive average areas, which indicates the growth rate of average area from one order to another, is highest in regions of Neogene sedimentary formations, where tectonics, structure and rock type are highly important in determining greater relief fragmentation. High values of the ratio of average areas are also found in the Azuga basin, situated fully in a mountainous region, where again tectonics play a greater role (Fig. 26C).

Another basic element specified in the law of average areas is the average area of the first-order basins, which plays a significant role in discharge phenomena and relief evolution. In the mountainous region of the Ialomiţa system, for instance, the most frequent areas of first-order basins average 10 — 12 ha, while in the Subcarpathian regions this value is halved, first-order basins of 4 — 6 ha being quite frequent (Fig. 26). It is obvious that the small first-order basin area reflects the highly fragmented relief, with reduced length of slope runoff and reduced formation time for floods. These characteristics, given the fairly steep slopes and highly friable rock, contribute markedly to accelerating the evolution of Subcarpathian drainage basins.

Considering the verification of the above laws in the case of the fifth-order basin of the Azuga at its confluence with the Prahova or in the case of the latter before its junction with the Azuga (Fig. 26C, D), both the law of summed areas and the law of stream numbers are followed quite closely, while the ratio of the progression determined by the average areas resulting from the quotient of the first two ratios determines perfectly the line uniting the measured mean values.

The model for establishing average areas, which has been verified for a fairly large number of basins, gives a regression line which intersects the lines of stream numbers and summed areas (Fig. 25). Of the characteristic points of the resulting triangle, that determined by the intersection of the lines given by the laws of summed areas and of average areas is the most significant, because the abscissa of this point gives a value for the order of the drainage basin analyzed. Accordingly, under the physiographical conditions specific to the Prahova basin, which determine the ratios of the three progressions, the numbers of basins of lower orders correspond to the fifth-order system being only 75% completed; in the case of the Ialomicioara, the figure is 65%. In the former case, the actual order of the drainage basin is thus 4.75, and in the latter, 4.65 (Fig. 25C, D).

In order to investigate the stage of evolution of fragmentation, the partial relationships between drainage basins of successive orders can be analyzed. Their variation, considered in relation to the physiographical conditions, can also provide information concerning the latter's evolution over time.

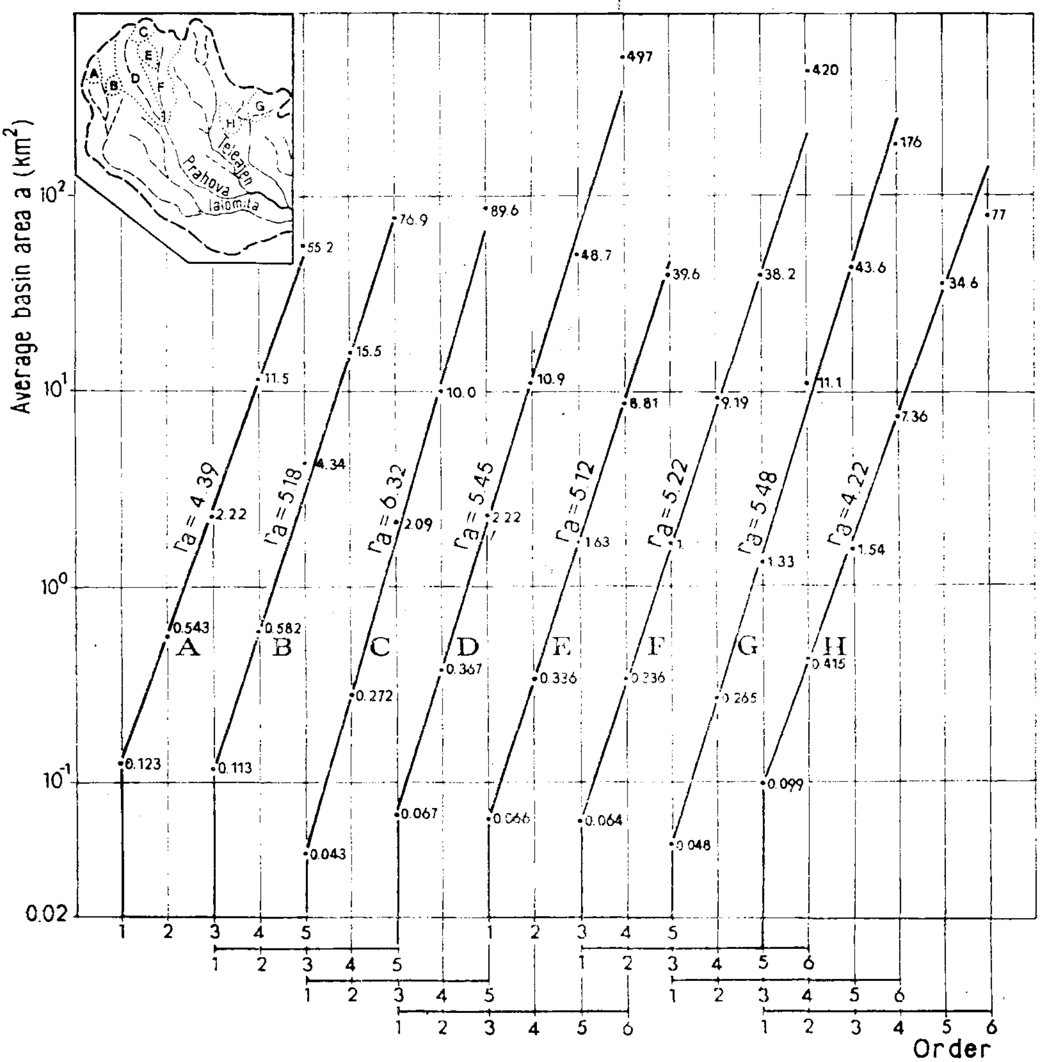



Fig. 26. Regression of average basin areas on order for the basins of (A) the Brătei upstream from its confluence with the Ialomiţa, (B) the eastern Ialomicioara at its confluence with the Ialomiţa, (C) the Azuga at its confluence with the Prahova, (D) the Prahova at its confluence with the Doftana, (E) the Prislop at its confluence with the Doftana, (F) the Doftana at its confluence with the Prahova, (G) the Cricovu Sărat at its confluence with the Lopatna, and (H) the Sărăţel at its confluence with the Lopatna.

*Interbasin areas.* Starting from research carried out by Schumm (1956) and Strahler (1964), orders may also be assigned to the areas between basins of various orders. Thus, in a drainage basin of a given order, after delimiting all first-order drainage basins, a number of areas will remain which lack gullies and which are drained by higher-order channels. Such areas may be termed interbasin areas of the first order ($A_{i1}$). In this case, the area of a fourth-order drainage basin will be given by

$$A_4 = \sum_{i=1}^{n} A_1 + A_{i1}$$

Likewise, on delimiting the second-order drainage basins, a second-order interbasin area will remain between them, drained only by higher-order channels. Within this area, first-order basins may exist. The same reasoning applies to a fourth-order basin; the areas between third-order basins become third-order interbasin areas ($A_{i3}$) being drained by streams of lower orders (first and second) which are nevertheless direct tributaries of the fourth-order main stream, and so on.

In practice, the various interbasin areas can be determined easily on the basis of the law of summed areas. To this end, it is noted

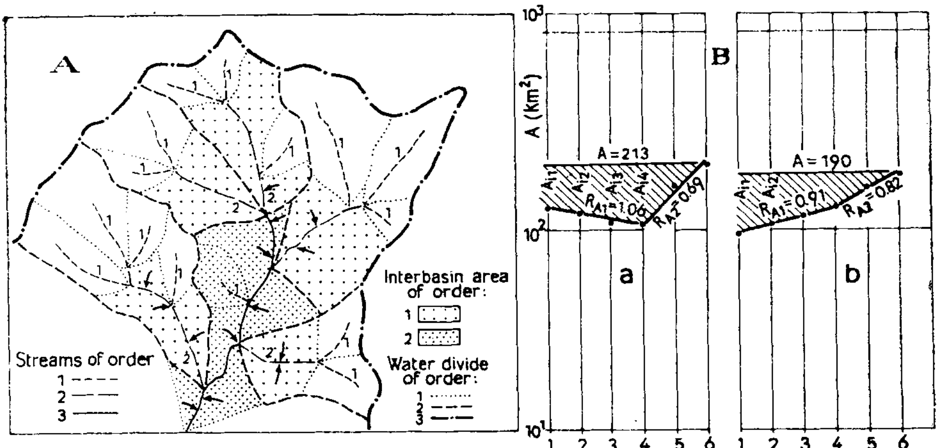



Fig. 27. Basin and interbasin areas : (A) distributions in a third-order basin ; (B) interbasin areas of successively higher orders for (a) the Vărbilău and (b) the Mislea, both at the confluence with the Teleajen.

that, for example, the sum of the first-order interbasin areas added to the sum of the drainage basin areas of the same order must equal the area of the whole drainage basin (Fig. 27B) or

$$A_{i1} = A_u - A_1, \quad A_{i2} = A_u - A_{i2}\ , \ldots, A_{i(u-1)} = A_u - A_{(u-1)}$$

It is thus obvious that the distribution of interbasin areas is closely related to the law of summed areas, and may be analyzed only in relation to it. Two characteristic situations can be singled out: first, cases in which the summed areas determine a single progression, which also defines a single progression for the interbasin areas, and second, cases in which the summed areas for successive orders yield two progressions, and the interbasin areas behave similarly (Fig. 27B). For the Ialomiţa system, this latter case is more frequent for higher-order basins in the Subcarpathian area, and seems to be related to accentuated relief fragmentation and to the increasing density of the drainage network following the transition to a new cycle of geomorphological evolution. In this respect, it is obvious that a decrease of the first-order interbasin area can occur only through the appearance of a number of new channels draining the area in question. Their area will thus be transferred from the first-order interbasin area to augment the first-order summed drainage basin area, and so on for the higher orders. This means that the points defining the law of summed areas will show increased ordinates, the magnitudes of the increases being inversely proportional to basin order. By virtue of the progressive reduction of the interbasin areas, two progressions for each drainage basin will be obtained at a given moment (Fig. 27B(a, b)). For the drainage basin of the Vărbilău stream, the drainage density is 4.55 km $km^{-2}$, while for the Mislea it is 3.98 km $km^{-2}$. It is evident that the lower the interbasin areas, the more fragmented and better drained the corresponding drainage basin, as happens in the Subcarpathian regions as compared to mountainous or plain areas in Romania.

It may be concluded that the areas of drainage basins of successive orders do not develop at random. Their evolution is governed by laws which can be ascertained and analyzed in relation to the aggregate of physiographical factors that determine the ways in which relief fragmentation occurs and in which a certain area is divided for drainage purposes among the streams of various orders. Further, the fact that drainage-basin areas show a tendency to reach a certain equilibrium indicates that they impart this behaviour to other phenomena to which they are related and which they determine.

Chapter V

# Perimeter of Drainage Basins

The perimeter of a drainage basin is defined as the horizontal projection of its water divide. It delimits the area of the drainage basin on the map, and is always smaller than the true length of the water divide; however, being determined more easily than the latter, it is always used for topographical purposes.

The water divide is the line linking the points of greatest height between two drainage basins, and separating their surface runoffs. It delimits the entire catchment area which is drained by the whole of a river network. Its parameters depend on the situation of a drainage basin within the major relief units. Thus, in mountainous areas, both the absolute and relative average heights of the water divide will be greater than in plain areas. To approximate the average height $\bar{H}$ of the water divide, one-half the sum of the average heights of the peak ($\bar{h}_p$) and of the saddles ($\bar{h}_s$) (Appolov, 1963) may be used:

$$\bar{H} = (\bar{h}_p + \bar{h}_s)/2 \tag{48}$$

where $\bar{h}_p = \Sigma h_p/n_p$ and $\bar{h}_s = \Sigma h_s/n_s$, $n_p$ being the number of peaks and $n_s$ the number of saddles.

*Mean slope of water divide.* As in the case of other length elements, the mean slope of a water divide may be estimated from the ratio of the divide's maximum height $H$, considered from the mouth to the highest point of the basin, to one-half its length:

$$I_P = 2H/P \tag{49}$$

Although strictly valid only for a single maximum in the perimeter altitude, this relationship is followed well by drainage basins of lower orders (first to fourth), since in this case the numbers of peaks and saddles are small. Thus, in the case of a first-order basin, there will be little difference between the water-divide slopes on the left- and right-hand sides, even if the basin is asymmetrical. In this case eqn. (49) can be used to calculate the average slope.

However, with increasing drainage-basin order, the regularity of the course followed by the water divide lessens. As a consequence, the lengths of the water divide on the two sides change in relation to the skewness of the drainage basin, to the numbers of peaks and saddles, to the slope, etc. In the case of high-order basins, the situation is complicated markedly by the existence of numerous saddles and peaks, which modify both the length and the slope of the water divide. In such instances it would be more appropriate to calculate the real length of the water divide on the basis of its longitudinal profile, i.e., taking into account the slopes of various sectors. However, if a certain approximation is admitted, eqn. (49) can be employed successfully.

Taking for instance the fourth-order drainage basin of the Bizdidel upstream from its confluence with the Fineasca, the following slope is obtained for the right-side water divide :

$$\tan \alpha_r = 516/9.4 = 55\ ‰$$

while on the left-hand side it is

$$\tan \alpha_l = 516/5.0 = 103\ ‰$$

The average slope of the whole perimeter is

$$\tan \alpha_P = 1032/14.4 = 72\ ‰$$

If the mean slope of the channel of the same drainage basin is calculated, it is found that the difference between this slope and that of the perimeter is only slight. Nevertheless, in most instances, the slope of the channel has a higher value than the slope of the perimeter of a given basin.

If the mean slope of a water divide is calculated as the ratio of the vertical distance between the extreme points to one-half the total length, the resulting value will be the more erroneous, the greater the height differences along the water divide. These differences must therefore be taken into account in more detailed studies.

*Law of summed perimeters.* The perimeter of a drainage basin depends on the latter's area and shape. If the perimeters of all drainage basins of a given order are calculated, their addition yields a sum of perimeters of the respective order. By repeating this operation for all orders and representing the values on semilogarithmic paper, it appears that an inverse relationship exists between basin order and the summed perimeter, to the effect that the higher the basin order, the smaller the sum of the perimeters. This relationship has been verified in a large number of cases, resulting in the following law : *the summed perimeters of drainage basins of successively higher orders tend to form a decreasing geometric series in which the first term is the summed perimeter of first-*

*order basins and the ratio is the ratio of successive summed perimeters* (Fig. 28). The ratios between the terms of successive orders are

$$P_1/P_2 = R_{P_1},\ P_2/P_3 = R_{P_2},\ P_3/P_4 = R_{P_3},\ \ldots,\ P_{u-1}/P_u = R_{P_{u-1}}$$

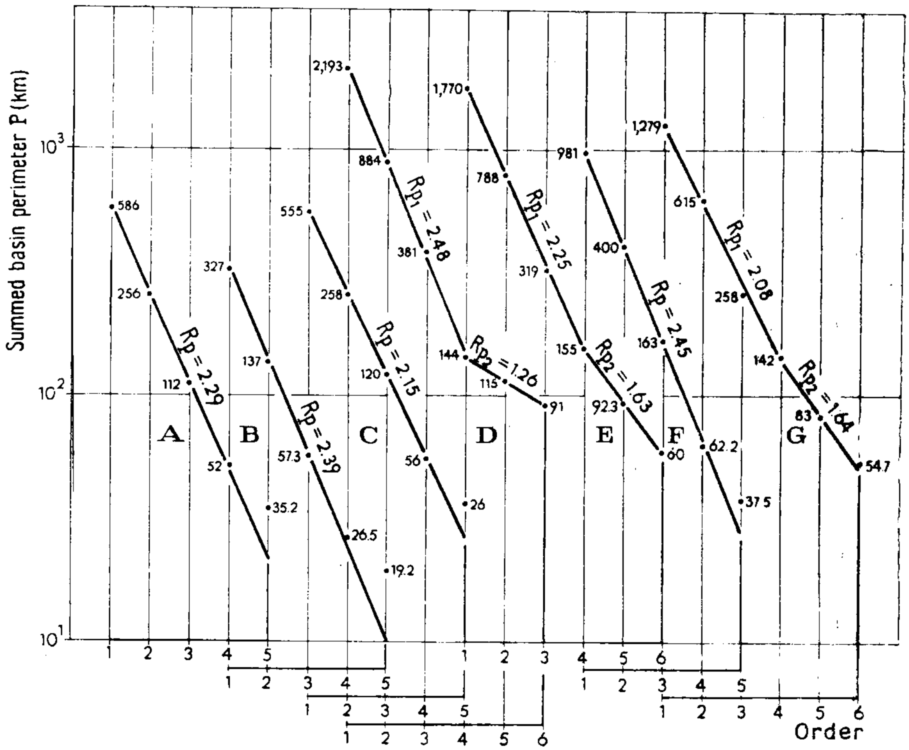



Fig. 28. Regression of summed basin perimeters on order for the basins of (A) the eastern Ialomicioara at its confluence with the Ialomiţa, (B) the Ruda at its confluence with the Cricovu Dulce, (C) the Bertea at its confluence with the Vărbilău, (D) the Vărbilău at its junction with the Teleajen, (E) the Mislea at its confluence with the Teleajen, (F) the Cricovu Sărat at its confluence with the Salcia, and (G) the Lopatna at its confluence with the Sărăţel.

The ratio of the geometric series may be obtained as the weighted mean of these partial ratios, or using one of the other methods mentioned previously.

Denoting the ratio of the series by $R_P$ and the first term by $P_1$, the term of rank $u$ will be

$$P_u = P_1/R_P^{u-1} \tag{50}$$

Since this decreasing geometric series has been well verified, there is again no need to rely on field measurements to obtain the value of $P_1$, which, like any other term, can be calculated on the basis of the formula

$$P_x = P_u \, R_P^{u-x} \tag{51}$$

or

$$\log P_x = F - Gx \tag{52}$$

where $F$ and $G$ are constants and $x$ is basin order.

In the case of the summed perimeters also, the value of the term of highest rank deviates from the law in the case of higher-order basins. This happens because higher-order basins generally develop on gentler slopes, which results in a greater number of crenulations of the perimeter, and hence in its elongation. Such a deviation can be perceived in the series of values obtained for the fifth-order basin of the Bertea river upstream from its confluence with the Vărbilău (Fig. 28C):

| | Order | | | | |
|---|---|---|---|---|---|
| | 1 | 2 | 3 | 4 | 5 |
| $P_{\text{meas.}}$ | | 258.4 | 120.3 | 55.9 | 36.5 |
| $P_{\text{calc.}}$ ($R_P$ = 2.15) | 555.5 | 258.4 | 120.2 | 55.9 | 26.0 |

The transition from mountainous and hilly regions to plain areas is also accompanied by an increase of the summed perimeters of drainage basins. Considering, for example, the case of the sixth-order basin of the Mislea upstream from its confluence with the Teleajen, or of the Lopatna upstream from its confluence with the Sărăţel (Fig. 28E, G), the values obtained delineate two lines, with different ratios. Thus, the subbasins of the first to fourth orders define a line with a much more marked slope compared to the line specific to the fifth and sixth orders.

In such instances, to avoid erroneous values for high-order basins, a second line with a different ratio must thus be established. Taking the case of the Vărbilău basin upstream from the confluence with the Teleajen, calculation of the values for the fifth and sixth orders using the equation given by the first to fourth fails to yield good results. A second line, with a different perimeter ratio specific only to the fifth and sixth orders, is necessary (Fig. 28D):

| | Order | | | | | |
|---|---|---|---|---|---|---|
| | 1 | 2 | 3 | 4 | 5 | 6 |
| $P_{\text{meas.}}$ | | 884 | 381 | 144 | 115.0 | 91 |
| $P_{\text{calc.}}$ ($R_P$=2.48) | 2 192 | 884 | 356.6 | 143.8 | | |
| $P_{\text{calc.}}$ ($R_P$=1.26) | | | | 143.8 | 114.1 | 90.6 |

Had the same ratio been applied to the last two orders, a perimeter of 23.4 km would have resulted for the sixth-order basin, which is utterly impossible given the area of this basin.

If necessary, the total perimeter ($\Sigma P$) for all the basins of successive orders within a given system can be determined using the formula

$$\sum_{i=1}^{u} P_i = P_u(1 - R_P^u)/(1 - R_P) \tag{53}$$

when there is a single line, or

$$\sum_{i=1}^{u} P_i = \sum_{i=1}^{k} P_i + \sum_{i=k+1}^{u} P_i = P_k(1 - R_{P_1}^k)/(1 - R_{P_1}) +$$

$$+ P_u(1 - R_{P_2}^{u-k})/(1 - R_{P_2}) \tag{54}$$

when two lines are delineated by the summed perimeters in a given system.

Generally, the law of summed perimeters applies differently in hilly and mountainous areas, where the relief energy is great and where the drainage basins, those of lower orders in particular, have well-defined perimeters without many crenulations. In contrast, in plain areas, where the relief energy is low and the water divide has more crenulations, drainage basins tend to acquire an elongated shape and to have lower summed-perimeter ratios.

*Law of average perimeters.* It has been seen that the summed perimeters of various orders form a decreasing geometric series whose terms can be calculated. It is known that the numbers of drainage basins of successive order also form a decreasing geometric series. The ratio of these two series yields a third, which represents the average perimeters of drainage basins of successive orders. Representation of the data on semilogarithmic paper thus yields an obvious dependence on order of the average perimeter of drainage basins (Fig. 29). The functional relationship between the two elements may be formulated as a law : *the average perimeters of drainage basins of successive orders tend to form an increasing geometric series in which the first term is the average perimeter $p_1$ of first-order basins and the ratio is the ratio $r_p$ of successive average perimeters.*

From the two preceding series, $P_1/N_1 = p_1 \ldots, P_u/N_u = p_u$. Substitution of the appropriate expressions for $P_u$ and $N_u$ (eqns. (50) and (27)) then yields

$$P_u/N_u = P_1 R_c^{u-1}/N_1\ R_P^{u-1}$$

Given that $P_1/N_1 = p_1$ and $R_c/R_P = r_p$, the general term of the average-perimeter series is obtained as

$$p_u = p_1\, r_p^{u-1} \tag{55}$$

Therefore, the ratio of the new series is given by the quotient between the ratios of the two previous series. The following formula can also be used to calculate the values on the ordinate :

$$\log p = f + gx \tag{56}$$

where $f$ and $g$ are constants and $x$ is drainage-basin order.

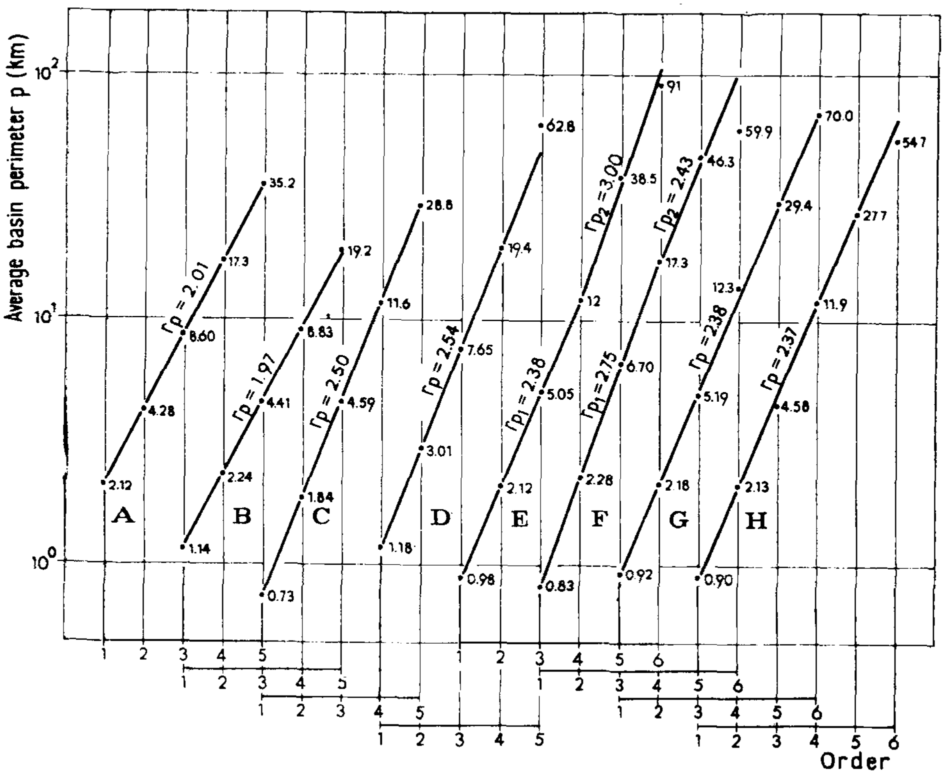



Fig. 29. Regression of average basin perimeters on order for the basins of (A) the eastern Ialomicioara at its confluence with the Ialomiţa, (B) the Ruda at its confluence with the Cricovu Dulce, (C) the Purcaru at its confluence with the Doftana, (D) the Teleajen at its confluence with the Telejenel, (E) the Vărbilău at its confluence with the Teleajen, (F) the Mislea at its confluence with the Teleajen, (G) the Cricovu Sărat at its confluence with the Lopatna, and (H) the Lopatna at its confluence with the Sărăţel.

Generally, the average perimeters measured for drainage basins of successive orders comply with the above law, the series ratio ranging from unity to the value of the confluence ratio. The most frequent values vary between 2.00 and 3.00 (Fig. 29). Note that the existence of a ratio of average perimeters equal to or smaller than unity is impossible, since, in this instance, a constant or decreasing series would result. A constant series would imply that all the basins of successively higher order have the same perimeter, which is impossible given the verified relationship between increasing drainage-basin area and perimeter. A decreasing series of average perimeters from lower to higher orders is again impossible in the classification system adopted, since the average first-order basin perimeter cannot be higher than the average second- or third-order basin perimeters.

The laws of the average and summed perimeters and their morphometrical models can be verified by graphical representation of the two series in relation to the law of stream numbers (Fig. 30), complemented

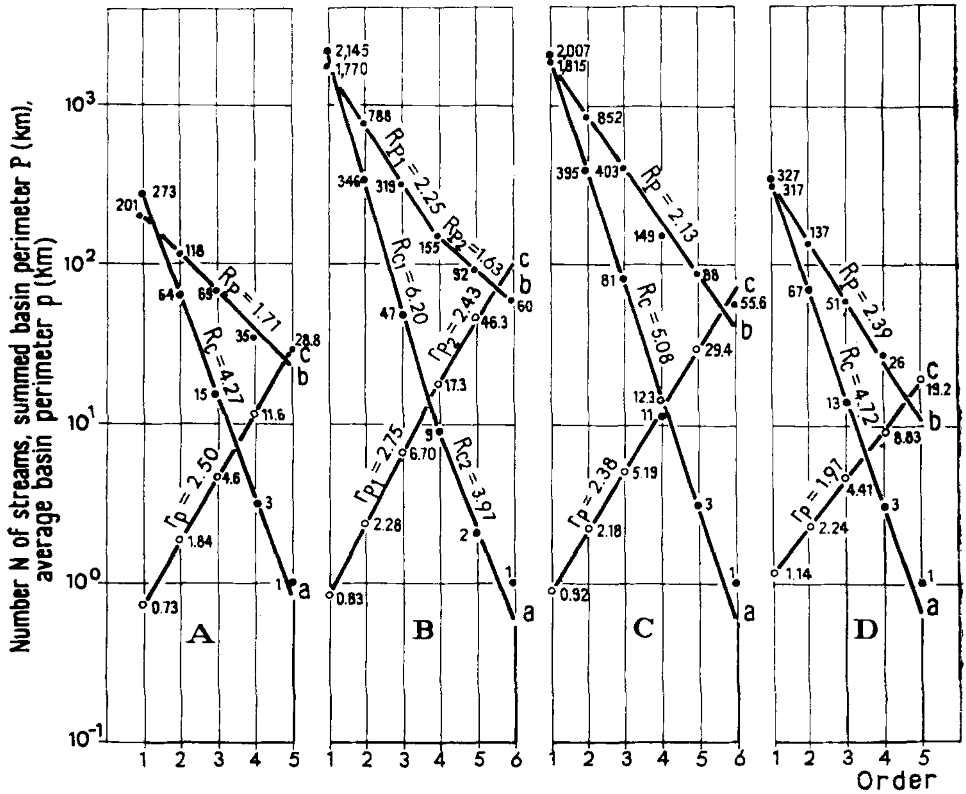



Fig. 30. Morphometrical models of basin perimeter for the basins of (A) the Purcaru at its confluence with the Doftana, (B) the Mislea at its confluence with the Teleajen, (C) the Cricovu Sărat at its confluence with the Lopatna, and (D) the Ruda at its confluence with the Cricovu Dulce (a, regression of numbers of streams ; b, regression of summed basin perimeters ; c, regression of average basin perimeters).

by analysis of individual values. Again, for drainage basins of successive orders, the two lines determined by the summed and average perimeters intersect at a point whose coordinates may be calculated. The abscissa of the point will define the order of a given drainage basin when the respective lines are well established.

Taking the drainage basin of the Purcaru stream, a left-side tributary of the Doftana (Fig. 30A), the values defining the two laws delineate the corresponding lines quite well, the slopes being related. The ratio of the series determined by the average perimeters is given by the quotient of the ratios of the stream-number series and of the series of summed perimeters of drainage basins of successive orders. The calculated parameters of the morphometrical model of perimeters for the Purcaru drainage basin are as follows (Fig. 30A):

| | Order | | | | |
|---|---|---|---|---|---|
| | 1 | 2 | 3 | 4 | 5 |
| $N_{meas.}$ | 273 | 64 | 15 | 3 | 1 |
| $N_{calc.}$ ($R_c$ =4.27) | 273 | 64 | 15 | 3.51 | 0.82 |
| $P_{meas.}$ | | 117.8 | 68.9 | 34.7 | 28.8 |
| $P_{calc.}$ ($R_P$=1.71) | 201.5 | 117.8 | 68.9 | 40.3 | 23.6 |
| $p_{meas.}$ | | 1.84 | 4.6 | 11.6 | 28.8 |
| $p_{calc.}$ ($r_p$ = 2.50) | 0.73 | 1.84 | 4.59 | 11.48 | 28.7 |

The series analyzed may delineate only one line (Fig. 30A, C, D), or two lines (Fig. 30B), in relation to the specific conditions in the basin considered which bear on the length of the perimeter. Generally, increasing perimeters are found in larger basins with a smaller relief energy : the presence of many crenulations increases the perimeter length. The same method is applied if the series of values indicate two different progressions within the same basin. In this case, the ratio of the series for the perimeters is first calculated for say the first- to fourth-order basins, and then for the higher orders. Single series are obtained for drainage basins situated in homogeneous physiographical conditions, as usually found for low-order basins. In the case of higher orders, it is very likely that a basin will have developed in two major relief units, and this will bear on the appearance of the corresponding lines. If a basin develops within a single relief unit, a single line will appear even if the basin is of high order. In the opposite case, the basin will have two lines which reflect the situations for the lower (first to fourth) and higher (fifth to eighth) orders (Fig. 28D, E, G, Fig. 29F).

If it is admitted that drainage basins of higher orders are the oldest, the law of summed perimeters can also provide information concerning the evolution of erosion and relief-fragmentation processes. This means that concomitant with increasing relief fragmentation by a river network, the numbers of basins of lower orders will increase. Hence, the sums of their perimeters also increase, the slope of the line becomes steeper, and the series ratio thus increases in value.

*Relationship between perimeter and drainage-basin area.* There is a relationship of direct proportionality between the increases in the areas of drainage basins of successive orders in their perimeters. More detailed analysis and correlation of the two elements show that their growth rates remain the same only when the drainage basins retain the same form factor from the source to the mouth of a given system. It is therefore necessary that geometrical similarity should exist between them. In reality, the shapes of basins of successive orders remain the same only in very rare instances. The shape changes from the source to the mouth to the extent to which a basin receives new tributaries of different sizes and shapes. An important role in the modification of drainage-basin shape, and hence in the growth rate of the perimeter in relation to area, is played by rock type, structure and tectonics, to which should be added the role of paleogeographical evolution.

Given that the relationship of proportionality is valid only for basins with the same form factor, irrespective of their size, these two elements may be correlated to establish their functional relationship quantitatively. This has been done for drainage basins with form factors ranging from 0.750 to 0.800, giving the relationship between perimeter $P$ and area $A$ as

$$P = k\,A^n \tag{57}$$

Since this is the equation of a parabola, the data may be represented on logarithmic paper to give a straight line (Fig. 31), the equation of

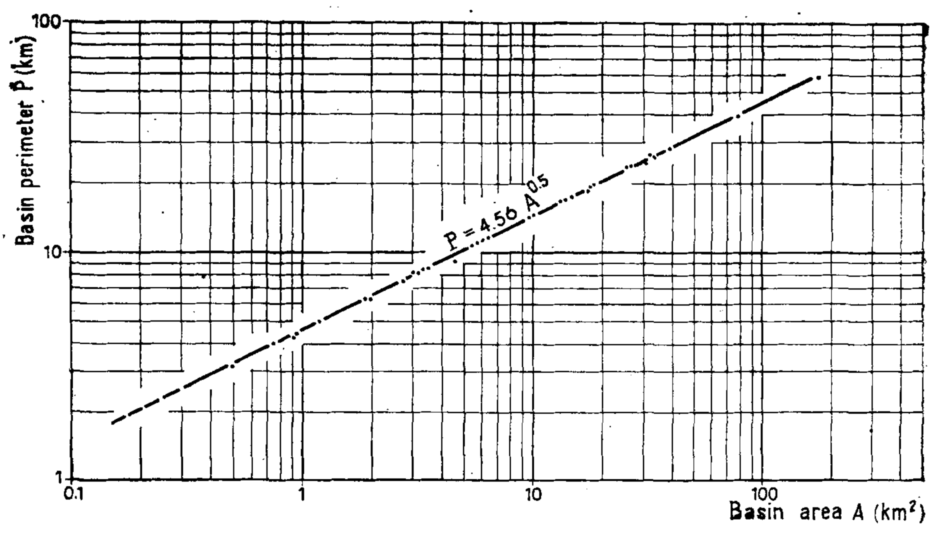



Fig. 31. Regression of basin perimeter on basin area for form factors in the range 0.75—0.80.

which can be calculated quite easily by the chosen-points method (Bloh, 1971) or by computing the parameters $n$ and $k$ from the relationships

$$n = (\log y_2 - \log y_1)/(\log x_2 - \log x_1)$$

and

$$k = y_1/x_1^n \text{ or } \log k = \log y_1 - n \log x_1$$

Substitution of the appropriate values into the foregoing relationships yields the result

$$P = 4.56 \; A^{0.5} \text{ or } P = 4.56 \sqrt{A}$$

From the dimensional viewpoint, the 0.5 exponent of area implies that $k$ is nondimensional: this is thus a case of isometry (Church and Mark, 1980). It is thus quite obvious that for basins having the same form factor there is a very close relationship between increases in area and in perimeter.

On the basis of the foregoing considerations, a general formula was sought to express the relationship between the perimeter of a drainage basin and the area it delimits, including the form factor. Values for a large number of cases were first represented on logarithmic paper taking this factor as a parameter. Finally, the relationship between the three variables was established as

$$P = 4A^{0.5} \; R_f^{-0.5} \text{ or } P = 4 \sqrt{A}/\sqrt{R_f} \tag{58}$$

Certain difficulties are encountered in drawing a nomogram to establish the perimeter as a function of area and form factor simultaneously, since if scales with the same moduli for area and perimeter are used, a very small increment is obtained for the form factor. In order to surmount this shortcoming, a nomogram with logarithmic scales having different moduli (bases) for the dependent and independent variables may be drawn, giving an increased variation of the form factor. By employing this type of graphical representation, the relationship between the variables mentioned, which using equivalent scales is almost indiscernible from the plotted curves, can easily be determined.

## Relationship between Average Perimeter and Average Area of Drainage Basins

If the average perimeters of drainage basins of various orders are represented on logarithmic paper in relation to the average areas of the corresponding orders, a series of straight lines are obtained relating the two variables according to the general expressions $p = ka^n$. The constant

$k$, determined for a fairly large number of curves, has values very close to 4 (Fig. 32). Values around 0.5 are found for the exponent $n$, indicating isometry from the dimensional viewpoint. Since an increase in drainage-basin area results in a proportionate increase in perimeter, this allometric relationship may be established quantitatively.

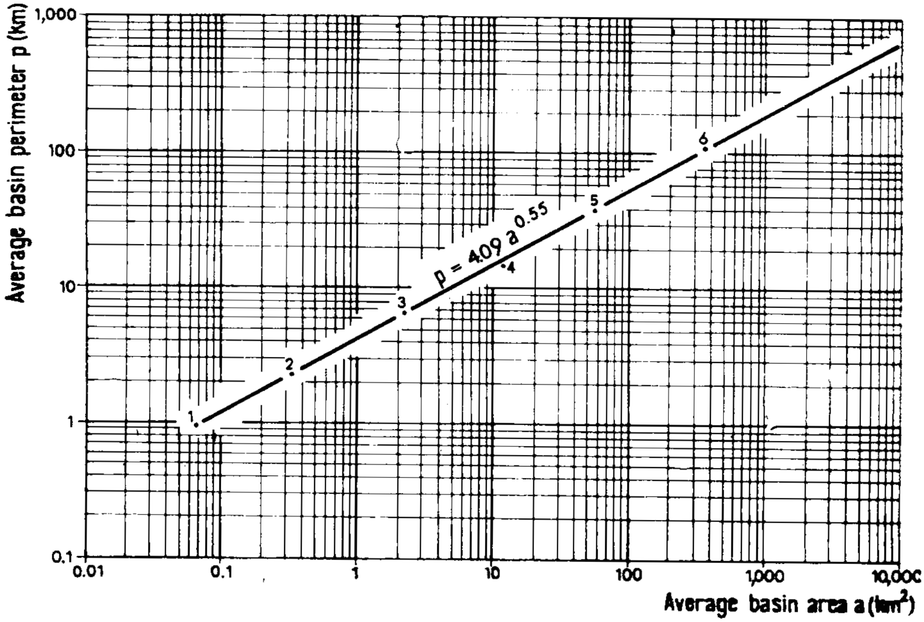



Fig. 32. Regression of average basin perimeter on average basin area for 16 sixth-order basins within the Ialomiţa drainage system.

As an example, the appropriate values were calculated for 16 sixth-order basins situated in both mountainous and hilly regions. These values were represented graphically, and, using the methods mentioned previously, the following relationship was established:

$$p = 4.09\ a^{0.55}$$

That the exponent is greater than 0.5 is a result of the specific conditions in the region analyzed. Thus, whereas for small values of area there is a direct relationship between the two elements, and the form factor tends to approximate unity, with increasing area a more marked increment of the perimeter of higher-order drainage basins occurs. The cause is the occurrence for larger basins of a greater number of crenulations in the perimeter, which increase its length. Furthermore, in a region with obvious tiering of the relief, drainage basins tend to be elongated as the outcome of paleogeographical evolution and of the present specific conditions. The data analyzed confirm the fact that while lower-order basins tend to have a form factor equal to unity, those of higher orders are increasingly elongated, and their form factor decreases accordingly.

## Length and Width of Drainage Basins

The notions of length and width are used frequently in assessing the size of a drainage basin.

*Drainage-basin length.* Although length is an important element in characterizing the size of a drainage basin, its use as a parameter may lead to loss of accuracy, because the method of establishing it is approximate in nature, the value obtained being related to the shape and size of the basin and to the scale considered.

Horton (1932) defined basin length as the straight-line distance from a basin's mouth to the point on the water divide intersected by the projection of the direction of the line through the source of the main stream. Ongley (1968) stated that it is difficult to draw such a line on the basis of the Horton—Strahler or Shreve classification systems, or in the case of highly asymmetrical basins, because tributaries may be longer than the main stream. Ogievsky (1952) showed that basin length could be established by measuring a line drawn from the mouth to the most distant point of the drainage area, on condition that it passes through the midpoints of lines drawn across the basin.

Schumm (1956) considered that the maximum length of a drainage basin should be measured parallel to the main drainage line. However, Maxwell (1960) showed that establishment of the maximum length has certain shortcomings. In order to avoid them he introduced the concept of drainage-basin diameter. This is defined as the length of the horizontal projection of a line drawn parallel to the drainage line starting from the basin mouth and intersecting the water divide by projection through the source of the main stream, such that (a) the main channel is divided into segments whose summed lengths on the opposite sides of the line are approximately equal, (b) the line should be parallel to that separating the slopes of the opposite sides of the basin, (c) it should divide the drainage-basin area approximately equally, and (d) it should be the greatest such diameter. Some compromise among these criteria may be necessary for highly irregular basins.

Apollov (1963) showed that if a drainage basin has a regular shape, its length can be evaluated on the basis of the straight line drawn from the mouth to the most distant point on the water divide. If a drainage basin is irregular, a unique characteristic basin length cannot be obtained. In such instances, the author suggests determination of the median. To this end, a celluloid disk may be employed which has a central orifice around which a number of concentric circles are drawn, with successive radii differing by a constant amount, say 1 cm. By placing the disk on a map of the drainage basin such that two opposite sides of the basin touch or are at equal distances from the same circle, various points may be marked through the central orifice which indicate the midpoints of the diameters considered. When a sufficiently large number of points have been marked, the best-fitting line drawn through them gives the median.

Ongley (1968), after pointing out the limitations of the various parameters defined above, proposed an objective method for establishing the length of a drainage basin, based on its vectorial axis. The premise of the method is that this axis is determined by the hydraulic and geometrical characteristics of a Horton—Strahler river network whose evolution is not influenced by tectonics or variations in lithology. To establish the vectorial axis of a drainage basin of order $x$, the initial and final points of the main stream (of order $x$) are linked by a straight line, giving a vector $S_x$. The same method is used to determine the individual vectors for the various subbasins of order $x-1$. The equivalent vector for the streams of this order is then obtained as the resultant, and the vectorial axis by combining the vectors $S_x$, $S_{x-1}$ graphically or trigonometrically in a vector diagram, using rectangular coordinates in which the $y$ direction represents north. The resulting vectorial axis starts from the basin mouth at a certain azimuth, and includes all the vectors for the rivers of order $x$ and $x - 1$ (Fig. 33). It is both a hydra-

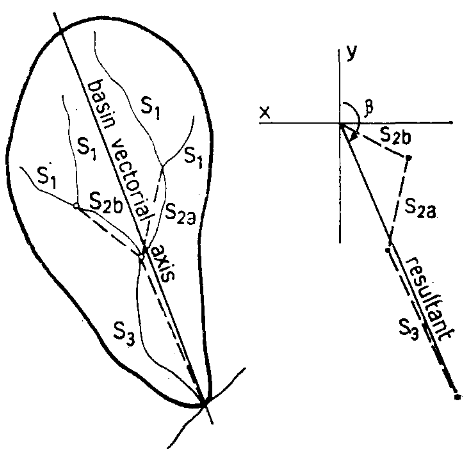



Fig. 33. Determination of basin vectorial axis using Ongley's method (1968).

ulic and a morphological axis of the drainage basin, since the computations involve the lengths and directions of tributaries of order $x-1$ (Ongley, 1968).

Differences are sometimes found in the basin lengths determined according to the method developed by Schumm and using the median. In most instances, the values obtained on the basis of the maximum length are greater than those using the median. Besides maximum basin length, the concept of average basin length is also used in the specialized literature. This is generally obtained by computation, in conjunction with the average basin width (see below).

*The average width* $B$ of a drainage basin may be calculated as the ratio of the area $A$ and the maximum length $L_m$, considered as the distance from the mouth to the most remote point on the water divide in the direction of the main stream, or the length of the median (Chebotarev, 1953):

$$B = A/L_m \tag{59}$$

In certain cases it may also be necessary to determine the average widths of the two sides of a basin, for which Apollov (1963) recommended the formulae $B_l = A_l/L_b$ and $B_r = A_r/L_b$, for the left- and right-hand sides respectively, where $L_b$ is the basin length.

Using data for a large number of rivers in the USSR, Sokolov (1962) established that basin width $B$ (determined as the ratio of basin area $A$ and the length $L$ of the main stream) and area $A$ are related according to $B = 0.32\ A^{0.5}$ for drainage basins larger than 250 km$^2$, and $B = 0.65\ A^{0.5}$ for smaller basins. This indicates that the latter basins are almost twice as wide as medium-sized or large ones.

*Average length and width.* The average length $\bar{L}$ and average width $\bar{B}$ of a drainage basin are calculated starting from the two basic elements of area $A$ and perimeter $P$. Considering the shape of a basin to be equivalent to a rectangle, the two sides are given by the roots of a second-degree equation in which the product of the roots is equal to the basin area ($A = \bar{L}\bar{B}$) and their sum is one-half the perimeter ($P/2 = \bar{L} + \bar{B}$). The equation giving these roots can be written as

$$x^2 - Px/2 + A = 0$$

from which the two sides can be derived (Alexeyev, 1975; Zăvoianu, 1978) as

$$\bar{L} = P/4 + \sqrt{(P/4)^2 - A} \quad \text{and} \quad \bar{B} = P/4 - \sqrt{(P/4)^2 - A} \tag{60}$$

In this case, in addition to establishing the length and width, a brief evaluation can be made of the shape of the basin in relation to the value under the root sign. If this is higher than zero, the basin is elongated. In fact, most drainage basins fall in this category. When the values under the root sign is equal to zero, the length of the basin is equal to its width, and the basin shape is equivalent to a square. If a value smaller than zero is found under the root sign, the shape of the basin will be closer to a circle. In this case, the length may be taken as the side of a square with the same perimeter as the basin. For cases when $A > (P/4)^2$, Alexeyev (1977) proposed the following relationship for calculating the average length $\bar{L}$:

$$\bar{L} = 4(A/P) \tag{61}$$

The maximum length of a drainage basin and the average length thus calculated are related in a way which reflects the tendency of the former to have a greater value. A smaller difference is found between the mean and median values.

The average length of a drainage basin thus computed correlates quite well with the length of the main stream, these two elements being directly related. The same holds for generally symmetrical basins, whose area increases in direct proportion to the length of the main stream. Deviations of points from the line occur if there are sudden increases of drainage-basin area due to the contributions of major tributaries.

Chapter VI

# Basin Shape

Assessment of a basin's shape or outline as being elongated, oval, fan-like or circular is only qualitative. Such an assessment can be used to explain the unfolding of certain hydrological processes, but remains subjective so long as it is not parallelled by a quantitative assessment. This is the more so since a series of morphometrical parameters and even the way in which floods are formed and move depend on basin shape. It is known that floods are formed and travel more rapidly in a round basin than in an elongated one, and moreover that floods in basins of the former type are stronger and have a higher velocity, and thus greater erosion and transport capacities. As a consequence, the suspended load is greater, and the evolution of such drainage basins is thus more rapid. This is illustrated by the differences between the hypsometric curves for two basins of equal area situated in homogeneous physiographical conditions but having differing forms: the volume under the topographical surface is smaller in the case of round basins. An elongated shape favours a diminution of floods because tributaries flow into the main stream at greater intervals of time and space.

Given the importance of basin shape, numerous researchers have sought, relying on the principle of geometrical similarity, to replace a number of descriptions such as "oval", "square" or "oblong" by more precise indices that can be used in mathematical formulations. Lee and Sallee (1970) showed that such an index must have three properties: (1) each shape should have a single index; (2) two different forms should never have the same index; and (3) two similar forms should be characterized by the same index. Attempts to describe basin shape and relate it to hydrological processes have been numerous, but only the most important work is referred to here.

*The form factor.* Horton (1941) considered that a normally developed basin should be pear-shaped, this being a sign that the basin has resulted from erosion processes on an initially inclined surface. A quantitative description of basin shape was developed by Horton as early as 1932, with the introduction of the form factor $R_f$, which represents the dimensionless ratio of the area $A_b$ of a drainage basin to the square of its maximum length $L_m$:

$$R_f = A_b/L_m^2 \qquad (62)$$

The form factor is equal to unity when the basin shape is a square, and decreases according to the extent of elongation. For a basin whose shape approximates a circle, the ratio is higher than unity, tending to the theoretical value of $4/\pi = 1.273$ for a perfect circle, although this is never found under natural conditions. Nevertheless, if a drainage basin does closely resemble a circle, its maximum length would normally be considered as being approximately equal to the circle diameter. This may be remedied by using a function which considers the drainage-basin length as being equal to the side of a square having the same perimeter as the basin (see Table 5 below).

*The compactness coefficient.* This coefficient ($m$) has been used in the Soviet specialized literature to describe drainage-basin shape. It represents the ratio of the actual basin perimeter $P$ to the perimeter $P'$ of a circle of equal area (Luchisheva, 1950):

$$m = P/P' \text{ or } m = 0.282\,P/\sqrt{A} \tag{63}$$

This coefficient is equal to unity when the basin shape is a perfect circle, increasing to 1.128 in the case of a square, and may exceed 3 for a very elongated basin (Table 5).

Other specialized Soviet work has used a ratio resembling Horton's form factor to express basin shape. This ratio is given by $F_t = A/L^2$, where $A$ is the basin area and $L^2$ the area of a square with a perimeter equal to that of the basin (Ogievsky, 1952; Apollov, 1963).

*Circularity ratio.* Miller (1953) introduced the circularity ratio $RC$, which represents the quotient between the area $A_b$ of a basin and the area $A_c$ of a circle whose circumference is equal to the basin perimeter:

$$RC = A_b/A_c \tag{64}$$

To establish this ratio, the area and perimeter of a drainage basin must thus be measured. The ratio is equal to unity when the basin shape is a perfect circle, decreasing to 0.785 when the basin is a square, and continues to decrease to the extent to which the basin becomes elongated. The circularity ratio can be derived directly from the formula

$$RC = 4\pi A/P^2 \tag{65}$$

where $P$ is the drainage-basin perimeter. In practice, the circularity ratio is never equal to unity, because the variation of physiographical factors and the general relief slope always impart an elongated shape to drainage basins.

Singh (1970) showed that circularity bears an inverse relationship to basin area, but the correlation is not strict because the circularity ratio is influenced by the slope, relief, structure and tectonics of geological formations. The only significant correlation established between

TABLE 5

**Limits of variation of form indices**

| No. | Name of index | Originator | Formula recommended | Limits of variation | | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|
| | | | | Circle | Square | Rectangles | | | | |
| | | | | | | $L = 2B$ | $L = 5B$ | $L = 10B$ | $L = 20B$ | $L = 40B$ |
| 1 | Form factor | Horton (1932) | $R_f = A_b/L^2_{max}$ | 1.277 | 1.000 | 0.501 | 0.200 | 0.100 | 0.050 | 0.024 |
| 2 | Compactness coefficient | Luchisheva (1950) | $m = 0.282 \frac{P}{\sqrt{A}}$ | 1.000 | 1.128 | 1.198 | 1.514 | 1.971 | 2.649 | 3.700 |
| 3 | Form factor | Ogievsky (1952), Apollov (1963) | $F_f = A/L^2$ | 1.277 | 1.000 | 0.501 | 0.200 | 0.100 | 0.050 | 0.024 |
| 4 | Circularity ratio | Miller (1953) | $RC = A_b/A_c$ | 1.000 | 0.785 | 0.696 | 0.436 | 0.257 | 0.142 | 0.073 |
| 5 | Elongation ratio | Schumm (1956) | $R_e = D_c/L_b$ | 1.275 | 1.128 | 0.799 | 0.505 | 0.357 | 0.252 | 0.176 |
| 6 | Lemniscate ratio | Chorley et al. (1957) | $R_l = P/P_m$ | 1.000 | 0.897 | 0.866 | 0.894 | 0.927 | 0.959 | 0.978 |
| 7 | Elongation degree | Diaconu and Lăzărescu (1965) | $R_a = A/L$ | 1.130 | 1.000 | 0.708 | 0.447 | 0.316 | 0.223 | 0.156 |
| 8 | Form factor | Zăvoianu (1978) | $RF = 16A/P^2$ | 1.277 | 1.000 | 0.877 | 0.555 | 0.328 | 0.181 | 0.093 |

this coefficient and any of the morphometrical parameters discussed in other Chapters concerns stream occurrence, with which the circularity ratio shows a negative correlation (correlation coefficient −0.52) (Singh, 1970).

The correlations of the circularity ratio with the form factor, calculated using the maximum basin length, the median length, the length of the main stream and the average length, are revealing. Marked scatter of the points is evident in the first three cases (Fig. 34A – C), whereas perfect correlation is found in the fourth (Fig. 34D). This de-

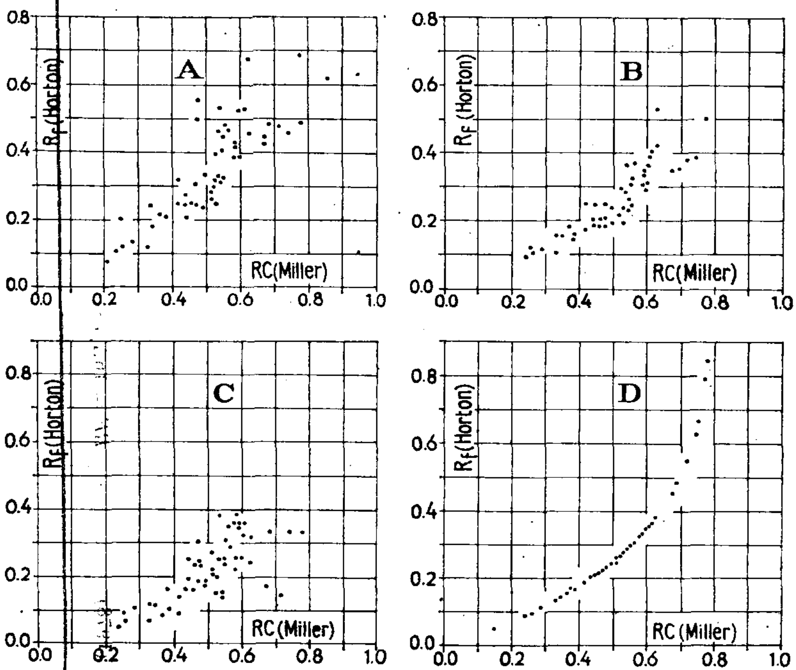



Fig. 34. Relationships between form factor $R_f$ and circularity ratio $RC$ established using (A) the maximum and (B) the median basin length, (C) the length of the main stream, and (D) the average basin length calculated from eqn. (60).

monstrates the superiority and greater objectivity of the average length calculated according to eqn. (60). In all of these correlations, the smallest values correspond to very elongated basins and the highest to basins whose shape resembles a square or circle.

*Elongation ratio.* Schumm (1956) proposed an "elongation ratio" $R_e$ to characterize basin shape, given by the quotient of the diameter $D_c$ of a circle of area equal to that of the basin, and the maximum basin length $L_b$ measured parallel to the axis of the main stream :

$$R_e = D_c/L_b \tag{66}$$

Taking $A = \pi D^2/4$, from which $D = (4A/\pi)^{1/2}$, the elongation ratio can also be written (Seyhan, 1977a) as

$$R_e = 2(A)^{1/2}/\pi^{1/2}\, L_b, \text{ or } R_e = 1.129\, \sqrt{A}/L_b \tag{67}$$

Seyhan (1975, 1976) found that rainfall and runoff were better correlated with the elongation ratio than with the form factor. The value of the former ratio varies from 1.275 when the basin shape is a circle, to 1.128 when it is a square, and decreases in proportion to increasing elongation, reaching a minimum of approximately 0.200. In calculating this ratio, the points show great scatter if the maximum length is computed as indicated by Schumm. However, if the average basin length is employed, the relationship between the circularity ratio and the elongation ratio is well established (Fig. 35a).

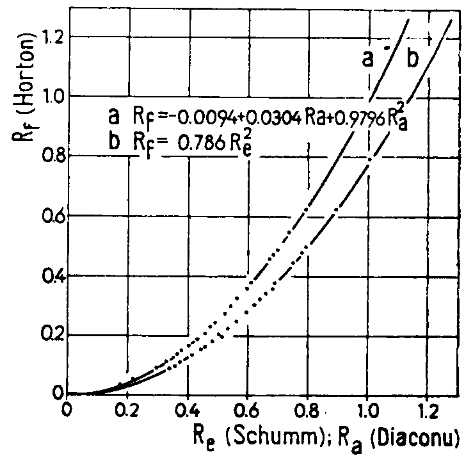



Fig. 35. Relationships between form factor $R_f$ proposed by Horton (1941) and elongation ratios proposed by (a) Schumm (1956) ($R_e$) and (b) Diaconu (1965) ($R_a$), calculated using the average basin length.

*Lemniscate ratio.* Starting from Horton's observation that the "ideal" form of a drainage basin developed on initially sloping ground closely resembles a "pear" or "drop" shape, Chorley et al. (1957) and subsequently Morisawa (1958) and Strahler (1964) noted that there is a close resemblance between such shapes and lemniscate curves. Thus, the polar form of the lemniscate equation is chosen to express the variation of basin shape :

$$\rho = L_b \cos p\, \theta \tag{68}$$

where $\rho$ is the radius from the outlet to the rim, $\theta$ the angle between the baseline and the radius under consideration, and $L_b$ the greatest diameter, analogous to the drainage-basin length, which becomes equal to $\rho$ when $\theta$ is zero. The coefficient $p$, given by the relationship

$$p = L_b^2\, \pi/4A \tag{69}$$

and termed the *rotundity coefficient*, itself provides an indication of the shape of a basin, in that when it is equal to unity the shape is a perfect circle, the value increasing to 1.27 in the case of a square basin and to 10—15 in the case of very elongated basins.

Using the value of the rotundity coefficient, it can be established that the perimeters $P$ of ideal lemniscate curves constitute a complete elliptical integral denoted by $E(K)$, which can be resolved using the tables adopted by Weast (1964), where $K$ is given by

$$K = \sqrt{(p^2 - 1)}/p$$

In practice, as shown by the quoted authors, the rotundity coefficient $p$ is first calculated, followed by $K$ and the value of arcsin $K$ (i.e.,$\sin^{-1}K$). Then the value of $E(K)$ is extracted from the tables, and the lemniscate perimeter is obtained as

$$P = 2L_b E(K) \tag{70}$$

Finally, the lemniscate ratio $R_1$ is obtained as the ratio between the lemniscate perimeter $P$ corresponding to the length and area of the basin analyzed, and the actual perimeter $P_m$. This ratio indicates the extent to which the shape of a basin approaches that of an ideal lemniscate. For a circle, its value is equal to unity, since for $p = 1$, $K = 0$, while $E(K)$ has a maximum value of 1.5708. To the extent to which the basin shape differs from a circle, the ratio decreases to approximately 0.86, after which the value increases again to reach unity for very elongated basins. This happens because for high values of the rotundity coefficient both $K$ and $E(K)$ tend to approximate unity, and the perimeter depends only on twice the length $L_b$. This inconvenience, the small range of variation of the lemniscate ratio and the more complicated calculations explain why this index is less used.

*Degree of elongation.* According to Diaconu and Lăzărescu (1965), drainage-basin shape can also be defined using various ratios between length $L$, average width $B$ and area $A$. Thus, the ratio $B/L$ indicates the degree of elongation to the effect that when values above unity are obtained, the shape of the basin becomes fanlike. As concerns the ratios $B/\sqrt{A}$ and $\sqrt{A}/L$ (note that $\sqrt{A}$ represents the side of a square of area equal to that of the basin), values below unity characterize elongated basins while those above unity, round basins. The ratio $\sqrt{A}/L$ amounts to 1.130 for a perfect circle, decreasing to unity for a square and to 0.156 in the case of very elongated basins whose average length is some 40 times greater than their average width (Table 5).

Tolentino et al. (1968) used a draingae-basin form index given by

$$k = P/2\sqrt{\pi A} \tag{71}$$

In 1976, W. L. Magette, V. O. Shanholtz and J. C. Carr (quoted by Seyhan, 1977a) gave the same relationship, terming the index the *compactness coefficient*. This expression is also identical to one being used in the Soviet literature as early as 1950 (Luchisheva, 1950) (Table 5).

From an analysis of all the indices which have been used to characterize drainage-basin shape, it may be observed that :

(a) all are dimensionless numbers deriving from ratios whose components have dimensions of length ;

(b) none is perfectly correlated with the others if it is calculated using the maximum or median basin length ;

(c) all are correlated well and relationship of correspondence between them can be established if they are calculated using the average basin length calculated according to eqn. (60) ;

(d) as reference figure, a circle or a right-angled quadrilateral equivalent are used.

On the basis of the foregoing considerations, it results that two major elements should be taken into account in describing the shape of a drainage basin : its area, and the length of the perimeter delimiting it. Application of the indices analyzed to a large number of basins has indicated a great variety of shapes, this variety being partly responsible for the great number of attempts to evaluate basin shape as faithfully as possible. In almost all cases, the area of a drainage basin can be considered in terms of an equivalent quadrilateral, or, more rarely, an equivalent circle. Chorley et al. (1957) showed that no real basin ever has the shape of a circle and, furthermore, that basins manifest no tendency to approximate this form. The circularity ratio thus provides a limited indication of the shape of a drainage basin, since most basins are elongated. It has been verified that the higher the order of a drainage basin, and hence the greater its area and perimeter, the greater the probability that it will be elongated. Concomitantly, the lower the order, the closer the shape of a basin to a square, oval or circle, the perimeter being straighter and having fewer crenulations. Given the general lack of resemblance to a circle, it is perhaps more appropriate to take as the reference figure a square whose perimeter is equal to that of the basin. Then, the ratio of the area $A$ of the basin to that of the reference square, given by $(P/4)^2$, will give a form factor with the square as reference (Zăvoianu, 1978) :

$$RF = A/(P/4)^2 \text{ or } RF = 16A/P^2 \tag{72}$$

This index is very close to that established by Horton, for which reason we have preserved the same name. The difference between them consists in the fact that the denominator is here considered to be the area of a square whose perimeter is equal to that of the basin, instead of the square of the maximum length.

Given the interdependence between the three variables $R_f$, $A$ and $P$, a nomogram can be drawn to establish the form factor more easily

(Fig. 36). Take a basin with an area of 280 km² and a perimeter of 110 km. To determine the form factor, a perpendicular line is drawn from 280 on the abscissa until it meets the value of the perimeter (110 km), and then, by constructing a horizontal line, the form factor (0.36) is found on the ordinate.

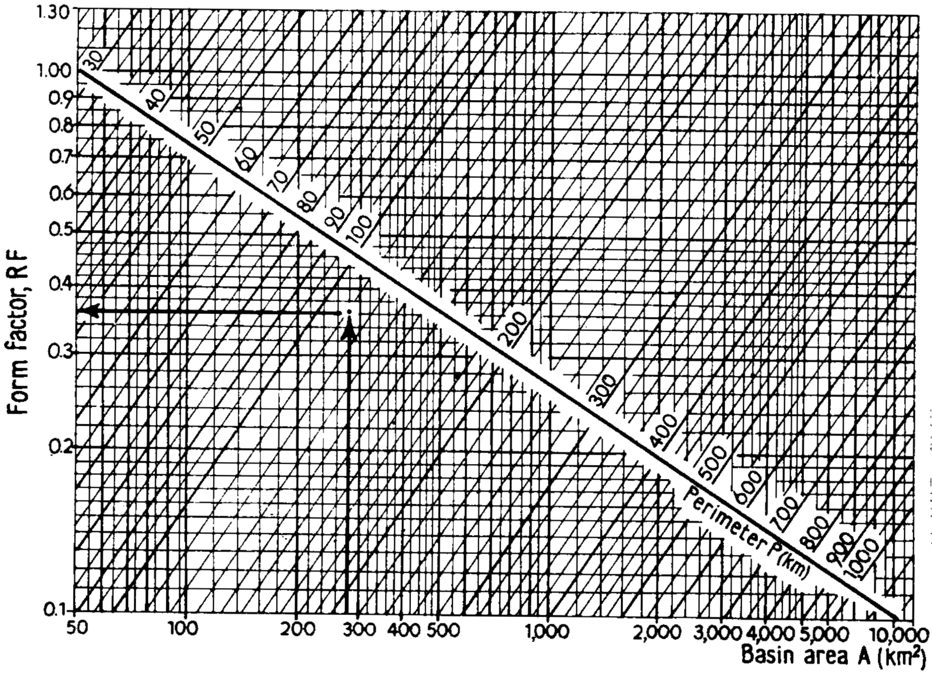



Fig. 36. Nomogram of form factor $RF$ proposed by Zăvoianu (1978) as a function of basin area $A$ and perimeter $P$.

The form factor in the present version is equal to unity when the basin shape is a square, and decreases to the extent to which the basin analyzed is more elongated. For round basins, the ratio becomes greater than unity, reaching 1.27 in the case of a perfect circle. There is a well-defined relationship between this ratio and that suggested by Miller, which allows one index to be inferred from the other (Fig. 37B) :

$$RC = 0.784\ RF \quad \text{or} \quad RF = 1.274\ RC \tag{73}$$

The present factor correlates well with the elongation ratio proposed by Schumm (1956), with the compactness coefficient, and with the form factor recommended by Luchisheva, etc., when the drainage-basin perimeter is considered in the computations (Fig. 37A).

The values established for 53 fifth-order basins yield an almost normal distribution around the mean value of 0.649. This is also evidenced by the frequency histogram and the probability curve, which has

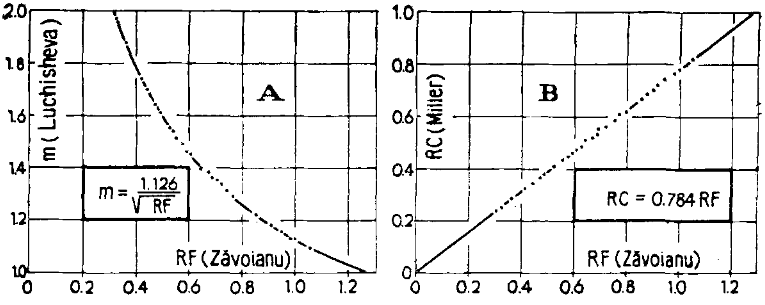



*Fig. 37.* Relationships between (A) compactness coefficient $m$ and form factor $RF$, and (B) circularity ratio $RC$ and form factor $RF$.

a coefficient of variation of 0.25 and a skewness coefficient of −0.08 (Fig. 38B). For a sample of 229 fourth-order drainage basins the mean value is 0.752, while the distribution of values around the average presents greater asymmetry than in the first case; thus, while the coefficient of variation has a value (0.23) very close to that for the previous sample, the skewness coefficient increases to −0.43 (Fig. 38A).

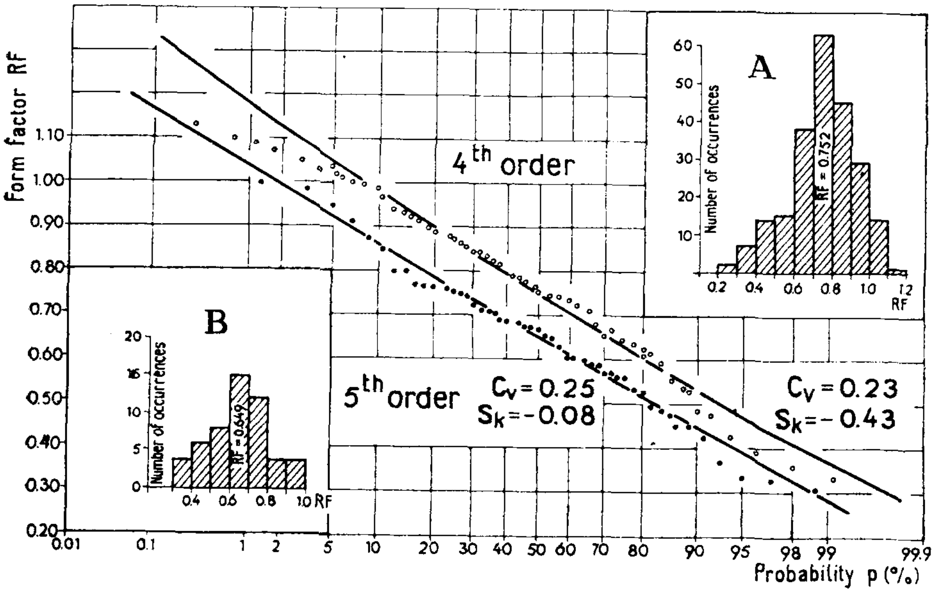



**Fig. 38.** Probability curves for form factor $RF$ and frequency distributions around the mean for (A) fourth-order and (B) fifth-order basins.

The variation of the physiographical conditions in a region plays an important role in determining basin shape. Thus, streams descending from a mountainous zone to hilly or plateau regions generally have more elongated basins compared to those whose sources lie in the latter two relief units. This proves, as stated by Horton, that all elongated basins must have appeared on initially inclined surfaces. During the evolution of basins and river networks, the form factor is liable to change. There are telling examples where basins have been deprived of or enlarged by areas of various sizes, which can change the form factor considerably.

*Law of average form factors*. Along a stream, the form factor will retain the same value if the increases of area and perimeter are proportional, but in most instances its value will change in relation to the contributions of the various tributaries. If form factors are calculated for the third- to eighth-order basins in the Ialomiţa system and the values represented on semilogarithmic paper, an inverse relationship is noted between the form factor and basin order (Fig. 39), reflecting

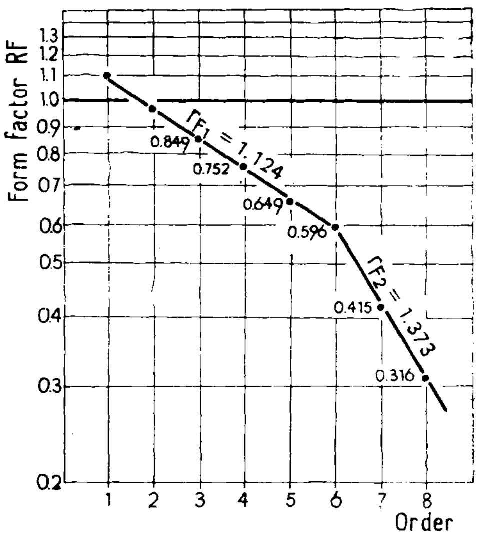



Fig. 39. Regression of form factors on order for sub-basins within the Ialomiţa basin.

the tendency of drainage basins to become more elongated with increasing order. Given the limited data, a tentative law may be formulated to express the observed relationship : *the average form factors of drainage basins of successively higher orders tend to form a decreasing geometric progression in which the first term is the form factor* $\overline{RF}_1$ *of first-order basins and the ratio is the ratio* $r_F$ *of successive average form factors*. The relationship can be written as

$$\overline{RF}_x = \overline{RF}_1 / r_F^{x-1} \tag{74}$$

This relationship may be considered to be well established for lower orders, based on form factors for 1113 third-order, 229 fourth-order, 53 fifth-order and 15 sixth-order basins. For the seventh and eighth orders the numbers of cases are too small (five and one, respectively) for much reliance to be placed in the corresponding values. However, as in the case of other relationships, it is very likely that a tendency may exist for the form factor to decrease more rapidly for high orders, hence yielding a second series with a much higher ratio (Fig. 39).

Even leaving the last two orders aside, the remaining values point to a tendency for the form factor to decrease less with increasing order for lower orders, so that for both the first and second orders its value is very close to unity. However, it is obvious that for these elementary runoff units an important influence on shape is also exerted by the slope of the topographical surface. Data are generally taken from maps, and when measuring the map projection of a basin and its area, the values obtained differ from reality to a greater extent, the steeper the slope. Thus, even if elementary basins are elongated in reality, their map projections may be round or almost square.

From an analysis of the values for several cases, the conclusion may be drawn that to the extent to which the form factor decreases, certain morphometrical elements deviate from their laws of development. For a given order, increases are noted in the length of the main stream, in the area and in other parameters of elongated basins in relation to round ones, indicating that the form factor is significant in drainage-basin evolution.

Chapter VII

# River Length

The length of a stream is the distance measured along the stream channel from the source to a given point or to the outlet, a distance which may be measured on a map or from aerial photographs. On large-scale maps, it is measured along the geometrical axis, or the line of maximum depth.

The source is considered as the location (or the point on the map) at which a stream starts, which as a rule coincides with the place where the smallest perennial stream appears (Chebotarev, 1964). For larger streams formed by the junction of two smaller streams with different names, the confluence point is usually taken as the source. In this case, in order to calculate the total length, the length of the largest tributary is added to the length of the main stream. As mentioned previously, in order to distinguish which is the main stream and which the tributary upstream from a confluence in a river network, Horton (1945) recommended the use of rules concerning the angle of junction and stream length. In the case of streams originating in lakes, the source is considered to be the point at which the stream flows out of the lake.

In detailed studies, especially in the case of small basins, the source is generally established in a subjective manner, taking as the initial point the first perennial stream found downstream from the water divide along the channel axis. In rainy periods, first-order streams will appear much closer to the water divide, which implies a corresponding change in position of the previously established point (Diaconu and Lăzărescu, 1965). In consideration of the Horton—Strahler classification principles, the source should be taken as the point in the first-order basin at which the first distinct channel appears or where stream flow begins. According to A. V. Broscoe (quoted by Strahler, 1964), this means that between the point thus established and the water divide there remains a distance more or less equal to the length of overland flow.

The mouth is the point at which a stream flows into another stream, or into a lake, sea or ocean.

The two points thus established should also be characterized by their altitudes. The length of a channel between its source and mouth may be measured using a curvimeter, which gives the total length, or by compass.

*River sinuosity.* Since rivers are of considerable significance in the evolution of a landscape, many local studies have relied on a number of quantitative indices characterizing the plane configuration of river channels. Subjectively, such channels can be considered as straight, sinuous, meandering or braided. Research carried out to date has led to the conclusion that owing to the diversity of physiographical conditions, straight channels do not exist. Even for shorter sectors, the straight-line distance between two points on a river does not exceed ten times the amplitude of meander (i.e., the deviation from linearity) between the points (Leopold and Langbein, 1962).

Calculation of the sinuosity coefficients currently used in the specialized literature requires knowledge of the channel length $L_c$, measured along the river course on a map projection, the shortest (straight-line) distance $L_a$ between the two extreme points considered, and the length $L_v$ of the valley axis, taken as the centre of the strip of land between the bases of the valley walls. In regions with rough relief, the valley walls descend directly to the water's edge. In this case, the valley length along a stream is equal to the length of the channel. When a stream has a well-developed floodplain, the length of the valley is measured along the axis dividing the strip between the bases of the walls into two equal parts (Fig. 40). The above measurements are used in calculating the following coefficients.

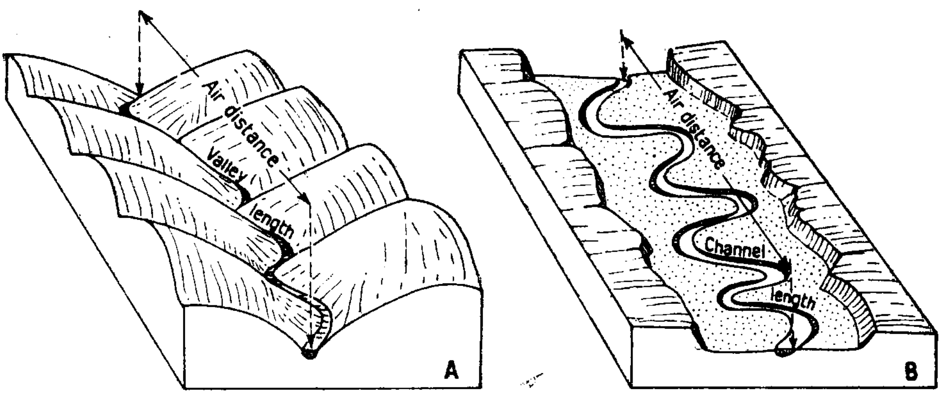



Fig. 40. (A) Topographical and (B) hydraulic sinuosity.

*Coefficient of topographical sinuosity.* This coefficient is obtained as the quotient between the length $L_v$ of the valley axis and the straight-line distance $L_a$ between the two extreme points (Mueller, 1968):

$$T_s = L_v/L_a \tag{75}$$

Topographical sinuosity is the result chiefly of the interaction of geological and geomorphological factors from which the present relief make-up, and hence the undulating course of valleys, has resulted in the course of time.

*Coefficient of hydraulic sinuosity.* By virtue of the laws of hydrodynamics, water courses, under the influences of slope, of the nature of the rock below the river bed, and also of suspended load, tend to meander along valley bottoms, resulting in sinuous courses which may deviate to a greater or lesser extent from the valley axis. A measure of this deviation, termed the hydraulic sinuosity, is given by the ratio of the channel length $L_c$ to the length $L_v$ of the valley axis:

$$H_s = L_c/L_v \tag{76}$$

The values obtained are higher than unity except in the case when the valley walls descend directly to the water's edge; in this case, the length of the valley axis is equal to the channel length, and the ratio is equal to unity. Leopold et al. (1964) showed that the coefficients of hydraulic sinuosity found in practice may take values as high as $H_s = 4$. On the basis of this coefficient, a distinction is made between sinuous rivers, with coefficients up to 1.5, and meandering rivers, with higher values.

*River sinuosity coefficient.* As in the previous cases, this is a dimensionless value, given by the ratio of the main-channel length $L_c$ to the straight-line distance $L_a$ between the two extreme points (Luchisheva, 1950):

$$K_s = L_c/L_a \tag{77}$$

In most work this coefficient is used to express global sinuosity, i.e., including both topographical and hydraulic sinuosity. A distinction between the latter two types is made only in detailed local studies, because it has been noted that topographical sinuosity is specific to regions with high relief energy, i.e., in an early stage of geomorphological evolution, whereas hydraulic sinuosity occurs mostly in flatlands or in regions where relief evolution has reached maturity.

For the Ialomiţa river, the sinuosity coefficient is 1.2 for the mountainous and hilly regions, a value accounted for primarily by topographical sinuosity. The course of the Ialomiţa in the piedmont plain has a coefficient of 1.40 between the confluences with the Vulcana and the Cricovu Dulce. The highest sinuosity coefficient for the Ialomiţa is found for the lower course, between the confluence with the Prahova and the outflow into the Danube, the value of 1.83 indicating a meandering river.

In a more detailed analysis, it is possible to evaluate the proportions of topographical and hydraulic sinuosity for a water course. Mueller (1968) suggested the following formula for evaluating the respective proportions *TSI* and *HSI*;

$$HSI = 100\,(K_s - T_s)/(K_s - 1) \tag{78}$$

$$TSI = 100\,(T_s - 1)/(K_s - 1) \tag{79}$$

These indices indicate the percentage of a stream's departure from a straight-line course due to hydraulic sinuosity in the first case, and to topographical sinuosity in the second.

*Braiding coefficient.* In many regions, geomorphological conditions determine a marked braiding of rivers, resulting in lateral streams which enclose a number of islands. In order to evaluate the degree of braiding, the braiding coefficient is calculated as the ratio of the sum of the lengths of the lateral streams to the length of the main stream in the sector studied (Chebotarev, 1953):

$$K_b = (l_1 + l_2 + l_3 + \ldots + L)/L \tag{80}$$

The value obtained shows by how many times the length of the secondary streams is greater than the length of the parent stream.

*Drainage scheme.* In order to follow more easily the drainage system or stream network in a given territory, as well as the positions of the main tributaries, a drainage scheme may be constructed. Luchisheva (1950) proposed a horizontal line on which distances are marked, the lengths of the parent stream and tributaries being represented on a given scale by straight lines making certain angles with the horizontal. For this scheme to be as close to reality as possible, it is desirable that in drawing it the positions both of the main stream and of the tributaries should be oriented in relation to the north. The tributaries must be represented on the given scale, but at the same time confluence angles should also be observed. Depending on how many elements it contains, such a scheme can show gauging stations or eventual modifications of the drainage network as required.

*Correction of river length.* To establish the length of a river channel, its map projection is usually measured. This is quite valid for plain regions where, in longitudinal profile, rivers have very small slopes. However, in regions with high relief, where the slopes of lower-order streams may have values up to and in excess of 400 $^0/_{00}$, for limited sectors the map projection may differ considerably from the real length of a channel. The difference between the real length and the map projection depends on the slope angle. Thus, when detailed studies are made of regions with steep slopes, it is necessary to correct the map projection $L_{map}$ in order to establish the real length $L_r$ of the river analyzed. To this end, the following formula can be used:

$$L_r = L_{map}/\cos \alpha \tag{81}$$

where $\alpha$ is the angle of slope of the stream, which may be measured by established methods. Since the slope is calculated in m km$^{-1}$ ($^0/_{00}$), the tangent of the angle is determined and there from, using tables, its cosine may be established easily.

For example, in the case of the Brătei basin, in the mountainous reaches of the Ialomiţa, the average slope of the first-order channel segments in the Horton—Strahler system is $453^0/_{00}$. Starting from the map projection of the respective segments, i.e., 74.4 km, this value of slope yields the real length as

$$L_r = 74.4/0.911 = 82 \text{ km}$$

This means that in reality the average length of the first-order streams is 0.337 km instead of 0.308 km as given by the map projection.

## Analysis of River Length using the Horton—Strahler System

The reversal effected by Horton (1945) in the river classification system then being used in Europe offered the possibility of a more detailed analysis of various morphometrical elements. Using this system, which progresses from small to big, from simple to complex, it was noted that the average length of stream segments increases proportionally to stream order. More detailed analysis of stream lengths after classifying channels according to the Horton—Strahler system yields a number of laws which are very helpful in analyzing drainage, relief fragmentation and even runoff and flooding processes.

*Law of summed lengths*. By summing the lengths of the stream segments of various orders within a drainage basin, a series of summed lengths in relation to order is obtained. Representation of these values on semilogarithmic paper indicates that the summed lengths of streams of various orders show an inverse relationship to order which has been verified for a very large number of cases (Fig. 41). A law relating the two elements may thus be formulated as follows: *the sums of the lengths of stream segments of successively higher orders tend to form a decreasing geometric series in which the first term is the summed length $L_1$ of first-order streams and the ratio is the ratio $R_L$ of successive summed lengths*. Given this relationship, the ratio of a particular progression may be obtained as the weighted mean of the measured ratios between successive orders, the weighting being done according to the summed lengths obtained, rather than according to the numbers of streams of each order. This is preferred because it may happen that not all of the streams counted are also measured, and in such cases weighting according to the number of streams would result in an incorrect value of the ratio.

Taking for instance the lengths of streams of successive orders in the sixth-order basin of the Doftana upstream from its confluence with the Prahova (Fig. 41C), the summed lengths for the various orders form the progression 664 : 339 : 140 : 88.7 : 44.1 : 40.

In order to calculate the ratio of this progression as the weighted mean of the partial ratios, we denote the latter by

$$L_1/L_2 = q_1,\ L_2/L_3 = q_2,\ L_3/L_4 = q_3,\ \ldots\ L_{u-1}/L_u = q_{u-1}$$

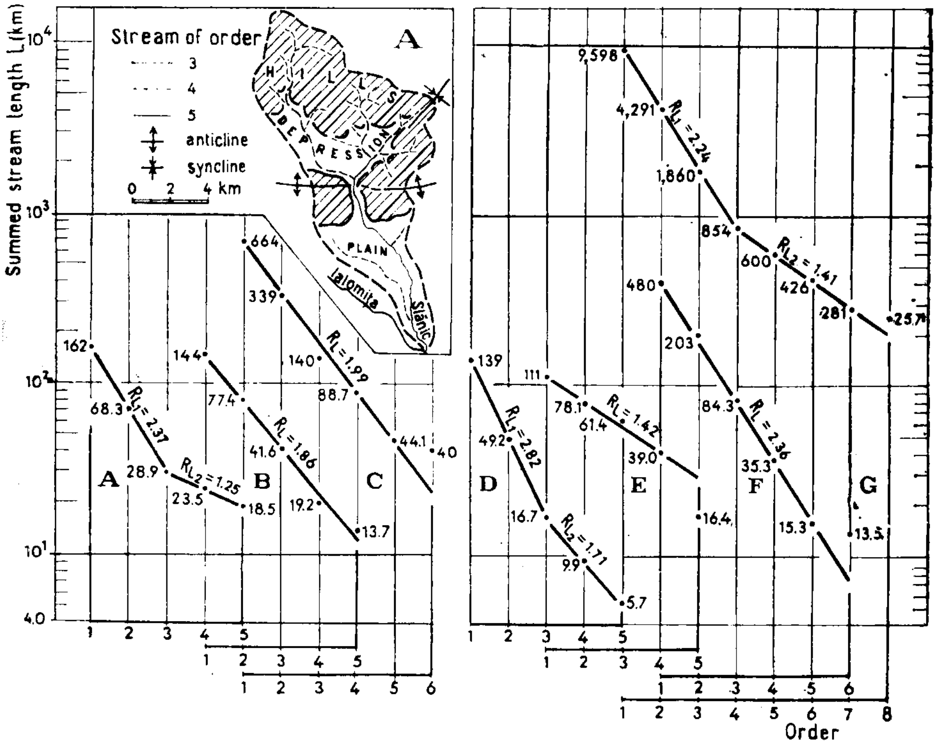



Fig. 41. Regression of summed stream lengths on order for the basins of (A) the Slănic at its confluence with the Ialomiţa, (B) the Azuga at its confluence with the Prahova, (C) the Doftana at its confluence with the Prahova, (D) the Ogrezeanca at its confluence with the Drajna, (E) the Dîmbu at its confluence with the Teleajen, (F) the Cricovu Sărat at its confluence with the Lopatna, and (G) the Ialomiţa at its confluence with the Danube.

and the sums of the corresponding lengths as

$$L_1 + L_2 = p_1,\ L_2 + L_3 = p_2,\ L_3 + L_4 = p_3,\ \ldots\ L_{u-1} + L_u = p_{u-1}$$

The ratio of the summed lengths, or the ratio of the geometric progression, is then established using the relationship

$$R_L = (q_1 p_1 + q_2 p_2 + q_3 p_3 + \ldots + q_{u-1} p_{u-1}) /(p_1 + p_2 + $$

$$+ p_3 + \ldots + p_{u-1}) \qquad (82)$$

In the present case, this yields

$$R_L = (1966 + 1159 + 360 + 267 + 92.7)/(1003 + 478.8 + 228.5 +$$

$$+ 132.8 + 84.1) = 1.99$$

Given this ratio, the terms of the progression can be computed for each separate order. For example, if the summed lengths of streams from the second order upwards have been calculated, the summed length of the first-order streams can be obtained by extrapolating the graphical correlation (Fig. 41) or by using the equation

$$L_1 = L_2 R_L \tag{83}$$

The general term of the progression can be calculated using the formula

$$L_u = L_1/R_L^{u-1} \tag{84}$$

By replacing $u$ by $x$, the summed length for any term is obtained in terms of the first-order value. In the case of the Doftana basin, analyzed above, the length of the main stream should be

$$L_6 = L_1/R_L^5 = 664/1.99^5 = 21.3 \text{ km}$$

In order to determine a term $x$ when the calculated value of the general term is known, the following relationship can be used :

$$L_x = L_u R_L^{u-x} \tag{85}$$

The relationship can also be solved as an equation of the form

$$\log L_x = H - Ix \tag{86}$$

where $H$ and $I$ are constants and $x$ is stream order.
For example, the summed length of the third-order streams within the Azuga basin upstream from the confluence with the Prahova is obtained from eqn. (85) as

$$L_3 = L_5 R_L^{5-3} = L_5 R_L^2 = 12 \times 1.86^2 = 41.5 \text{ km}$$

The law of summed lengths has been investigated and verified for a large number of drainage basins of various orders. In particular, it is followed closely in both mountainous and Subcarpathian regions, even though there are marked variations of rock type and structure, and active tectonics. For higher-order basins (from sixth to eighth), which are mostly elongated basins determined by paleogeographical evolution and which have developed in flat areas, the values obtained for higher-order streams deviate from the law, being systematically higher than those predicted by the correlation for the lower orders. This is because as soon as a stream enters flatland, lateral erosion prevails due to the decrease of slope, followed by braiding and meandering of the water course which, as has been seen, increase the sinuosity coefficient. An increase of stream length results, which explains the deviation from the law established for the drainage system in mountainous and hilly areas.

For basins developing in two major relief units, and hence having a significant number of higher-order streams in plain regions with small relief energy, there will obviously exist two lines with different slopes. Thus, for the basin of the Ialomiţa at its confluence with the Danube, the summed lengths of streams of successive orders give two geometric progressions with different ratios, the first being valid for lower-order (from first to fourth) and the second for higher-order streams (fifth to eighth) (Fig. 41G). The noticeable discontinuity in slope occurring after the fourth order, corresponding to the transition from a higher to a smaller ratio, is a consequence of the differing geomorphological conditions under which the higher and lower basins of the Ialomiţa have evolved. The ratios of the two progressions can be calculated using one of the methods mentioned earlier. For greater reliability, the average of the partial ratios weighted by the lengths of the streams considered may be used in the computations.

The summed-lengths ratios calculated for the 53 fifth-order streams in the Ialomiţa basin are normally distributed around a mean of 2.17. The lowest values of the ratio are found for the basins in the piedmont region and the subsidence plain (Dîmbu, $R_L = 1.42$; Crivăţ, $R_L=1.14$; Sărata, $R_L = 1.50$), while the highest values belong to a series of basins in the Subcarpathian region (western Ialomicioara, $R_L = 2.91$; Zîmbroaia, $R_L = 3.05$). For the 14 sixth-order basins, the average ratio is 2.20, very close to that for the fifth-order basins, but the variation is much lower (between 1.61 for the Sărata basin and 2.58 for the Vărbilău basin). The spatial distribution of the ratio indicates values for fifth-order basins which have developed in mountainous regions ranging from 1.57 (Muşiţa basin) to 2.22 (Prahova basin at the confluence with the Azuga). The transition to hilly regions marks an increase in the ratio, generally to values between 2.25 and 3.00.

Thus, an analysis of regional variations in the relationship between the summed length of streams and their order indicates a number of particular features resulting from the diversity of geographical factors, but this does not refute the tendency of these parameters to show law-like behaviour.

*Total stream length*. The total length $\Sigma L$ of the streams of all orders in a given drainage basin can be calculated from the sum of the corresponding decreasing geometric progression (as also used for the law of stream numbers):

$$\Sigma L = L_u(1 - R_L^u)/(1 - R_L) \tag{87}$$

In the case of the fifth-order basin of the Azuga at its confluence with the Prahova (Fig. 41B), the total length is

$$\Sigma L = 12(1 - 1.86^5)/(1 - 1.86) = 12(-21.26)/ -0.86 = 297 \quad \text{km}$$

as compared to the value of 296 km given by direct measurements, i.e., a difference of 1.0 km. This small error (0.3%) indicates that the method may be used safely when the necessary data are known. However, why should it be necessary to employ this formula if all the streams have been measured in order to establish the summed-lengths ratio? Verification of the law of summed lengths for a large number of cases proves that for sufficiently large basins the summed length of first-order streams may be omitted, since this is the most expensive to measure. In this situation, the series ratio $R_L$ can be calculated on the basis of orders from the second upwards, and then $L_1$ and $L_u$ can be established using eqns. (83) and (84), after which the results are introduced directly in the formula.

If the summed length $L_1$ of the first-order streams is known, the total of the terms can also be obtained from the formula

$$\Sigma L = L_1(1/R_L^u - 1)/(1/R_L - 1) \tag{88}$$

This method applied to the drainage basin of the Azuga at its confluence with the Prahova yields the value

$$\Sigma L = 144\,(0.538^5 - 1)/(0.538 - 1) =$$

$$= 144\,(0.0451 - 1)/(0.538 - 1) = 298 \text{ km}$$

This verification of the formula permits its use as an alternative to the previous one, depending on the data available.

Equation (87) can be employed successfully instead of that developed by Horton (1945, p. 293, formula 16) to calculate the total length of the streams in a given basin. The method proposed here is faster and more accurate, since in Horton's formula it is first necessary to calculate the summed length of the first-order streams in order to obtain their average length, or to use the method of inverse order (Horton, 1945). The present formula requires only directly measured values, and only three (instead of four) values need to be calculated.

As noted above, in the case of large basins which extend over two or three major relief units, with the passage from hills to plains the lengths of streams of various orders tend to give rise to a second geometric progression with a much lower summed-lengths ratio. In computing the total length of all streams for such basins it is absolutely necessary to take both progressions into account. In the case of the Ialomiţa basin, for which the first series is valid for streams of the first to fourth orders and the second for streams of the fifth to eighth orders, the total length is accordingly

$$\sum_{i=1}^{8} L_1 = \sum_{i=1}^{4} L_i + \sum_{i=4+1}^{8} L_i$$

or, using the corresponding formulae,

$$\sum_{i=1}^{8} L_i = L_4 (1 - R_{L_1}^4)/(1 - R_{L_1}) + L_8 (1 - R_{L_2}^4)/(1 - R_{L_2}) \qquad (89)$$

Substitution of the corresponding values for the Ialomiţa basin yields

$$\sum_{i=1}^{8} L_i = 854 (1 - 2.24^4)/(1 - 2.24) + 216(1 - 1.41^4)/(1 - 1.41) =$$

$$= 16\,651 + 1555 = 18\,206 \text{ km}$$

which compares with the value of 18 167 km given by direct summation of the measured values. The very small difference between the two values demonstrates the applicability of the formula.

*Law of average lengths.* Given the laws of stream numbers and of the summed lengths of streams of successive orders in a given drainage basin, the law of average lengths can easily be deduced. The ratios of the successive values defining the two series form a new geometric series, denoted by Horton (1945) as

$$L_1/N_1 = l_1,\ L_2/N_2 = l_2,\ L_3/N_3 = l_3, \ldots,\ L_{u-1}/N_{u-1} = l_{u-1},\ L_u/N_u = l_u$$

The law thus established states that: *the average lengths of stream segments of successively higher orders in a basin tend to approximate an increasing geometric series in which the first term $l_1$ is the average length of first-order segments* (Horton, 1945 ; Strahler, 1952 b) (Fig. 42). The ratio $r_l$ of successive average lengths may again be established by calculating the weighted mean of the partial ratios or their arithmetic mean, or by establishing the slope of the line drawn through the measured points (Fig. 42). Alternatively, if the values obtained for the two previous series show little scatter, the ratio of the average-length series can be calculated very easily given that the new series can also be written as :

$$L_1/N_1,\ (L_1/N_1)(R_c/R_L),\ (L_1/N_1)\ (R_c^2/R_L^2),\ \ldots\ (L_1/N_1)\ (R_c^{u-1}/R_L^{u-1})$$

Since $L_1/N_1 = l_1$, the series becomes

$$l_1, l_1(R_c/R_L),\ l_1(R_c^2/R_L^2),\ \ldots,\ l_1(R_c^{u-1}/R_L^{u-1})$$

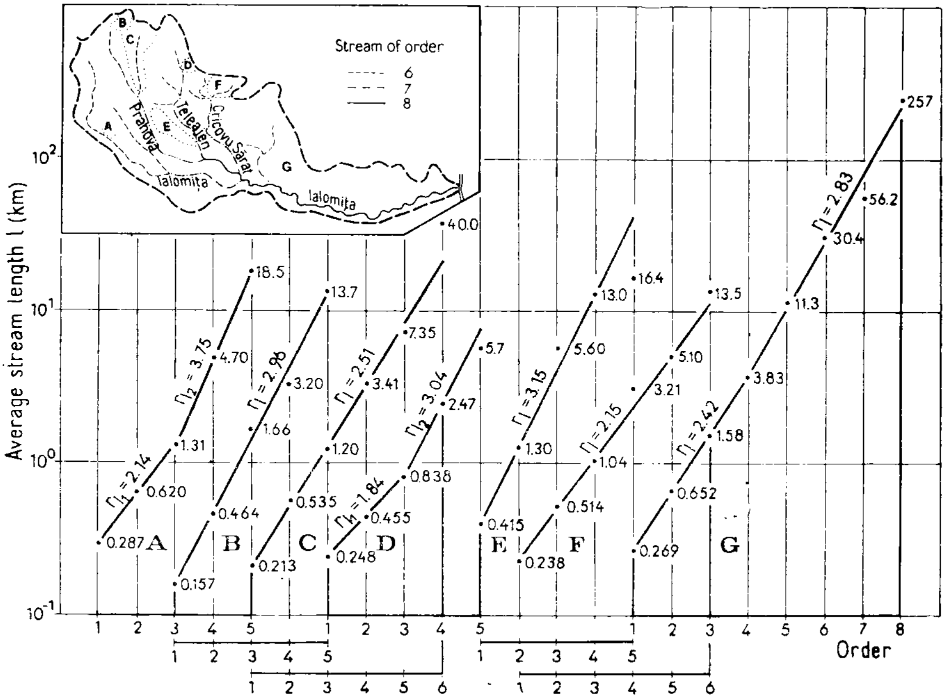



Fig. 42. Regression of average stream lengths on order for the basins of (A) the Slănic at its confluence with the Ialomiţa, (B) the Azuga at its confluence with the Prahova, (C) the Doftana at its confluence with the Prahova, (D) the Ogrezeanca at its junction with the Drajna, (E) the Dîmbu at its confluence with the Teleajen, (F) the Cricovu Sărat at its confluence with the Lopatna, and (G) the Ialomiţa at its confluence with the Danube.

If the progression of average lengths is calculated from the ratio of the two previous series, its ratio will be equal to the ratio of the two ratios, and thus $R_c/R_L = r_l$. The ratio of the average lengths is thus calculated from the quotient of the two ratios, and the new series can be written as

$$l_1,\ l_1 r_1,\ l_1 r_l^2,\ l_1 r_l^3,\ \ldots,\ l_1 r_l^{u-1}$$

From the methodological viewpoint, in order to establish the average lengths of streams of successive orders it is advisable to start from the smallest order measured and then calculate the lengths of the other orders from the relationship

$$l_u = l_1 r_l^{u-1} \tag{90}$$

or

$$\log l_x = h + ix \tag{91}$$

where $h$ and $i$ are constants and $x$ is stream order.

The average lengths of stream segments of successive orders show smaller deviations from the law of geometric variation than do the summed lengths. Nonetheless, in this case also, the streams of the highest order show the greatest deviations from the relationship. However, for large basins, the overall scatter is less. This is because, as also remarked by Leopold and Langbein (1962), stream length evolves according to probability laws. Thus, the higher the number of cases in a sample, the closer the mean of the sample to that of the whole population. Since the number of cases decreases in geometric progression from smaller to higher orders, it is natural that for the higher orders, proportionally to the decrease in the number of cases, the scatter should be greater.

The ratios of average lengths for individual fifth-order basins within the Ialomiţa system range from 1.34 for the Băltăţeşti basin in the Podeni depression, to 3.45 for the Mislea basin upstream from the confluence with the Cosmina. The mean value for the 53 fifth-order basins analyzed is 2.37, which is greater than the ratio of the summed lengths. Sometimes, the values of the average-lengths ratio indicate a disproportionately large difference between the lengths of streams of higher orders (Fig. 42). However, as a rule, the law of average lengths leads to a single geometric progression, although a second one can sometimes be sketched for higher-order basins, with little difference between the ratios. The fact that no clearcut zoning can be deduced from the spatial distribution of the average-lengths ratio indicates that drainage basins behave as distinct systems as well as evolving in relation to specific conditions.

## Morphometrical Model of Stream Length

By establishing the two series for the numbers of streams of successive orders and for their summed lengths, the law of average lengths and its ratio can easily be deduced with minimum computations by representing the two progressions for the same sample of cases on the same graph. This yields two lines intersecting at the point where the equations of the two series have common roots (Figs 43(a, b) and 44(a, b)). As has been seen, the ratio of the two series gives an increasing geometric progression which represents the average lengths of streams of successive orders. Its representation on the same graph will give a new line which intersects the former two, thereby yielding a scalene triangle (Figs. 43(c) and 44(c)) for the usual semilogarithmic coordinates, or a spherical triangle in normal coordinates. Of the three apexes of this triangle, that formed by the lines representing the summed and average lengths has special significance : its abscissa gives a value for the overall order of the system considered. The value of the abscissa may be established using eqns. (84) and (90). After taking logarithms, and noting

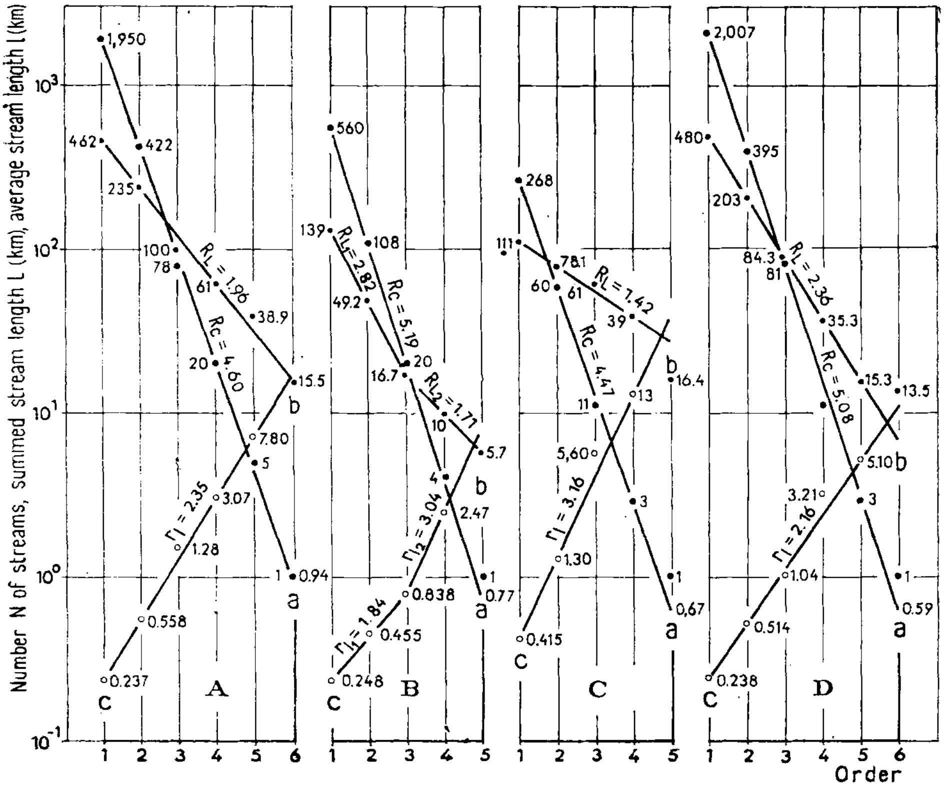


Fig. 43. Morphometrical models of stream length for the basins of (A) the Doftana at the Teșila gauge station, (B) the Ogrezeanca at its confluence with the Drajna, (C) the Dîmbu at its confluence with the Teleajen, and (D) the Cricovu Sărat at its junction with the Lopatna (a, regression of numbers of streams ; b, regression of summed stream lengths ; c, regression of average stream lengths).

that for the point of intersection $L_u = l_u$, the following relationships can be written :

$$\log L_u = \log L_1 - (u - 1) \log R_L$$

$$\log l_u = \log l_1 + (u - 1) \log r_l$$

which yield the result

$$(u - 1)\ (\log R_L + \log r_l) - \log L_1 + \log l_1 = 0$$

or

$$u = (\log L_1 - \log l_1)/(\log R_L + \log r_l) + 1 \tag{92}$$

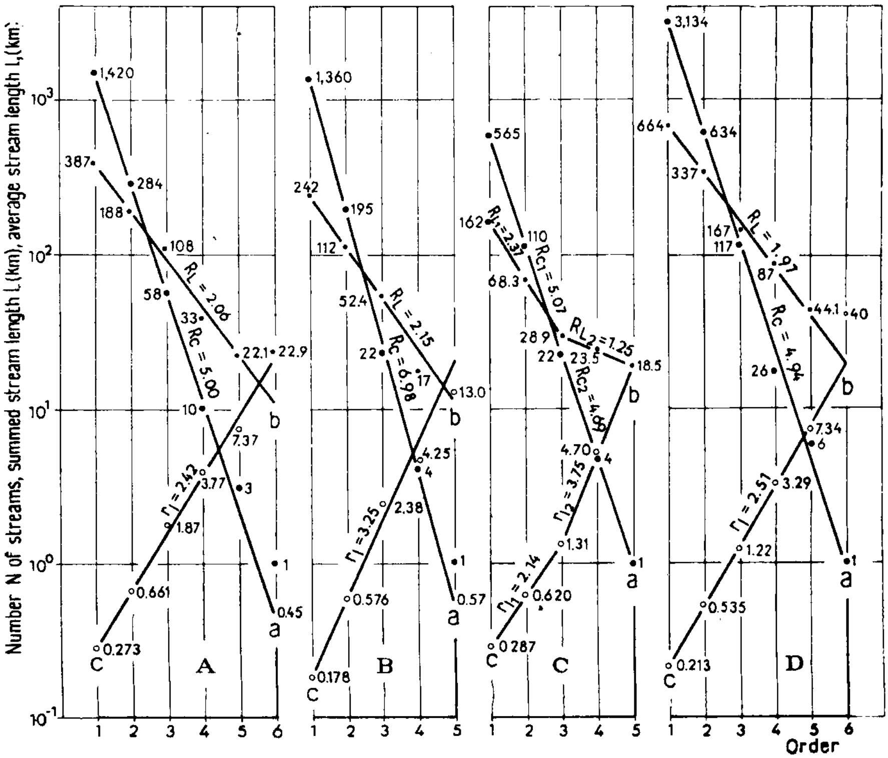


Fig. 44. Morphometrical models of stream length for the basins of (A) the Ialomiţa at its confluence with the western Ialomicioara, (B) the Vulcana at its junction with the Ialomiţa, (C) the Slănic at its confluence with the Ialomiţa, and (D) the Doftana at its confluence with the Prahova (a, regression of numbers of streams; b, regression of summed stream lengths; c, regression of average stream lengths).

For example, substitution of the appropriate values for the Cricovu Sărat upstream from its confluence with the Lopatna (Fig. 43D) gives

$$u = (2.68124 + 0.62342)/(0.37291 + 0.33445) + 1 = 5.67$$

Therefore, the basin of the Cricovu Sărat upstream from this confluence has an order of 5.67.

A number of relationships exist between the equations established for the three progressions defining drainage. As already shown, $r_l = R_c/R_L$, and hence if two of these parameters are known, it is easy to calculate the third. The confluence ratio $R_c$ varies between 2 and fairly high values ($\sim 10$), depending on basin order, shape, lithology, etc. The law of summed lengths being a decreasing and that of average lengths an increasing geometric progression, it results that the ratio $R_L$ of successive summed lengths must always be smaller than $R_c$. If $R_L$ were equal to $R_c$, then $r_l$ would be equal to unity and the average lengths would form a constant series, which is impossible in reality. Likewise, if $R_L$ were greater than $R_c$, their ratio would be below unity and the law of average lengths a decreasing progression. The summed-lengths ratio cannot be smaller than or equal to unity, since in this case the law of summed lengths would no longer be a decreasing progression : hence $1 < R_L < R_c$. If a constant $R_c$ is assumed in the case of a drainage basin of a given order, note that to the extent to which the ratio of summed lengths decreases, that of average lengths increases. Therefore, for natural drainage basins, $r_l$ will be higher than unity (Milton, 1966) but cannot exceed the confluence ratio : $1 < r_l < R_c$.

Analysis of the above relationships shows that the series are interrelated to such an extent that verification of the law of average lengths presupposes knowledge of the other two. Establishment of the law of average lengths alone, without considering the law of summed lengths, is insufficient, especially as regards the slope of the regression line. Representation of the law of summed lengths is necessary precisely to allow verification of the law of average lengths and an accurate evaluation of its ratio and the corresponding regression line.

Analysis of a large number of cases has led to the conclusion that the ratios of successive summed and average lengths may be related in various ways, depending on conditions specific to the basin considered. In the case of the Brătei upstream from its confluence with the Ialomiţa, where the basin has developed entirely in a zone of Precambrian crystalline schists in the Leaota mountains, the ratios of the summed and average lengths are very similar : 2.06 and 2.03, respectively. For certain basins which have developed on Neogene sedimentary formations in hilly areas, and especially for basins on the borderline between the Subcarpathians and the piedmont plains, there is a decrease of the summed-lengths ratio and an increase of the average-lengths ratio, as found for the Valea Sultanului upstream from its confluence with the Ursei (Fig. 45A), the Vulcana upstream from its confluence with the Ialomiţa, and the Slănic upstream from its confluence with the Ialomiţa (Fig. 44B, C). This reflects increases in the average lengths of streams of all orders. For the Dîmbu at its confluence with the Teleajen, the confluence ratio (4.47) indicates that in the piedmont plain of Ploieşti there are fewer streams of each order than in the Subcarpathians or even the Carpathians. The corresponding ratio of the summed lengths of streams of successive orders is 1.42, which indicates that there is no great difference between these lengths, whereas the average-lengths ratio of 3.16 indicates a great difference between successive orders. Under these conditions, although the number of streams is smaller than in other regions, their average length is greater. This is due to the lower relief energy and hence lower general slope in this region, favouring the elongation of stream segments.

Application of the above system of data analysis to several drainage basins throws light on a number of characteristic situations. For basins situated on rock with an almost constant resistance to erosion and with homogeneous physiographical conditions over the entire basin area, the three laws are all followed closely and each is represented by

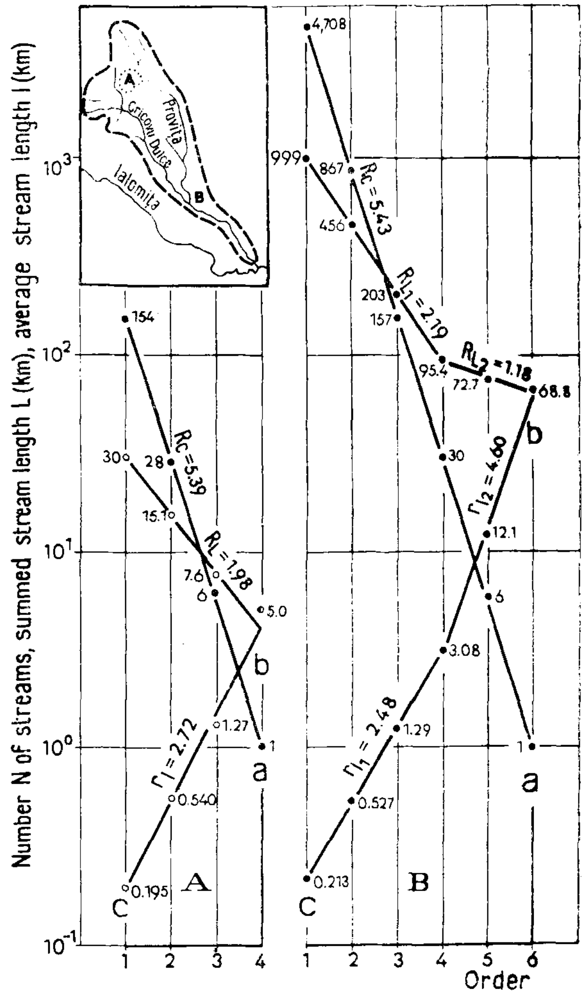



Fig. 45. Morphometrical models of stream length for the basins of (A) the Sultanu at its confluence with the Ursei, and (B) the Cricovu Dulce at its confluence with the Ialomiţa (a, regression of numbers of streams; b, regression of summed stream lengths; c, regression of average stream lengths).

a single line (Fig. 43A, C, D). If, however, a basin extends over two major relief units with rock types differing markedly as concerns resistance to erosion, or has a relief suffering considerable tectonic influence, the points show greater scatter. In most instances, they tend to group in such a way as to define one line for the lower and another for the higher orders. This tendency is demonstrated most clearly by the summed lengths (Figs. 43B, 44C, 45B and 46A), which explains why it is necessary to represent and analyze this series of values in every case. Frequently, if only the laws of stream numbers and of average lengths were analyzed, the tendency of the points to define two lines might well go unnoticed (Fig. 46A). The law of summed lengths, however,

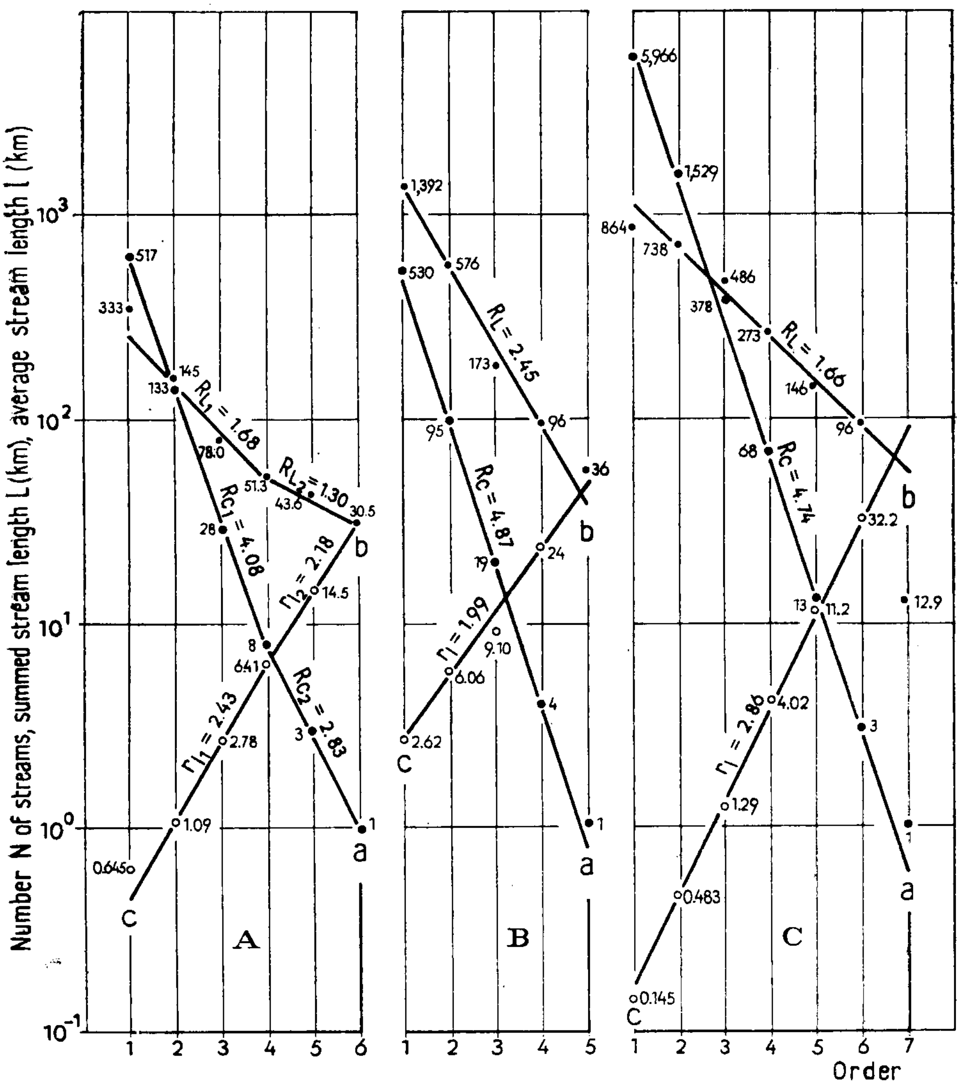



Fig. 46. Morphometrical models of stream length for (A) the Rio das Antas on the Brazilian plateau, (B) the Wardha in the Vidarbha region, India, and (C) the Allegheny in the United States (a, regression of numbers of streams; b, regression of summed stream lengths; c, regression of average stream lengths) (after Christofoletti, 1970; Pofali, 1979; Morisawa, 1962).

shows this tendency clearly. In the case of the Rio das Antas basin in the Brazilian plateau, which has an area of 423 km$^2$, if only the laws of stream numbers and of average lengths were analyzed, a single line could be established with very small error for either of the series (Christofoletti, 1970). However, the series given by the summed lengths of the stream segments of various orders shows considerable deviations (in the same direction) for the fourth- and fifth-order stream segments. Their tendency is to define a new line with a slope smaller than that established by the first to fourth orders. In such instances, the apparent error is greatly reduced by drawing a second line with a ratio much smaller than that established for the lower orders. Such results indicate that when a basin's relief energy is low, its stream segments tend to become longer while their number decreases. A more thorough analysis of the ways in which the points defining the numbers and average lengths of the stream segments are grouped in the preceding exapmle also reveals the possibility of drawing two lines with a point of inflection at the fourth order (Fig. 46A).

Even when the rocks and physiographical conditions in a basin are homogeneous, two regression lines may still be found owing to a change in the rate of evolution of the basin as a result of modifications wrought by the flows of matter and energy or by tectonic movements. The rate of evolution may also be altered by man's intervention, for example, through massive deforestation.

There are instances when the law of stream numbers gives a single line while the laws of summed and of average lengths give two lines with differing ratios (Fig. 45B).

In comparing a large number of cases from the specialized literature with our measurements, we faced certain difficulties in establishing data for the first order on large-scale maps. In most instances, both the summed and average lengths of first-order streams deviate from the corresponding lines, although these are the most numerous data and hence should have the most accurate values. The causes of these deviations concern the accuracy of cartographic representation and the way in which the length of first-order stream segments is established. Whereas the initial point of a second-order stream is easy to plot on a map, this is much more difficult for first-order streams. Plotting of the initial points of such streams from the inflections of contours in relation to scale is a painstaking operation. For these reasons, even if the number of stream segments is established correctly, although it too may sometimes be overestimated, the length may easily be greater (Fig. 46A) or smaller (Fig. 46C) than the calculated value. This is also demonstrated by the law of average lengths, which gives much higher values for the first-order streams in the former case, and much smaller in the latter, in comparison to the values obtained by computation. In such instances, it is easier to establish the lengths of stream segments from the second order upwards and therefrom, on the basis of the calculated ratios, the length of the first-order streams whose direct measurement, besides being the most difficult, yields values of uncertain accuracy.

In almost all the cases analyzed, the three progressions have three points of intersection on the graph and form a triangle. However, the values may be such that the three progressions have only two points of intersection and no longer form a triangle (Fig. 46B). This situation occurs more frequently in the case of models of other morphometrical elements.

*Relationship between river length and drainage-basin area.* An increase in drainage-basin area brings about a directly proportional increase in stream length.

Using data obtained by Filenko (1950), Makkaveyev (1955) established a relationship between mountain stream lengths $L$ in the Crimea and the corresponding drainage-basin areas $A$ : $L = 2.51 \sqrt{A}$. For mountain streams in the Caucasus the relationship is $L = 2.37 \sqrt{A}$, and for the largest plain rivers in the USSR, the People's Republic of China and the United States, it is $L = 2.9 \sqrt{A}$.

From measurements of stream length and the adjacent areas effected at some 400 gauging stations in the United States (Langbein, 1947), Hack (1957) found that the two variables are related according to $L = 1.4\ A^{0.6}$ (units of miles), i.e., the length of a given stream is proportional to the respective area to the power of 0.6.

Morisawa (1962) established close relationships between the logarithms of basin areas and of the corresponding average or cumulated average lengths.

For the river network in Romania, Diaconu (1971) found the relationship $L = 1.32\ A^{0.595}$. By processing data for several hundred basins in the USSR, Sokolov (1962) also found a relationship similar to that proposed by Hack. Thus, for drainage basins with areas larger than 250 km$^2$, $L = 3.1\ A^{0.5}$, while for smaller basins, $L = 1.54\ A^{0.5}$. The exponent of the area is thus the same for all basins, but for those with areas smaller than 250 km$^2$ the constant is halved.

There is a close relationship also between the logarithm of the length of the main stream from its source to a given point and the logarithm of the adjacent area (Strahler, 1964). This relationship has also been verified for rivers in Romania, although deviations from it are found depending on the basin form factor. Thus, for equivalent areas, the length of the main stream is greater in elongated basins in comparison with round basins.

The close relationship between average stream-segment length and average basin area for a given order may be demonstrated by plotting the respective values on semilogarithmic paper (Schumm, 1956). As an example, Fig. 47 gives mean values for 16 sixth-order basins situated in mountainous, hilly and piedmont regions of the Ialomiţa basin. The equation of the line, $l = 1.09\ a^{0.55}$, indicates that stream length is positively allometric with respect to basin area (Church and Mark, 1980). However, the coefficient of 1.09 cannot be compared with those in the previous equations, since this relationship refers not to the length of the main stream but to the average lengths of stream segments of successive orders and the average areas of basins of various orders in the Horton—Strahler classification system.

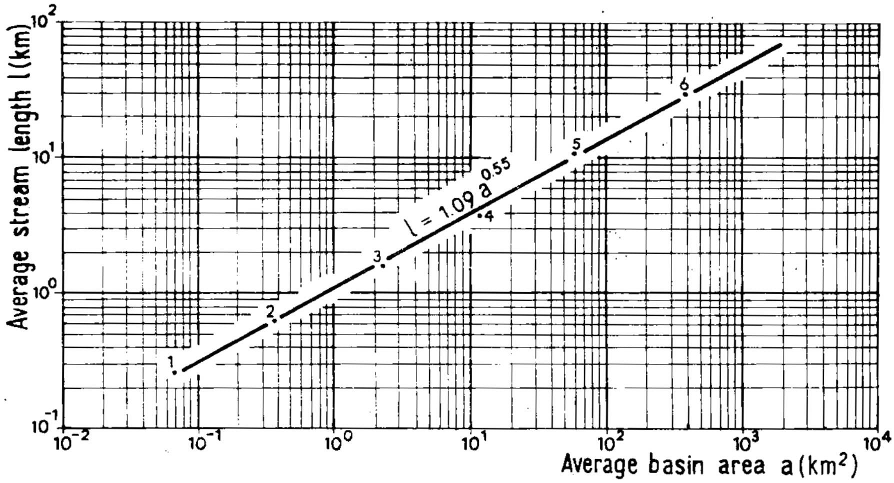



Fig. 47. Relationship between average stream length $l$ and average basin area $a$ for a sample of 16 sixth-order basins in the Ialomiţa basin.

## Average Length of Overland Flow

It is known that during torrential rainfall or a thaw, as soon as the soil's infiltration capacity and the retention capacity of the plant cover have been exhausted, on sloping ground the excess water will flow towards the channel network along the line of steepest slope. The distance covered from the water divide to the nearest channel represents the length of overland flow, an important variable on which runoff and flood processes depend. The horizontal projection of this distance may be obtained from maps.

The length of overland flow is an important element in the detailed study of runoff processes in drainage basins and especially of the concentration time $T_c$, which is the time needed for water arriving at the surface to travel in flood period from the remotest point to the nearest gauging station. It is further possible to distinguish between the interval $T_{c1}$ required for *overland flow* and that ($T_{c2}$) required for *stream flow* once formed in the channel network, to travel to the gauging station (Lambert, 1975). Therefore

$$T_c = T_{c1} + T_{c2} \tag{93}$$

Taking into account the locations of the above processes, it is obvious that overland flow depends on the length of the slope and on the nature and state of the surface, being influenced by the way in which the land is used and the type of cultivation. Overland flow disappears shortly after rainfall, water being absorbed by the soil, retained by the plant cover, or evaporated. The differences between overland flow and stream flow result both from their differing hydraulic regimes and from the fact that during overland flow infiltration takes place over the entire basin area. Stream flow persists from several hours to several days after rainfall, depending on the size of the channel concerned. It also depends on the distance travelled by water in the channel, and hence on the size of the basin, and also on the slope and state of the channels, which may be influenced by river engineering work.

The study of overland flow is very important because it affects not only the water regime of a river network but also the long-term evolution of drainage basins. The length of overland flow depends primarily on the degree of relief fragmentation, and hence on the drainage density. According to Horton (1945), this parameter "is in most cases approximately half the average distance between the stream channels and hence is approximately equal to half the reciprocal of the drainage density":

$$l_0 = 1/2D_d \tag{94}$$

Since $D_d = \Sigma L/A$, the relationship may also be written as

$$l_0 = A/2\Sigma L$$

Using eqn. (107) to establish drainage density on the basis of Horton's laws, this becomes

$$l_0 = A_u(1 - R_L)/2L_u\,(1 - R_L^u) \tag{95}$$

To take account of the slopes of the surface and of the river network, Horton (1945) proposed the formula

$$l_0 = 1/2D_d\,\sqrt{1 - (S_c/S_g)^2} \tag{96}$$

where $S_c$ represents the slope of the stream channels and $S_g$ the mean slope of the ground.

Chapter VIII

# Stream and Drainage Densities

Because of the continual interaction between surface runoff and the factors responsible for the resistance of a drainage area (rock, soil, plant cover), drainage density is an important feature of drainage systems from both the geomorphological and hydrological viewpoints. From the former point of view, a river network is an element of the landscape, its spatial distribution on a drainage basin influencing relief fragmentation and to a great extent the types and intensities of certain geomorphological processes. For hydrologists, drainage density is significant in that it plays an important role in surface-runoff processes, influencing the intensity of torrential floods, the concentration, the sediment load and even the water balance in a drainage basin. Being such an important parameter, drainage density has been studied by many researchers, who have dealt with both the definition of the concept and quantitative methods of calculation, and the relationships of causality and interdependence with the factors determining it. Drainage density is assessed in terms of the number of streams on the one hand, and their length per unit area on the other.

## Stream Density

Stream density is an important morphometrical indicator which can provide further information concerning the response of drainage basins to runoff processes. It is defined as the ratio of the total number of streams to the drainage area, i. e., it represents the number of streams per unit area (Horton, 1932 ; A. Malicki (1937), quoted by Wilgat, 1968 ; Apollov, 1963). In introducing his classification system, Horton (1945) also defined the similar notion of stream frequency, as "the number of streams, $F_s$, per unit area", or

$$F_s = N/A \tag{97}$$

where $N$ is the total number of streams in a drainage basin of area $A$. Subsequently (R. D. Freitas (1952), quoted by Christofoletti, 1969), the term "stream frequency" was replaced by that of stream density $D_s$, in analogy to drainage density.

With the completion of Horton's classification system by Strahler (1952a), the evaluation of this parameter underwent a slight change. Since any water course ends with a first-order stream segment, the total number of first-order streams in the Strahler system is equal to the total number of streams of all orders in the Horton system (Shreve, 1966). In the former case, the stream frequency $F_s$ calculated from eqn. (97) becomes the ratio of the number $N_1$ of first-order streams to the drainage area $A$ :

$$F_s = N_1/A \tag{98}$$

As advances were made in such studies, the concept of stream density was applied and analyzed more thoroughly (Zăvoianu, 1972b; Christofoletti and Oka-Fiori, 1980) so as to make it compatible with other morphometrical indices such as drainage density (Christofoletti, 1978).

The number of streams of all orders using Strahler's classification system is higher than the total number of streams in the Horton system, and this makes it necessary to introduce the notion of stream-segment density $D_s$ as the ratio of the total number $\Sigma N_i$ of stream segments to the given drainage area $A$ :

$$D_s = \Sigma N_i/A \tag{99}$$

The stream-segment density may be calculated by considering the decreasing geometric progression formed by the numbers of stream segments of successively higher orders, whose sum is given by eqn. (29). The situations analyzed demonstrated that the total number of stream segments can be calculated quite satisfactorily in this manner, with an error of 3 – 5 %. Then, in order to calculate the stream density when the basin area is known, the only operation to be performed is to introduce this value into the equation (Zăvoianu, 1972b)

$$D_s = \Sigma N_i/A = N_u(1 - R_c^u)/A(1 - R_c) \tag{100}$$

Since difficulties of computation arise only for the expression $(1 - R_c^u)/(1 - R_c)$, for a given value of the confluence ratio this quotient may be established using appropriate tables. For instance, the stream density in the basin of the Azuga upstream from its confluence with the Prahova is given by

$$D_s = N_u(1 - R_c^u)/A(1 - R_c) = (1.0/89.6)(1-5.50^5)/(1 - 5.50) = 12.5$$

the terms $N_5 = 1.0$, $R_c = 5.50$, $u = 5$ and $A = 89.6$ being known. The fact that the term $N_u$ of the progression is in this case equal to unity facilitates the calculation. If the term of the highest rank differs from unity, and there are many such situations, then this must obviously be considered in the computations. In the case of the Bizdidel upstream from its confluence with the Ialomiţa, for instance, for which $N_u = 0.958$, $R_c = 5.25$, $u = 5$ and $A = 95.5$, a density of 9.4 stream segments per unit area is calculated.

A brief analysis of any topographical map reveals that the density of stream segments in a zone bears a direct relationship to the degree of relief fragmentation, and also reflects the number of valley segments per unit area. Research carried out in the Piracicaba and São Pedro areas (Brazil) has shown that, at least for these regions, deep geological formations have no control over the spatial distribution of streams (Christofoletti and Oka-Fiori, 1980). The quoted authors noted that stream density is a hydrological phenomenon related to the characteristics of surface geological formations. For the conditions specific to Romania, the spatial distribution of stream-segments density is influenced by the resistance to erosion and the permeability, tectonics and structure of the rock types in a basin, as reflected in the hydrological behaviour of surface formations and the degree of geomorphological evolution, and by climatic conditions.

Although a direct relationship exists between the total number of stream segments and drainage area (Seyhan, 1976), a well-defined inverse relationship between stream-segment density and drainage area exists only in cases in which the number of streams remains constant for various values of areas. In establishing the correlation between these two variables, therefore, the number of stream segments must necessarily be taken into account as a parameter, as in the case of drainage density.

Statistical processing of data for a sample of 53 fifth-order basins situated in widely varying physiographical conditions revealed the highest densities (22.4 $km^{-2}$) to be characteristic of the Subcarpathian region, and the lowest ones (1.20 $km^{-2}$) of the piedmont flatlands. The sample analyzed shows an average of 10.4 stream segments per unit area, with a variance of 29.4 and a standard deviation of 5.43. The skewness coefficient ($S_k = 0.16$) indicates a slight positive asymmetry, while the coefficient of variation is 0.52.

## Drainage Density

Attempts to find an index expressing the density of river networks were made as early as 1894, when A. Penck (quoted by Morariu et al., 1962) defined the density $D$ of a river network as the ratio of the length $l$ of the main stream to the number $n$ of sectors counted between confluences with the more important tributaries :

$$D = l/n \tag{101}$$

Here the density is thus given only in terms of the average length of the measured sectors : the smaller the value, the higher the density of the network.

L. Neumann (1900) (quoted by Morariu et al., 1962) defined the density of a river network as the ratio of its length to the area of its drainage basin. According to Neumann, inequalities in network densities are caused by variations in the type of substratum, in surface gradient, in the plant cover and in precipitation, which plays the greatest role. Horton (1932) noted higher drainage densities in areas with higher precipitation, and very low densities in basins with great permeability.

Feldner (1903) (quoted by Morariu et al., 1962) proposed that drainage density should be expressed as the ratio of the drainage area to the number of interfluves. This ratio represents the basin area pertaining to an interfluve : the lower its value, the higher the density, and vice versa.

The method proposed by Neumann and complemented by various calculation techniques gained in popularity with the passage of time. A number of authors, such as Surken and E. P. Senkov (quoted by Morariu et al., 1962), suggested as a method of computation that the drainage area be divided into a number of regular geometrical figures, preferably squares with a side of 1 or 2 km. The length of the network in each square is then measured using a curvimeter or dividers, and divided by the square's area. Finally, the resulting values are assigned to classes and each square is hatched or coloured accordingly. The values obtained for each of the squares can also be used to construct lines of equal density, giving a picture of the spatial distribution of density (Morariu and Savu, 1954). Instead of equal-area squares, basins of various orders may be taken as units, the resulting values then being grouped by basin order (Morariu et al., 1956 ; Grigore, 1979).

The progress made by Horton (1932, 1945) in river morphometry, and in particular the inversion of the previous classification system, also lent impetus to research concerning drainage density. The term was used increasingly, and although objections did exist, the same definition was maintained. Langbein (1947) showed that drainage density is inversely proportional to the length of overland flow, and that its reciprocal provides a measure of the average distance between rivers.

Neumann's method has sometimes been criticized for failing to offer an accurate definition of the density of a drainage system. To remedy this, Wilgat (1968) proposed a distance method for evaluating drainage density, which consists in measuring the distance from a divide to the nearest water course, and then drawing isolines through points with the same values. This implies that the smaller the distance from the nearest water course, the greater the density, and vice versa. The method provides a good representation of river-network density (Wilgat, 1968).

Chebotarev (1953) showed that drainage density can also be determined as the ratio of average stream length $l$ to the average adjoining area $a$. If in a drainage basin of area $A$ there is a certain number $n$ of streams or stream segments with a total length $\Sigma L$, then $l = \Sigma L/n$ and $a = A/n$, or $\Sigma L = nl$ and $A = na$. On substituting these expressions, the formula for drainage density becomes

$$D_d = \Sigma L/A = nl/na = l/a \tag{102}$$

The reciprocal of the density then represents the average area associated with a unit of average length:

$$1/D_d = a/l \tag{103}$$

The average length $l_0$ of overland flow can also then be calculated as

$$l_0 = 1/2D_d = a/2l \tag{104}$$

Equations (102)—(104) are valid also for the Horton—Strahler and even Shreve classification systems, with $l$ denoting the average length of a stream segment and $a$ its average adjoining area.

The accuracy required in drainage-density measurements depends on the goal pursued, and is limited by the scale of the map used. Horton (1945) suggested that all streams (perennial, intermittent and ephemeral) should be considered. In subsequent work the river network was completed by including valleys deduced from contour lines (Strahler, 1953; Carlston, 1963; Bowden and Wallis, 1964; Gregory, 1968; Orsborn, 1970).

To obtain more accurate information about drainage density, in addition to faithful mapping of channel networks on large-scale maps or aerial photographs, account must be taken of the time during which a network fulfils the runoff function: thus, there are channel networks with perennial runoff, with seasonal runoff and with temporary runoff. A fully accurate value of the drainage density is obtained by considering all these networks, but for certain purposes the densities may be calculated for each of the categories.

*The density of perennial drainage* ($D_{dp}$), which has long been referred to simply as the density of a river network, is now related only to perennial streams (Hirsch, 1967; Lambert, 1975), whose existence is related to the capacity of underground water to ensure perennial runoff. It is thus inappropriate to calculate the length of overland flow on the basis of the perennial drainage density, since the values obtained will be exaggerately high.

*The density of seasonal drainage* ($D_{ds}$) is calculated using the sum of the lengths of streams whose runoff is seasonal (Hirsch, 1967; Lambert, 1975), i.e., streams which become dry in the hot season once they have exhausted the water reserves in the soil and underground layers they drain (Lambert, 1975). The phenomenon of seasonal dryness depends greatly on the permeability of deposits and on the climatic conditions within a geographical zone. If the lengths of the perennial and seasonal networks are added, the resulting value of the drainage density is greater, and the length of overland flow obtained therefrom is smaller.

*The density of temporary drainage* ($D_{dt}$), which Lambert (1975) termed the density of topographical drainage or the density of thalwegs, refers to the length of channels performing runoff functions only during rainfall and snowmelt. We propose the term temporary drainage for the sake of terminological consistency, since, as Lambert showed, each type of drainage density corresponds to a certain frequency distribution of runoff.

From the foregoing considerations, the overall drainage density is obtained as

$$D_d = (\Sigma L_p + \Sigma L_s + \Sigma L_t)/A \tag{105}$$

where $\Sigma L_p$, $\Sigma L_s$ and $\Sigma L_t$ are the total lengths of perennial, seasonal and temporary streams.

The three types of streams may be distinguished on a map by appropriate marking or the use of different colours. A series of comparisons can then be made which lead to interesting conclusions. For instance, the ratio between $D_{ds}$ and $D_{dp}$, indicating by how many times the seasonal is higher than the perennial drainage density, provides useful information concerning the rate of drying in a region (Lambert, 1975). The lengths of overland flow calculated taking all three categories of streams into account have the lowest values and are also closest to reality; hence it is important to establish the overall drainage density as accurately as possible if it is to be used in calculations of concentration time and the length of overland flow.

However, measurements of the total stream length are painstaking and time-consuming. To eradicate these shortcomings while maintaining the quality of the results, several methods for rapid calculation of drainage density have been suggested and verified.

## Measurement of Drainage Density

*Use of Horton's laws.* Employing the data given by the laws of stream numbers and of average lengths, Horton (1945) proposed the following relationship for the calculation of drainage density:

$$D_d = l_1 r_b^{u-1}(\rho^u - 1)/A(\rho - 1) \tag{106}$$

where $l_1$ is the average length of first-order streams, $r_b$ is the bifurcation ratio, $u$ the order of the main stream, and $\rho$ the quotient between the length ratio $r_l$ and the bifurcation ratio $r_b$. To facilitate the computations, Horton constructed plots giving the factor $(\rho^u - 1)/(\rho - 1)$ for various orders of the main stream. The relationship incorporates a number of physiographical factors and is based primarily on the laws governing the development of the river network in a given drainage basin. It again requires determination of the total stream length in a basin.

Strahler (1957) showed that the summed lengths of streams of successive orders are inversely proportional to order, yielding a geometric progression from which the law of summed lengths is established (Fig. 41). Once such a progression has been verified, the sum of its terms can be calculated using eqn. (87), yielding a new formula for the calculation of drainage density in a given basin (Zăvoianu, 1972b):

$$D_d = \Sigma L/A = L_u(1 - R_L^u)/A(1 - R_L) \qquad (107)$$

This expression is a considerable simplification of the previous formula established on the basis of Horton's laws and is thus easier to use. Values of the term $(1 - R_L^u)/(1 - R_L)$ may be tabulated or plotted, as done by Horton (1945) for his relationship. For instance, in the case of the Azuga basin at the confluence with the Prahova, for which the law of total lengths gives $L_u = 12$ km and $R_L = 1.86$, the drainage density is

$$D_d = 12(1 - 1.86^5)/89.6(1 - 1.86) = (12/89.6)\,(-21.6/-0.86) =$$
$$= 3.31 \text{ km km}^{-2}.$$

The drainage density obtained by direct measurements (3.30 km km$^{-2}$) is very close to this value, which demonstrates the applicability of

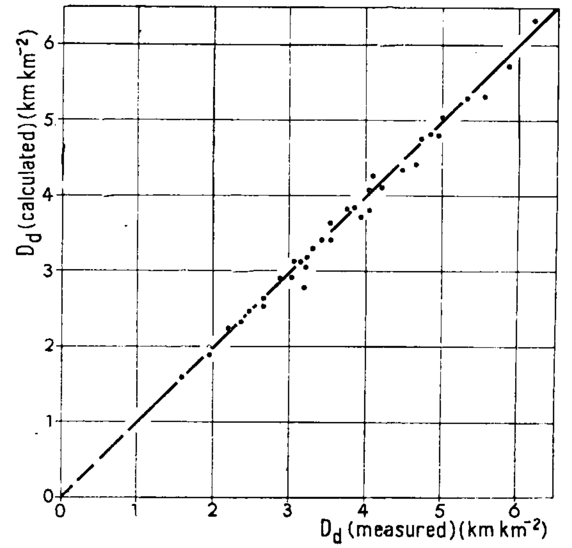



Fig. 48. Relationship between drainage density $D_d$ calculated from eqn. (107) and obtained by direct measurements for 53 fifth-order basins.

the formula. The ratio of the calculated and directly measured drainage densities will obviously be unity in the ideal case. A different value implies that the ratio for the law of summed lengths has not been evaluated with sufficient accuracy and should be recalculated. The respective values established for 53 fifth-order basins (Fig. 48) show very small differences, and hence that the formula is accurate. The following conclusions may be drawn from the analysis of eqn. (107):

a very good correspondence exists between the measured and calculated values for drainage basins with form factors between 0.45 and 1.00;

for elongated or very elongated basins with form factors lower than 0.45, the correspondence is less good, since in these cases there is a marked disproportion between the length of the main stream and the summed lengths of streams of lower orders: if the former value is considered, especially in calculating the weighted-mean ratio, the terms which define the general orientation of the lines will be ignored;

ignoring the length of the main stream results in a greater difference between the real and calculated densities: points situated on the right-hand side of the correlation line give greater main-stream lengths.

A good example of the last point is provided by the Pîrscov stream, a left-side tributary of the Ialomiţa, for which a large difference (11.9%) is found between the density calculated on the basis of eqn. (107) (2.81 km km$^{-2}$) and that given by direct measurements (3.19 km km$^{-2}$). As a matter of fact, in this basin, whose form factor is 0.306, the length of the main stream is 15 times greater than that calculated on the basis of the law of average lengths (Fig. 48).

*The intersection-line method.* The bases of this method, although it was used originally only for determining the length of contour lines, were laid by Wentworth (1930) and Horton (1932). Langbein (1947) used the method for calculating the length of contour lines for use in determining basin slope, but Carlston and Langbein (1960) first applied it to determine drainage density. The method consists in placing a grid on the drainage network and counting the number of points at which the drainage network intersects the transverse lines. The drainage density is then given by the ratio of the number $N$ of intersections to the transverse length $L$. Complemented by new elements, the method has been employed and recommended more recently, for example, by McCoy (1971) and Mark (1974).

Using the equivalence established by Wentworth (1930) between his formula for basin slope and that developed by Finsterwalder, Verhasselt (1961) noted that if equidistance is removed from these formulae, the following relationship is obtained:

$$l/A = n/0.6366 \tag{108}$$

where $l$ is the network length in relation to an area $A$, and $n$ the ratio of the number $N$ of intersections with the grid to the total length $L$ of the grid lines in area $A$. Hence $D_d = l/A = (N/L)(1/0.6366) = 1.5708(N/L)$, which is precisely the formula recommended by Mark (1974). Starting from variants of this method, several formulae have been developed for drainage density:

$$D_d = 1.414\ N/L \text{ (Carlston and Langbein, 1960)} \tag{109}$$

$$D_d = 1.8 + 1.27\ N/L \text{ (McCoy, 1971)} \tag{110}$$

$$D_d = 1.571\ N/L \text{ (Mark, 1974)} \tag{111}$$

$$D_d = 1.83\ N/L \text{ (Christofoletti and Filizola, 1978)} \tag{112}$$

*Intersection-points method.* Verhasselt (1961) recommended a fast method for establishing stream lengths and areas which offers the advantage of reducing substantially the time required for the calculation while errors are negligible. It has been noted that if a grid of small squares with known sides is placed on a drainage network, a quanti-

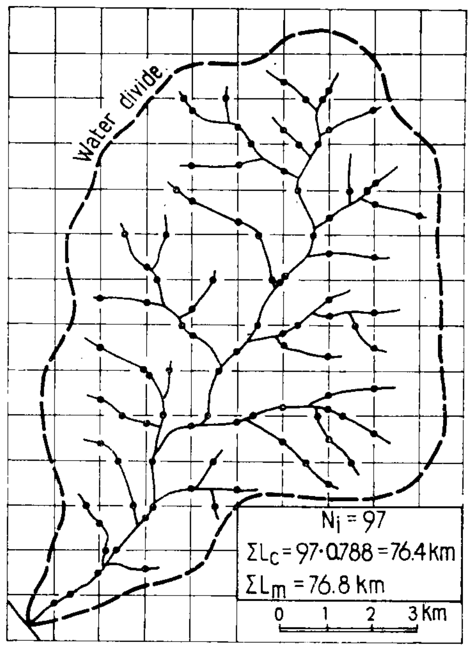
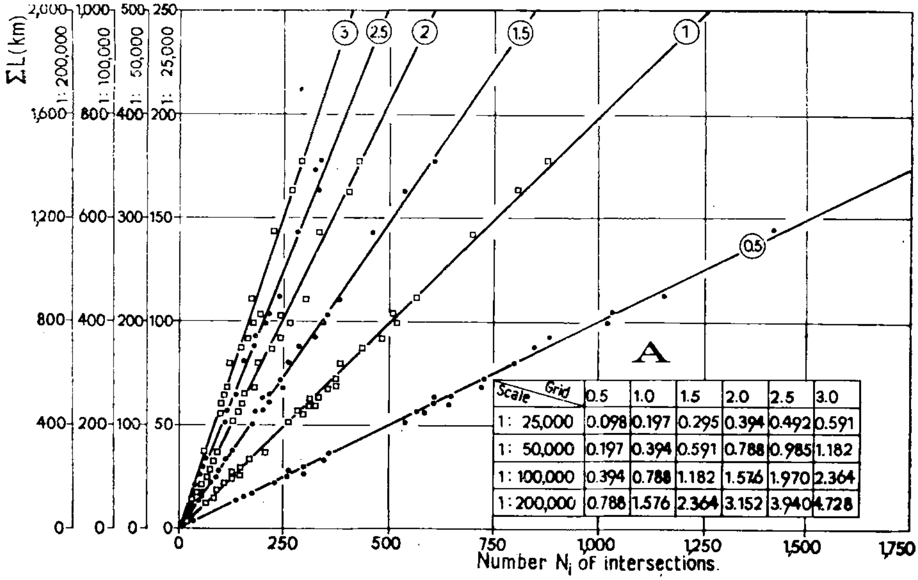



Fig. 49. Determination of network length using the intersection-points method: $N_i$, number of intersection points; $\Sigma L_c$, calculated network length; $\Sigma L_m$, measured network length.

tative relationship exists between the number of points at which the network intersects the sides of the squares and the total network length (Fig. 49): measurements have shown that the number of intersection points increases in direct proportion to the total network length. Once this relationship has been determined, the only operation to be performed is to place a grid with known sides on the network to be measured and count the points at which the network intersects the sides of the squares (Zăvoianu, 1967). The value obtained allows determination of the total length of the river network in the measured area, and the former divided by the latter then gives the drainage density.

Before applying the method, the most appropriate size for the squares must be established. To this end, various square sections with a side of 10 cm, representing an area of 6.25 km², were marked out on 1 : 25 000 maps in regions with various drainage densities. Networks of squares with various sides (0.5, 1.0, 1.5, 2.0, 2.5, and 3.0 cm) were then applied as described and the numbers of intersection points counted. In unclear situations or when a stream fell exactly on a line, the grid was imagined to be displaced by 1 mm, always




| Scale \ Grid | 0.5 | 1.0 | 1.5 | 2.0 | 2.5 | 3.0 |
|---|---|---|---|---|---|---|
| 1: 25,000 | 0.098 | 0.197 | 0.295 | 0.394 | 0.492 | 0.591 |
| 1: 50,000 | 0.197 | 0.394 | 0.591 | 0.788 | 0.985 | 1.182 |
| 1: 100,000 | 0.394 | 0.788 | 1.182 | 1.576 | 1.970 | 2.364 |
| 1: 200,000 | 0.788 | 1.576 | 2.364 | 3.152 | 3.940 | 4.728 |

Fig. 50. Relationship between total length $\Sigma L$ (km) of drainage network and number $N_i$ of intersection points for square grids with sides of 0.5, 1.0, 1.5, 2.0, 2.5 and 3.0 cm, on various scales, and table of coefficients $a$ in the equation $\Sigma L = aN_i$.

in the same sense (Verhasselt, 1961). The total length (km) of the river network was also measured for each map section using dividers having a 3 mm opening, and the values thus obtained were plotted versus the corresponding numbers of intersection points. The results (Fig. 50) show that the two variables are related by linear functions of the form $Y = mX$. The equations of the lines were established analytically and correlation coefficients $r$ were also calculated. Of the six grids used, that having a side of 1 cm gave the best value of the correlation coefficient, followed by those with sides of 1.5, 2.0 and 2.5 cm. The grid with a side of 0.5 cm yielded the lowest correlation coefficient.

Errors thus occur when the squares of the grid are either too large and do not encompass the network adequately, or too small and yield an excessive number of intersection points. Given the best size of the squares and the slope of the corresponding correlation line, the only operation to be performed is to count the intersection points of the network with the appropriate grid. The error amounts to 3% at most, while the time required is five times shorter than for direct measurements. The method may be applied on various map scales, taking the scale into account when converting the measured length into kilometres, and even to other networks (e.g., the density of roads in a given region).

The drainage densities calculated for 53 fifth-order basins within the Ialomiţa drainage system according to the above method range from 1.16 km $km^{-2}$ for the Crivăţ stream to 6.19 km $km^{-2}$ for the Ogrezeanca stream, the mean value being 3.60 km $km^{-2}$. The variance for the whole sample is 1.15, the standard deviation being 1.07 and the coefficient of variation 0.30 ; the skewness coefficient is 0.05. Generally, it is obvious that the lowest values occur in flatland regions, followed by those characteristic of mountainous zones, while the highest values are recorded for the Subcarpathian region. Mountain drainage basins have densities ranging from 2 to 4 km $km^{-2}$. In the Subcarpathian region, where there is a wide variety of rock type and structure, the drainage density has values between 2.0 and 6.2 km $km^{-2}$. The physico-mechanical properties of the rock in this region, together with man's action, favour marked relief fragmentation and increases of slope steepness, and hence an intensification of surface runoff and an increase of suspended load. Indeed, suspended load moves out of this region at a rate in excess of 10 t $ha^{-1}y^{-1}$, which contributes to the rapid degradation of sloping ground.

Given the fact that the greatest role in drainage density is played by lower-order streams with temporary runoff, a map of drainage density will be similar to a map showing horizontal relief fragmentation. From this viewpoint, horizontal fragmentation, taken as the length of the valley network per unit area, as used in geomorphology (Tufescu, 1966) differs from drainage density in terms not of magnitude but of connotation. The former is based on the length of the valley network, the latter on the lengths of channels. Considering that, as concerns fluvial processes, the formation of a valley requires a channel and that on maps the length of an elementary channel is obtained from the length of the corresponding valley, as in the contour crenulation method, it is obvious that the two terms refer to the same notion. It would thus perhaps be more useful to establish horizontal fragmentation from the width of interfluves and thereby avoid duplication of the drainage density.

## Dependence of Drainage Density on Physiographical Conditions

Being an important parameter in the characterization of river systems, drainage density has been the subject of many regional studies intended to simplify calculation methods, to establish its relationship to environmental conditions and to other morphometrical elements, or to determine how it is affected by man's activities. However, no study has expressed quantitatively all the factors on which drainage density depends.

In a detailed study, Strahler (1956b) showed that drainage density is inversely proportional to local relief for first-order basins, and depends on the Horton, Reynolds and Froude numbers. Strahler explained the mechanism whereby drainage density is related to accelerated erosion as the surrounding conditions are modified, demonstrating that the geometry of drainage systems tends to adjust continuously so as to achieve a steady equilibrium state between the processes resulting from transfers of mass and energy and the forms thereby created.

Research has shown that drainage density depends on a number of morphometrical elements, on climatic conditions and on the hydrological behaviour of the terrain, although, as stated above, no formula has yet been developed which includes all these factors.

*Dependence on morphometrical elements.* This should be studied starting from the very relationship which gives drainage density, which indicates that the latter depends on both the length of the river network and the drainage area. Melton (1958a) showed that drainage density presents no systematic variation in relation to drainage area. Gregory (1971) referred to such a variation but the correlation coefficient established by him was only $-0.46$. Starting from studies carried out previously by several researchers, Pethick (1975) also sought to find a relationship between the two elements. He found an inverse correlation for drainage basins in several regions of the globe, of the form

$$D_d = 6.675\ A^{-0.3366} \tag{113}$$

with a correlation coefficient of $-0.77$. The low correlation coefficient and the great scatter of the values demand consideration of the stream length also, since this enters into the calculation of drainage density. Indeed, if network length is taken as a parameter, families of lines are obtained in a logarithmic plot which define the points perfectly. Given that the dependence of drainage density on area and network length can be written as

$$D_d = f(A^{-1}\Sigma L) \tag{114}$$

then the value of the exponent $(-1)$ indicates that the families of lines with $\Sigma L$ as a parameter will make an angle of 45° with the abscissa (Fig. 51). Thus the modulus for each family of lines may be established from a logarithmic plot and a nomogram drawn for the calculation of drainage density as a function of the two variables. To calculate the modulus of the logarithmic scale for $\Sigma L$ as a parameter, a constant area is maintained on the abscissa for two or three sufficiently distant values; the values of the drainage density for various stream lengths are then obtained on the ordinate.

From an analysis of the formula used to calculate drainage density and the arrangement of points on the above nomogram, the following conclusions can be drawn :

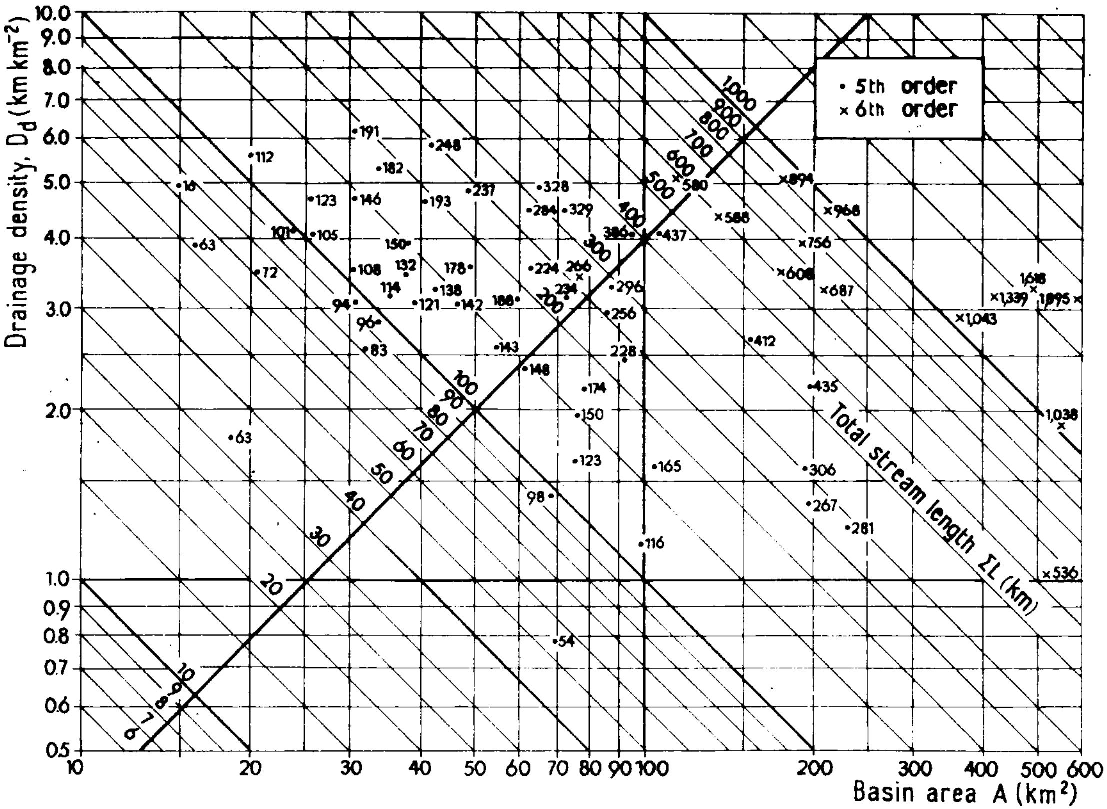



Fig. 51. Nomogram for determination of drainage density $D_d$ as a function of basin area $A$, with network length $\Sigma L$ as a parameter.

there is an inverse relationship between drainage density and basin area only when stream length remains constant (Horton, 1945); in this case, the larger the drainage area, the lower the drainage density, and vice versa; and of two or more basins with the same stream length, the density will be higher in that with smaller area;

if the area remains the same, drainage density and stream length are directly proportional; thus when two or more basins have the same area, the drainage density will be greater in the basin with greater stream length.

*Influence of geographical factors on drainage density.* Regional variations of drainage density are explained largely by differences in infiltration capacity (Horton, 1945) due to variations in permeability or transmissibility (Carlston, 1963). Rock resistivity and permeability are also included here. A low drainage density is favoured by rock showing high resistance to erosion or a highly permeable substratum (Strahler, 1956b) forming a relief with gentle slopes covered by dense vegetation.

According to Melton (1957), there is an inverse relationship between drainage density and Thornthwaite's effective precipitation index $P/E$, with a correlation coefficient of 0.943. However, Bandara (1974) found (for Sri Lanka) a positive correlation, with a correlation coefficient of 0.776.

The dependence of drainage density on hydrological factors ($H$), on slope ($S$) and on vegetation cover ($V$) was investigated by Hirsch (1967). To this end, after selecting study samples, Hirsch established a variation scale for each factor. Thus, five classes of hydrological behaviour were established, the value 1 corresponding to ground for which infiltration proceses are quite extensive and surface runoff is poor, and the value 5 to bare, impermeable rock with almost 100% runoff, the remaining values corresponding to intermediate cases. The same method was employed to quantify slope and vegetation. Although useful, these classifications are obviously somewhat subjective. Nonetheless, it has been shown that among the variables considered, there is a general relationship of the type

$$D_{\mathrm{d}} = aH \pm bS \pm cV - d \qquad (115)$$

in which the constants $a$, $b$, $c$ and $d$ differ from one region to another according to local conditions (Table 6). It should be remarked that in all the cases analyzed, the hydrological characteristics of a region are the most important in determining drainage density, followed by slope and vegetation cover.

Drainage density also changes with the passage of time, depending on the stage of geomorphological evolution of a region, on long-term changes in climatic conditions, and, more recently, as a consequence of man's action. There are many instances in which man, acting upon the environment to satisfy his needs more easily, has disturbed the equilibrium established among components of the environment over

TABLE 6

**Formulae for calculating drainage density in relation to hydrological characteristics *(H)* slope *(S)* and vegetation cover *(V)* for various regions in France (after Hirsch, 1967)**

| Region | Map scale | Formula | Remarks |
|---|---|---|---|
| Reims | 1 : 250 000 | $D_d = 0.74H + 0.33S - 0.21V + 0.53$ | |
| Troyes | 1 : 250 000 | $D_d = 0.77H - 0.46S + 0.18V + 0.68$ | |
| Cévennes | 1 : 500 000 | $D_d = 1.25H + 0.15S + 1.00V - 3.04$ | |
| Roubion | 1 : 50 000 | $D_d = 0.78H + 0.40S + 1.47V - 5.26$ | seasonal drainage |
| | | $D_d = 0.77H + 0.36S + 0.82V - 3.93$ | perennial drainage |

the course of time. Such actions may trigger chain reactions leading to accelerated erosion (Strahler, 1956b).

Drainage density also depends on discharge and on suspended load (Abrahams, 1972). An attempt to express the relationship between stream flow and drainage density was made by Carlston (1963), who, starting from Jacob's model and data for 13 basins in the United States, arrived at the conclusion that base flow $Q_b$ (cubic feet per second per square mile) is related to drainage density according to

$$Q_b = 14\ D_d^{-2} \tag{116}$$

In addition, a direct relationship of the form

$$Q_{2.33} = 1.3D^2 \tag{117}$$

was established for the maximum flow (return period 2.33 years).

F. W. Trainer (1969) (quoted by Dynowska, 1976), considering one-year flows in 10 basins with an area of approximately 5000 km$^2$, also found an inverse relationship between base flow and drainage density.

On the basis of data collected at several gauging stations in the Beskidy Mountains, Dynowska (1976) drew the conclusion that there is no relationship between drainage density and base flow expressed in absolute units, even for drainage basins situated in homogeneous geographical conditions. However, there is a good relationship between drainage density and the coefficient of base-flow supply (%), for which the regression curve yields a hyperbolic equation:

$$H_b/H_t = 12.328/D_d^2 + 25.97 \tag{118}$$

The coefficient of base-flow supply is the ratio of the base flow $H_b$(l s$^{-1}$km$^{-2}$) to the total runoff $H_t$ (l s$^{-1}$km$^{-2}$). The correlation coefficient $r$ between the two variables in eqn. (118) is 0.72 and is significant at the 5% level of confidence.

The preceding observations also demonstrate that drainage density has higher values in regions where streams are highly loaded with alluvia. This means either that drainage density bears on the amount of suspended load, in which case the relationship between them would be very important, or that both drainage density and the amount of suspended load are the result of the same factors.

Bratsev (1964) showed that there are instances when drainage density is higher in plains than in mountains, because stream lengths in the former regions may be 80–100% greater due to meandering. If it is intended to employ density in defining runoff, meandering must be omitted in calculating stream length because it bears no relation to the amount of water drained. This can be done by applying a correction which reduces river length in proportion to the extent of meandering. Then there is a direct relationship between runoff and drainage density (Bratsev, 1964).

*Relationship between drainage density and density of stream segments.* In a study of the geometrical properties of a sample of 156 mature drainage basins, Melton (1957) found a direct relationship between stream density and drainage density, which can be expressed by the formula

$$D_s = 0.694\ D_d^2 \qquad (119)$$

This formula was established with a correlation coefficient of $+0.97$. The errors obtained in applying this formula to other basins indicate that the value of the constant differs from one region to another in relation to the extent to which the basin topography is adjusted to environment conditions.

For example, for the basins analyzed in Romania it has been noted that an important role in determining stream and drainage density is played by rock type, and in particular by differences in the degree of consolidation and resistance to erosion. This may be seen by delimiting areas containing rocks in the same category of resistance and establishing the drainage and stream-segment densities within each category (see Table 7 below). A plot of the values obtained shows that these are ordered according to rock type, with distinct situations for consolidated and nonconsolidated rock (Fig. 52). For consolidated rock with a geological resistance ranging from 4 to 9, both drainage density and stream density are inversely related to the degree of rock resistance. Thus, the greater a rock's resistance to erosion and the greater its compactness and angle of internal friction, the smaller the number of streams per unit area and the lower the drainage density. To the extent to which these physico-mechanical properties are less marked, the values of the two latter elements increase, reaching a maximum in the case of the Cîndeşti gravel (Fig. 52). For nonconsolidated rocks, whose resistance coefficient ranges from 1 to 4, there is a direct relationship between geological resistance and stream or drainage density. For these rocks, which represent Pliocene or even Pleistocene

TABLE 7

**Schematic classification of lithological formations in the basin of the Ialomiţa stream, according to their resistance to perforation and shear**

| No. | Lithological formation | Category | $A$ (km$^2$) | $N$ | $L$ (km) | $D_s$ (km$^{-2}$) | $D_d$ (km km$^{-2}$) |
|---|---|---|---|---|---|---|---|
| 1 | Mica and Paleozoic sericite—chlorite slates | 9.0 | 55.0 | 331 | 146.6 | 6.00 | 2.66 |
| 2 | Teleajen strata (clay—marl slates, Albian—Vraconian curvicortical sandstones), Lower Cretaceous | 8.0 | 248.0 | 1 789 | 685.7 | 7.25 | 2.77 |
| 3 | Marls, lime marls, slates (Neocomian), lime sandstone and coarse sandstones, Bucegi conglomerates (Albian), lime breccias | 7.0 | 693.3 | 5 829 | 2098.0 | 8.40 | 3.03 |
| 4 | Şotrile flysch (lime sandstones, gray marls and gray—greenish clays), Pucioasa strata (marls, clays, breccias, etc.) | 6.0 | 87.3 | 956 | 329.0 | 11.00 | 3.77 |
| 5 | Sandstones, marls, plasters, conglomerates (Miocene), marl and sandy clays | 5.0 | 183.0 | 3 072 | 909.0 | 16.80 | 4.96 |
| 6 | Clays, sands, gravels of the Cindeşti strata | 4.0 | 25.0 | 511 | 138.0 | 20.40 | 5.52 |
| 7 | Clays, marls, sands, rare gravels (Pliocene) | 3.0 | 296.0 | 4 730 | 1328.0 | 16.00 | 4.48 |
| 8 | Sands, clays, Romanian Quaternary gravels | 2.0 | 160.0 | 1 321 | 600.0 | 8.30 | 3.75 |
| 9 | Gravels, sands, Upper Pleistocene loess deposits | 1.5 | 93.0 | 304 | 242.0 | 3.27 | 2.60 |
| 10 | Loess deposits | 1.0 | 128.0 | 177 | 110.0 | 1.38 | 0.86 |

formations, the values of the two elements increase from the loess and terrace deposits to the Cîndeşti gravel.

Therefore, the drainage and stream densities do not observe a vertical zonality, the highest values being recorded in the Subcarpathians and not in the mountainous areas, the amount of suspended load also

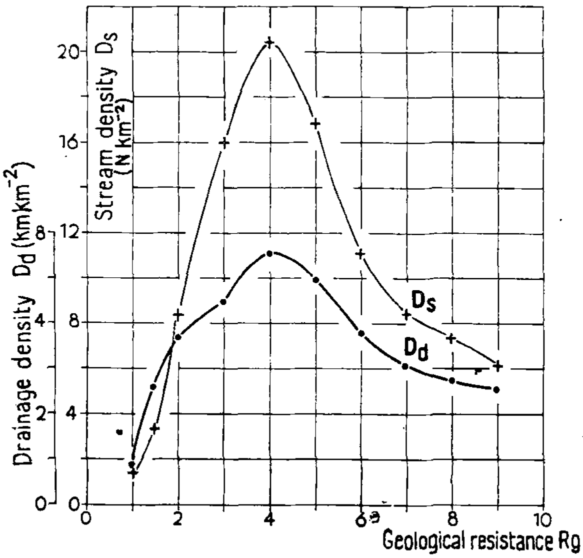


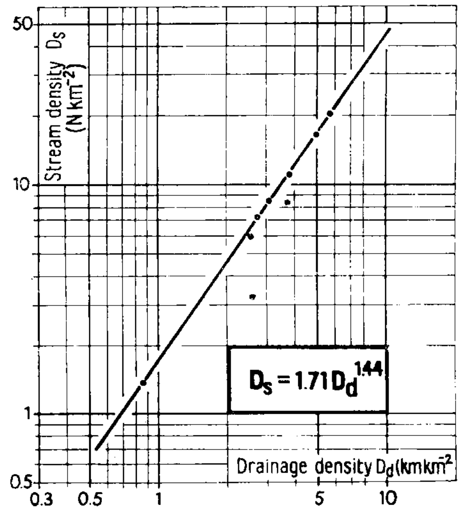

Fig. 52. Variations of drainage density $D_d$ and stream-segment density $D_s$ in relation to geological resistance for third-order basins situated on homogeneous rocks.

being greatest in the former region. Generally, in regions at high altitudes which are characterized by very hard rock such as the schists of the Leaota series, the drainage density is small. Descending to the Subcarpathian region, the number of stream segments increases progressively, reaching a maximum at the Cîndeşti gravel in the Subcarpathians where the relief presents the highest degree of fragmentation. The border area between the Subcarpathians and the plains is marked by a steady decrease in the number of stream segments and in drainage density, which reach their lowest values in the flatlands characterized by loess or alluvial deposits having small relief energy and slope but high permeability.

The preceding correlations of drainage density and stream density with resistance to erosion are followed with one exception, concerning the values for the Romanian Quaternary formations (sands, clay, gravel) (Fig. 53) occurring in regions with low relief energy, where the number of stream segments decreases very greatly and is no longer proportional to river length. The latter also increases because of the higher sinuosity and meandering coefficient, entailing an increase of drainage density. The relationship between stream density $D_s$ and drainage density $D_d$ for this region is given by the formula

$$D_s = 1.71\ D_d^{1.44} \text{ or } D_d = 0.690\ D_s^{0.702} \tag{120}$$

Fig. 53. Relationship between stream-segment density $D_s$ and drainage density $D_d$ for basins situated on lithological formations (see Table 7).




Therefore, there are proportional increases in the values of the two elements for these basins situated on lithological formations. This has a great bearing on discharge and suspended load, on the times required for flood formation and transmission, and on the rate of geomorphological evolution.

Chapter IX

# Vertical Distances

The vertical distances traversed by channels and drainage basins, which largely determine the latter's potential energy, are highly important as concerns both runoff processes and drainage-basin evolution. Their measurement is absolutely necessary in calculating the theoretical runoff energy of precipitation, the theoretical linear energy, the energy of slopes, etc. The hydraulic energy (kinetic and potential) of a mass of water in motion also depends on the fall $f$ traversed in releasing it.

In terms of geomorphological evolution, the potential energy $E_p$ of any material point in a drainage basin represents the energy amassed in reaching the present altimetric position from a lower plane or from sea level. For a mass $M$, it is equal to the mechanical work required to raise $M$ to a height $h$ against the force of gravity $g$. Therefore

$$E_p = \text{mechanical work} = Fh = Mgh \tag{121}$$

This presumes that the potential energy of a stone or rock on a slope or mountain peak is a measure of the mechanical work effected by internal and external forces in bringing it into this position. A mass of water, regardless of the point in which it is found, possesses potential energy in relation to any lower level, energy which although dependent on the vertical distance, is independent of the path followed by the mass of water in traversing it.

## Stream Vertical Distances

The vertical distance from a stream's head to the local base level (Broscoe, 1959; Strahler, 1964) appears in the specialized literature under various other names, such as stream relief (Fok, 1971) or stream fall (Yang, 1971). We here adopt the term stream fall in order to distinguish the vertical distances traversed by river networks from those of drainage basins, the term stream relief being more appropriate to the latter.

Stream vertical distance has several aspects :

*the main-stream fall* represents the vertical distance from the stream head to a gauging station, to a downstream confluence with a stream of an immediately higher order, or to sea level; it is given by the difference in elevation (m) between the two extreme points;

*the fall of a stream segment* is given by the difference in elevation between the two extreme points of the segment;

*The Horton—Strahler fall of a stream segment* is measured from the beginning to the end of a given order $x$.

For streams of a given order, the elevation differences between the points of origin and of confluence with another stream of the same order vary within fairly narrow limits. This is not surprising if it is borne in mind that these elevation differences depend on the resistance of the river bed to erosion, on the stage of paleogeographical evolution and on the amount of water running through the respective channel. Discharge, in turn, depends on basin order, and the greater the discharge, the higher its erosion and transport capacities. Hence channels tend to descend continuously such that the vertical distance between the initial and final points depends on the channel order.

*Law of summed stream falls.* By marking the altitudes of the extreme points of all streams in a given basin, it is possible to calculate the altitude difference or fall for each segment. On summing the individual values for the streams of each order (denoting the sums obtained by $F_1$ for the first order, by $F_2$ for the second, and so on), a series of values is obtained whose representation on semilogarithmic paper as a function of stream order indicates an inverse relationship (Fig. 54). From the cases analyzed, a rule can be derived which states that *the summed falls of streams of successively higher orders tend to form a decreasing geometric progression in which the first term $F_1$ is the summed fall of first-order streams and the ratio is the ratio $R_F$ of successive summed falls.* The calculation method for the ratio being known, the term of order $u$ in this progression is given by

$$F_u = F_1/R_F^{u-1} \tag{122}$$

In analyzing the ratio $R_F$ of this progression for a large number of basins in comparison with the confluence ratio $R_c$ given by the law of stream numbers, several situations are noted : the ratio of the summed stream falls may be higher, equal to, or smaller than the confluence ratio. In the second instance, the lines given by the two series are parallel. The values obtained by calculation for sixth-order streams in the Ialomiţa basin are very close to those given by field measurements, and define a line from which only the values for higher-order streams (fifth and sixth) deviate (Fig. 54A, D, F — H). In most instances, the values calculated for these orders lie above the line, resulting in a concave form, a fact remarked by Schumm (1956) with reference to the law of stream numbers.

Equation (122) can also be written in exponential form :

$$\log F = K - Mx \qquad (123)$$

in which the ordinate has a logarithmic and the abscissa an arithmetic scale.

The ratios of summed stream falls found for a sample of 53 fifth-order basins range from 2.18 to 7.36, with an average value of 4.73. The variance of the sample is 1.2, with a standard deviation of 1.10, a coefficient of variation of 0.23, and a skewness coefficient of 0.21.

The sum of the terms of the series can be calculated using the formula

$$\Sigma F = F_u(1 - R_F^u)/(1 - R_F) \qquad (124)$$

*Law of average stream falls.* First formulated by Fok (1971) as the law of stream relief, this law is expressed by the exponential equation

$$\log f_x = k - mx \qquad (125)$$

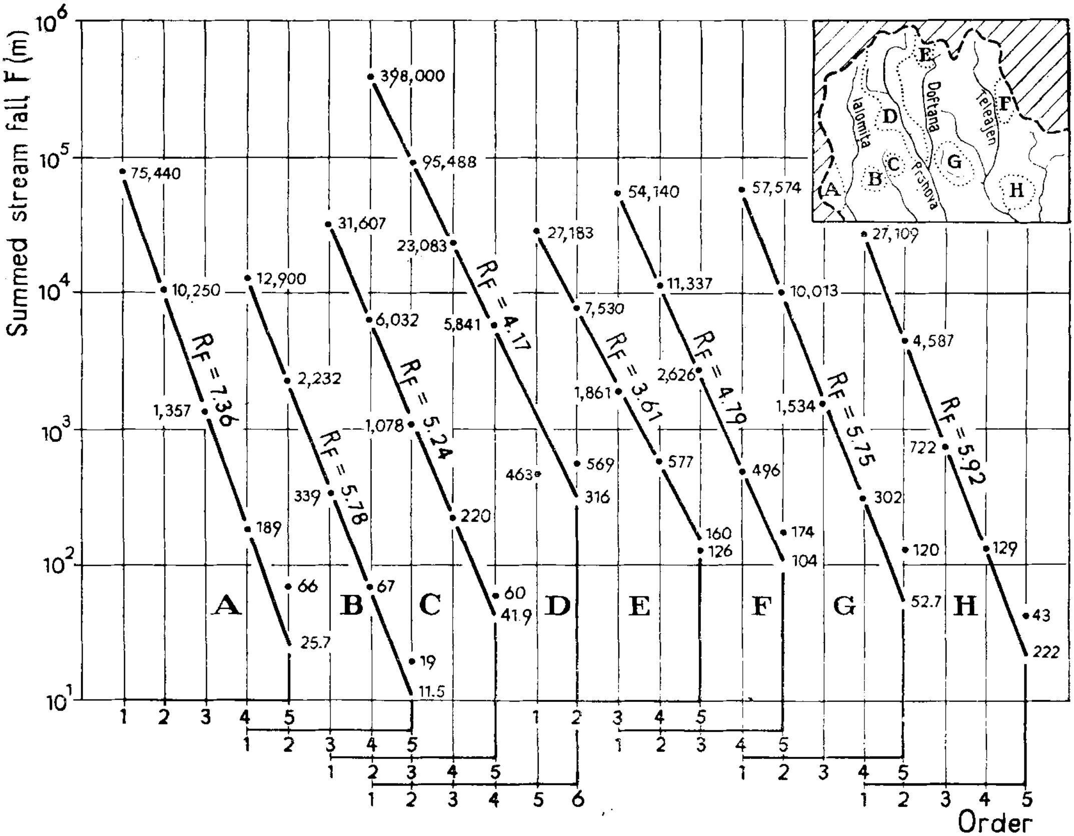
where $f_x$ is the average stream relief for streams of a given order $x$ in a basin, and $k$ and $m$ are constants. Yang (1971) adopted the concept of stream fall with reference to the same law, as employed in the present work.

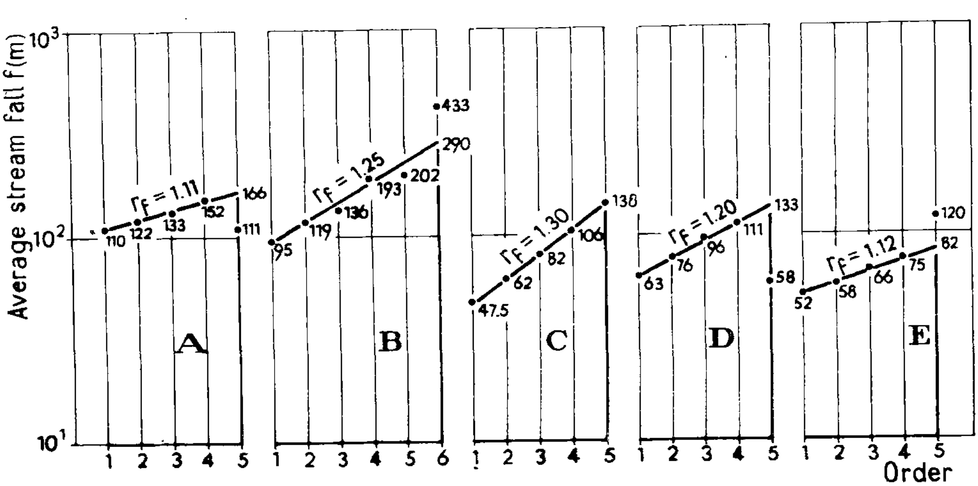
It has been seen that the summed stream falls for streams of successively higher order tend to form a decreasing geometric progression whose ratio can be established by known methods. The average falls between the heads and mouths of streams of various orders in a given basin may be calculated from the ratios of corresponding terms in the summed-falls and stream-numbers series. As expected and verified by measurement, *the average stream falls of streams of successively higher orders tend to form a geometric progression in which the first term is the average fall $f_1$ of first-order streams and the ratio is the ratio $r_f$ of successive average falls* (Fig. 55).

---

Fig. 54. Regression of summed stream falls on order for the basins of (A) the Vulcana at its confluence with the Ialomiţa, (B) the Tisa at its confluence with the Strîmbu, (C) the Sultanu at its confluence with the Tisa, (D) the Prahova at its confluence with the Doftana, (E) the Muşiţa at its confluence with the Doftana, (F) the Drajna at its confluence with the Ogrezeanca, (G) the Mislea upstream from its confluence with the Cosmina, and (H) the Bucovel at its confluence with the Iazu.

Fig. 55. Regression of average stream falls on order for the basins of (A) the Doftana at its confluence with the Muşiţa, (B) the Doftana at its confluence with the Prahova, (C) the Ogrezeanca at its confluence with the Drajna, (D) the Bertea at its confluence with the Vărbilău, and (E) the Mislea at its confluence with the Cosmina.


Summed stream fall, F (m)
$10^6$
$10^5$
$10^4$
$10^3$
$10^2$
$10^1$
A
B
C
D
E
F
G
H
$R_F = 7.36$
$R_F = 5.78$
$R_F = 5.24$
$R_F = 4.17$
$R_F = 3.61$
$R_F = 4.79$
$R_F = 5.75$
$R_F = 5.92$
75,440
10,250
1,357
189
66
25.7
12,900
2,232
339
67
19
11.5
31,607
6,032
1,078
220
60
41.9
398,000
95,488
23,083
5,841
463
569
316
27,183
7,530
1,861
577
160
126
54,140
11,337
2,626
496
174
104
57,574
10,013
1,534
302
120
52.7
27,109
4,587
722
129
43
222
Ialomita
Doftana
Teleajen
Prahova
Order


Average stream fall f (m)
$10^3$
$10^2$
$10^1$
A
B
C
D
E
$r_F = 1.11$
$r_F = 1.25$
$r_F = 1.30$
$r_F = 1.20$
$r_F = 1.12$
110
122
133
152
166
111
95
119
136
193
202
290
433
47.5
62
82
106
138
63
76
96
111
133
58
52
58
66
75
82
120
Order

To establish the equation of the line defined by the points corresponding to each order, the values of $F_u$ and $N_u$ from eqns. (122) and (27) may be employed. In this case, the ratio of the new progression is given by

$$f_u = F_u/N_u = F_1 R_c^{u-1}/N_1 R_F^{u-1}$$

Since $F_1/N_1 = f_1$ and $R_c/R_F = r_f$, the general term of the new progression may also be written as

$$f_u = f_1 r_f^{u-1} \tag{126}$$

The two lines given by the summed and average falls always intersect. The abscissa of this point, which may be established from the equations of the two lines, represents the order of the basin considered and should be the same as the order determined by means of the laws of areas or lengths.

The ratio of the average-falls progression indicates the tendency of the series: if the value is higher than unity, the geometric series is increasing; if it is below unity, the progression is decreasing. The ratios of the progressions established for the previous sample of 53 fifth-order basins range from 0.569 to 1.69, with a mean of 1.072. The variance for the same cases is 0.048, with a standard deviation of 0.22, the probability distribution curve having a coefficient of variation of 0.20 and marked positive skewness ($S_k = 0.51$).

The law of average falls is not always strictly obeyed, owing primarily to variations in rock type and structure as concerns resistance to erosion, a major factor determining stream falls. There are very many cases in which a channel leaves a rock layer or traverses the borderline between two layers having quite different resistances to erosion. In such situations, small falls may appear which change the longitudinal river profile and hence the usual distribution of the average falls. It is obvious that the higher the lithological homogeneity, the more closely will the law of average falls be followed. The expected correlation will not be matched perfectly for basins having marked tectonics and a varied lithology and structure.

From the viewpoint of evolution, it can be assumed that an increasing progression of average falls, with smaller values for lower-order and greater values for high-order streams, implies the existence initially of a sloping relief, with a continuous process of network branching and channel formation. Once the branching process has achieved a certain equilibrium in relation to the conditions specific to a given basin, continuous deepening of the network takes place over time, affecting primarily the lower orders, such that the average falls for successive orders may become more or less identical. As this process continues, with the falls of higher-order streams decreasing as a result of accumulation in the main channels, while erosion continues in lower-order basins, a decreasing geometric progression of the average falls will be found, indicating a more advanced stage of evolution. The sequence of phases occurs as described above only in the evolution of a drainage system without neotectonic movements, whose effects may thus be identified from a detailed analysis of the law of average falls.

## Drainage-Basin Vertical Distances

In addition to the geographical coordinates which indicate the position of a drainage basin on the globe, a series of characteristic points or planes related to its altitude or height must be known in order to define its spatial position accurately. These characteristics are important because they determine the amounts of matter and energy captured by unit area and the rates of transfer within a basin.

In order to define these elements, account must be taken of two reference planes to which all processes and phenomena taking place at the surface of a basin are related, either directly or indirectly: namely, sea level and the horizontal plane passing through the mouth of the basin. The latter, although not stable in all cases, usually does not show excessive variation and can thus be taken as a local base level.

Any point or plane in a basin has a certain altitude in relation to sea level:

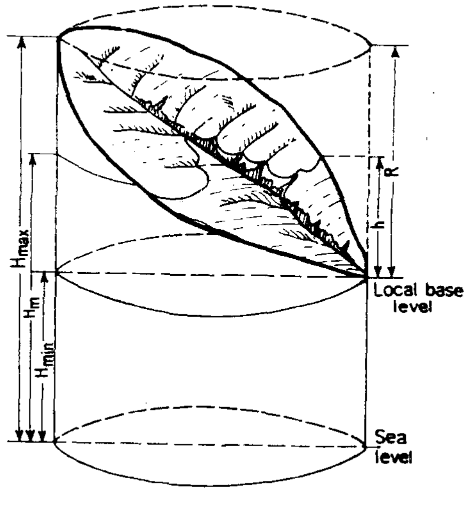



Fig. 56. Scheme showing characteristic vertical-distance elements for a drainage basin: $H_{max}$, maximum altitude; $H_m$, mean altitude; $H_{min}$, minimum altitude; $R$, maximum height or total relief; $h$, mean height.

*the mean altitude* $H_m$ is the mean vertical distance of the basin surface above sea level (Fig. 56);

*the maximum altitude* $H_{max}$ is the vertical distance from sea level to the highest point in the basin area; and

*the minimum altitude* $H_{min}$ is measured from sea level to the lowest point of the basin (Fig. 56).

If a point or plane in a basin is related to the horizontal plane which passes through the basin's lowest point, we speak of a *height*. As a rule, the lowest point of a basin is its mouth, which corresponds to the minimum-altitude plane. The following basin height elements should be mentioned.

*The basin height R*, also termed the total or maximum basin relief (Schumm, 1956), is given by the difference between the maximum and minimum basin altitudes (Fig. 56):

$$R = H_{\max} - H_{\min} \tag{127}$$

The basin height determines the potential energy of a basin and of the processes taking place at its surface. To avoid confusion, the term "maximum basin height" is used henceforth where it is necessary to make a distinction between basin height and mean basin height.

*The mean basin height h* is given by the difference between the mean and minimum basin altitudes:

$$h = H_{\mathrm{m}} - H_{\min} \tag{128}$$

Note that this and the preceding height element can be determined quite easily if the altitudes of the characteristic planes are known.

*Mean basin altitude.* This requires more detailed consideration. The idea of calculating mean basin altitude is by no means of recent origin. In 1816, A. Humboldt calculated the mean altitudes of peaks and passes using altitudes determined by barometer. In 1842, he sought to compute the mean altitudes of the continents, using a series of parallel profiles whose mean altitudes were weighted proportionally to their length. However, the principle of establishing the mean altitude of an area was set in 1854 by C. Koritska, who showed that the relief must be divided into horizontal sections assimilated to truncated cones. The mean altitude is then obtained by dividing the relief volume by the average area of the cone bases (Baulig, 1959). This principle was used subsequently by L. Neumann in 1886 and in many other cases, although the method was painstaking. Martonne (1940) described a further procedure for calculating the mean altitude of a region and subsequently worked for more than a year to determine the mean altitude of France and of its major natural regions. Langbein (1947) calculated the mean, maximum and minimum altitudes for about 350 gauging stations in the United States.

Soviet scientists (Luchisheva, 1950; Chebotarev, 1953) have recommended that the mean altitude be calculated from the weighted mean of the sum of partial volumes between contour lines, a procedure which has been used increasingly in the specialized literature:

$$H_{\mathrm{m}} = (a_1h_1 + a_2h_2 + \ldots + a_nh_n)/A \tag{129}$$

where $a_1, \ldots, a_n$ are the areas between successive adjacent contour lines, $h_1, \ldots, h_n$ the half-sums of the altitudes of successive adjacent lines, and $A$ is the drainage-basin area. There is also a shorter form of the expression, namely,

$$H_{\mathrm{m}} = \Sigma A_i h_i / A \tag{130}$$

where $A_i$ is the area between two contour lines and $h_i$ their mean altitude.

The mean altitude can also be calculated as the mean of the extreme altitudes, but the results are generally not so accurate :

$$H_{\mathrm{m}} = (H_{\mathrm{max}} - H_{\mathrm{min}})/2 \tag{131}$$

However, this method yields good results when applied to lower-order basins with a more homogeneous relief.

To assess the accuracy of the above two formulae for various conditions, a comparative study was effected by determining the mean altitudes of a large number of basins from the third to the sixth order. The following conclusions were drawn. The higher the basin order, the greater the error in using eqn. (131), the value obtained from the mean of the extreme altitudes being greater than those given by the area-weighted mean. Nonetheless, for fifth-order basins, the values determined by the two methods are correlated well (correlation coefficient 0.990) : for this order, the mean altitude can thus be evaluated from the two extreme altitudes using the formula

$$H_{\mathrm{m}} = 1.08H - 137 \tag{132}$$

where $H$ is the mean of the extreme altitudes. For fourth-order basins (Fig. 57), the correlation coefficient increases to 0.997, the difference between the values determined by the two methods is smaller, and the following formula can be employed :

$$H_{\mathrm{m}} = 1.01H - 34 \tag{133}$$

The correlation established for the third-order basins indicates still smaller differences between the values calculated by the two methods, so that in many instances they are identical (Fig. 58).

The lines determined for the third to fifth orders indicate that to the extent to which the order decreases, the coefficient $a$ in the general equation $y = ax - b$ approximates unity more closely, while $b$ comes very close to zero. It is thus obvious that for lower orders (first to third), the mean altitude can accurately be taken as the mean of the extreme values. This result is fully explained by the fact that lower-order basins generally have a homogeneous surface whose small area means that random disturbing events, which would distort the values obtained, will be very unlikely. Moreover, lower-order basins also show greater lithological and physiographical homogeneity. Thus, instead of using the weighted mean, whose calculation is time-

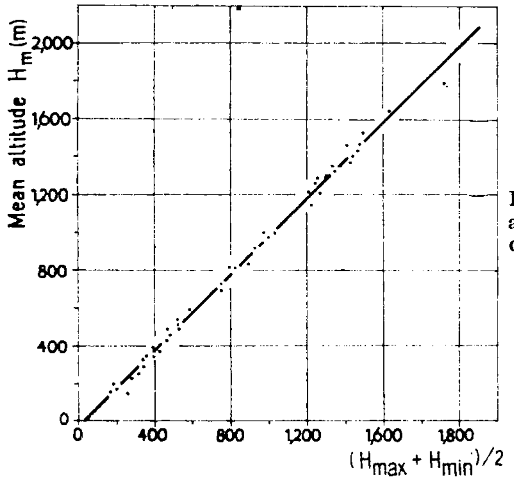



Fig. 57. Relationship between mean altitudes of fourth-order basins as determined from eqn.(129) and as mean of extreme altitudes.

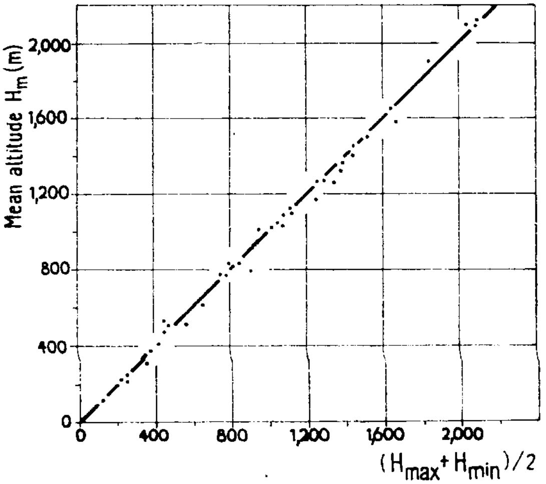



Fig. 58. Relationship between mean altitudes of third-order basins as determined from eqn. (129) and as mean of extreme altitudes.

consuming, the mean altitudes of first- to third-order basins may be calculated with reasonable accuracy as the mean of the extreme altitudes.

The mean altitude can also be calculated as the mean ordinate of the hypsometric curve (Remenieras, 1960), or by dividing the latter's integral (the volume) by the basin area (Merlin, 1965). In either case, construction of the hypsometric curve requires calculation of the same data as used in eqn. (129).

Kamalov (1965) proposed the following method for calculating the mean altitude of a valleyside or sloping area. A line $AB$ equal to the horizontal projection of the valley is drawn on the map and is then divided into equal parts. Through the corresponding points, denoted by $L_1, L_2, \ldots, L_n$, lines $a\,b$ are drawn on the map perpendicular to $AB$ and in turn divided into equal parts. The mean altitude of a line $ab$ perpendicular to $AB$ at point $L_1$ is calculated from the formula

$$\bar{h} = \sum_{i=1}^{n} h_i/n \tag{134}$$

where $n$ is the number of points taken on $ab$, and $h_i$ is the absolute height at point $i$ on line $ab$ perpendicular to the projection of the length of the slope. Once the mean altitude of each line $ab$ has been obtained, the mean altitude of the valleyside or area is calculated from the formula

$$H_{\mathrm{m}} = \sum_{i=1}^{m} \bar{h}_j/m \tag{135}$$

where $m$ is the number of lines perpendicular to $AB$, and $\bar{h}_j$ the mean altitude of line $j$. This method is quite similar to that of intersection points as used in determining drainage density.

*Law of summed mean altitudes.* If the mean altitudes of drainage basins of successive orders in the Horton—Strahler system are determined and the values are summed by order, the resulting series of values ($H_1$, sum of mean altitudes of first-order basins, $H_2$, sum of mean altitudes of second-order basins, etc.), plotted on semilogarithmic paper versus basin order yield a straight-line relationship which is followed by the data for a large number of basins situated in various physiographical conditions (Fig. 59(b)). An inverse relationship thus exists between the summed mean altitudes and basin order, such that *the sums of mean altitudes of drainage basins of successively higher orders tend to form a decreasing geometric progression in which the first term is the sum $H_1$ of the mean altitudes of first-order basins and the ratio is the ratio $R_H$ of successive summed mean altitudes.* The term of rank $u$ can be calculated from the formula

$$H_u = H_1/R_H^{u-1} \tag{136}$$

or, in exponential form, from

$$\log H = N - Ox \tag{137}$$

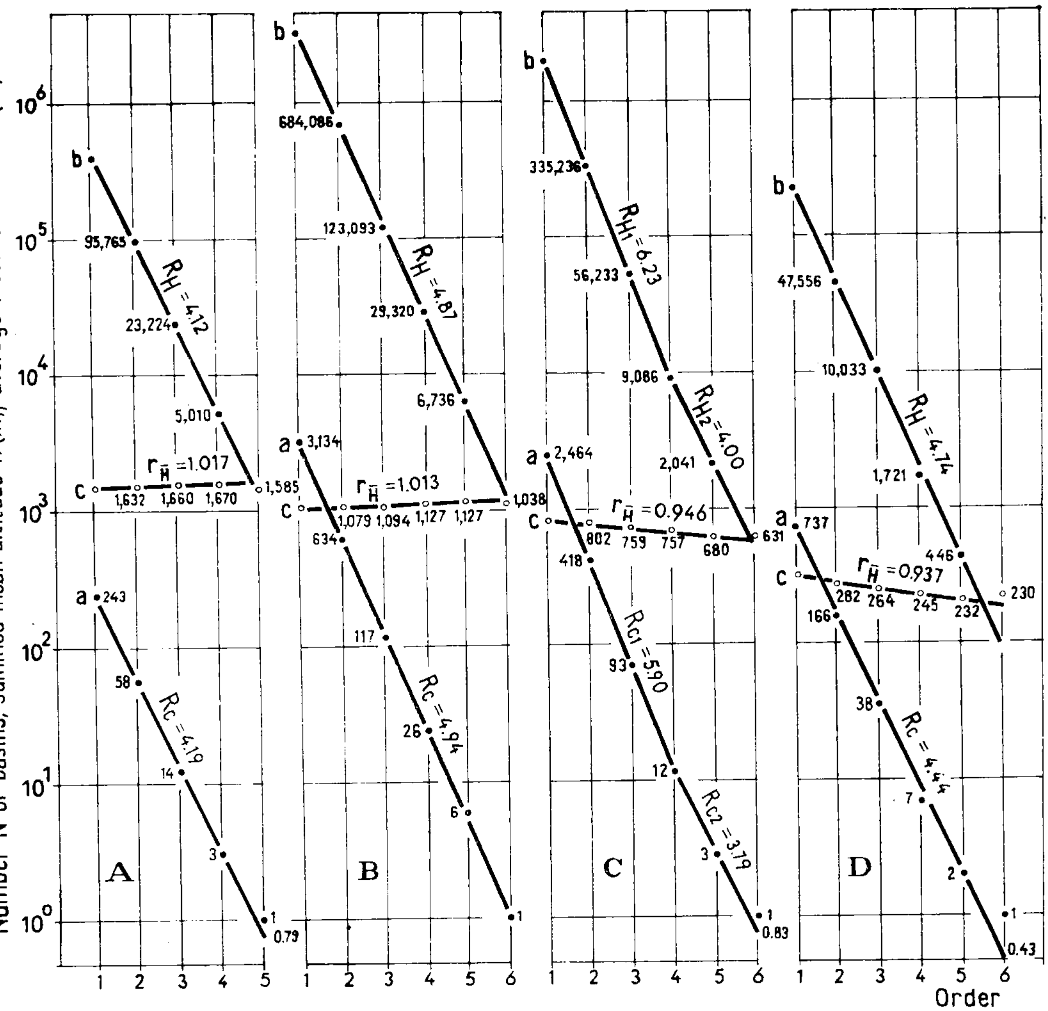



Fig. 59. Morphometrical models of mean basin altitude for the basins of (A) the Brătei at its confluence with the Ialomiţa, (B) the Doftana at its confluence with the Prahova, (C) the Vărbilău at its confluence with the Teleajen, and (D) the Bucovel at its confluence with the Teleajen (a, regression of numbers of streams; b, regression of summed mean altitudes; c, regression of average mean altitudes).

while the sum of the terms is

$$\Sigma H = H_u(1 - R_H^u)/(1 - R_H) \tag{138}$$

*Law of average mean altitudes.* By dividing the terms of the above series by the corresponding numbers of basins, the average

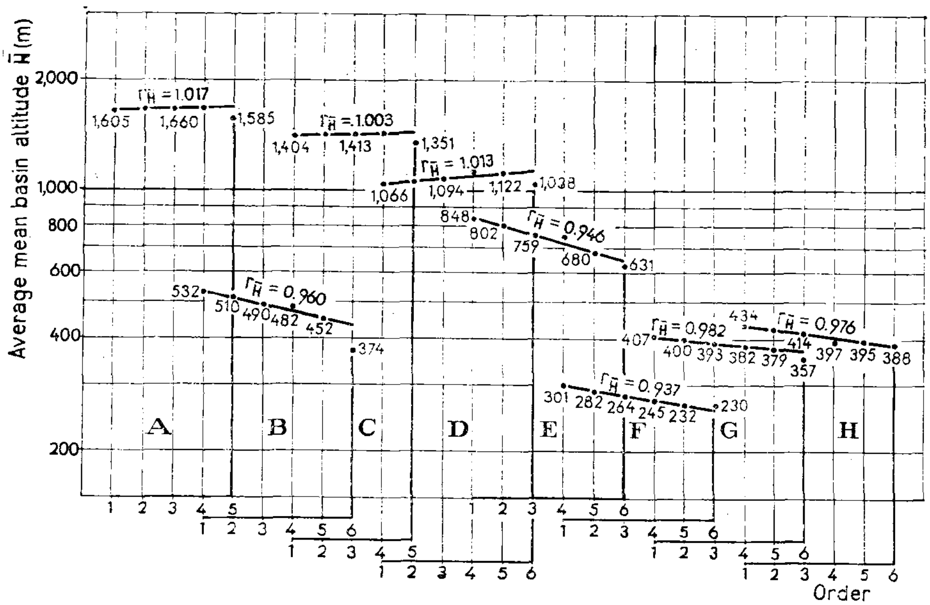



Fig. 60. Regression of average mean basin altitudes on order for the basins of (A) the Brătei at its confluence with the Ialomiţa, (B) the Cricovu Dulce at its confluence with the Ialomiţa, (C) the Azuga at its confluence with the Prahova, (D) the Doftana at its confluence with the Prahova, (E) the Vărbilău at its confluence with the Teleajen, (F) the Bucovel at its confluence with the Teleajen, (G) the Cricovu Sărat at its confluence with the Lopatna, and (H) the Lopatna at its confluence with the Sărăţel.

mean altitudes of basins of successive orders are obtained as

$$H_1/N_1 = \bar{H}_1,\ H_2/N_2 = \bar{H}_2,\ H_3N_3 = \bar{H}_3, \ldots, H_u/N_u = \bar{H}_u$$

The resulting series of terms, $\bar{H}_1$ being the average mean altitude of the first-order basins, $\bar{H}_u$ that of the basin of highest order, also depends on basin order, such that *the average mean altitudes of basins of successive orders tend to form a geometric series in which the first term $\bar{H}_1$ is the average mean altitude of first-order basins and the ratio is the ratio $r_{\bar{H}}$ of successive average mean altitudes* (Fig. 60). The general term of the series can be calculated from the formula

$$\bar{H}_u = \bar{H}_1 r_{\bar{H}}^{u-1} \tag{139}$$

in which $r_{\bar{H}}$ is given by the ratio $R_c/R_H$, or using the expression

$$\log \bar{H} = n + ox \tag{140}$$

where $n$ and $o$ are constants.

The above laws, verified with the help of the model of mean altitudes (Fig. 59), show various interrelationships which are highly significant. The ratio of the progression determined by the summed mean altitudes is generally very close to the confluence ratio. However, there are three cases giving differing tendencies of the line defined by the average mean altitudes. If the ratio $R_H$ of the series determined by the summed mean altitudes is smaller than the confluence ratio, the average mean values will form an increasing geometric progression with a ratio higher than unity, although very close to it (Figs. 59A, B and 60A, C, D). If the ratio of the summed mean altitudes is equal to the confluence ratio, their quotient will determine an average mean altitude series whose ratio will be unity. Finally, when the summed-altitudes ratio is higher than the confluence ratio, the average mean series will be decreasing and will have a ratio below unity, although very close to it (Figs. 59C, D and 60B, E — H).

*Law of summed maximum basin heights.* The vertical distance between the points of highest and lowest altitude in a basin area has been termed variously in the course of time. Strahler (1952 b, 1964) referred to it as the maximum basin relief, while Schumm (1956) termed it total relief, a term also used by Melton (1957) and Morisawa (1962). By relief, Strahler meant the difference in height or, more precisely, in altitude between two points. In order to avoid confusion between morphometrical and geomorphological relief, it is suggested that basin height be employed in the former sense.

Total or maximum basin relief, defined as the difference in altitude between the highest point on the water divide and the lowest one at the river mouth, was measured by Schumm (1956) "along the longest dimension of the basin parallel to the principal drainage line", while J. C. Maxwell (1960) (quoted by Strahler, 1964) measured it "along the basin diameter, an axial line found by use of rigorously defined criteria".

By calculating, as shown above, the maximum heights of basins of various orders in a given drainage system and summing the values for each separate order, a series of values is obtained which, represented on semilogarithmic paper in relation to basin order, demonstrates that *the summed maximum heights of basins of successively higher orders tend to form a decreasing geometric progression in which the first term is the sum $R_1$ of the heights of first-order basins and the ratio is the ratio $R_R$ of successive sums* (Fig. 61(b)). The ratio can be calculated as the weighted mean or by employing one of the methods mentioned previously. The general term of this decreasing progression is given by

$$R_u = R_1/R_R^{u-1} \tag{141}$$

or, in exponential form, from

$$\log R = T - Ux \tag{142}$$

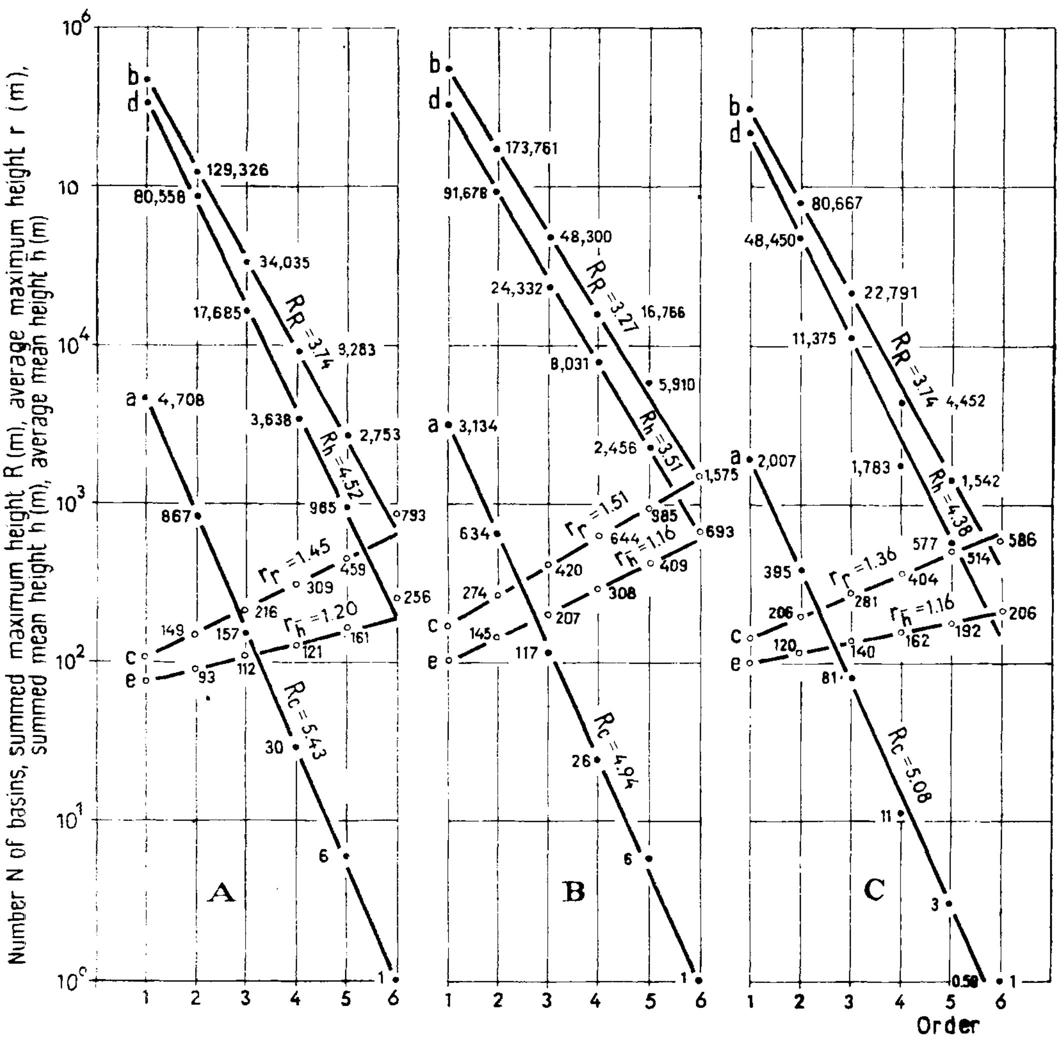


Fig. 61. Morphometrical models of maximum and mean basin heights for the basins of (A) the Cricovu Dulce at its confluence with the Ialomiţa, (B) the Doftana at its confluence with the Prahova, and (C) the Cricovu Sărat at its confluence with the Lopatna (a, regression of numbers of basins; b, regression of summed maximum heights; c, regression of average maximum heights; d, regression of summed mean heights; e, regression of average mean heights).

where $T$ and $U$ are constants and $x$ is basin order. The sum of the terms is given by

$$\Sigma R = R_u(1 - R_R^u)/(1 - R_R) \tag{143}$$

*Law of average maximum basin heights.* If the above series of values is divided by the corresponding series of numbers of basins, a new law is derived, as given by Morisawa (1962) : *the average maximum* *heights of basins of successively higher order tend to form an increasing geometric progression in which the first term is the average maximum heights $r_1$ of first-order basins and the ratio is the ratio $r_r$ of successive average maximum heights* (Fig. 61(c)). The equation relating the variables can be written as

$$r_u = r_1 r_r^{u-1} \tag{144}$$

or, in exponential form,

$$\log r = t + ux \tag{145}$$

If the two measured series of values considered in the computations themselves yield good geometric progressions, then the average maximum-heights ratio $r_r$ obtained by dividing the confluence ratio $R_c$ by the summed maximum-heights ratio $R_R$ will describe the new series accurately (Fig. 61A – C).

*Law of summed mean basin heights.* Although important, the maximum basin height does not faithfully define the position of the whole basin area relative to the plane of the mouth. To reduce the effect of high peaks on the water divide, basin height can also be calculated as the difference between the mean altitude of the whole basin area and that of the basin's lowest plane, giving the mean basin height. With greater approximation, relief can also be determined as the altitude difference between a point on the water divide and the nearest channel along the overland flow line. This procedure is recommended only for determining the height of valleysides. Since basin area is the most important element controlling the amounts of matter and energy received, the mean height is a more useful parameter in field activities. It is obtained merely by establishing the difference between the mean basin altitude and the altitude of the lowest point (the mouth).

If this operation is performed for a sufficiently large number of basins in a given drainage system and the resulting values are summed by order, the following rule can be established: *the summed mean heights of basins of successively higher orders tend to form a decreasing geometric progression in which the first term is the sum $h_1$ of the mean heights of first-order basins and the ratio is the ratio $R_h$ of successive sums* (Fig. 61(d)). The general term is given by the equation

$$h_u = h_1 / R_h^{u-1} \tag{146}$$

or

$$\log h = V - Zx \tag{147}$$

where $V$ and $Z$ are constants and $x$ is basin order. The sum of the terms for the whole series is given by

$$\Sigma h = h_1(1 - R_h^u)/(1 - R_h) \tag{148}$$

*Law of average mean basin heights.* Again, taking ratios of the summed mean heights and the numbers of cases for each order, a new series of values is obtained for the average mean heights, from which it results that *the average mean heights of basins of successively higher orders tend to form an increasing geometric progression in which the first term is the average mean height* $\bar{h}_1$ *of first-order basins and the ratio is the ratio* $r_{\bar{h}}$ *of successive averages* (Fig. 61(e)). The equation of the series can be written as

$$\bar{h}_u = \bar{h}_1 r_{\bar{h}}^{u-1} \tag{149}$$

or

$$\log \bar{h} = v + zx \tag{150}$$

where $v$ and $z$ are constants. Analysis of measured data in relation to the models used for determining the laws of maximum and mean heights indicates that the corresponding progressions are followed closely in practice, the number of cases deviating from the rule being fairly small (Fig. 61A – C).

Chapter X

# Slope

Slope may be defined as the tangent of the angle of inclination of a line or plane defined by a land surface. It is the result of a complex and continuous interaction between internal and external forces acting upon the Earth's surface. It depends on rock and climatic conditions, which may in certain regions be constant over long periods of time, and on the thickness, texture and mobility of surface layers of soil, organic matter, etc., which in turn depend on climate (Baulig, 1959). In a drainage system, valleyside and channel slopes control directly the potential and kinetic energy of water flows and thus the intensity of runoff, erosion and transport processes. These factors all tend towards a state of equilibrium in relation to overall local geographical conditions. The sine of the slope angle indicates the magnitude of the component of the gravitational surface acting to produce movement of solid bodies, water or soil particles down a slope (Strahler, 1956a). Slope also plays an important role in river processes. Neither the formation of runoff, the movement of floods, the power potential of river courses, the modelling and evolution of river channels, nor the erosion and transport processes occurring in the latter, can be approached without knowing the slopes of the land surface and river network.

## Slope of Ground Surface

The velocity of overland flow of surface water resulting from precipitation and snowmelt is determined by, among other factors, the slope of the ground surface. This explains why calculations of this morphometrical element, which is accordingly of great practical significance, have been a focus of research since the end of the last century. In 1890, S. Finsterwald and K. Peuker (quoted by Baulig, 1959) independently proposed a formula for determining the mean slope of a ground surface from map measurements, the slope of an area between two contour lines being given as

$$\tan\alpha = FL/A \qquad (151)$$

where $F$ is the contour interval, $L$ the average length of the contour lines, and $A$ the enclosed area in horizontal projection. In 1909, Emm. de Martonne recommended a similar method for calculating the mean slope of mountain massifs, employing the product of the mean length of two contour lines and their difference in elevation, divided by the area enclosed by them.

Wentworth (1930) proposed a simpler method for calculating the mean slope of the ground surface, using instead of the lengths of contour lines, the number $N$ of contours per unit length, determined using a square grid:

$$\tan\alpha = FN/0.6366 \tag{152}$$

where $F$ is the contour interval; the statistical coefficient of 0.6366 is the mean value of $\sin\alpha$ for all possible angles $\alpha = 0 - 90°$ between the contour lines on the map and the lines of the grid.

Horton (1932) recommended the intersection-line method for calculations of slope. Its purpose is to determine the mean distance $l'$ between contour lines. To this end, as shown by Langbein (1947, p. 137), the area "is subdivided into squares of equal size by lines forming the boundaries between adjacent squares. The number of contours crossed by each subdividing line is counted and the lengths of the lines are scaled. Then the average scale-distance $l'$ between contour crossings on the subdividing lines is

$$l' = \Sigma l/N \tag{153}$$

where $N$ is the number of contours crossed and $\Sigma l$ is the total length of the subdividing lines". Knowing the mean value of $\sin\alpha$ for all angles from 0° to 90°, namely, 0.6366, the average horizontal distance $L$ between contours will be

$$L = 0.6366l'$$

and the mean slope of the area is

$$\bar{S} = F/0.6366\ \Sigma l/N = 1.571\ FN/\Sigma l \tag{154}$$

where $F$ is the contour interval or difference in elevation. Tests have demonstrated the applicability of this method (Langbein, 1947; Verhasselt, 1961).

For a more rapid evaluation of valleyside slopes when the scale of the map and the difference in elevation between contour lines are known, a graph of slopes derived from topographical maps can be used, or similar, more detailed graphs can be drawn, according to convenience (Grigore, 1979).

## Mean Slope of Drainage-Basin Surface

In the specialized literature, the following formula is used to calculate the mean slope of a drainage basin:

$$S_b = h[(l_0 + l_n)/2 + l_1 + l_2 + \ldots + l_{n-1}]/A \tag{155}$$

where $h$ is the (constant) difference in elevation and $l_0, l_1, \ldots, l_n$ are the lengths of the contour lines considered; $A$ is the basin area (Luchisheva, 1950). In shorter form, this formula is being used currently in the specialized literature (Ogievsky, 1952; Roşca, 1959; Diaconu and Lăzărescu, 1965; Vladimirescu, 1978):

$$S_b = \Delta H \, \Sigma L / A \tag{156}$$

where $\Delta H$ is the difference in elevation and $\Sigma L$ the length of the contour lines considered.

For lower-order basins with reasonably homogeneous relief, the basin slope $S_b$ can also be established as the ratio of the maximum height to the basin length $L$:

$$S_b = (H_{max} - H_{min})/L \tag{157}$$

In studies performed in the United States (Schumm, 1956; Strahler, 1964), the relief ratio is used, defined as the ratio of the difference between the maximum and minimum altitudes to the greatest basin length measured along the main valley line, and thus similar to $S_b$ from eqn. (157). Using data for seven drainage basins, Schumm (1956) established good relationships between this ratio and drainage density, river slope, elongation ratio and suspended load. Morisawa (1962) showed that the relief ratio decreases with increasing stream order for drainage basins which are lithologically homogeneous: the logarithm of the relief ratio tends to show a linear relationship to stream order.

Since the maximum altitude of a basin, given by the highest point on the water divide, and the maximum altitude of the main river in the basin differ only slightly, the relief ratio approximates to a very great extent the slope of the parent stream, with which it is well correlated. However, for higher-order and less homogeneous basins, the relief ratio cannot be considered to indicate the true mean slope of the basin area. For more accurate calculations, an equation including the lengths of contour lines, the difference in elevation between them and the basin area must be used.

*Law of summed mean basin slopes.* If the mean slopes of drainage basins of successive orders are determined and the values obtained are summed by order ($S_{b_1}$, sum of slopes of first-order basins, $S_{b_2}$, sum of slopes of second-order basins, and so on up to the mean slope of the highest-order basin), by plotting on semilogarithmic paper the resulting values are found to vary such that *the sums of the mean slopes of basins of successively higher orders tend to form a decreasing geometric progression in which the first term is the sum $S_{b_1}$ of the mean slopes of first-order basins and the ratio is the ratio $R_{S_b}$ of successive summed slopes.* This relationship has been verified in a large number of cases and is illustrated graphically in Fig. 62(b). The general term of the progression is

$$S_{b_u} = S_{b_1}/R_{S_b}^{u-1} \tag{158}$$

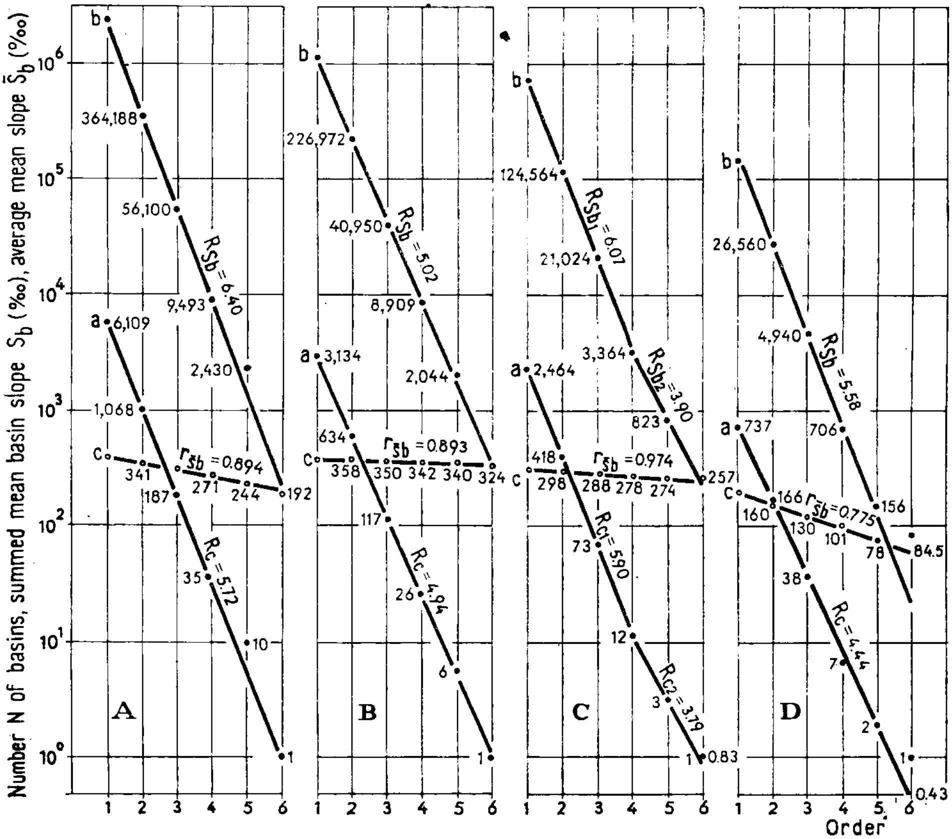



Fig. 62. Morphometrical models of mean basin slope for the basins of (A) the Ialomiţa at its confluence with the Cricovu Dulce, (B) the Doftana at its confluence with the Prahova, (C) the Vărbilău at its confluence with the Teleajen, and (D) the Bucovel at its confluence with the Teleajen (a, regression of numbers of streams ; b, regression of summed mean slopes ; c, regression of average mean slopes).

For basins developing in two relief units, although the sums of the mean basin slopes determine two geometric progressions with different ratios, the average mean values may not reveal such behaviour (Fig. 62C). Hence it is necessary to examine the variations of the sums for each separate order, this being apparently quite important from the geomorphological viewpoint.

*Law of average mean basin slopes.* The average mean slopes, obtained by dividing the terms of the preceding series by the corresponding numbers of basins, again form a geometric series in which *the average mean slopes of basins of successively higher orders tend to form a geometric progression in which the first term is the average mean slope $s_{b_1}$ of first-order basins and the ratio is the ratio $r_{s_b}$ of successive average mean slopes.* The ratio of the new series is given by the quotient

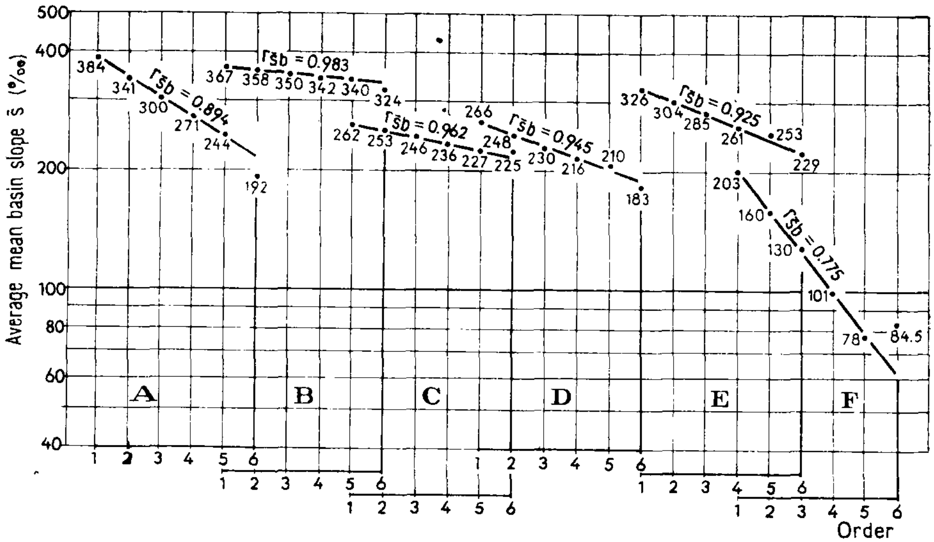


Fig. 63. Regression of average mean basin slopes on order for the basins of (A) the Ialomiţa at its confluence with the Cricovu Dulce, (B) the Doftana at its confluence with the Prahova, (C) the Drajna at its confluence with the Teleajen, (D) the Mislea at its confluence with the Teleajen, (E) the Cricovu Sărat at its confluence with the Lopatna, an l (F) the Bucovel at its confluence with the Teleajen.

of the ratios of the two constituent series, i.e., $r_{s_b} = R_c/R_{s_b}$, and the general term is

$$s_{b_u} = s_{b_1} r_{s_b}^{u-1} \tag{159}$$

The values determined for a large number of basins in the Ialomiţa system, situated in various geographical conditions, indicate that the highest average mean slope values are characteristic of basins in mountainous regions, followed by those in hilly and plain regions. For a basin which has developed in a mountainous relief unit, the ratio of the average mean slopes is close to unity, and hence the difference between the average mean slopes of successive-order basins is small (Fig. 63B — D). If a basin has developed over both mountainous and hilly or plain regions, as is the case for the basin of the Ialomiţa upstream from its confluence with the Cricovu Dulce, or for the border area between the Subcarpathians and the plain region, a smaller ratio of the progression is found, and hence there are differences between the average mean slopes of basins of successively higher orders (Fig. 63A, F). In the case of the Bucovel, a basin which has developed on an anticline, the upward movement of the crust has similarly resulted in a greater difference between the slopes of streams of successively higher orders, and hence in an acceleration of erosion and transport processes.

## Mean Channel Slope

Towards the base of a valleyside, overland flow enters the channel network, the flow of water being governed from now on by another regime, that of stream flow. This is determined by channel hydraulics, in which a very important role is played by the mean slope, defined as the tangent of the angle of inclination. This dimensionless parameter is given by the ratio of the fall $\Delta H$ between two points to the horizontal distance $L$ between them:

$$\tan \alpha = \Delta H / L \tag{160}$$

In field studies, the mean slope can be measured by means of a clinometer or theodolite, while in the laboratory topographical maps with contour lines are used.

There are numerous factors determining channel slope. It depends on the stage of geomorphological evolution of the region considered, on the rocks traversed by the channel and their structure and tectonics, and on basin area, which controls the water flow and hence the granulometry of sediment in the channel. As far back as 1877, G. K. Gilbert pointed out that stream slope is inversely proportional to water flow and to the basin area determining it while it is directly proportional to the size of sediment in the channel (Hack, 1957). Gilbert also noted the relationships between slope, lithology and tectonics.

Of the morphometrical elements of a channel network, slope is the most dynamic, tending to achieve equilibrium relatively rapidly in relation to local physiographical conditions determining the base level and the resistance of the substratum to the action exerted by water and suspended load. Slope governs most of the morphohydrographical processes taking place in channels, such as the rates of erosion, transport and deposition processes, and the amount of supended load. In a lithologically homogeneous area the magnitude of the channel slope is determined by the water flow rate, which in turn depends on the basin area.

## Channel Slope as a Function of Order

Being a highly characteristic element of a water course, slope has long been defined and used for practical purposes. Originally, the mean slope of a river was considered from its source to its mouth or along a particular sector, and there were no studies assessing the average mean slope of the entire stream network in a given basin. A first step along these lines was made by Horton (1945), who formulated a law of average slopes for streams of successive orders. The question was taken up by Strahler (1952b), Schumm (1956) and Hack (1957), among others, the law being applied in numerous situations. For instance, Hack (1957) established an equation relating channel slope to the ratio of sediment particle size and drainage area. Morisawa (1962) formulated the law of slopes in both exponential terms and as a geometric progression.

*Law of average mean channel slopes.* The average mean slopes of streams of successive orders may be determined from the laws of average lengths and of average falls, i.e., by applying the definition of mean slope as the ratio of the fall between the initial and final points of a stream segment to its length. The procedure is thus simply to take the ratios of corresponding terms in the two series (Fig. 64). Plotted on

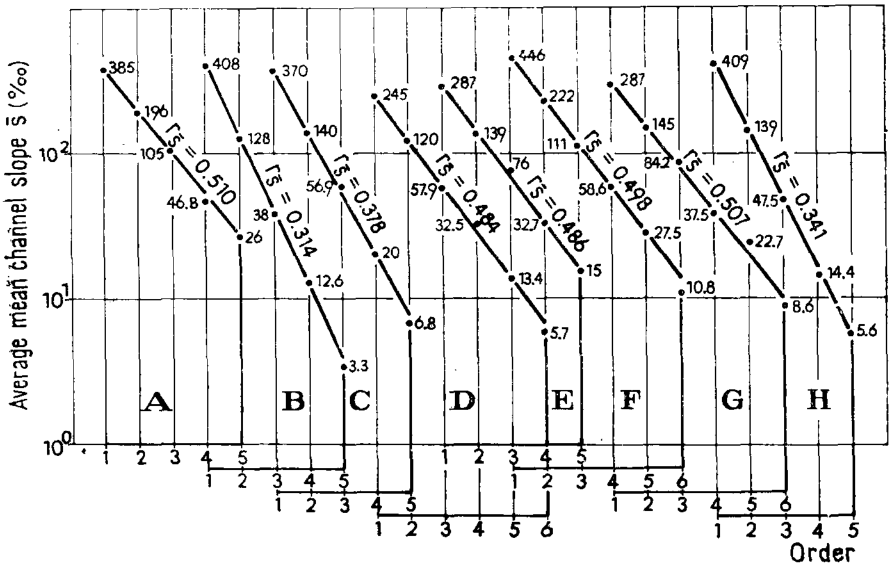



Fig. 64. Regression of average mean channel slopes on order for the basins of (A) the western Ialomicioara at its confluence with the Ialomiţa, (B) the Strîmbu at its confluence with the Tisa, (C) the Ursei at its confluence with the Cricovu Dulce, (D) the Ialomiţa at its confluence with the Cricovu Dulce, (E) the Purcaru at its confluence with the Doftana, (F) the Doftana at its confluence with the Prahova, (G) the Vărbilău at its confluence with the Teleajen, and (H) the Cosmina at its confluence with the Mislea.

semilogarithmic paper in relation to stream order, the values obtained indicate an inverse relationship between the two variables : *the average mean slopes of stream segments of successively higher orders tend to form a decreasing geometric progression in which the first term is the average mean slope* $s_1$ *of first-order streams and the ratio is the ratio* $r_{\bar{s}}$ *of successive average mean slopes or the quotient between the average-falls and average-lengths ratios* (Fig. 64). The general term of the progression is thus

$$s_u = s_1 r_s^{u-1} \tag{161}$$

The laws mentioned earlier yield this result directly. The general term for the average falls is given by eqn. (139) and for the average lengths by eqn. (90) (Fig. 65(I)A, B). Their ratio, given that $f_1/l_1 = s_1$ and $r_f/r_l = r_s$, gives

$$s_u = f_u/l_u = f_1 r_f^{u-1}/l_1 r_l^{u-1} = s_1 r_s^{u-1}$$

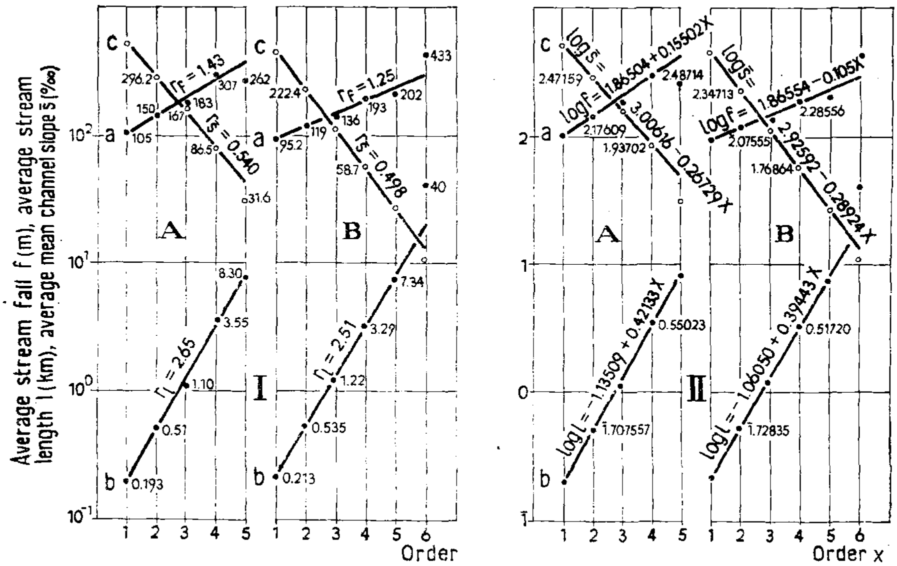



Fig. 65. Determination of average mean channel slope for stream segments of successively higher orders starting (I) from the properties of the geometric progressions for average stream falls and lengths, and (II) from exponential regression of measured data, for the basins of (A) the Prislop at its confluence with the Doftana, and (B) the Doftana at its junction with the Prahova (a, regression of average stream falls ; b, regression of average stream lengths ; c, regression of measured average mean channel slope).

The same result is obtained starting from the laws of summed lengths and of summed falls given by eqns. (84) and (122), respectively (Zăvoianu, 1974) (Fig. 66(I)A, B).

If the exponential forms of these laws are used, the same results are obtained but one of the methods mentioned previously must be employed to establish the constants. For example, to apply the chosen-points method, logarithms are established for the ordinates of pairs of measured points situated on each of the lines for the average falls, average lengths and average mean slopes. These points should preferably represent the same orders in all three cases. As a rule, the points for the highest order deviate from the lines and thus should not be used. Starting from the general exponential equations, a system of equations with two unknowns is established for each line. The constants are determined using the method of reduction or substitution, and the equation of each line may then be written (Fig. 65(II)A, B). The smallest-squares method can also be employed for more accurate calculation of the constants. Given the exponential equations for the average stream falls and lengths, the equation of the law of average mean slopes can also be determined by taking the ratio of the two equations. The same method is applied employing the summed falls and lengths, the final result being the same (Fig. 66(II)A, B).

Of all the laws considered so far, the law of average mean slopes is that followed most closely. This demonstrates that the channel slopes of a drainage network show the most marked tendency to reach a state of equilibrium as a result of mass and energy transfers in relation to the specific conditions in a basin. Accordingly, the average mean slopes of streams of successive orders in fifth- and sixth-order basins determine lines whose slope varies from one basin to another in relation to physiographical conditions (Fig. 64). Basins developing in the Carpathian region can thus be distinguished from those developing in the Subcarpathian and plain regions, such as the Ialomiţa basin upstream from the confluence with Cricovu Dulce (Fig. 64D). The deviation for the sixth-order streams within this basin is due to the fact that mountain streams have slopes several times steeper than those of Subcarpathian or flatland ones. Some streams also have discontinuities of slope caused by highly varying resistance to erosion of the rocks over which they flow, or by local structural features and tectonics. The average mean slope ratio $r_{\bar{s}}$ for the fifth-order basins considered ranges from 0.289 to 0.693, with a mean of 0.458. A low ratio implies a steep dip in the line connecting successive average mean slopes. Large differences between the slopes of streams of successive orders suggest rapid erosion, leading eventually to smaller slope differences and a state of equilibrium.

*Mean slope of entire channel network in a drainage basin.* As has been seen, determination of the mean slope of a stream segment or a river course from its source to a given point is a simple operation when the fall and the stream length are known. In fact, most studies which deal with this topic resort to the mean slope of the parent stream as a measure of the processes taking place in the basin considered. However, in a number of cases the mean slope of the main stream differs considerably from the mean slopes of other streams within a basin.

In any drainage basin, the main stream extends to the lowest altitude. However, as has been stated, there is an inverse relationship between flow rate and slope, which means that the main stream is also characterized by the smallest slope. Thus the mean slope of the main stream is not at all representative of the mean slopes of its tributaries, since the latter increase progressively, as evidenced by the laws analyzed. Now, the appearance and evolution of many hydrological and geomorphological processes are due to the entire channel network with all its features, and not simply to the main stream. Hence it is necessary to determine the mean slope of the entire river network in a given basin. From the situations analyzed, it results that the mean slopes of the channels in a given basin determine the mean slope of the basin area. There is a permanent tendency towards equi-

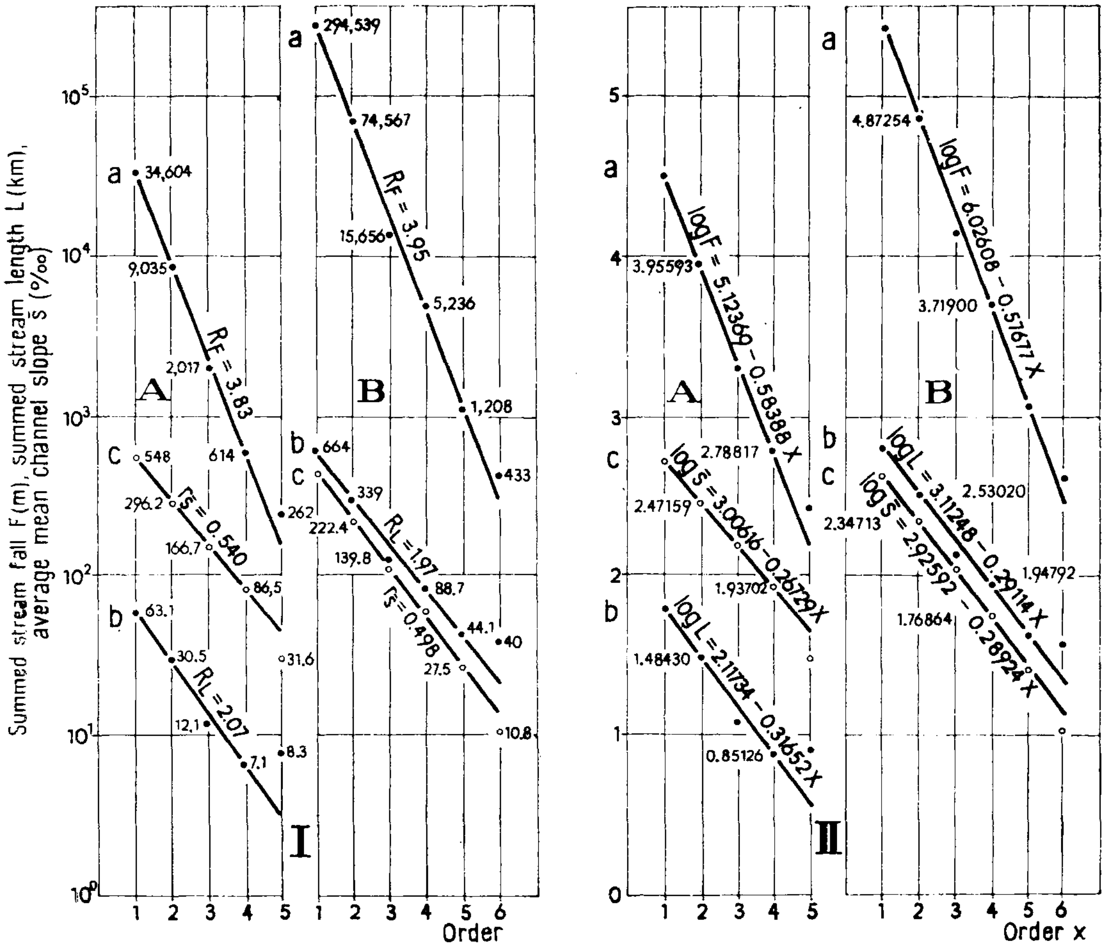



Fig. 66. Determination of average mean channel slope for stream segments of successively higher orders starting (I) from the properties of the geometric progressions for summed stream falls and lengths, and (II) from exponential regression of measured data, for the basins of (A) the Prislop at its confluence with the Doftana, and (B) the Doftana at its confluence with the Prahova (a, regression of summed stream falls; b, regression of summed stream lengths; c, regression of measured average mean channel slopes).

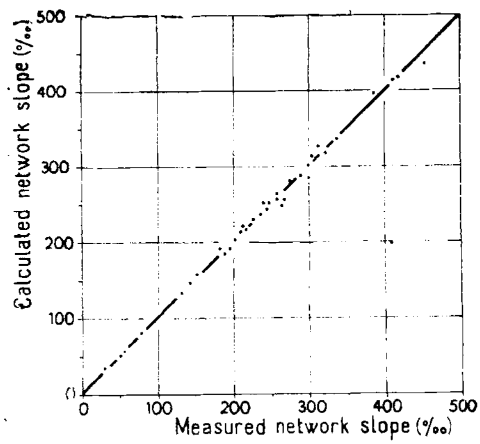



Fig. 67. Realationship between mean slopes of entire stream networks determined by calculation and by direct measurements.

librium between these two elements in relation to physiographical factors, primarily geological resistance to erosion.

The mean slope of an entire drainage network can easily be calculated using the Horton—Strahler ordering system. The starting point is again to determine mean slopes by using stream falls and lengths. It is known that the summed falls and summed lengths of streams of successive orders form two geometric progressions whose general terms are given respectively by eqns. (122) and (84). For these two progressions, the sums of the respective terms can be determined using eqns. (124) and (87). The mean slope of the whole drainage network is then calculated simply by taking the ratio of the two sums:

$$\bar{S}_n = \Sigma F/\Sigma L = [F_u(1 - R_F^u)/(1 - R_F)]/[L_u(1 - R_L^u)/(1 - R_L)] =$$

$$= F_u(1 - R_F^u)(1 - R_L)/L_u(1 - R_L^u)(1 - R_F) \qquad (162)$$

Therefore, the mean slope of the entire river network in a basin of order $u$ depends on the fall $F_u$, on the length $L_u$ of the segment of highest order, on the falls ratio $R_F$ and on the ratio $R_L$ of summed stream lengths in the basin. Since these values are known already from the previous progressions, application of the formula is a simple matter.

To determine whether the values given by eqn. (162) provide an adequate approximation for the mean slope of an entire river network, calculated values were compared with measurements for fifth-order basins within the Ialomiţa system. Good agreement is obtained between the values determined by the two methods (Fig. 67), demonstrating the applicability of eqn. (162).

*Relationship between mean slope of basin surface and mean slope of channel network.* If an elementary channel appears in an area, which will generally be inclined, its slope will evolve as a function of rock type and the amount of water it drains. The channel represents the line along which the modelling agent (water) acts with the greatest intensity, eventually diminishing the difference in height between the extreme points. Concomitant with the reduction in channel slope, valleyside slope processes carry new material downwards. Movement of such material out of the basin occurs quite rapidly, being governed by the duration and intensity of streamflow. These processes being continuous, accumulated modifications in the channel slope also bear on the processes occurring at the surface of the adjacent valleysides. However, the response of these regions, i.e., of the basin surface, is slower, and for streams of higher orders it depends on an aggregate of physiographical factors. Nevertheless, the basin surface tends to reach a state of equilibrium in relation to the mean slope of the river network draining it, although this process depends to a very great extent on the basin's homogeneity, its stage of evolution, the type of constituent rocks, and so on.

The mean slope of the basin surface has been correlated with the mean slope of the river network for the fifth- and sixth-order basins in the Ialomiţa system. The correlation indicates a direct relationship between these two elements, i.e., a decrease in the mean slope of the network entails a decrease in the mean slope of the basin area. Considering the sixth-order basins, those which have developed in mountainous regions have mean surface slopes greater than their mean river-network slopes. The basins of the Mislea and the Cricovu Sărat upstream from the Lopatna, which lie entirely within the Subcarpathian region, have mean river slopes greater than their mean surface slopes. This indicates extensive erosion and transport, which, in tending towards equilibrium, will result in increases in the mean slope of these basins. Nevertheless there are basins, such as those of the Ialomiţa upstream from its confluence with the Cricovu Dulce, of the Lopatna upstream from the Sărăţel, and of the Sărăţel, for which the two slopes are almost equal.

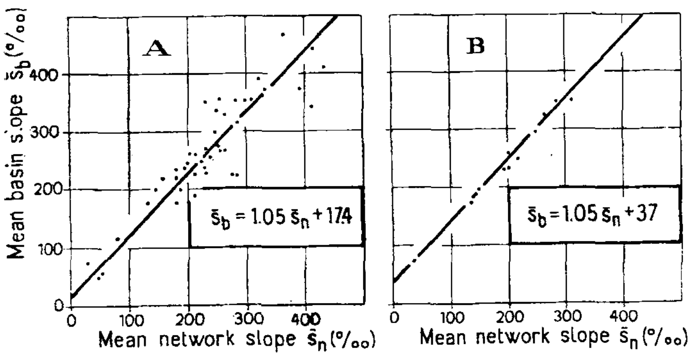



Fig. 68. Relationship between mean basin slope $\bar{s}_b$ and mean network slope $\bar{s}_n$ for (A) a sample of 53 fifth-order basins, and (B) the fifth-order basins in the Ialomiţa system upstream from the confluence with the Cricovu Dulce.

The two elements for the fifth-order basins considered (Fig. 68A) are related linearly with a high correlation coefficient of 0.897. It is noteworthy that precisely those values deviating from the relationship can be explained in terms of local physiographical factors : rock type, structure, tectonics and neotectonics. The great diversity of geographical conditions in the Ialomiţa basin could only generate considerable deviations of the values for some of its subbasins. For more homogeneous subbasins, such as that of the Ialomiţa at its confluence with the Cricovu Dulce (Fig. 68B), the correlation coefficient of 0.994 indicates, as expected, a closer relationship between the two slopes, reflecting their tendency to achieve a state of equilibrium in relation to local geographical conditions.

*Relationship between mean slopes of entire channel networks and drainage basins and geological resistance.* Of the factors controlling the various morphometrical elements, that most clearly expressed is the resistance of the geological substratum. Indeed, numerous elements, such as the areas of basins of various orders, stream lengths and falls, and, as a result, the mean slopes of channel networks, are directly related to the geological resistance and tectonics of a region.

A direct relationship exists between the mean slope of the entire channel network in a given drainage basin and its geological resistance (Fig. 69A). That two curves appear here is fully explained by the fact that two markedly different groups of rocks make up the geological surface of the catchment area considered. In the case of consolidated rock, which is highly resistant to erosion, the channel-network slopes vary between 435 and 185‰ (Fig. 69A(b)). In the case of nonconsolie dated rock, i.e., rock whose resistance is lower than 5, the mean slop- of the channel network varies from 34‰ (the Pîrscov basin) to 280‰ (the Cosmina basin at the confluence with the Mislea) (Fig. 69A(a)).

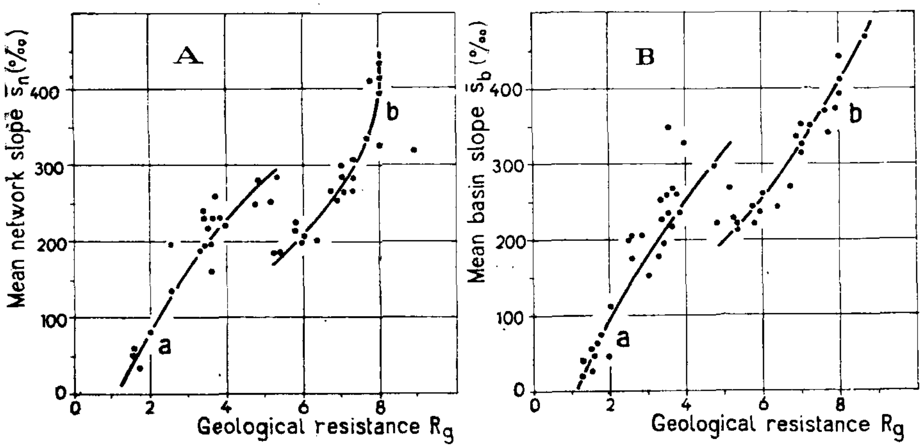



Fig. 69. Relationships of (A) mean network slope $\bar{s}_n$ and (B) mean basin slope $\bar{s}_b$ to category of geological resistance $R_g$ for a sample of 53 fifth-order basins (a, nonconsolidated rock; b, consolidated rock).

The curves of the mean slope for fifth-order basins also vary in relation to the category of resistance to erosion. For consolidated rock, the slopes range from 467‰ (the Brătei basin) to 215‰ (the basin of the Purcaru, a left-side tributary of the Doftana) (Fig. 69B(b)). For nonconsolidated rock, the greatest slopes are found for basins which have developed on the Cîndeşti gravel, such as the basin of the Rudei, a left-side tributary of the Cricovu Dulce, with a slope of 327‰ (Fig. 69B(a)), while the lowest values (21‰) are recorded in the Crivăţ basin.

On superimposing the two graphs, it becomes obvious that the mean channel slope is generally less than the mean basin slope, except for the limits of the range of variation. At the upper limit the difference between the absolute values is nonetheless small, while at the lower limit the two curves virtually coincide.

The smaller channel slopes indicate that deep erosion in channels is more intense at the upper limit for nonconsolidated and at the lower limit for consolidated rock. This phenomenon is characteristic of the Subcarpathian region, where intense erosion and transport take place. Over geological time, there is also an obvious tendency of the mean basin slope to decrease with decreasing mean channel-network slope. From the geomorphological viewpoint, the situations are characteristic of early and late stages in relief evolution.

Examination of the sixth-order drainage basins demonstrates a similar relationship between the two elements. In the case of the mean channel slope, two curves can again be drawn for the two major groups of constituent rocks (Fig. 70A). However, the values for the mean basin slope define only one curve with a slight inflection corresponding to the border area between nonconsolidated and consolidated rock (Fig. 70B).

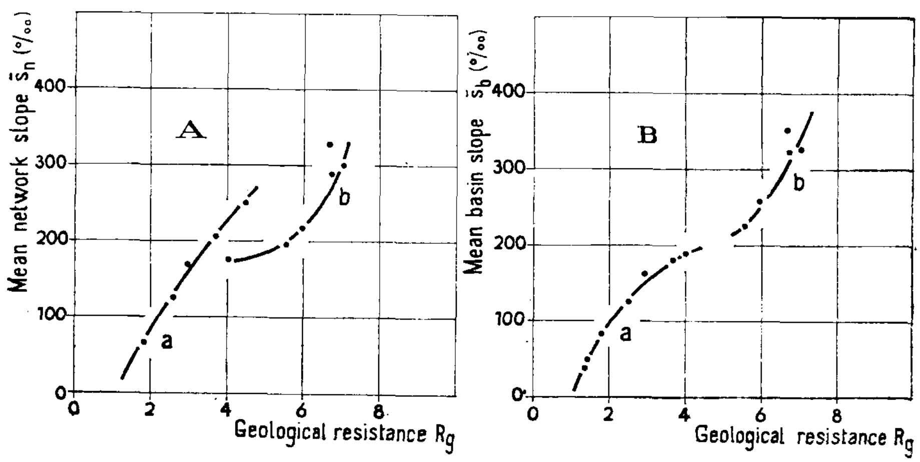



Fig. 70. Relationships of (A) mean network slope $\bar{s}_n$ and (B) mean basin slope $\bar{s}_b$ to category of geological resistance $R_g$ for a sample of sixth-order basins (a, nonconsolidated rock; b, consolidated rock).

It may therefore be concluded that the geological substratum, through its constituent rocks, tectonics and structure, influences to a very great extent numerous morphometrical characteristics of drainage basins.

*Relationship between average mean channel slopes and average areas of basins of successive orders.* It is known that channel slope depends directly on discharge: the greater the latter, the greater the hydraulic energy of stream flow and hence the higher the erosion and transport capacities. The latter two processes tend to diminish the fall between a stream's spring and mouth, and hence also its general slope. In turn, the discharge of a stream is directly proportional to basin area. It can thus be anticipated that an inverse relationship exists between the slope of a drainage network and the area of the corresponding basin

(Hack, 1957). In a more detailed analysis, studying this relationship as a function of drainage-basin geology, Hack noted a fairly marked influence of the latter. Thus, various curves linking average mean slope and area may be drawn with distinct parameters in the case of limestone regions, marl areas and other geological formations.

The average mean slopes of stream segments and the average areas of basins of successive orders are thus related inversely (Fig. 71). A plot of the two sets of values on logarithmic paper indicates a hyperbolic relationship of the form $S_n = K/A^m$, whose parameters $m$ and $K$ can easily be calculated from the coordinates of any two points :

$$m = (\log y_2 - \log y_1)/(\log x_1 - \log x_2)$$

and

$$K = y_1 x_1^m$$

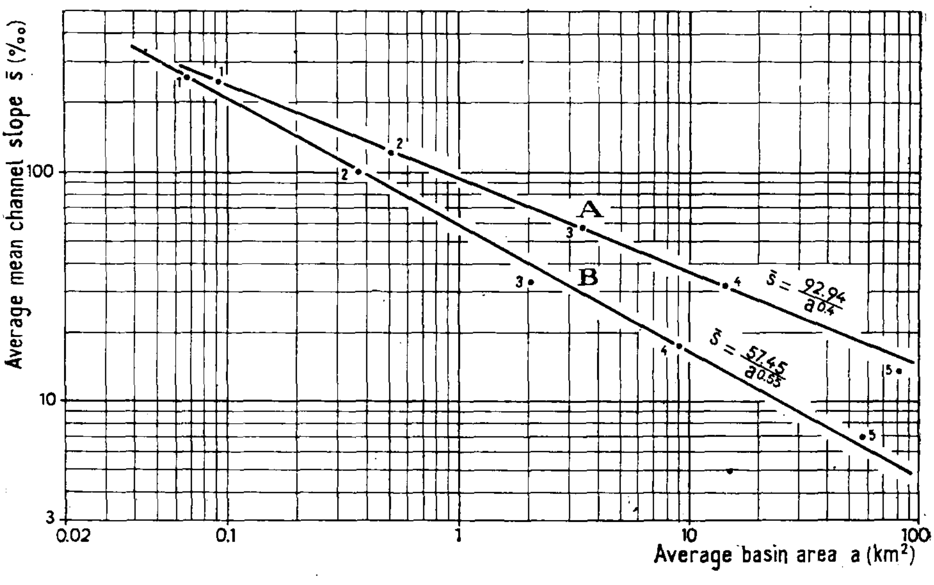



Fig. 71. Relationship between average mean channel slope $\bar{s}$ and average basin area $a$ for basins of successively higher orders for the basins of (A) the Ialomiţa at its confluence with the Cricovu Dulce, and (B) the Cricovu Dulce at its confluence with the Ialomiţa.

The expression can alternatively be written in logarithmic terms for the sake of greater simplicity. The equation is followed fairly closely in practice, with the exception of values for the highest-order basin. Deviations from the curve are due mostly to geological formations which, as has been seen, exert considerable control over morphometrical elements and hence drainage-basin area.

# Chapter XI

# Hypsometric Curves and Longitudinal Stream Profiles

## Hypsometric Curves

The basic idea of plotting hypsometric curves dates from 1854, when Karl Koritska showed that in order to determine mean relief altitude the relief must be divided into horizontal sections. However, the graphical representation of the curve is due to Albert de Lapparent, who, in 1883, developed a system of coordinates to represent the areas enclosed by several contour lines by a single curve (Baulig, 1959). The principle was improved and used with good results by John Murray in 1888 for sea beds, and by Langbein (1947) and Strahler (1952b) for drainage basins. Today, it is used frequently in the study of relief and even for analyzing the distribution of other elements in relation to altitude.

A hypsometric curve is a graphical representation showing on the abscissa the basin areas situated above various altitudes. If necessary, the basin areas can be given as percentages of the total. The hypsometric curve has also been termed the drainage-basin relief graph (Vladimirescu, 1978). Being a global representation, such a curve has the disadvantage of containing little or no information about certain significant relief features, in particular slope discontinuities, platforms and scarps, which are not at the same altitude throughout the basin area (Baulig, 1959).

*Hypsometric curves in absolute units.* The hypsometric curve of a drainage basin or relief unit is plotted by drawing the perimeter, marking chosen contour lines and establishing the enclosed areas planimetrically. This can be achieved in two ways.

(1) The planimeter is applied to the area between the basin perimeter and the contour line of highest altitude, then to the area between the highest and second-highest contour lines, and so on up to the area between the penultimate and final contour lines, the latter enclosing the point of minimum altitude. The partial areas $a_i$ thus obtained can be used to construct a graph of the basin area distribution by altitudinal increment, in which the area (km$^2$ or percentage) is represented in the form of a frequency-distribution histogram. By uniting the points at the horizontal midpoints of the class intervals,

a frequency curve can be obtained. The forms of such curves distinguish basins with the greatest proportions of small-, medium- or high-altitude areas (Appolov, 1963). For instance, the histogram of partial areas with a class interval of 100 m for the Doftana basin (Fig. 72) has the largest areas at moderate altitudes, between 900 and 1200 m, the mean altitude of the basin (1038 m) falling in this interval.

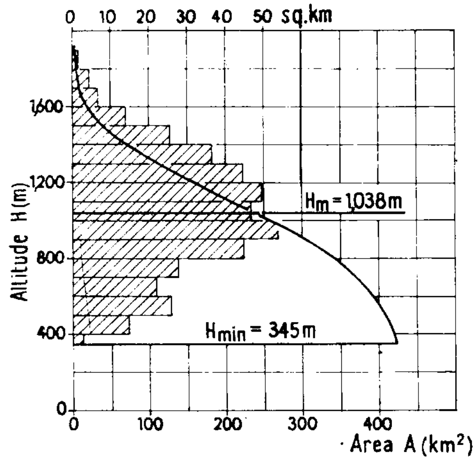



Fig. 72. Hipsometric curve and histogram (frequency distribution) of partial areas for the Doftana drainage basin.

The frequency-distribution representation of the areas can be used to construct a hypsometric curve of cumulative absolute frequencies (Réméniéras, 1960) whose abscissa shows the area fo a basin which is found above a given altitude. If the relative frequencies of areas are used instead of absolute values, the result will be a similar hypsometric curve but for relative frequencies. Again, percentages of the total area can be plotted on the abscissa and the result is a percentage hypsometric or probability curve.

(2) The planimeter is applied to the area between the contour line of highest altitude and the basin perimeter, then to the area between the second contour line and the basin perimeter, and so on up to the area of the whole basin (Strahler, 1952b). In this case, the only operation to be performed in obtaining the hypsometric curve is the graphical representation of the measured values. Starting from this method, partial areas can also be established from the differences between successive planimetric measurements.

The form of the graph showing the distribution of partial areas by altitudinal increment is also reflected in the hypsometric curve. Thus, if the peak of the former distribution corresponds to small altitudes, the hypsometric curve will be concave (Fig. 73a), and if it is found at high altitudes, the curve will be convex (Fig. 73c).

Hypsometric curves present a number of advantages in relief studies. They permit determination of the mean altitude of a drainage basin or region and can be used in calculating partial and total relief

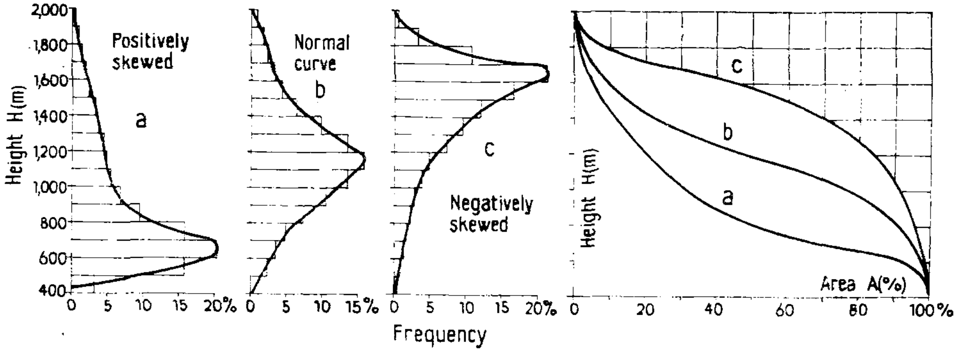



Fig. 73. Relationship between asymmetry of partial-area frequency-distribution curve and form of hypsometric curve.

volumes. To calculate the mean altitude, the area below the hypsometric curve is divided by the integral basin area (Vladimirescu, 1978). The mean altitude of the basin is also the mean ordinate of the hypsometric curve (Réméniéras, 1960).

Langbein (1947) also referred to the usefulness of hypsometric curves in hydrological and erosion investigations. In very detailed studies, Strahler (1952b) and then Schumm (1956) demonstrated the multiple possibilities of interpretation offered by hypsometric analysis and the relationship between the form of the hypsometric curve and the stage of geomorphological evolution of a drainage basin.

*Percentage hypsometric curves.* Through the work of Strahler (1952b), the interpretation of hypsometric curves took an entirely new course. Strahler showed concretely how a detailed analysis of such curves can provide valuable indications of the stage of evolution of a particular region or drainage basin.

In qualitative analyses of erosional forms, plotting of hypsometric curves in absolute units is not always advisable, because areas of various sizes cannot be compared, the slope of the curve obtained differing greatly in relation to scale. To remove this inconvenience, it is recommended that dimensionless parameters be used, whose values no longer depend on the absolute scale of the topographical forms considered (Strahler, 1952b).

For the hypsometric analysis of a drainage basin of any size, a section limited by two horizontal planes, one of which passes through the basin's mouth and the other through its highest point, is taken as the reference figure. These planes may change position in the course of time, should the reference points become displaced in altitude. To plot percentage hypsometric curves, either the percentage method or the grid-square method may be used.

*The percentage method* is similar to that employed for determining hypsometric curves in absolute units, i.e., the basin is delimited, the necessary contour lines are chosen and the areas between them measured planimetrically following one of the preceding procedures. At this stage, the values for the areas above a certain height, obtained directly or by summation, have dimensions of area. To convert them into dimensionless values which are always comparable irrespective of drainage-basin shape or size, they are simply divided by the total basin area $A$.

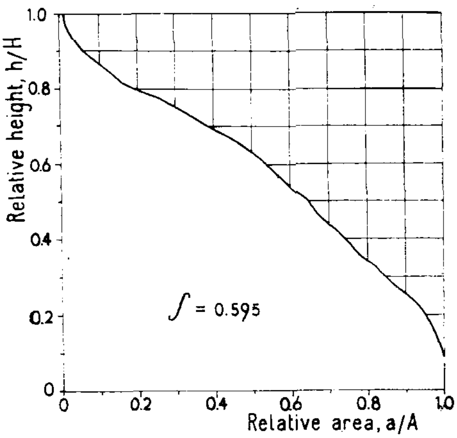



Fig. 74. Percentage hypsometric curve.

The values thus obtained increase from 0.0 for the highest point of the basin to 1.0 for the local base level, above which the entire basin area lies. Hence the abscissa of the hypsometric curve will comprise values ranging from 0.0 to 1.0 irrespective of the basin's size (Fig. 74). To further increase comparability, dimensionless values may also be used for the ordinate. This is done by determining the maximum basin height $H$ as the difference in altitude between the extreme points. The heights $h_i$ above which various basin areas $a_i$ lie are then divided by this value, giving ratios between 0.0 for the base plane and 1.0 for the highest point of the basin. Therefore the coordinates of the resulting hypsometric curve are ratios of areas (abscissa) and heights (ordinate).

*The grid-square method* was developed by Haan and Johnson (1966) to facilitate the plotting of hypsometric curves. It is based on the use of a net of points uniformly distributed with a certain density. After placing the chosen net (drawn on tracing paper) over the map of a particular basin, the numbers of points falling between successive contour lines are counted and their percentages calculated in relation to the total number of points in the basin area. The number of points between two contour lines being proportional to the area enclosed by them, the calculated percentage values may be used directly to construct a percentage hypsometric curve very close to that obtained using planimetric methods but with a computation time reduced by four to ten times.

As demonstrated by Strahler (1952b), percentage hypsometric curves are of great help in calculating the relief volume of a drainage basin, i.e., the volume enclosed between the topographical surface and the horizontal plane passing through the basin mouth. To this end, it suffices to take the products of the partial areas and the mean heights of the contour lines enclosing them, and sum the resulting values from the highest point to the reference base plane. Strahler then calculated the ratio of the resulting relief volume to the volume of the reference section between the upper and lower reference planes, obtaining a relative hypsometric integral representing the ratio of the area below the hypsometric curve to that of the rectangle delimited by the maximum values on the abscissa and ordinate (Fig. 74). Note that the curves established in absolute units and the percentage curves for the same drainage basin have the same shapes and characteristics points; the percentage curve is simply preferable when comparative studies are made.

In analyzing a large number of hypsometric curves for drainage basins situated in differing geographical conditions, the question has arisen of whether or not the higher plane of the reference section or volume, from which the topographical surface has developed through the modelling action of eroding agents, should be considered as horizontal. This is obviously highly important as concerns the hypsometric curves. For small, lower-order basins which have developed in homogeneous conditions and which can be considered to have started their evolution from a quasihorizontal area, the error resulting from this assumption will be small, and the hypsometric integral can be related to the area of the corresponding rectangular reference figure. For higher-order basins in regions with rough relief, the hypsometric integral should be determined carefully, since only in very rare instances can the initial relief be considered to have been horizontal. It is then necessary to determine the volume of the reference figure corresponding as closely as possible to the real initial situation. If the plane from which modelling action began cannot be established and the hypsometric curve must be related to the horizontal summit plane of the basin, the curve will not give the true distribution of areas by altitudinal step.

This becomes even more obvious on comparing the hypsometric curves for several sixth-order basins, for instance, for the Ialomiţa upstream from its confluence with the Cricovu Dulce, whose area-distribution curve has positive skewness, and for the Cricovu Dulce at its confluence with the Ialomiţa, or the Teleajen at its confluence with the Drajna, whose curves show an almost symmetrical distribution. The hypsometric integral for the Ialomiţa $\left(\int_p^m = 0.245\right)$ suggests that its basin relief is in a stage of advanced maturity. However, the hypsometric integrals for the neighbouring basins (the Cricovu Dulce and Teleajen), 0.415 and 0.475, respectively, indicate a much younger relief (Fig. 75). Analysis of the geomorphology and paleogeographical evolution of the three basins in relation to their geographical conditions

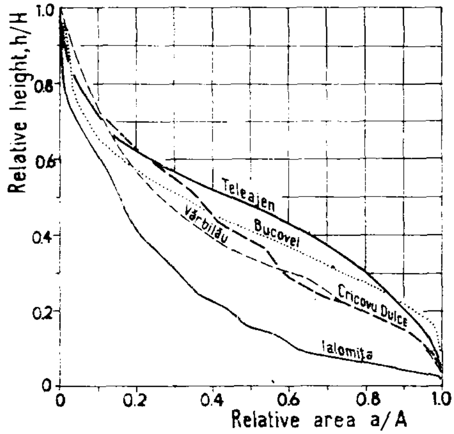



Fig. 75. Percentage hypsometric curves for sixth-order subbasins in the Ialomița basin.

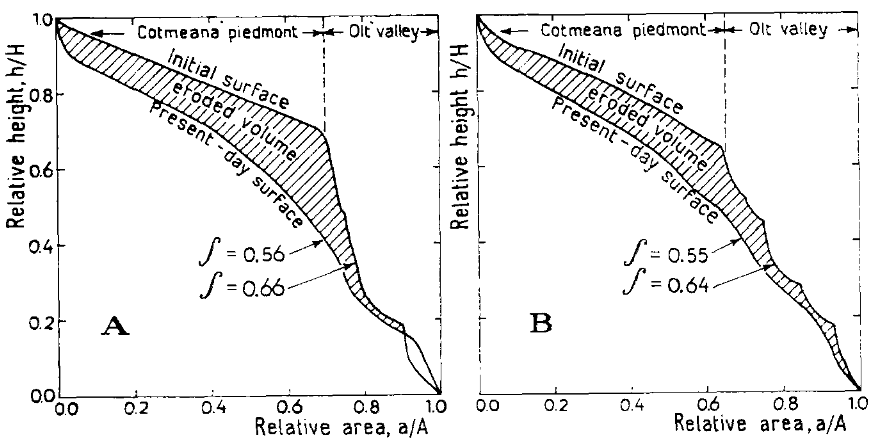



Fig. 76. Percentage hypsometric curves for (A) the Stăneasca and (B) the Trepteanca basins.

precludes such a flat differentiation between the Ialomița and the other two basins in terms of stage of evolution, even if reference is made to the mountainous and Subcarpathian zones.

For basins which have developed in a piedmont area or for basins whose initial relief can be reconstituted, the relative hypsometric integral should be determined by reference not to the volume of the rectangular figure defined by the two extreme coordinates but to the area of the figure corresponding to the initial relief. The method is then more time-consuming, but allows more accurate calculation of the hypsometric integral and of the amount of material eroded.

For comparison, two basins in the Getic Piedmont, drained by left-side tributaries of the Olt, were chosen for which the initial area on which modelling started could be reconstituted (Fig. 76). In the lower parts of these basins, the initial piedmont surface has been removed by the erosive action of the Olt and replaced by terraces. Concomitant with the deepening of the Olt and the formation of these terraces, the streams in the piedmont formations began to deepen and branch in proportion to the lowering of the base level. Even if the terraces continued to be modelled subsequent to their carving and fragmentation, this would have been the result mainly of valleyside rather than hydrological processes. Thus by reconstituting the initial surface of the piedmont and terraces, their initial hypsometric curve can be determined. To this end, the contour lines of the initial surface are plotted using the representative altitudes of the interfluves. Then, using a planimeter and the known methods, the corresponding curve is established, which represents the reference figure for the present hypsometric curve. Graphical representation of the two curves allows comparison and calculation of the initial and present relief volumes, their difference giving the amount of material removed from basin through erosion.

For instance, the areas encompassed by the initial and present hypsometric curves for the Trepteanca basin give hypsometric integrals, related to the area of the rectangular reference figure, of 0.64 and 0.55, respectively (Fig. 76). However, if the area enclosed by the present curve is related to the initial area, the integral obtained is 0.859, which means that of the initial volume of 17.35 km$^3$, only 2.446 km$^3$ has been carried away, i.e., 14% and not 45% as indicated by relating the hypsometric integral to the area of the entire rectangular figure determined by the coordinates.

Knowing the age of the initial area and data such as the above, an average rate of erosion and even a mean specific runoff of sediment yield can be evaluated. The method for representing the values can be chosen in relation to purpose: absolute or percentage values may equally be employed on one or both coordinates (Strahler, 1952 b).

*Application of the principle of hypsometric analysis to determination of water-balance elements.* Owing to their versatility, hypsometric curves can be used to study various phenomena. In Romania, the principle has been employed in establishing the parameters required for optimum siting of erosion-control projects on valleysides (Podani, 1978). In climatic studies, hypsometric methods can be used for snow surveys and to determine distributions of precipitation, runoff and evapotranspiration.

Indirectly, the principle can be applied to determine the basic elements of the water balance in a drainage basin or a particular region, starting from the formula

$$X = Y + Z \pm \Delta U \qquad (163)$$

where $X$ is precipitation, $Y$ runoff, $Z$ evapotranspiration and $\Delta U$ the variation of water reserves over a one-year period in the area analyzed (Diaconu, 1971b). Since the last term averages out over long periods of time, the formula can be simplified.

The terms in the formula are not equally well known. Over the last few decades, during which the bases of hydrometeorological networks have been established in many countries, systematic measurements of precipitation have been effected at large numbers of gauging stations. These have provided data over a sufficient period to indicate adequately the spatial distribution and seasonal variation of precipitation. For estimating total runoff, there already exists a network of basic gauging stations which have established data over a similarly long period for the major elements of the water regime.

A component of the water balance which is less well-established by direct measurements is evapotranspiration. Generally, this element is determined using empirical formulae which, although considering temperature and other climatic elements, are nevertheless approximative. Determination by direct measurements is painstaking and has been done sporadically, the number of lysimeter stations being too small to allow accurate regional generalizations. Given these circumstances, if precipitation and runoff have been determined, evapotranspiration is better estimated by the difference between these two elements. To this end, the principle of hypsometric analysis can be used to study the altitudinal distribution of the elements of the water balance in relation to relief.

This method has been applied to determine the mean evapotranspiration in the area corresponding to the 1 : 200 000 Tîrgovişte map sheet for Romania. First, areas by altitudinal increment were determined and summed appropriately to give the hypsometric curve (Table 8).

*Precipitation*. Estimation of this term in the water-balance equation raises a number of questions related both to the spatial distribution of rain gauges and to the amount and accuracy of direct data. Knowing the locations of the gauging stations, the complete series of data for long-established stations are used to extend the incomplete series by means of known statistical methods, and the mean for the entire period may then be calculated. To facilitate the computations and the comparison of precipitation and runoff, the same time periods should be considered for both. In the present example, the interval 1950—1975 was chosen, for which data are available for both elements. An equation may be established representing the dependence of precipitation on altitude (Fig. 77), allowing calculation of the amount (mm) of precipitation received at each altitude step, which may be converted into units of $m^3s^{-1}$ to facilitate comparison with runoff. The cumulative volumes from the maximum to the minimum altitude may be represented graphically to give a cumulative precipitation curve analogous to the hypsometric curve (see Fig. 79 below).

*Runoff*. Knowing the amount of water received, runoff must also be established, and to this end use is made of data provided by gauging stations for basins of various mean altitudes. After analysis of the data in order to determine possible modifications of the runoff regimes due to river engineering activities, the data for stations with incomplete series are correlated with those for stations with complete series and

TABLE 8

**Elements of the water balance for the Tîrgovişte map sheet (1 : 200 000 scale)**

| Altitude (m) | Area ($km^2$) | Precipitation ($m^3s^{-1}$) | Runoff ($m^3s^{-1}$) | Evapotranspiration ($m^3s^{-1}$) |
|---|---|---|---|---|
| 2000—2100 | 1 | 0.034 | 0.027 | 0.007 |
| 1900—2000 | 3 | 0.100 | 0.079 | 0.021 |
| 1800—1900 | 5 | 0.165 | 0.128 | 0.037 |
| 1700—1800 | 11 | 0.359 | 0.267 | 0.092 |
| 1600—1700 | 25 | 0.807 | 0.572 | 0.235 |
| 1500—1600 | 47 | 1.505 | 1.021 | 0.484 |
| 1400—1500 | 75 | 2.382 | 1.555 | 0.827 |
| 1300—1400 | 115 | 3.612 | 2.264 | 1.348 |
| 1200—1300 | 168 | 5.218 | 3.131 | 2.087 |
| 1100—1200 | 232 | 7.125 | 4.091 | 3.034 |
| 1000—1100 | 325 | 9.847 | 5.362 | 4.485 |
| 900—1000 | 468 | 13.949 | 7.133 | 6.816 |
| 800—900 | 629 | 18.467 | 8.952 | 9.515 |
| 700—800 | 880 | 25.337 | 11.496 | 13.841 |
| 600—700 | 1280 | 35.979 | 15.056 | 20.923 |
| 500—600 | 1840 | 50.388 | 19.131 | 31.257 |
| 400—500 | 2576 | 68.582 | 23.396 | 45.186 |
| 300—400 | 3 364 | 87.136 | 26.872 | 60.264 |
| 200—300 | 4778 | 118.490 | 31.152 | 87.338 |
| 100—200 | 5823 | 139.455 | 33.180 | 106.275 |
| 50—100 | 5824 | 139.472 | 33.181 | 106.291 |

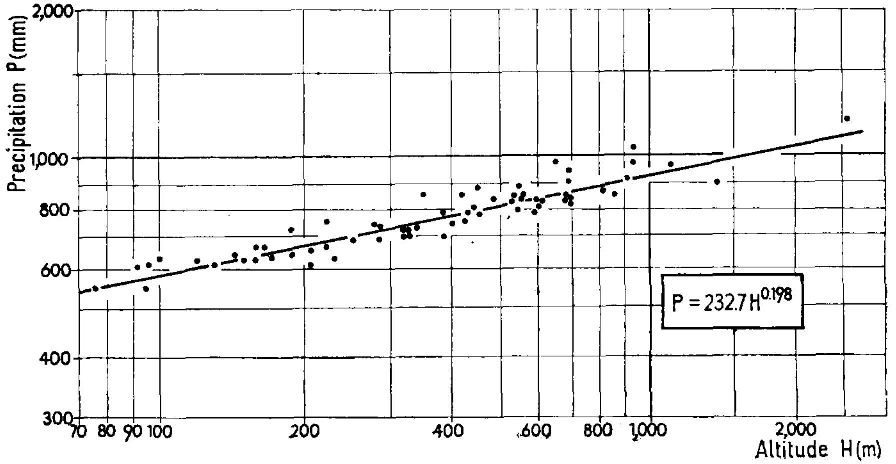



Fig. 77. Relationship between precipitation and altitude for the Tîrgovişte map sheet (1 : 200 000 scale, Gauss projection).

situated on the same river or in similar geographical conditions. After completion of the series and calculation of the mean flow rates in the given period, mean specific discharges $q$ (l $s^{-1} km^{-2}$) are calculated as

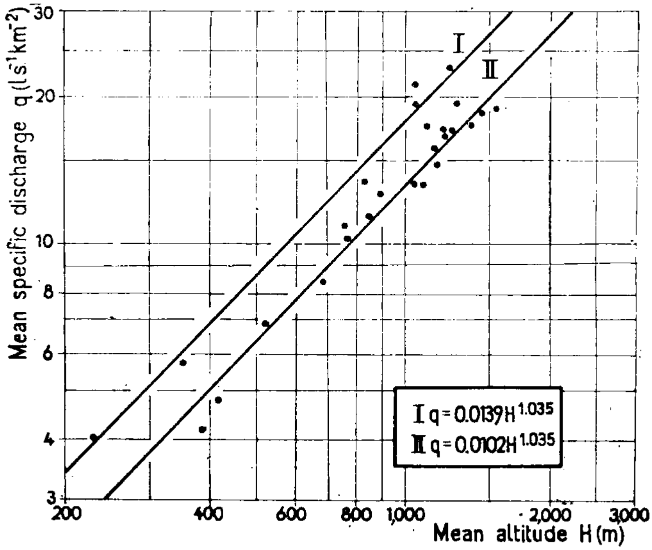



Fig. 78. Relationship between mean specific discharge and mean altitude of drainage basins.

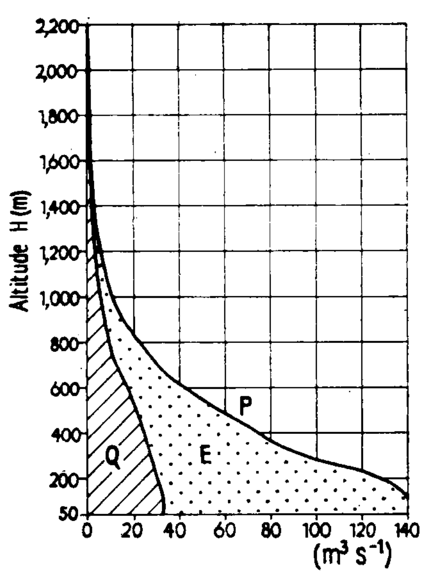



Fig. 79. Hypsometric curves for precipitation $P$ and discharge $Q$ for the Tîrgoviște map sheet (1 : 200 000 scale).

the ratios of the mean flow rates to the corresponding basin areas. The values obtained are correlated with the mean drainage-basin altitude and one or several lines are determined in relation to the conditions specific to particular basins or regions (Fig. 78).

Calculation of runoff volumes by altitudinal increment can be done only after the validity domain of each line has been established. Using the data obtained, cumulated from the maximum to the minimum altitude, a hypsometric curve can be plotted which, represented on the same figure as the corresponding curve for precipitation, gives a clear indication of the amounts of water received as precipitation which run off at various altitudes (Fig. 79, Table 8).

*Evapotranspiration.* From the water-balance equation, evapotranspiration can be calculated by taking the difference between the precipitation and runoff curves (Table 8). It should be noted that evapotranspiration shows clear vertical zonality, in the present example increasing from the highest peaks in the Southern Carpathians to the plains. In the area studied, of the total amount of water received as precipitation, only 23.79 % reaches the river network, the remainder (76.21 %) returning into the atmosphere by evapotranspiration (Zăvoianu, 1980).

## Longitudinal Stream Profiles

Being important in both geomorphological and hydrological studies, results obtained from the analysis of longitudinal stream profiles have attracted the attention of many researchers. Thus, in the last five decades, remarkable results have been recorded concerning both profile representation and analysis of the genetic factors generating various types of profiles (Jovanović, 1940; Makkaveyev, 1955), and the estimation of form (Ivanov, 1951). A quantitative expression has been found relating stream slopes, basin area and channel sediment size, accounting for the similarity of profile forms in areas with the same type of rock (Hack, 1957). The types of formulae which can be used to approximate the form of entire profiles, or of their sectors, were studied by Broscoe (1959), and detailed analyses have been made of equilibrium channel profiles (Birot, 1961) and of the similarity between the stages of evolution of longitudinal stream profiles and thermal systems (Devdariani, 1963).

The longitudinal profile of a stream, in the form of a plot of altitude as a function of distance along the course, will depend on numerous determining factors among which important roles are played by the initial relief, by rock type, by the stage of paleogeographical evolution and by positive or negative tectonic movements.

Generally, the form of a profile, and especially its degree of concavity, is the result of several factors, of which the following may be mentioned:

the concentration of the discharge collected by a drainage network into single channels, which determines the transport capacity of the main stream;

the wearing away of the sediments in a given channel through the action of water, resulting in decreased sediment particle size and thus greater ease of transport; and the decreasing slope of valleysides from the source to the mouth in most drainage basins, which is a major factor determining the size distribution of material supplied from the valleysides to the channels (Birot, 1961).

In a detailed study, Hack (1957) made a number of important contributions to the analysis of longitudinal profiles. He showed that rivers crossing regions formed from the same rocks have longitudinal profiles of similar shape. He further noted an inverse relationship between stream slope and discharge : since its transverse section increases with increasing discharge, resulting in a proportional reduction of the energy spent in overcoming friction, a stream is able to maintain its capacity to transport material as its slope decreases. However, lack of data for discharge led Hack to seek a relationship between slope and basin area, which showed them to be inversely proportional, the parameters of the equation depending on the constituent rocks in the basin considered. The relationship thus differs between calcareous, granodiorite, sandstone and other zones, the channel slope being well correlated with the ratio of channel-sediment mean size to drainage-basin area. For rivers with channel sediments of the same size, the channel slope is inversely proportional to drainage area and to discharge. For basins of the same area, the slope is directly proportional to the channel sediment size (Hack, 1957). For regions which are geologically similar, there is generally a good relationship between channel slope and length, which also presupposes similarity of the longitudinal profiles.

Hack found that the profiles of stream segments with channel sediment of constant size could be described by equations of the form

$$H = C - K \log_e L \tag{164}$$

where $H$ is the altitude of a point, $L$ the distance along the stream, $e$ the base of natural logarithms, and $K$ and $C$ are constants : the constant $K$ can be determined as the product of the stream length from source to mouth and its slope at the latter point.

As shown by Broscoe (1959), four types of regression equations can be considered in constructing longitudinal profiles representing altitude $H$ as a function of distance $L$ along a stream course :

the *simple linear form* which, plotted directly, gives a straight line expressed by the equation

$$H = a - bL \tag{165}$$

where $a$ and $b$ are constants ;

the *exponential form*,

$$\log H = a - bL \tag{166}$$

which gives a straight line if altitude is plotted on a logarithmic and length on an arithmetic scale ;

the *logarithmic form*,

$$H = a - b \log L \tag{167}$$

giving a straight line if altitude is plotted on an arithmetic and length on a logarithmic scale; and finally,

the *power form*, when both variables are on logarithmic scales, given by the equation

$$\log H = \log a - b \log L \tag{168}$$

It may be noted that eqn. (168) has no solution for $L = 0$. To remedy this defect, Broscoe (1959) proposed that at the end of each profile a constant value $L_g$ be added which should represent the horizontal distance of overland flow, obtained by extending the stream length to a reference point on the water divide. There will thus be a corresponding constant $H_g$ defined as the vertical distance from the reference point to the head of the stream. The origin of the axes is thus displaced accordingly, the ordinate becoming $H + H_g$ and the abscissa $L + L_g$. Introducing the two constants, the form of the above equations will change accordingly and any of the situations presented can be solved (Strahler, 1964).

The analysis of longitudinal profiles, although of great importance in geomorphology and hydrology, is nevertheless painstaking, because of the great complexity of channel phenomena resulting from the interaction of geomorphology, hydrology and hydraulics. Numerous factors contribute to the achievement of a certain form of longitudinal profile, and no methods have yet been developed for determining the share of each. Numerous attempts have also been made to express mathematically the equilibrium profile which would be reached by a water course having the capacity to transport only material received from the valleysides and only to the base level.

The method chosen for the study of longitudinal profiles depends on the goal pursued and the data available. Thus, altitude and distance data for the points along a stream may be plotted directly on arithmetic scales to give a graphical representation of the actual longitudinal profile (Fig. 80). If tributaries are plotted on the same graph as the main stream, it is possible to locate the sources of all the tributaries at the origin without taking the confluence points into account; alternatively, the confluences may be observed, and the source points will then have various coordinates depending on the confluence points and tributary lengths and altitudes (Fig. 80).

Owing to the varying lengths of the tributary profiles, neither form of representation is very useful in comparative studies. To surmount this difficulty, several procedures of graphical representation have been attempted. Since tributaries are genetically related to the main stream, their profiles should be reduced by considering the latter as the fundamental profile (Fig. 81). Thus, in a very detailed study in which the systems of representing longitudinal profiles in metric coordinates were analyzed and their disadvantages enumerated, Jovanović (1940) recommended the comparison of profiles only after at least one

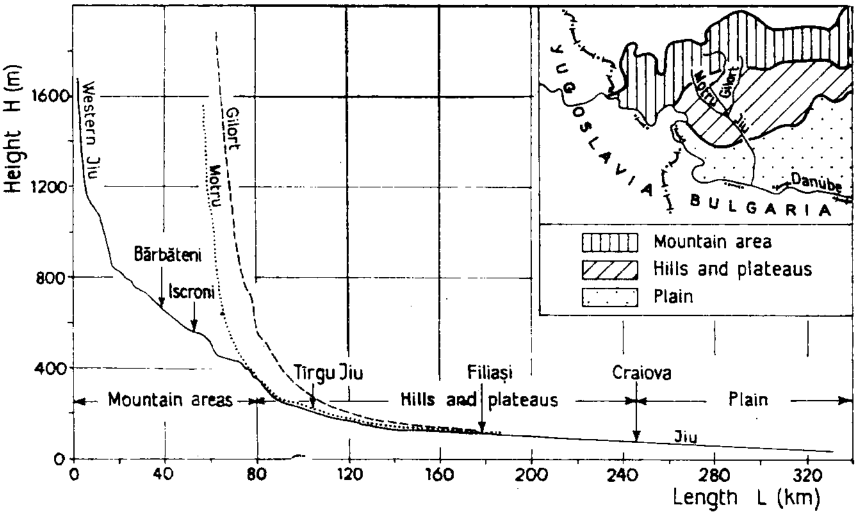



Fig. 80. Longitudinal profile of the Jiu and its main tributaries (arithmetic coordinates).

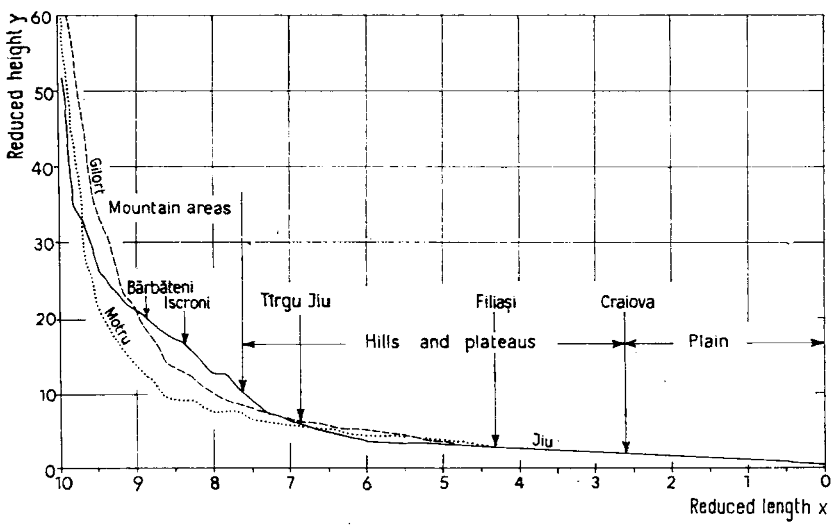



Fig. 81. Reduced decimal profile of the Jiu and its main tributaries.

of their coordinates has been normalized by multiplication by a constant, so that all profiles analyzed are reduced relative to the same total length. The transformation can be made by using a *percentage coefficient* $C_p = 100/L$, where $L$ is the total stream length. The distances from and heights above the mouth of the points on a profile are multiplied by this coefficient. The same result can be obtained using a *decimal coefficient* $C_d = 10/L$, or, more simply, by dividing the percentage coordinates by 10. Applying the decimal coefficient, profiles on reduced decimal coordinates are obtained. Use of the decimal coefficient for the heights (on the ordinate) allows slope values (‰) to be obtained directly from the ratio of the differences in the coordinates of pairs of chosen points.

Using the above systems of representation, profiles differing greatly in length can be compared without impairing detail. The coefficient of decimal reduction for the tributary is determined by taking the decimally reduced coordinate $X_0$ of the confluence point and the length $L_t$ of the tributary:

$$C_{dt} = (10 - X_0)\, L_t \tag{169}$$

Once the coefficient of decimal reduction for the tributary has been determined, the latter's profile is established by considering its confluence with the main stream as the base level. Hence, to establish the profile configuration in this system of representation, the decimally reduced coordinates of the tributary are added to the coordinates of the confluence point.

To characterize longitudinal profiles, Ivanov (1951) proposed construction of a rectangle of height $H$ and length $L$ enclosing a profile at its extremes, the area thus formed being divided by the profile into two parts. The ratio of the areas below ($w_i$) and above ($w_s$) the profile yields an index $n$ of the profile form:

$$n = w_s/w_i \tag{170}$$

If the index exceeds unity, the longitudinal profile is concave; if it is below unity, the profile is convex; and if the index is equal to unity, the profile is straight. The index also gives the exponent of a parabola passing through the extreme points of the profile.

Another method of representing longitudinal profiles consists in plotting distances on the abscissa as percentages of the total length of each stream, but without modifying the altitudes of the corresponding profiles (Ujvári, 1959). In this case, although the representation is distorted, the profiles can be compared more directly; this method is advantageous in particular for determining discontinuities in slope.

For the ordering system adopted in the present work, both the falls and lengths of streams of successive orders observe a law-like variation. Consequently, a comparative study of profiles is also more easily carried out by considering streams of the same order. Given the average falls and lengths for a particular basin, an average profile of the drainage network can be drawn around which individual profiles will vary in terms of both length and altitude. To plot such a profile, the coordinates of the initial and final points for each order must be determined. For the lengths, the successive points are obtained simply by summing the average lengths of the streams of the preceding orders. The altitude coordinates are obtained starting from the altitude of the final confluence or mouth of the stream of highest order $x$ (Fig. 82). To this value, the fall of the highest-order stream is added to give the ordi-

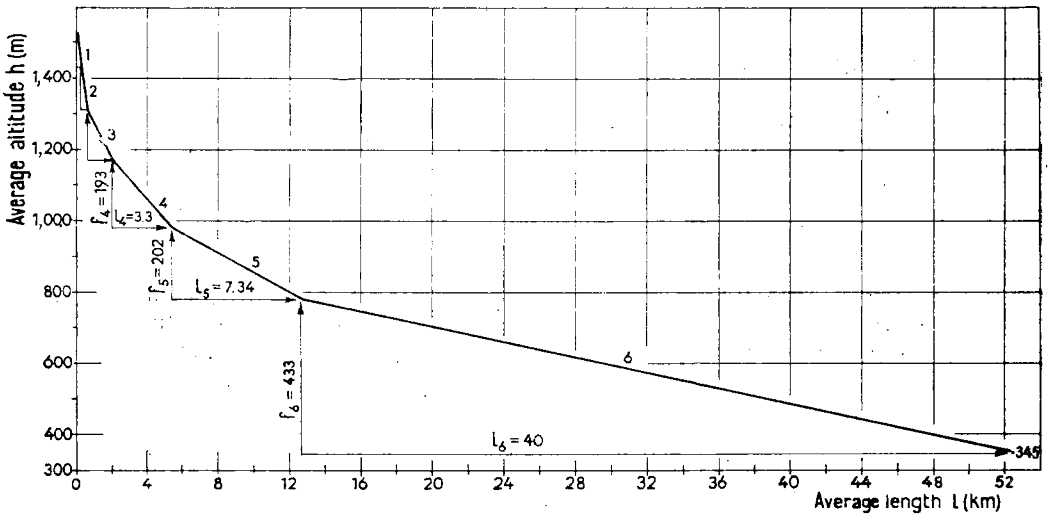



Fig. 82. Schematic longitudinal profile of the river network in the Doftana basin, using the average falls and lengths of stream segments.

nate of the final point for streams of order $x - 1$. To this, the average fall for streams of order $x - 2$ is added, and so on until the source point is reached. By uniting the resulting points by straight lines, the average mean slopes of stream segments of successive orders are obtained, the profile also giving the average falls and lengths of each order within the corresponding sections (Fig. 82).

Chapter XII

# Relationships between Morphometrical and Hydrological Features

Since fluvial forms are the result of a long process of drainage-basin evolution under the action of water, it is natural that interrelationships should exist between morphometrical parameters and discharge and suspended load. The determination of these interrelationships is of great importance in making regional generalizations and forecasts, and in deriving hydrological data for rivers for which no direct measurements are available. The most important morphometrical parameters which are related to hydrological features include the area, mean altitude and mean slope of a drainage basin, and the length and slope of the main stream.

After climatic conditions, the *area* of the drainage basin is the main element on which water discharge depends. As early as 1884, Kestlin remarked on the relationship between these elements and established a mathematical formula, valid for small rivers, for the calculation of discharge in relation to area:

$$Q = 16\alpha A \ (\mathrm{m^3 s^{-1}}) \tag{171}$$

where $\alpha$ is a coefficient depending on basin length.

In 1927, D. I. Kocherin used a formula of the type $q_{max} = K/A^n$ for the determination of maximum discharge, with parameters differing from one region to another. In the formulae proposed subsequently by Shevelev in 1937, by Sokolovsky in 1937 and 1945, by Ogievsky in 1938 (all quoted by Ogievsky, 1952), and by Leopold and Miller (1956), Hack (1957) and Hirsch (1962, 1963), among others, for the determination of mean and maximum discharge, basin area remains a basic element.

By processing of data collected by gauging stations in Romania, an expression has been established for the dependence of the maximum specific runoff $q_{max}$ having a probability of occurrence of 1 %, on drainage-basin area $A$ and mean altitude $H_m$ (Mociorniţa, 1964):

$$q_{max}\,1\% = f(H_m/\sqrt{A}) \tag{172}$$

An obvious indirect relationship is also to be noted between the logarithm of the maximum specific discharge having a probability of occurrence of 1%, and that of drainage-basin area (Mustaţă, 1964; Platagea and Platagea, 1965).

Together with other elements, basin area appears as an independent variable in the definition of other causal relationships, such as that between channel slope and drainage area (Hack, 1957).

For the geographical conditions of Romania, the *mean altitude* is the element which determines the altitudinal tiering of hydrometeorological processes and phenomena, i.e., mean specific runoff, seasonal and monthly runoff, the runoff coefficient, minimum discharge, and the amounts of precipitation and evapotranspiration. Even the mean slope of drainage basins depends on their mean altitude (Diaconu, 1966; Ujvári, 1972). Almost all meteorological phenomena, from air temperature to the degree and duration of cloud cover, are directly related to altitude. Furthermore, the mean specific runoff of suspended load is directly proportional to altitude up to 400 – 600 m, corresponding to the Subcarpathian region, and inversely proportional to altitude for drainage basins situated at greater altitudes (Diaconu, 1971a).

The ratio of the length of the main stream to the square root of basin mean slope determines the total duration of floods, while the ratio of main-stream length to the square root of river slope determines the rising period (Diaconu, 1971b).

As regards the classification system considered in the present work, fairly many studies have thrown light on the dependence of hydrological features (multiannual mean discharge during spring floods, and duration of spring high water) on basin order (Rzhanitsyn, 1960). The average mean slopes of streams of successive orders and of the entire river network in a given basin are accordingly implicated in the formation and movement of floods.

When direct data are insufficient, a series of morphometrical parameters are used in regional generalizations to forecast discharges with various probabilities of occurrence (Hirsch, 1962, 1963).

The list of examples could certainly be continued, although most of them have been cited extensively in the specialized literature. It should nevertheless be added that application of the present ordering system often sheds new light on the interrelationships between the morphometrical and hydrological characteristics of drainage basins.

## Relationship between Mean Discharge and Basin Order

It is already known that a direct relationship exists between the average area of drainage basins and their order. An early first step towards establishing a relationship between water discharge and drainage-basin order was made by Horton (1945). His studies have been continued by a number of researchers who have made important contributions to defining and applying regionally the relationship between discharge, order and area. Worthy of mention are the work of Rzhanitsyn (1960), who paid very great attention to mean specific discharge in relation to the order of stream segments, and the studies carried out by Hirsch (1962, 1963) on discharge forecasting.

In Romania, Platagea and Popa (1963) likewise established a good relationship between the mean and maximum discharges of streams and their order for the area between the Ialomiţa and the Trotuş, despite the great diversity of geographical factors.

*Law of summed mean discharges.* Denoting the sum of the mean discharges for all first-order basins by $Q_1$, that for the second order by $Q_2$, and so on up to the basin of highest order, a series results in which the number of terms equals the order of the main stream. Plotting of these sums on semilogarithmic paper (Fig. 83(b)) indicates a close relationship between the series of values obtained and basin order,

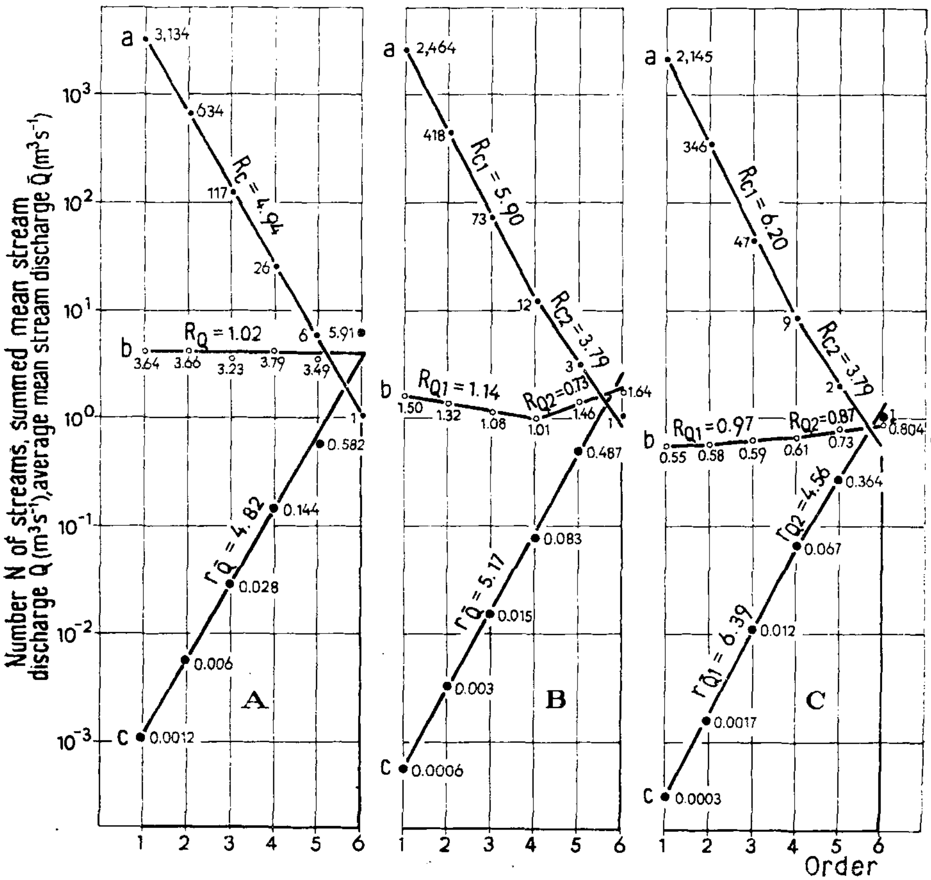



Fig. 83. Morphometrical models of mean stream discharge for the basins of (A) the Doftana at its confluence with the Prahova, (B) the Vărbilău at its confluence with the Teleajen, and (C) the Mislea at its confluence with the Teleajen (a, regression of numbers of streams ; b, regression of summed mean discharges ; c, regression of average mean discharges).

such that *the summed mean discharges of streams of successively higher orders tend to form a geometric progression in which the first term is the sum $Q_1$ of the mean discharges of first-order basins and the ratio is the ratio $R_Q$ of successive summed mean discharges*. The general term of the progression is given by the formula

$$Q_u = Q_1/R_Q^{u-1} \tag{173}$$

Analysis of a large number of such progressions shows that the ratio $R_Q$ is generally very close to unity. Higher values would indicate a decreasing, and lower values, an increasing series.

If a drainage basin develops in two major relief units with differing geographical conditions which influence runoff, then a second progression with a different ratio results for higher orders (Fig. 83B, C). This situation is clearly exemplified by the Vărbilău, whose upper and middle basins develop in a region of excessive humidity where precipitation greatly exceeds evapotranspiration, and whose lower basin extends in a hilly region with a slight moisture deficit. Combined with overall geographical factors, these conditions determine two discharge lines with differing ratios, the values for the first to fourth orders giving a decreasing geometric series, while the fifth and sixth orders, with a ratio below unity, determine an increasing geometric progression (Fig. 83B). In the case of the Mislea basin, which has developed over both Subcarpathian and piedmont regions, the corresponding progressions are both increasing and have ratios below unity (Fig. 83C).

*Law of average mean discharges*. The summed mean discharges of successive orders constitute a series of values defining a geometric progression. By taking the ratios of corresponding terms in this series and that of basin numbers, the average mean discharges of basins of successive orders are obtained, i.e.,

$$Q_1/N_1 = \bar{Q}_1,\ Q_2/N_2 = \bar{Q}_2,\ \ldots,\ Q_u/N_u = \bar{Q}_u$$

Representation of these values on semilogarithmic paper indicates a linear relationship with basin order, according to which *the average mean discharges of streams of successively higher orders tend to form an increasing geometric progression in which the first term is the average mean discharge $\bar{Q}_1$ of first-order streams and the ratio is the ratio $r_{\bar{Q}}$ of successive average mean discharges* (Figs. 83(c) and 84). The ratio of the new series is given by the quotient of the ratios of the two constituent series :

$$r_{\bar{Q}} = R_c/R_Q$$

this being the simplest, fastest and most precise way of determining the ratio. The general term is given (Platagea and Popa, 1963) by

$$\bar{Q}_u = \bar{Q}_1 r_{\bar{Q}}^{u-1} \tag{174}$$

From the analysis of a large number of drainage basins (Fig. 84A—G), it is found that the law of average mean discharges is followed much more closely than the law of summed mean discharges, especially for lower-order streams. The former law shows that discharge

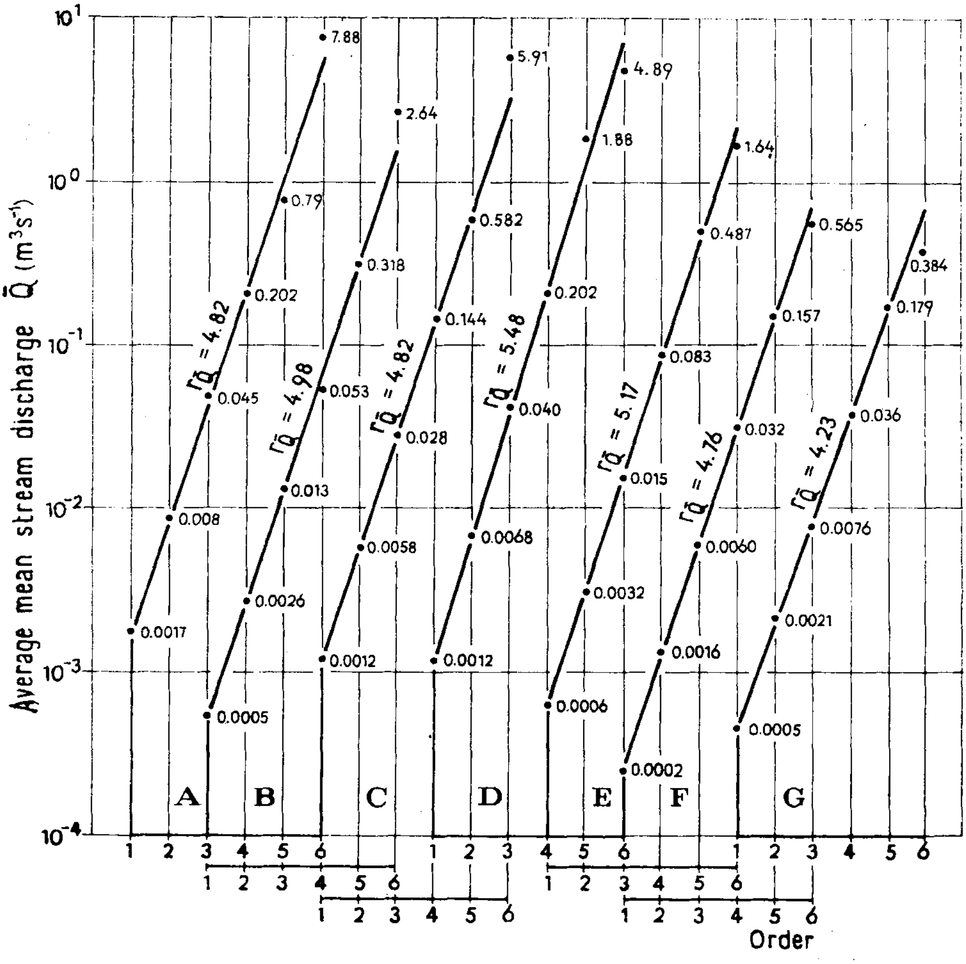



Fig. 84. Regression of average mean discharges on order for the basins of (A) the Prahova at its confluence with the Doftana, (B) the Cricovu Dulce at its confluence with the Ialomiţa, (C) the Doftana at its confluence with the Prahova, (D) the Teleajen at its confluence with the Drajna, (E) the Vărbilău at its confluence with the Teleajen, (F) the Lopatna at its confluence with the Sărăţel, and (G) the Bucovel at its confluence with the Teleajen.

increases exponentially as a function of stream order. The ratios obtained for the progression determined for sixth-order basins range from 6.39 (Mislea basin) to 4.23 (Bucovel basin at the confluence with the Teleajen). These high values indicate large differences between the mean discharges of basins of successive orders : for the Mislea basin, the mean discharge for second-order streams is 6.39 times greater than that for first-order streams.

There are cases when the series of basin numbers and of summed mean discharges each determine two lines, whereas the average values yield a single series with a constant ratio (Fig. 83B). This is why an analysis of the summed mean discharges should always be made, since in many cases these yield significant information which is not apparent from the average values. In other cases, the line for the average mean discharges does show a discontinuity corresponding to those indicated by the basin numbers and summed mean discharges (Fig. 83C).

Analysis of the series of values determined for sixth-order basins reveals the influence of geographical conditions on the distribution of discharge. Thus, basins which have developed to a great extent in mountainous areas present the highest values of average mean discharge for first-order streams, while the smallest values correspond to hills and plains. For instance, whereas the first-order streams in the Prahova basin upstream from the confluence with the Doftana have an average mean discharge of some 1.7 l $s^{-1}$, the corresponding value for the Mislea, Cricovu Sărat and Lopatna basins is only some 0.2 l $s^{-1}$, i.e., the average discharge is almost eight times lower. Certainly, the area of lower-order basins is smaller in hilly areas, but other geographical factors also play a considerable role in determining the mean discharge of first-order basins.

*Relationship between average mean discharge and average area of basins of successive orders.* From the foregoing considerations, it results that both the average area and the average mean discharge are directly proportional to basin order : an increase of order corresponds to a proportional increase of area, and hence of the corresponding discharge. The two values are well correlated for sixth-order basins (Fig. 85), the positions of the lines also reflecting the mean basin altitude and specific geographical conditions. Indeed, the coefficient $K$ in the equation $\bar{Q} = Ka^n$ (where $\bar{Q}$ is average mean discharge and $a$ is average area) is well correlated with average mean altitude for first-order basins, indicating that an increase of mean basin altitude entails an increase of the coefficient $K$.

As concerns the sixth-order basins of the Drajna and Cricovu Sărat respectively at their confluences with the Teleajen and Lopatna (Fig. 85E, G), the average area of the first-order subbasins is 4.2 ha for the former and 4.8 ha for the latter. However, the calculated average mean discharges of the first-order streams are 0.24 l $s^{-1}$ for the Cricovu Sărat and 0.48 l $s^{-1}$ for the Drajna, the average mean altitudes of the corresponding first-order basins being 407 and 737 m, respectively. The positions of the respective lines (Fig. 85) reflect the influences both of mean altitude and of physiographical factors on discharge. For instance, considering only the average mean discharges of first-order basins, the highest value (1.8 l $s^{-1}$) is found for the Prahova basin upstream from the confluence with the Doftana, for an average area of 6.7 ha (Fig. 85C). For Subcarpathian basins, the average mean discharge of first-order streams decreases (Fig. 85E – H), the highest values being found for the Mislea and Lopatna basins : some 0.2 l $s^{-1}$ for an average area of 4.6 ha.

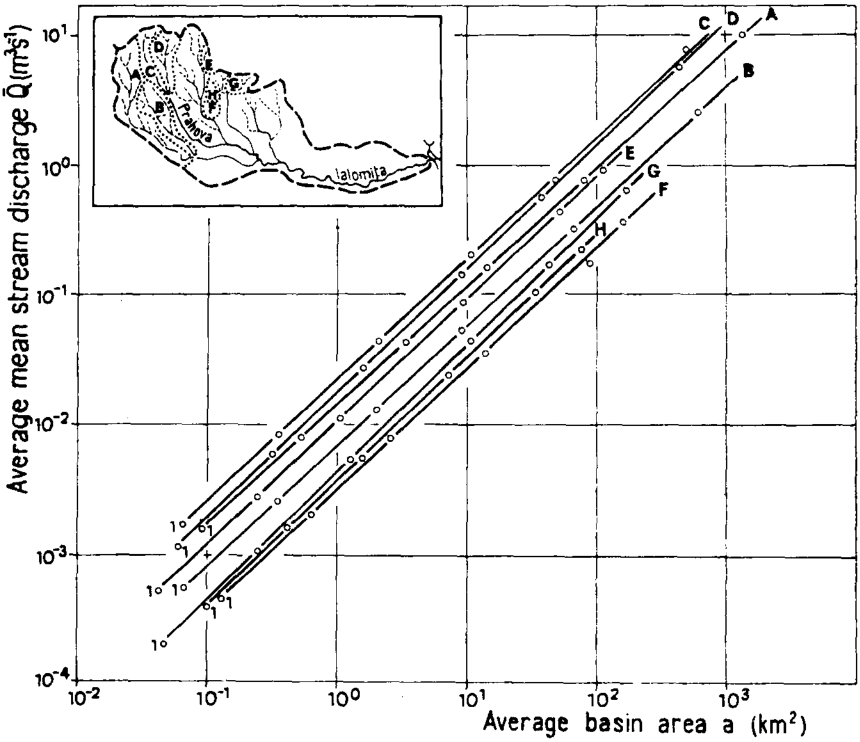



Fig. 85. Relationship between average mean stream discharge and average area for the basins of (A) the Ialomiţa at its confluence with the Cricovu Dulce, (B) the Cricovu Dulce at its confluence with the Ialomiţa, (C) the Prahova at its confluence with the Doftana, (D) the Doftana at its confluence with the Prahova, (E) the Drajna at its confluence with the Teleajen, (F) the Bucovel at its confluence with the Teleajen, (G) the Cricovu Sărat at its confluence with the Lopatna, and (H) the Sărăţel at its confluence with the Lopatna.

*Relationship between mean slope of drainage network and mean discharge for basins of successive orders.* One of the advantages of the classification system adopted in the present work is that the average mean slope of a drainage network can be calculated both for each separate order and for an entire basin, using eqns. (161) and (162), respectively. From the law of average mean slopes, it is obvious that the highest values are usually encountered in the smallest fingertip tributaries, which fulfil the drainage role only temporarily. To the extent to which order increases, the average area, and together with it the discharge, also increases. A greater volume of water has a greater transport capacity, and hence a greater capacity to diminish slope.

This is quite obvious on representing the average mean slopes of streams of successively higher orders in relation to the respective average mean discharges. An inverse relationship is found between the two variables, indicating an equation of hyperbolic form according to which an increase of discharge entails proportional decreases of the mean slopes of stream segments of successive orders (Fig. 86). However,

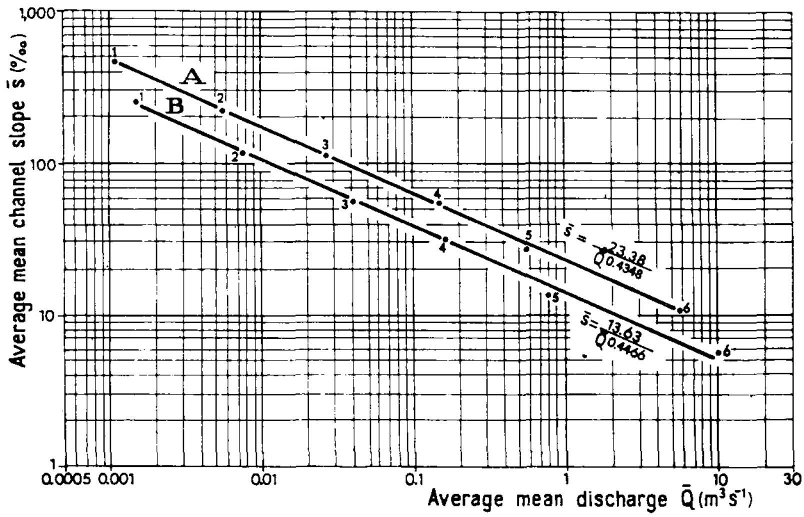



Fig. 86. Relationship between average mean channel slope and average mean stream discharge for the basins of (A) the Doftana at its confluence with the Prahova, and (B) the Ialomiţa at its confluence with the Cricovu Dulce.

although the form of the relationship between the two elements is constant, the parameters of the function differ from one basin to another under the influence of physiographical factors.

To establish the hyperbolic equation, the data are plotted on logarithmic paper to define a straight line corresponding to an equation of the type

$$S_n = K/\bar{Q}^m \qquad (175)$$

The parameters $m$ and $K$ are easily calculated from the coordinates of two points on the line. For example, for the sixth-order basin of the Doftana at its confluence with the Prahova, the corresponding relationships are

$$m = (\log 340 - \log 17)/(\log 2 - \log 0.0022) = (2.53148 -$$

$$- 1.23045)/(0.30103 + 2.65758) = 0.43974$$

and

$K = y_1 x_1^m$ or $\log K = \log y_1 + m \log x_1$

$\log K = 1.23045 + 0.43974 \times 0.30103 = 1.36283$

whence $K = 23.06$. The equation of the line is then obtained by substituting these values in eqn. (175).

*Determination of mean discharge in relation to mean altitude and area of drainage basins.* Socioeconomic development of any region implies extensive river engineering projects and a more rational management of water resources. To elaborate such projects very precise hydrological data are required which should cover the whole range of relevant phenomena. Although quite accurate, the data supplied by gauging stations cannot fully meet the information needs of design and construction activities in the domains of industrial and river engineering. The solution is either to extrapolate existing values to an entire region, or to resort to relationships between morphometrical and hydrological elements.

Since there is much specialized work dealing with the forecasting of maximum and minimum discharges having various probabilities of occurrence, and since the corresponding equations are valid only for the regions for which they were established, we here take as an example the nomogram for determining mean discharges in the Ialomiţa basin in relation to subbasin mean altitude and area, these being the chief morphometrical elements controlling the variation of discharge for Romania's climatic and geographical conditions.

It has been seen that in Romania, mean altitude is a very important factor influencing the vertical distribution of climatic elements, and hence water discharge. There is thus a direct relationship between discharge and this morphometrical parameter (Mociorniţa, 1964; Diaconu, 1966). Another element determining the mean discharge of a river is the area of its drainage basin, it being known that for similar geographical conditions there is a direct relationship between increases of basin area and of mean discharge. To express the interrelationships between these three variables in the form of an equation or a nomogram, it was first established by pairwise correlation that mean discharge $\bar{Q}$ is related logarithmically both to area $A$ and mean altitude $H_{\mathrm{m}}$, but that the logarithmic scales have different moduli. The relationship is thus best expressed by a formula of the type

$$\bar{Q} = kH_{\mathrm{m}}^{p}A^{n} \tag{176}$$

in which $k$, $p$ and $n$ are constants differing from one region to another. A nomogram for calculating mean discharges was established by taking mean discharge as a variable depending on mean altitude $H_{\mathrm{m}}$, with area $A$ as a parameter. All three variables have logarithmic scales, but the first two have the same modulus while the modulus

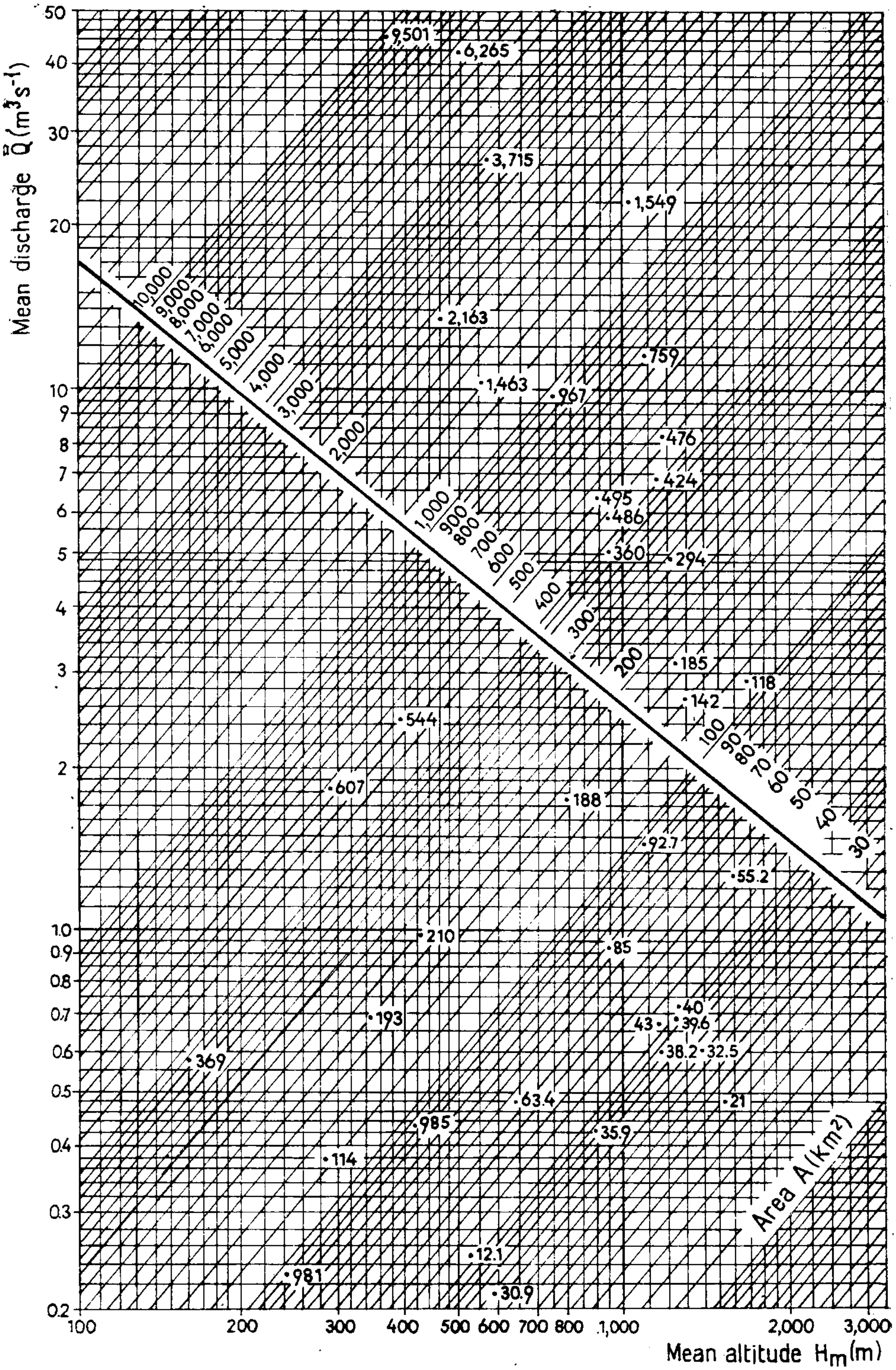



Fig. 87. Nomogram for determining the mean discharge of streams in the Ialomiţa basin as a function of drainage-basin area $A$ and mean altitude $H_m$.

of the parameter differs. To establish the latter, basin area values supplied by all 24 gauging stations in the Ialomiţa basin possessing discharge data for the period 1950—1975 were placed on the grid corresponding to the first two variables (Fig. 87). Using known methods (Bloh, 1971), the values were taken by pairs and the resulting twelve equations were treated logarithmically and divided into two groups. The values for each group were summed, and two equations were obtained whose solution gave the parameters $k$, $p$ and $n$. The following formula was thus obtained:

$$\bar{Q} = 2.65 \times 10^{-6} \, H_{\mathrm{m}}^{1.24} \, A^{0.9986} \tag{177}$$

The value of the constant $n$, which is very close to unity, indicates that the slope of the parameter scale is almost 45° and that the equation can be generalized. This formula or the nomogram yield the mean discharge with a maximum error of $\pm 10\%$ for any subbasin whose mean altitude and area are known. The accuracy of the values obtained depends greatly on the accuracy of the two morphometrical parameters employed. The general formula has been verified for other basins and regions in Romania, but with different constants as a function of local geographical factors. Using the values calculated for chosen streams, a map of discharges can be plotted, the class interval of the scale depending on the degree of detail desired.

*Determination of total duration of floods in relation to mean altitude and area of drainage basins.* Among the hydrological data of direct practical significance, particular importance is attached to those used in calculations of typical floods or of floods corresponding to maximum discharges with various probabilities of occurrence. In analyzing floods to establish their characteristic features, a number of difficulties are encountered stemming from the large variety of flood forms. Singular floods, which can be schematized more easily, are rather infrequent, while those with several main peaks, or with a main peak and several additional minor peaks depending on genetic factors and propagation time, constitute the majority of cases. To define the characteristic features, floods originating from strong singular rainfalls and which produce the largest discharges and water volumes may be chosen (Diaconu et al., 1961). The duration of singular floods is delimited by the points marking the moment when discharge starts to increase and the end of surface feed. Values of the maximum and basic yearly discharges, the durations of high and low water, the total duration, the flood volume and finally the form coefficient were determined for a sample of singular floods in Romania on the basis of data supplied by 152 gauging stations (Diaconu, 1971b). Analysis of these values indicated their dependence on a number of morphometrical characteristics of the corresponding river networks and drainage basins. Thus, the total duration $T_t$ and the rising time $T_r$ were found to be directly proportional to basin area $A$ and river length $L$, and inversely proportional to mean basin slope $S_b$ or mean main-stream slope $S_r$ (Diaconu et al., 1961; Mustaţă, 1964):

$$T_t = f(L/\sqrt{S_b}) \quad \text{or} \quad f(A/\sqrt{S_b S_r})$$

and

$$T_r = f(L/\sqrt{S_r}) \text{ or } f(L) \tag{178}$$

Of the multitude of possible relationships between the features of typical floods and morphometrical elements, the dependence of total flood duration $T_t$ on mean altitude $H_m$ and basin area $A$ may be chosen as the most representative for Romania. Research carried out by C. Diaconu (unpublished results) has allowed graphs to be established which relate these three elements . Thus, for the same area, total flood duration decreases with increasing altitude, and for the same mean altitude, it increases with increasing area. So far, three such graphs of total flood duration have been established for three areas in Romania.

Continuing this research, we have established mathematical expressions for corresponding dependences of total flood duration on mean altitude as the independent variable, with basin area as a parameter. To this end, mean altitude and area are plotted on logarithmic scales with differing moduli, while the total duration has a square-root scale. The relationship between the three variables can be expressed by a formula of the type

$$\sqrt{T_t} = -m \log H_m + n \log A + p \tag{179}$$

On the basis of the existing data, we plotted nomograms for the calculation of total flood duration for the three areas in Romania delimited by Diaconu during the research mentioned above. Following the methods recommended by Bloh (1971), we calculated the constants $m$, $n$ and $p$ in each case and finally obtained the following equations:

$$\sqrt{T_t} = -2.10741 \log H_m + 3.29686 \log A + 8.64508 \text{ (for the first area)}$$

$$\sqrt{T_t} = -2.21358 \log H_m + 3.11692 \log A + 8.70067 \text{ (for the second area)}$$

$$\sqrt{T_t} = -2.34353 \log H_m + 3.48702 \log A + 5.88076 \text{ (for the third area)}$$

The data employed refer to basins with areas larger than 70 km$^2$ and mean altitudes above 100 m; these equations are not recommended for other values. The nomogram for the calculation of total flood duration in the second area above is presented in Fig. 88.

Similar interrelationships can also be established for other flood characteristics, which facilitates the drawing of standard hydrographs for runoff sections not instrumented by gauging stations.

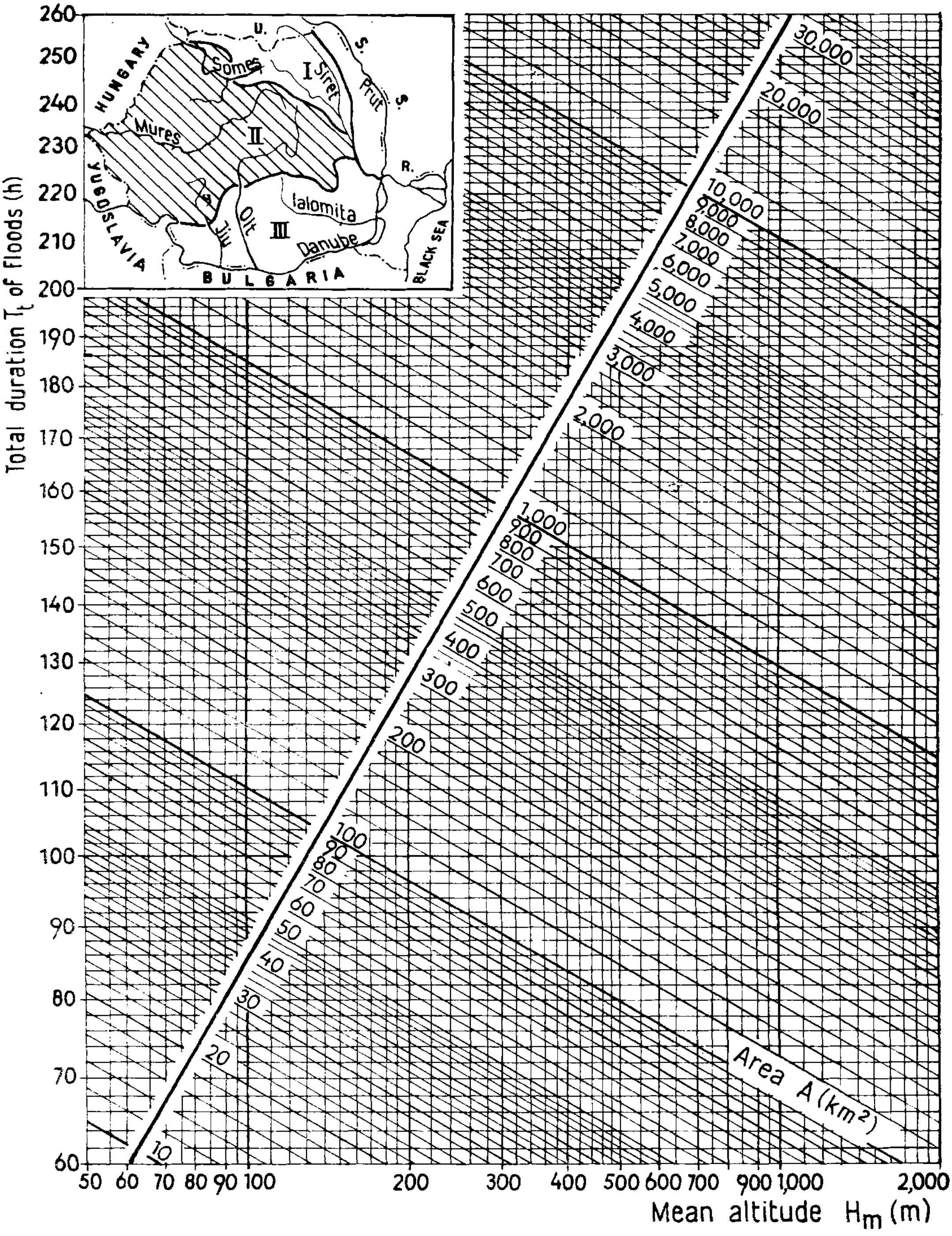



Fig. 88. Nomogram for calculating the total duration $T_t$ of floods in one region of Romania as a function of drainage-basin area $A$ and mean altitude $H_m$.

Chapter XIII

# Morphometrical Models of the Ialomiţa Basin

Morphometrical analysis of drainage basins shows that many morphometrical elements indicate a law-like tendency to reach a steady-state exponential variation in relation to basin order. The parameters of the resulting geometric progressions depend on overall physiographical conditions. Knowing the interrelationships between these elements, a series of morphometrical models may be constructed to characterize the present situation in a given basin. We here take as an example the basin of the Ialomiţa (a left-side tributary of the Danube), which extends over the southern side of the Southern Carpathians. Its relief includes a wide range of forms, the altitude varying from 2505 m at the Omu peak to about 7 – 10 m where the Ialomiţa flows into the Danube.

Extending over approximately 10 000 km$^2$, the basin has an average length of 325 km and an average width of 31 km. The present relief and channel network bear the mark of the tectonic movements which shaped the axis of the Southern Carpathians. Paleogeographical evolution has left traces of successive geological ages, Subcarpathian and plain regions being found in addition to the youngest Carpathian relief.

As soon as they appeared, the main geological formations making up the skeleton of the three relief units started to undergo continuous modelling by eroding agents, whose action at any time depends on the general conditions controlling the amounts of matter and energy moving within the system. Variations of the inputs and outputs of matter and energy thus influence modelling processes, and hence the achievement of dynamic morphometrical equilibrium. There have certainly been periods during which fragmentation and erosion processes leading to slope equilibrium were stronger than at present, as evidenced by the deposits found in the piedmont plains and fluvial terraces. Of the factors determining these processes, mention may be made of increased water discharge, the gradual uplift of the Carpathians and Subcarpathians, and the lowering of the subsidence plains. Age-long modelling processes have continuously modified the river network in achieving the present configuration of the drainage network in the Ialomiţa basin. There have been numerous detailed studies in which the relief of the Ialomiţa basin from the geomorphological viewpoint has been treated as a main or subsidiary topic (e.g., Vâlsan, 1915, 1939; Popp, 1939; Mihăilescu, 1963, 1966, 1969; Niculescu, 1962; Badea and Niculescu, 1964; Tufescu, 1966; Coteţ, 1973; Roşu, 1973). The temporal and spatial variations of the movements of matter and energy within the basin show vertical zonality. Therefore, the elements of the climatic and water regimes can be analyzed only if this is taken into account.

The basin has a mean discharge of 44 $m^3s^{-1}$, the greatest proportion (38.5%) occurring during the spring, followed by summer (23.6%), winter (22%) and autumn (15.9%). The maximum discharge during the floods of July 1975 was recorded at the Coşereni gauging station: 1400 $m^3s^{-1}$, an extremely high value. Besides catastrophic discharges, the basin is characterized by frequent spring high waters and summer floods, which largely determine the basin's morphometrical elements.

Discharges of suspended load are also important indicators of degradation processes, in that they indicate the amount of material leaving a basin area. At Coşereni the mean multiannual flow of suspended load is 189 kg $s^{-1}$; however, values as high as 33 000 kg $s^{-1}$ have been recorded during floods. Calculations have shown that the Ialomiţa carries annually into the Danube some 3 790 000 t of suspended or dissolved material, which is tantamount to 2 520 000 $m^3$ of eroded soil (Ujvári, 1972). This means that the basin area becomes lower on average by 0.3 mm per year due to the removal of this extraordinary amount of material.

The morphometrical laws established in the preceding chapters are obeyed reasonably well for the subbasins within the Ialomiţa system, their parameters having evolved continuously in close relation to the conditions specific to the main geographical regions. It is also obvious, analyzing the history of man's action on the environment, that land tilling, grazing, deforestation, etc. have naturally led to alterations of the equilibrium previously existing in nature, entailing a redistribution in time and space of the quantities of mass and energy circulating within the river system. As a consequence, changes have occurred in the morphometrical elements of the system so as to bring it into a new steady state corresponding to the new conditions. For instance, the frequency curves of basin area for various orders show strong positive skewness, indicating a tendency to relief fragmentation and increases in the numbers of basins of each order. Today, man's action upon the environment is increasingly far-reaching, and thus it is very difficult for equilibrium to persist until the morphometrical elements have also attained a steady state.

Of course, the available data are insufficient to establish the morphometrical models specific to successive steady states in order to reconstitute their historical evolution. However, by processing the data supplied by maps and aerial photographs, the present state of any drainage basin can be represented by a system of morphometrical models, each component characterizing an element of the basin by means of known formulae. Such models have been constructed for basins of various orders and situated in various physiographical conditions, but we here describe, by way of example only, the system of morphometrical models for the eighth-order Ialomiţa basin.

The first element to be analyzed in such a system of morphometrical models is the degree of branching of the river network. In the case of the Ialomiţa basin, it is obvious that the numbers of rivers of successive orders define two geometric progressions with different ratios: for the first- to fourth-order streams the ratios is 5.43, while for the higher orders (fifth to eighth) it is only 3.99. Taking into consideration the main geomorphological units, the altitudinal zoning of various hydrometeorological elements, the relief energy and the make-up of the geological substratum, the existence of two progressions is explained fully. Brief examination of the spatial distribution of the river network shows that most of the lower-order streams develop in mountainous and Subcarpathian regions. These regions thus define the position of the line and the ratio of the geometric progression for the lower-order streams (Fig. 89A). The progression is also related to the runoff regime, to the effect that as a rule lower-order streams fulfil the runoff function only temporarily, while those of higher orders are mostly perennial. The higher-order streams develop chiefly in Subcarpathian and flatland regions, whose different physiographical features account for the different ratio of the progression of stream numbers. The transition to the second line, with a gentler slope, corresponds to the transition to a zone with lower relief energy. The marked decreases in the amount of precipitation, in the mean slope of the ground surface, in the runoff coefficient and in the sediment yield, etc., found in these regions explain why their relief fragmentation, and hence the number of streams, is much lower. The total number of stream segments in the Ialomiţa basin, calculated using eqn. (29), is 43 779. Relating this value to the area of the basin, a mean density of 4.4 segments per square kilometer is obtained.

The summed lengths of the streams of each order tend to define a decreasing geometric progression, where again two branches with differing ratios are evident. The first is characteristic of lower-order and the second of higher-order streams (Fig. 89A). The total length of the entire drainage network in the Ialomiţa basin determined using eqn. (89) is 18 206 km. From the total length, the mean drainage density can also be calculated as 1.89 km $km^{-2}$.

The model of the average lengths of streams of successive orders is obtained by taking the ratios of corresponding terms in the preceding two series. The ratio of the new series is the quotient of the ratios of the constituent series. It is remarkable that in many instances when the series of stream numbers and summed lengths both yield two progressions with differing ratios, the average stream lengths yield only a single line. This is why the summed lengths of streams of successive orders must always be analyzed, since this series may be quite significant from the geomorphological viewpoint. In the present case, the progression of the average lengths shows a high ratio for the branch determined by the higher-order streams, while their summed lengths increase to a lesser extent. The higher-order streams in the Ialomiţa basin develop mostly in piedmont areas and subsidence plains; the number of streams diminishes considerably while their length increases

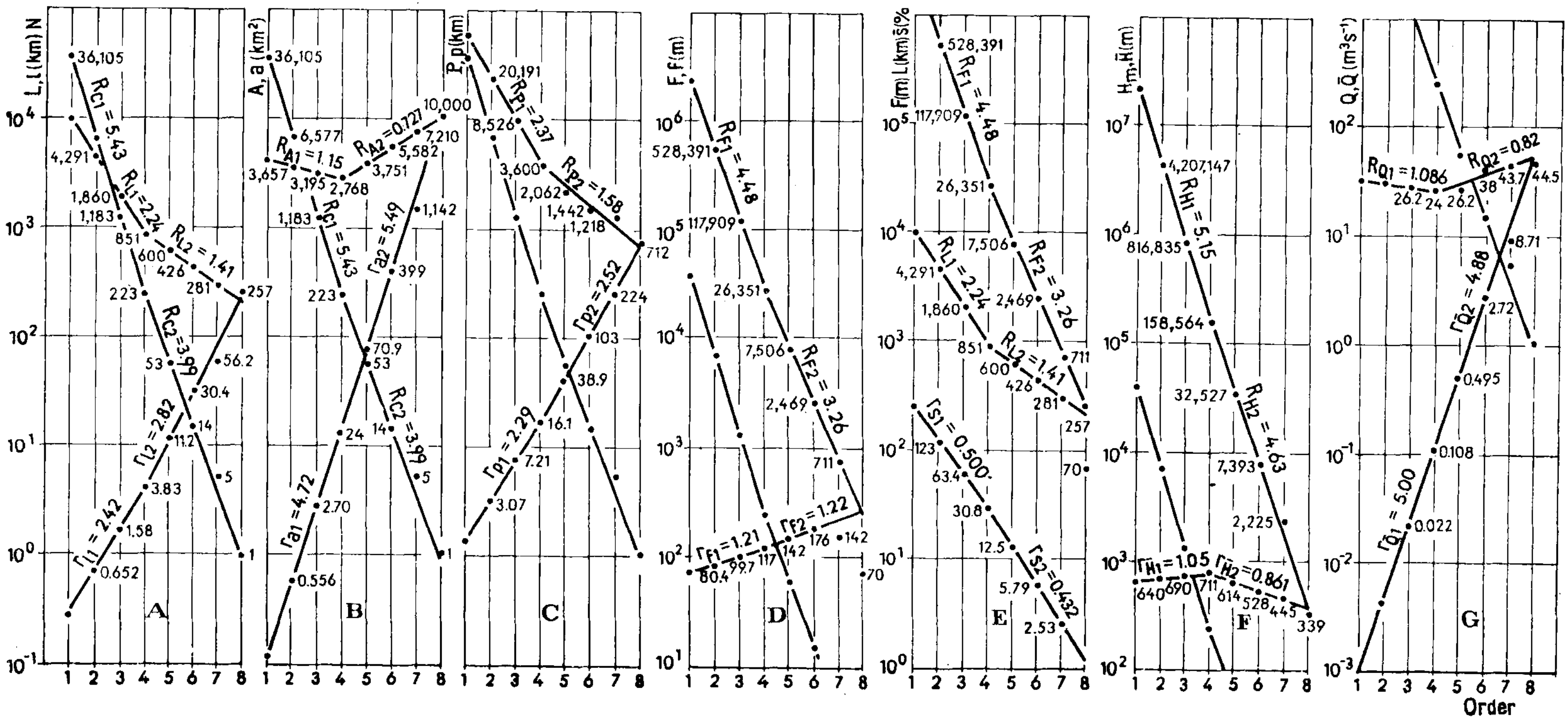



Fig. 89. The system of morphometrical models for the Ialomiţa drainage basin : (A) model of stream lengths ; (B) model of basin areas ; (C) model of basin perimeters ; (D) model of stream falls ; (E) model of mean stream slopes ; (F) model of mean altitudes ; (G) model of mean discharges.

as a consequence of meandering and a high sinuosity coefficient, accounting for the higher average length.

To establish the model of the areas of basins of successive orders, the areas of basins of each order must be summed. In this case two geometric progressions are again obtained, for higher- and lower-order basins, respectively. That the summed areas of the lower-order basins yield a decreasing geometric progression reveals strong fragmentation, the continuous decrease of basin area implying corresponding decreases of interbasin area (Fig. 89B). Since the latter is closely related to the total network length, it may be inferred that increased fragmentation leads to increased drainage density and to shorter lengths of slope runoff, and hence to an increase in the runoff coefficient and to acceleration of the formation and propagation of floods. For the average areas, the difference between the progressions for the higher and lower orders is so small that it can be detected only when analyzed in relation to the summed areas (Fig. 89B).

The two progressions obtained both for the summed and for the average basin perimeters for successive orders confirm the same differentiation between lower- and higher-order basins as above (Fig. 89C). The frequency curve for perimeter length as a function of basin order demonstrates the same marked positive skewness as shown by the basin areas, again attesting to the tendency to relief fragmentation, increases in the number of low-order basins and in their total perimeter, resulting in decreased average values.

The model of falls provides very important evidence concerning erosion and modelling processes within the Ialomiţa drainage system. The larger the falls of streams of successive orders, in relation to stream length, the steeper the channel slopes and hence the greater the erosive power of the stream flows. The summed falls for the basin yield decreasing geometric progressions, while the average falls form an increasing geometric progression whose ratio is very close to unity. It is noteworthy that for this model, the average falls of streams of the seventh and eighth orders are much lower than they should be. This also reflects differences in the channel slopes, especially for the lower reaches of the Ialomiţa (see below), and is perhaps due chiefly to subsidence processes that have occurred in the area in the course of time. These lowering processes have greatly diminished the fall between the initial point (at the confluence of the Prahova with the Teleajen) and final point of the eighth-order segment of the Ialomiţa. As a consequence, the fall obtained for this segment deviates considerably from the law observed faithfully by the first six orders (Fig. 89D).

As has been seen, the model of average mean slopes for streams of successive orders can be established in several ways. In the present example, we use ratios of the summed stream falls and lengths for successive orders (Fig. 89E). In this model, the closer the ratio to unity, the smaller the differences between the average mean slopes of streams of successive order, and vice versa. It is obvious that the greatest slope will be characteristic of first-order streams, and the lowest, of the main stream. For the eighth-order segment of the Ialomiţa, as for the fall, the real slope is much less than it should be: the real mean slope is $0.27\,^0/_{00}$, whereas according to the law of average mean slopes it should be $1.07^0/_{00}$, or almost four times greater. This deviation is fully explained in terms of the paleogeographical evolution of the lower reaches of the Ialomiţa. It has serious hydrological implications, the concomitant sudden decrease in transport capacity having obvious consequences in the piedmont and subsidence plains, which are accordingly subject to strong alluvial deposition in river channels and to lateral erosion, leading to the formation of meanders and ox-bow lakes. For the seventh order streams, although the average mean fall is again smaller than indicated by the model, the average length is also proportionally smaller, so that the average mean slope of this order has the expected value.

The model of mean slopes is generally followed very closely, and mean slope can be considered as being the most dynamic morphometrical element, responding immediately to changes in the flow of matter moving through a channel network. The energy of a water course acts jointly to lower its fall and increase its length. The details of the processes implicated in this tendency are complicated and lie outside the scope of this book, which deals only with the quantitative assessment of the resulting forms.

The model of average mean altitudes for drainage basins of successive orders is established similarly, starting from the series of summed mean altitudes and the numbers of basins of successive orders (Fig. 89F). For the Ialomiţa basin, the values indicate an increasing geometric progression with a ratio very close to unity for the lower-order streams, and a decreasing progression for the higher orders. The evolution of mean altitude is fairly slow, and once a steady state has been reached it remains unaltered for a long time. However, analysis of the model can provide certain information concerning upwards or downwards movements affecting an entire drainage basin or parts of it.

The average mean discharges of rivers of successive orders in the Ialomiţa basin clearly define two increasing geometric progressions corresponding to lower and higher orders, as also found for the summed values. As a rule, the model of mean discharges yields most information concerning drainage behaviour in relation to physiographical conditions; analyzed in relation to the models of basin areas and numbers, it provides an indication of the degree of fragmentation and even of the mean discharges corresponding to interbasin areas of various orders. The model of mean discharges can be correlated with others to determine the interrelationships among the highly complex phenomena underlying drainage-basin evolution in time and space.

The list of morphometrical models could be enlarged by including other elements of hydrological or geomorphological interest. Thus, for example, models of the confluence angle, or of the relief volumes of drainage basins of successive orders, can be constructed similarly, all such elements obeying well-defined laws of evolution which can be determined and applied as required. The models analyzed both for the Ialomiţa basin and for a series of smaller basins have demonstrated that the present structure of drainage basins is the result of a long process of evolution. Such basins have developed, and continue to do so, according to well-defined laws reflecting the tendency of all morphometrical elements to achieve a steady state. In this process of evolution, the morphology of drainage basins should always be considered as the result of the mutual action of relief, rock, vegetation, soil and hydrometeorological factors. In this context, the rate of evolution and the magnitudes of the morphometrical parameters are greatly influenced by the amounts of mass and energy the system receives from and loses into the environment.

Further research should attempt to assess the role of environmental factors in defining morphometrical elements and to determine the relationships between these two groups of variables. Such information would be of great use in the implementation of river engineering projects and related activities intended to promote socioeconomic development.

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# Subject Index