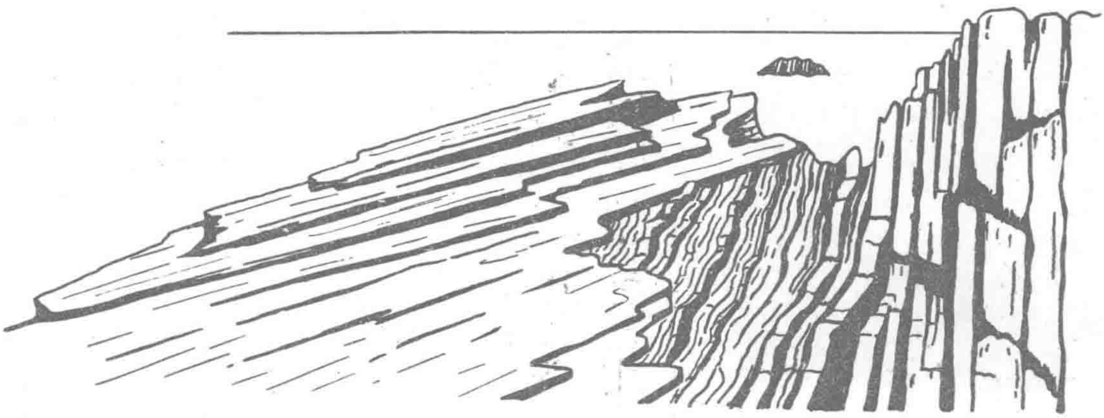

Principles

NEW YORK · JOHN WILEY & SONS, INC.   London · Chapman & Hall, Limited

To Charles Schuchert

whose life and work are an abiding inspiration to the authors of this work and to a host of other stratigraphers.

"He mapped the ancient seas and fathomed the geologic past."

Preface

UPON THE RETIREMENT OF CHARLES SCHUCHERT at Yale in 1924 the senior author of this book inherited a graduate course in the stratigraphy of North America. The first few weeks of the course were then devoted to principles and the remainder to a systematic account of the stratigraphic record in this continent. In following years the discussion of principles was gradually expanded at the expense of the rest of the course, and after the junior author joined the Yale staff in 1946 *Principles of Stratigraphy* became a full year course, shared equally by Dunbar and Rodgers.

The urgent need of a textbook for a course of this scope had long been felt, but its preparation first became feasible when we could share the labor. During the first year we divided the subject matter so that one of us would lead in the discussion of certain topics and the other would lead in other topics, but both of us met every class and took part in the discussions. In the second year we reversed the assignments, each leading in the presentation of the subject matter taught by the other during the first year, and again we both attended all class meetings. In the third year we reverted to the original distribution of subject matter and each began preparing a complete manuscript on the topics discussed, each instructor meeting the class in turn. As the book shaped up, each criticized the parts written by the other, and both subject matter and methods of treatment were modified in the light of experience. To an unusual degree, therefore, the present work represents collaboration of two authors with different backgrounds of experience and, in some instances, of different points of view and even of conviction.

Stratigraphy is a rapidly growing science in which there is still much to learn. Where wide differences of opinion still exist among stratigraphers we have tried to present conflicting views, feeling sure that future discoveries rather than any fiat on our part will determine which views are correct.

The literature on stratigraphy is vast, complex, in part conflicting, and rapidly expanding. We are acutely aware that we have not fully covered the field. To do so would require several volumes of text. We have felt it best, therefore, to present such matters as would give the student a grasp of general principles and would introduce him to the special literature in which he can pursue the subject further.

ACKNOWLEDGMENTS

In the preparation of this work we have received help from many sources. We are especially indebted to John E. Sanders who read the entire manuscript and made many helpful suggestions. Richard F. Flint likewise read five of the chapters, and Preston E. Cloud a sixth. Maurice Ewing generously placed at our disposal important data regarding the work of turbidity currents, and numerous other friends helped with specific problems. Our greatest debt is to the many students with whom the ideas here set forth have been debated and discussed over the years; the pool of their knowledge has broadened our understanding even as their thoughtful questions have sharpened our thinking on many problems.

Illustrations supplied by several institutions and numerous friends are identified in the figure legends; we wish to express to each of these our grateful thanks for this indispensable aid, as also to Werner F. Gossels and Shirley P. Glaser for assistance in drafting certain of the figures.

Our thanks go also to Mrs. Lorna Hodgson for her careful preparation of the typescript, and to Clara Mae LeVene for making the index.

New Haven, Connecticut
April 15, 1957

CARL O. DUNBAR
JOHN RODGERS

Introduction: the scope of stratigraphy

STRATIGRAPHY—literally writing about strata—is the study of stratified rocks. The term stratum (L. *sternere,* to spread) refers to planar units of rock that were originally spread as sheets over a surface of accumulation. The principles of stratigraphy were worked out primarily in sedimentary rocks, but they may be useful also in the study of such layered igneous rocks as ash falls and lava flows, and of metamorphic rocks insofar as they reflect an original sedimentary character.

The study of sedimentary rocks has three main aspects. The first is *sedimentary petrography,* the study of the rock material as such, its composition, texture, and structure. The second is *sedimentation,* the study of the processes by which sediments are formed, transported, and deposited. The third is *stratigraphy* proper, which deals with the overall relations of the stratified rocks, areal and temporal, and with the history they record.

Stratigraphy is necessarily based upon knowledge of sedimentary petrology and the principles of sedimentation. Here as elsewhere in geology, the present is the key to the past, and we can only infer the conditions under which ancient sedimentary rocks were formed when we understand how their modern counterparts are being produced. But stratigraphy goes beyond these basic disciplines in dealing with the broader relations of the layered rocks in the Earth's crust. In attempting to summarize the general principles used in stratigraphy we have necessarily introduced into this book much material belonging to sedimentary petrology and sedimentation, but no pretense has been made of covering those fields fully; each has its own vast literature and special devotees. Fortunately the literature in both sedimentary petrology and sedimentation has been summarized in recent standard works, and to these reference is frequently made in this book.

Stratigraphy in its restricted sense may be further subdivided into three phases.*

* Speaking humorously, we might in imitation of the trio petrography, petrology, and petrogeny, christen these subdivisions stratigraphy, stratilogy, and stratigeny; but in seriousness, the wholesale coining of such new compound terms from the classical languages is not a trend we wish to encourage.

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First is the description of the strata as they occur in sequence in local areas—a necessary if somewhat tedious procedure that provides the basic data for all further interpretation. Second is the correlation of these local sections—the determination of their mutual time relations and their place in the standard scale that forms the framework of historical geology. To many geologists these two parts make up the whole of stratigraphy, which is considered profoundly boring by all but those immediately concerned with the specific descriptions and correlations. But we believe that these, important and indispensable as they are, are but the means to a further end that constitutes the real core and interest of stratigraphy—namely, the interpretation of the stratigraphic record, both the rocks and their contained fossils, in terms of the past history of the Earth. Thus, though we have endeavored in this work to cover fully the subjects of description and correlation, we have devoted more than half the book to the discussion of principles and methods by which the stratigrapher interprets the data of description and correlation and builds up out of them a living picture of the geologic past.

The subject of stratigraphy has both an immense practical value and a broad philosophical interest. The stratified rocks contain the vast fuel deposits of the Earth's crust—all the coal and petroleum and much of the fissionable atomic fuel; they provide the reservoirs for ground water; and they include economic resources of many other kinds—stratified iron ore and numerous metalliferous deposits in sedimentary rocks, phosphate deposits, potash, sodium and other salts, gypsum, and limestone. Even where metalliferous deposits occur in intrusive igneous rocks or at their contacts in metamorphic aureoles the study of the surrounding framework of sedimentary rocks is commonly useful, even necessary, in working out the regional history that will permit an understanding of ore genesis and localization. From the broader philosophical point of view, stratigraphy provides the basis for understanding the past history of local regions and of the whole Earth—the changing patterns of land and seas, fluctuations of climate, even the history of the evolution of life on the Earth.

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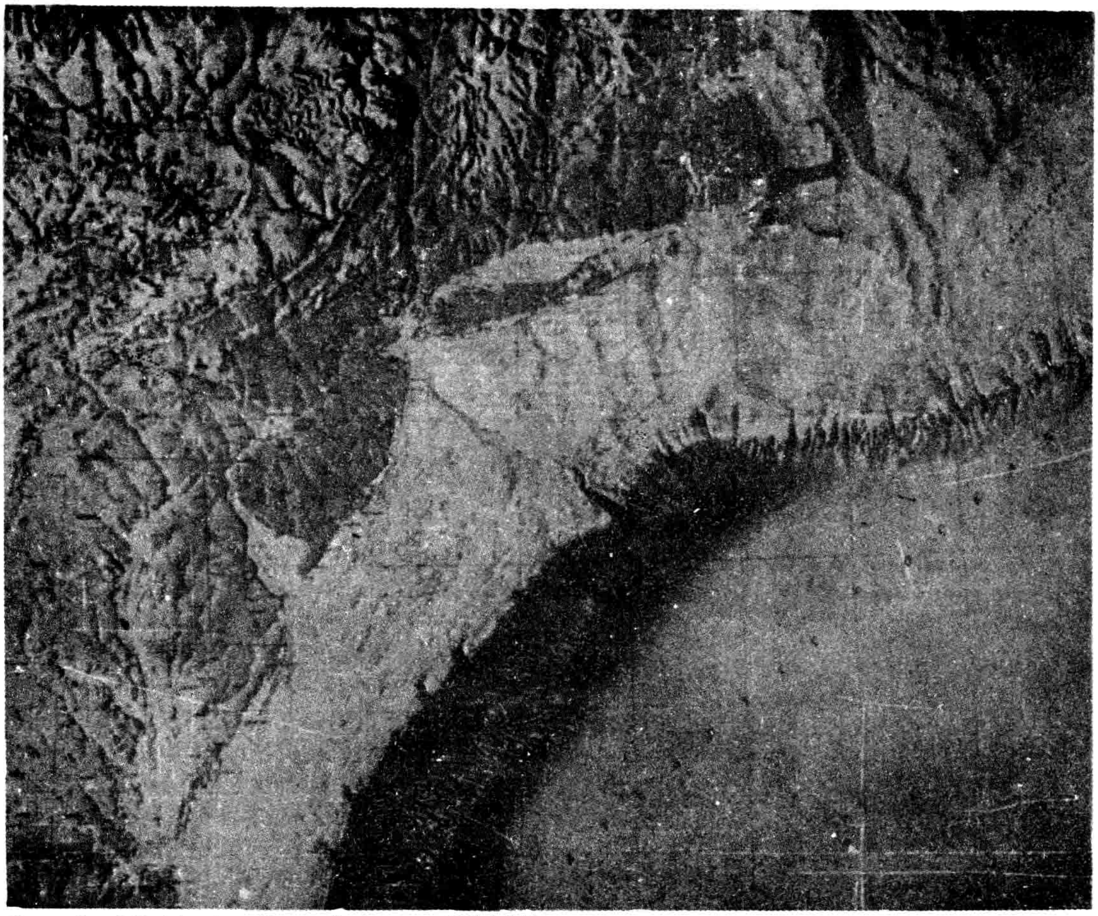
ENVIRONMENTS OF DEPOSITION



Plastic Relief Map by Aero, Service Corporation, Philadelphia

Figure 1. Relief model of part of the northeastern United States showing the continental shelf in its relation to the land and the continental slope. Vertical scale exaggerated 5 times.

1. Sedimentary processes

A SEDIMENTARY ROCK is composed of material eroded from some pre-existing terrane, transported to its place of accumulation, and deposited there. Such material is the grist of the sedimentary mill, and its ultimate nature is determined by what goes into the mill at the source, by what happens to it in transit and while exposed at the site of deposition, and by modifications during and after burial.

INFLUENCE OF THE SOURCE

The source area determines what goes into the mill. The Little Colorado River, for example, flows out of the Painted Desert in Arizona choked with mud as red as the Triassic redbeds from which it is derived. Ausable River, on the contrary, leaves the Adirondacks carrying clean white sand derived from the Potsdam formation (Cambrian), and the Wabash River, flowing through the fertile plains of Indiana, is roily with fine dark mud rich in organic matter.

Climate and relief also play an important role in determining the nature of the sediment at its source. A granite mass in a subarctic region will crumble into gravel and sand largely made of fresh feldspar, but in a warm humid climate, where deep weathering prevails, it will yield chiefly clay minerals and quartz. Even in the humid tropics, however, the same granite will produce feldspathic sediment if the slopes are steep and the rainfall is torrential so that physical erosion outstrips chemical decay (Krynine, 1935a). And if a deeply weathered surface be exposed to rapid erosion, as along the margins of a rising fault block, a mixture of much-decayed material and of fresh feldspar may be produced, like that of the Newark group (Triassic) of the Connecticut Trough (Krynine, 1950).

A mass of sediment deposited near its source will necessarily bear a strong impress both of the source rock and of the environment of the source area. Correct interpretation of a sedimentary deposit may be complicated, however, if the source rock itself bears the distinctive impress of an earlier sedimentary cycle. The till in the Central Lowland of Connecticut, for example, is locally as red as the underlying Newark beds (Triassic) from which it was largely derived, but over the crystalline uplands, both east and west, the till is entirely gray. In this instance, of course, the red color was not formed under glacial conditions but was inherited from a Triassic environment. In similar fashion, if the red mud of the Little Colorado River were spread over a nearby arid basin it would remain red even though the present environment in Arizona does not produce red color. Unless the source area were known and considered, the geologist of some future age would completely misinterpret the environment under which such a deposit was formed. Concentrations of clean sand, such as the St. Peter sandstone (Ordovician) of the Mississippi Valley and

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the Ridgeley sandstone (Devonian) of the Appalachian region, can hardly be the product of a single erosion cycle; their immediate sources must have been in vast sandy deposits concentrated during previous cycles.

TRANSPORTATION

Sedimentary material is normally transported in one of three ways—in solution, in suspension, or by bottom traction. The chief agents are flowing water, wind, and ice.

Transportation by Streams

Solution

The more soluble products of weathering go into solution and are carried away by ground water or by surface runoff to the streams and lakes and ultimately to the sea. At any stage in this journey, however, chemical reaction with other materials in solution may take place, precipitation may occur because of evaporation or of other changes in the physical-chemical equilibrium, or certain materials may be extracted by organisms and built into skeletons or living tissue. If none of these changes occur, the dissolved substances remain in solution. Scarcely any minerals are completely insoluble, and colossal amounts of all the common ions of which minerals and rocks are formed are present in streams, in lakes, and especially in sea water.

Suspension

Particles that do not settle readily to the bottom in a fluid are said to be held in suspension. The roily water of streams in flood owes its turbidity to mud carried in this way.

Influence of size and shape of particles. Particles of clay size settle very slowly and will remain in suspension for many hours, even in standing water. Large grains settle faster than small ones and those of high specific gravity more rapidly than light ones. Spherical particles likewise settle faster than irregular ones of equal mass because they offer less frictional resistance.

Rubey (1931, p. 28) made critical experiments to determine the rate of settling of sediment in pure, still water, with the results shown in Table 1.

TABLE 1. SETTLING VELOCITIES OF SEDIMENT IN STILL FRESH WATER. FROM RUBEY (1931).

	mm/sec
Very fine sand	>3.84
Coarse silt	0.96–3.84
Medium silt	0.24–0.96
Fine silt	0.06–0.24
Very fine silt	0.015–0.06
Coarse clay	0.00375–0.015
Medium clay	0.0009375–0.00375
Fine clay	<0.0009375

At these rates it would require about 2 hours for very fine sand to settle 100 feet, whereas fine clay would require about a year; to reach the ocean floor at a depth of 12,000 feet, very fine sand would need about 10 days and very fine clay would require more than 100 years. Neeb (1943, p. 94–95) reported that fine volcanic ash from the great eruption of Tamboro in 1815 is still settling out in some of the deep basins of the East Indies.

The settling velocity in still water is determined by the ratio of two forces, a downward force equal to gm (where g is the acceleration of gravity and m is the mass of a given particle), and an opposing or upward force, f (the frictional resistance), caused by the viscosity of the fluid. For particles smaller than fine sand the mass is small as compared with the viscous resistance of water and for such particles Rubey (1931, p. 17–31; 1933) showed that the settling velocity varies with the square of the diameter (Fig. 2, steeper curve), but for particles larger than coarse sand the mass is so great that viscous resistance is negligible and the settling velocity varies with the square root of the diameter (Fig. 2, flatter curve). For particles of intermediate size—fine to coarse sand—both mass and viscosity are important, and the settling velocity is intermediate between that for finer and coarser grades. The heavy curve in Figure 2 represents a general formula for all sizes.

The role of turbulence. In moving water, turbulence is an additional factor tending to keep sediment in suspension. In this connection two kinds of flow must be distinguished. At low velocities a fluid moves by a smooth gliding of fila-

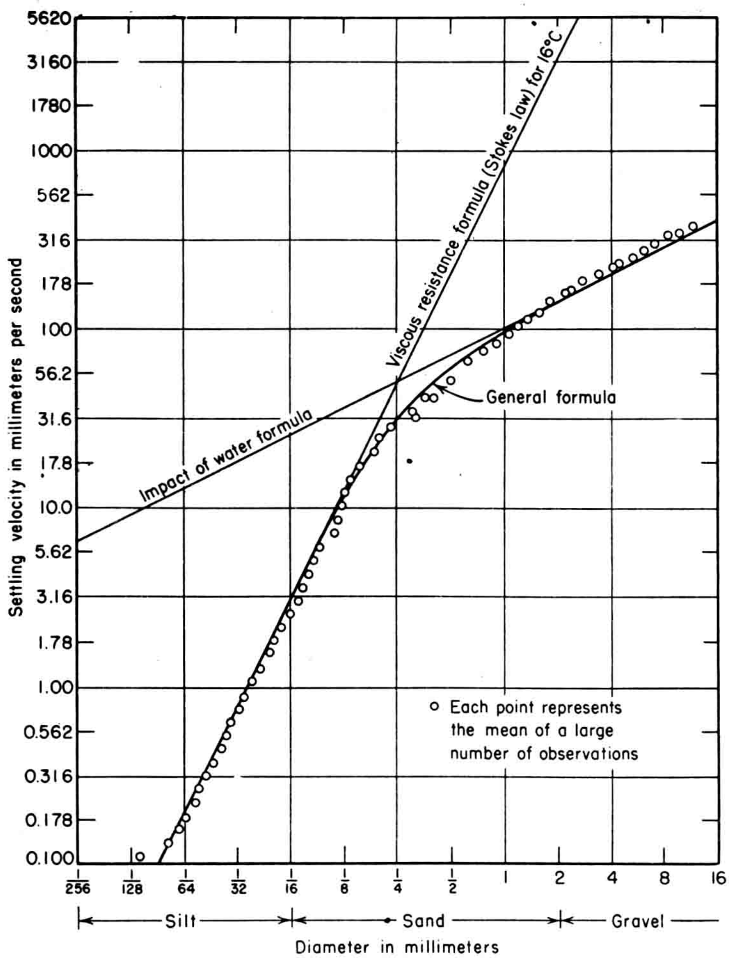

Figure 2. Settling velocity of quartz particles in still water, plotted on double logarithmic scale. From Rubey (1933). Experimentally determined velocities are represented by small circles.

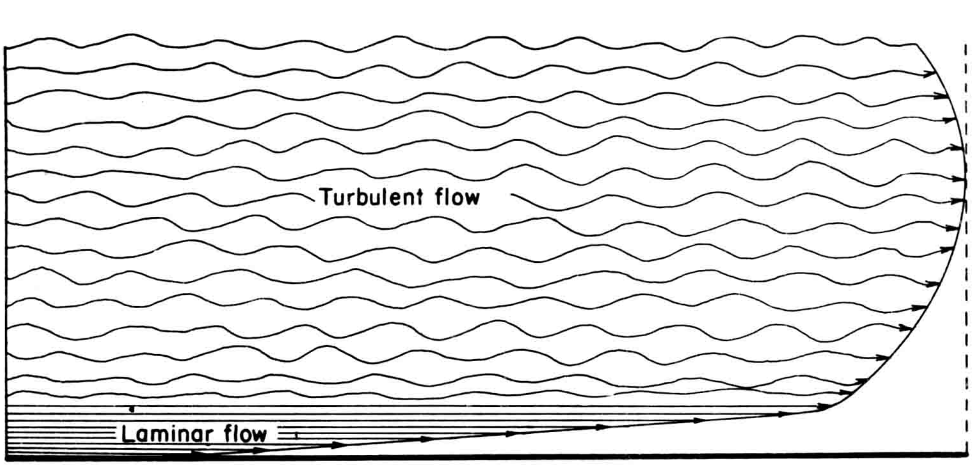

Figure 3. Laminar and turbulent flow in a stream. Velocity is indicated by the relative length of the flow-lines. Adapted from Rubey (1938).

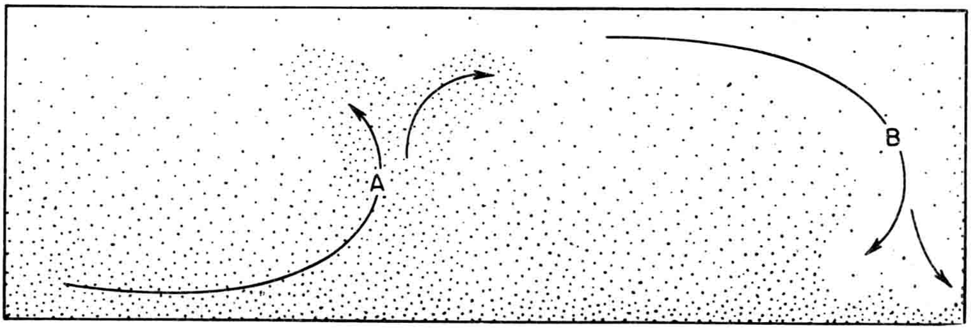

Figure 4. Role of turbulence in keeping sediment in suspension. The concentration of sediment is greater near the bottom and progressively less upward. The ascending eddy (A) therefore carries more sediment per unit volume than the descending eddy (B).

ments of current past one another, but at higher velocities the motion becomes irregular and distinctly eddying (Fig. 3). The first is *laminar,* the second *turbulent* flow. In laminar flow, particles of sediment settle as readily as in still water, but in turbulent flow, they are given repeated upward boosts that retard their settling. Of course, the upward eddies are on the whole balanced by downward eddies, and if the sediment were uniformly distributed throughout the current their effects would cancel out, the one carrying sediment downward as fast as the other carries it upward. But since solid particles constantly sink through the surrounding fluid, regardless of its motion, there is always a greater concentration of sediment near the bottom (Fig. 4). Thus ascending currents carry more sediment per unit volume than descending currents.

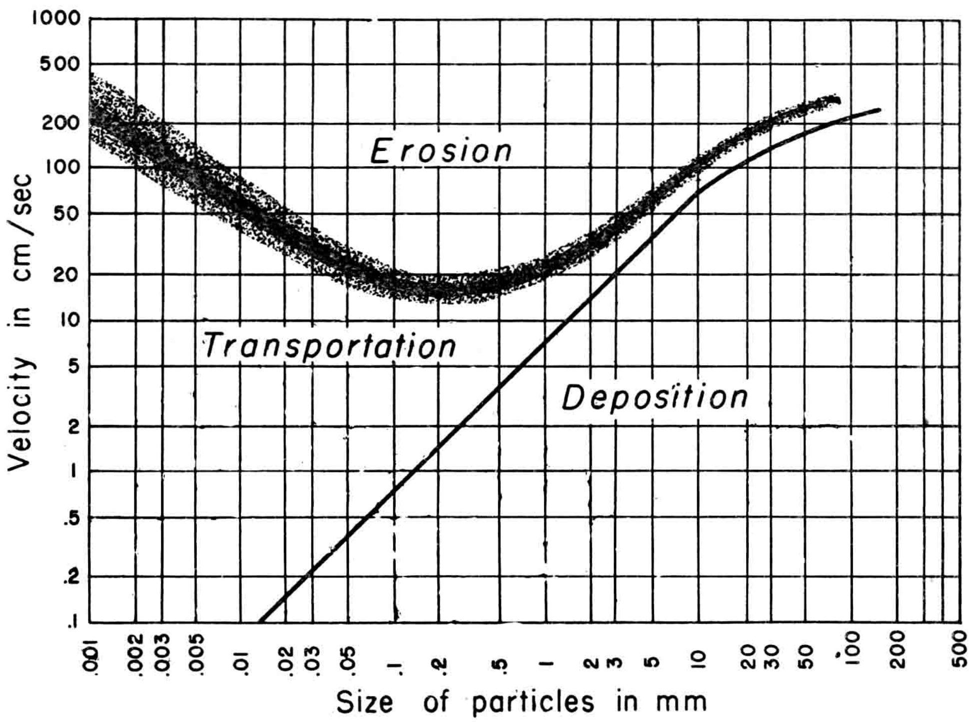

Figure 5. Graph showing velocities at which different size grades of sediment will be eroded, transported, or deposited; plotted on double logarithmic scale. After Hjulström (1935, p. 298).

Friction between a current and the surface of its confining channel causes a steep drop in the velocity gradient near the contact, where laminar flow is maintained. If the average velocity is considerable, the laminar flow may be limited to a mere film covering the surface of the materials on the bottom (whether they be solid bedrock or loose sediment), but in the movement of fine particles this layer plays an important role.

Resistance to movement after deposition. Hjulström (1935, 1939) showed that, after loose sediment of uniform size has come to rest, the velocity required to erode it is relatively high for the finest size grades, falls to a minimum for particles about 0.5 mm in diameter, and then increases again for coarser size grades (Fig. 5). There are two reasons for this apparent anomaly. First, a surface of uniform fine sediment has only microscopic relief, and individual particles project but little above the general level. Their exposed summits therefore lie within the bottom film of laminar flow or project but slightly above it and offer little surface for the turbulent current to work on. Second, the force of cohesion exerted by the clay minerals that are concentrated in the finest size grades inhibits the movement of individual particles. This explains the anomaly that silt and clay are more easily transported than sand but, once deposited, may be more difficult to move again.

Bottom traction

Coarse sediment is largely moved on the bottom where individual particles go leaping and tumbling along. Unhappily, the precise manner in which those at rest are picked up and set in motion, and the laws governing their movement, are not clearly understood. Gilbert's classic experiments on the transportation of debris by running water (Gilbert, 1914) were based on size-graded sediment in troughs having a flat bed of uniform cross section. While they provide a great mass of valuable data, the conditions represented in these experiments scarcely approach the complexity that exists in nature. Rubey's critical analysis of the force required to move particles on a stream bed (Rubey, 1938) shows the difficulty of the problem and gives references to the extensive and highly technical literature on the subject. Only the more general considerations need be presented here.

The analysis is simplified in the following discussion by assuming that the particles are spherical, but it should be remembered that in nature most of them are more or less irregular and that this adds complexity to their behavior.

Rolling. The simplest form of bottom traction occurs where spherical particles rest on a smooth surface. Here the force of the water is applied directly against the upstream side of the particles. Because of friction on the bottom and because the current striking its summit flows faster than that below, a particle tends to roll.

The influence of size is illustrated if we consider two particles, one having twice the diameter of the other (Fig. 6). The force exerted against such particles is mvk, in which m is the mass of water intercepted in a unit of time, v is its velocity, and k is a constant. The constant is needed because toward the margins of the sphere the water is deflected and only a small component of its force is exerted in the direction of movement. The mass of water intercepted is proportional to the cross section of the sphere, πr^2, but the mass of the particle to be moved by this force varies with

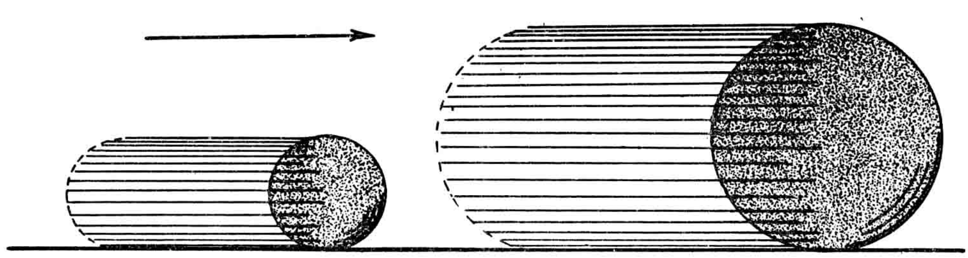

Figure 6. Influence of size of particles on velocity required for rolling on bottom. Horizontal lines represent direction of flow.

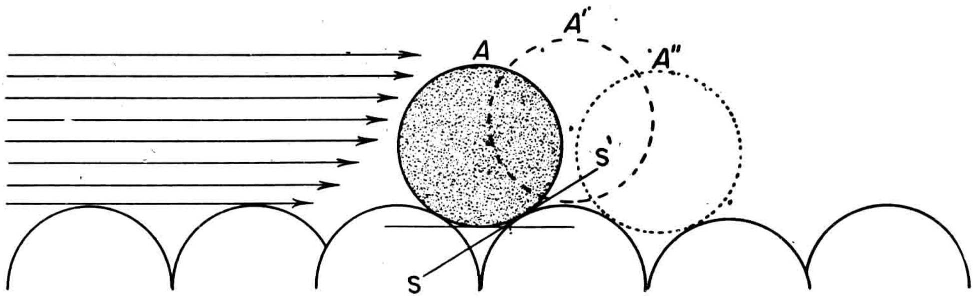

Figure 7. Beginning of saltation of a sand grain (A) resting on a surface of comparable grains.

the volume of the sphere, $\frac{4}{3}\pi r^3$. Accordingly, the larger of the two particles intercepts 4 times as much of the current but weighs 8 times as much as the smaller (assuming, of course, that they are made of the same material and that the velocity of the current is equal for both). This accords well with the common experience that sand is moved more readily than gravel and that as the velocity of a stream declines the coarsest particles of sediment come to rest first.

Saltation. In a stream the relations are rarely, if ever, so simple. Figure 7 illustrates a more normal condition in which particles of sediment are rolled over others. Here a particle, A, cannot be moved directly forward but must be pushed up an incline, s———s′, to a position A′ in order to get over the particle in front of it. Moreover, it is partly shielded from the current by the particles behind it. Both these circumstances tend to inhibit movement. Other forces are brought into play, however, for on such an irregular surface the flow is inevitably turbulent, and the eddy formed in the lee of an exposed particle produces suction. Moreover, the velocity of flow is not the same at all depths in a stream (Fig. 3, p. 5). Because of friction, it is least near the bed and the sides and rises to a maximum near the center of the channel and some distance above the median depth. The change per unit depth is rather gradual until the bed is approached and then the velocity gradient becomes very steep. Because the pressure diminishes with increase of velocity, a pressure difference is set up in the zone of steep velocity gradient and this produces a "hydraulic lift" (Hjulström, 1935, p. 267–270).

It is impossible to measure these competing forces experimentally, because any sort of gauge introduced will of itself create further disturbances. Rubey (1938) has given a theoretical analysis of the problem.

The force required to roll the particle A up the slope s———s′ to position A′ (Fig. 7) greatly exceeds that needed to roll it forward and down to position A″. Moreover, at position A′ it is elevated into current flowing more rapidly than at its original position or at A″. As a result, from position A′ it tends to leap forward into suspension until it has time to settle to the bottom again where the process is repeated. Consequently, the particle does not roll steadily along the bottom but proceeds in a series of leaps. Gilbert (1914, p. 26–30) termed this *saltation* and concluded that it is the chief mode of bottom traction in streams.

Isolated larger particles resting on finer, as a pebble or boulder lying on sand, tend to roll, and their movement is aided by excavation of sand in front of them by turbulent eddies.

Competence versus capacity

Gilbert (1914, p. 35) made an important distinction between the *competence* of a stream, which is a measure of the size of particles it can move, and the *capacity,* which is a measure of the total load it can transport. Competence depends largely on velocity, and capacity more on volume. For example, a small mountain torrent may be competent to move large boulders while lacking the capacity to transport a large quantity of sediment. The Mississippi River, on the contrary, lacks competence, in its lower reaches, to move boulders, but it has the capacity to carry the colossal load of some 500,000,000 tons of fine sediment that it delivers yearly to the Gulf of Mexico (Fisk and others, 1954, p. 80).

Load

The quantity of sediment actually being transported by a stream is its *load*. It is usually expressed in terms of weight (or of volume) of the material transported through a given cross section in a given unit of time, and normally refers to the solid material in suspension plus that moved along the bottom (some authors include material in solution as well). The material carried by rolling and saltation may be distinguished as the *traction load*.

The load of a stream of given velocity is influenced by the size of the loose particles available. A mountain brook, for example, may babble along crystal clear over boulders it is incompetent to move. Its load is zero. A second stream of identical size and velocity flowing over uniformly sorted gravel just within its capacity will carry a certain amount, but a third stream flowing over uniform sand will carry a larger load, and a fourth flowing over very fine sediment will carry a still larger load. Ideally, there is a capacity load for each size grade of sediment, the maximum being attained where all the available particles are of very fine size. In nature the sediment is normally heterogeneous, and a considerable range of size grades appears in the load.

Once fully loaded, a stream is unable to pick up additional sediment, however much may be available, and it may then flow along over a valley floor composed entirely of loose material.

As the velocity of a loaded stream decreases, both its competence and its capacity are reduced and it becomes *overloaded*. As competence declines, the coarsest particles are dropped first. The stream may then expend more of its energy on the remaining particles and may even pick up additional sediment, exchanging coarse for fine. The proportion of coarse material will vary from stream to stream and from place to place in a single stream. Rubey (1938, p. 139) concluded that, in a stream free to pick up much sand and gravel as its velocity is increased, the capacity load per unit of width will vary roughly as the cube of the velocity.

By-passing and total-passing

Since fine particles in motion are carried more readily than coarse ones of similar composition, and light ones more readily and more rapidly than comparable heavier ones, it is clear that even in a steady current a mass of heterogeneous sediment will not move as a unit. Even within the size range for which the current is competent, the finer particles will, in general, move faster and farther than the larger and heavier ones and, as the velocity eventually declines, the latter will come to rest while the former continue to move. Thus there is a constant passing of some particles by others during transportation. This phenomenon, termed *by-passing* by Eaton (1929, p. 714), is the means whereby a body of sediment becomes size-graded and the different components are winnowed out and segregated into deposits of gravel, sand, and silt. It is also probably the chief factor involved in the wear and rounding of gravel in transit.

Total-passing occurs where the current is able to transport all its load and none comes to rest.

Effects of stream transport on sediment

During transportation particles of sediment are modified by many factors. Kuenen (1956) recognizes seven distinct processes that operate to shape particles and reduce them in size. (1) Splitting is involved when a particle breaks into two or three subequal pieces. (2) Crushing occurs when a weak or small particle, caught between large ones, is pulverized. (3) Chipping involves the breaking of small flakes from the sharp edges of angular pebbles. It is important in the early stages of rounding but becomes negligible when the edges lose their sharpness. (4) Cracking produces tiny surface fractures by concussion when pebbles collide. Minute wedges may then be loosened between adjacent cracks and drop out. (5) Grinding is a form of abrasion produced when one pebble, pressed against another or against a rocky stream bed, is pushed bodily along. The result is similar to that involved in grinding thin sections. (6) Chemical attack, involving both rock decay and solution, is especially important where sediment lies ex-

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