Course material for Geology 109 - Geomorphology - 1. 1950-1954

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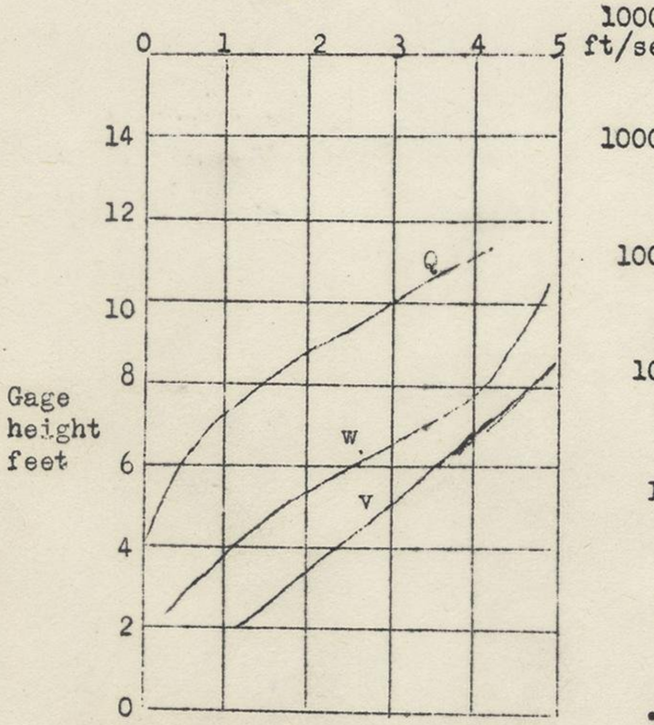
GEOLOGY 109
GEOMORPHOLOGY

Dimensions and competence of running water. Supplement I, 1953, p. 1

Introduction. Three papers have appeared on the subject of running water which appear to show marked progress in understanding of some problems. Two of these not only clarify some of the basic points of the physics of streams but also point the way to solution of many important problems of sediment transport. The third, deals with particle size distribution on an alluvial fan.

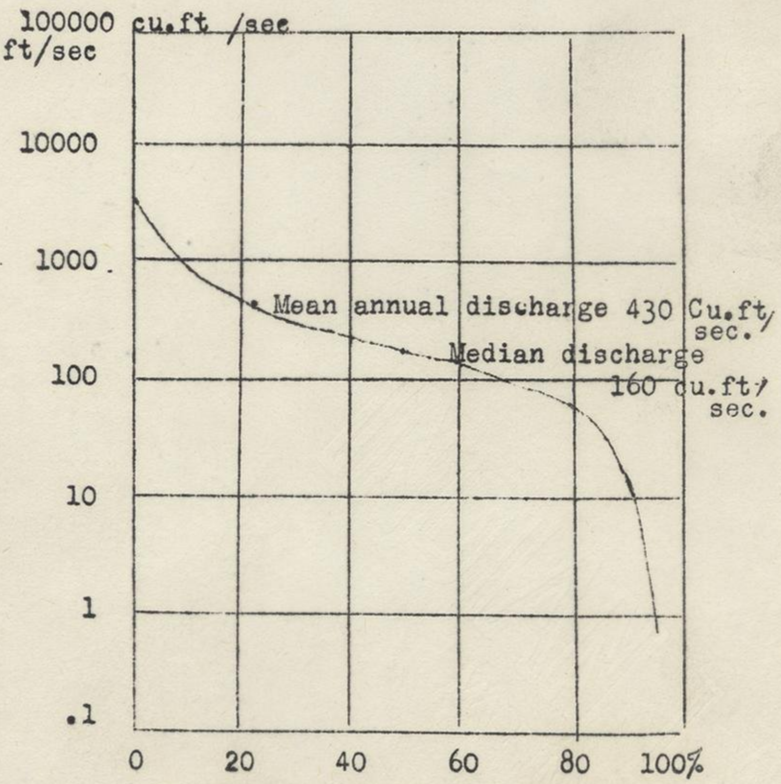
Discharge of streams. The fundmental quantity measured by hydraulic engineers is the discharge of streams. To find this figure they first discover a suitable cross section of the channel. This is subdivided into segments of known dimensions, then the average velocity of flow is found in each segment giving its discharge and the final sum of the segments is the Discharge (Q) = average width of channel (w) X average depth (d), X average velocity, (v) or Q = w.d.v. British engineering units are employed, cubic feet per second, and feet. Since the discharge of all rivers varies constantly it is necessary to connect each actual measurment to the gauge reading of water level in the river at that time. Most discharge determinations are read from a curve (Fig. 1) which indicates this relationship. Next a curve (Fig. 2) must be prepared which shows the percent of days that any given discharge is equalled or exceeded. The mean discharge is also computed as the aritmetical average of all recorded daily discharges. This quantity is generally larger than the median discharge which is equalled or exceeded exactly 50% of the time.

Fig. 1



Relation of quanities to gage reading in feet.
0 200 400 600 800 ft. width
0 10000 20000 30000 40000 cu.ft./sec.

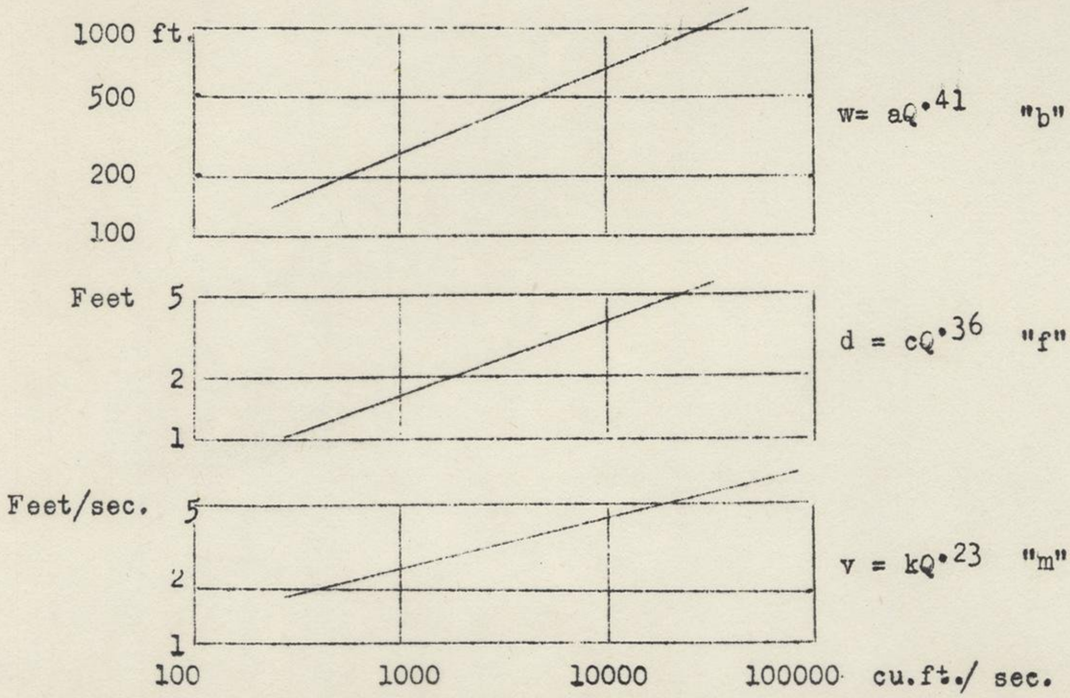
Fig. 2



Typical frequency or flow-duration curve of a river

<u>Inter-relations</u> of quanities. Platting on log-log paper demonstrates, as shown in Fig. 3, that w, d, and v are simple <u>power functions</u> of Q, the primary determination. In mathematical expressions $Q = wdv = aQ^b \times cQ^f \times kQ^m = ackQ^{(b+f+m)}$ From this it is evident that the sum of the exponents of Q must be unity and the product of the numberical constants must be the same. An average of 20 river sections studied gave b = 0.26, f = 0.40, m = 0.34 but the values of the constants varies much more widely than do the exponents. Evidently the values are related to the materials of the stream beds and possible to other factors. The limits of variation are unknown. Depth increases with discharge faster than does width.

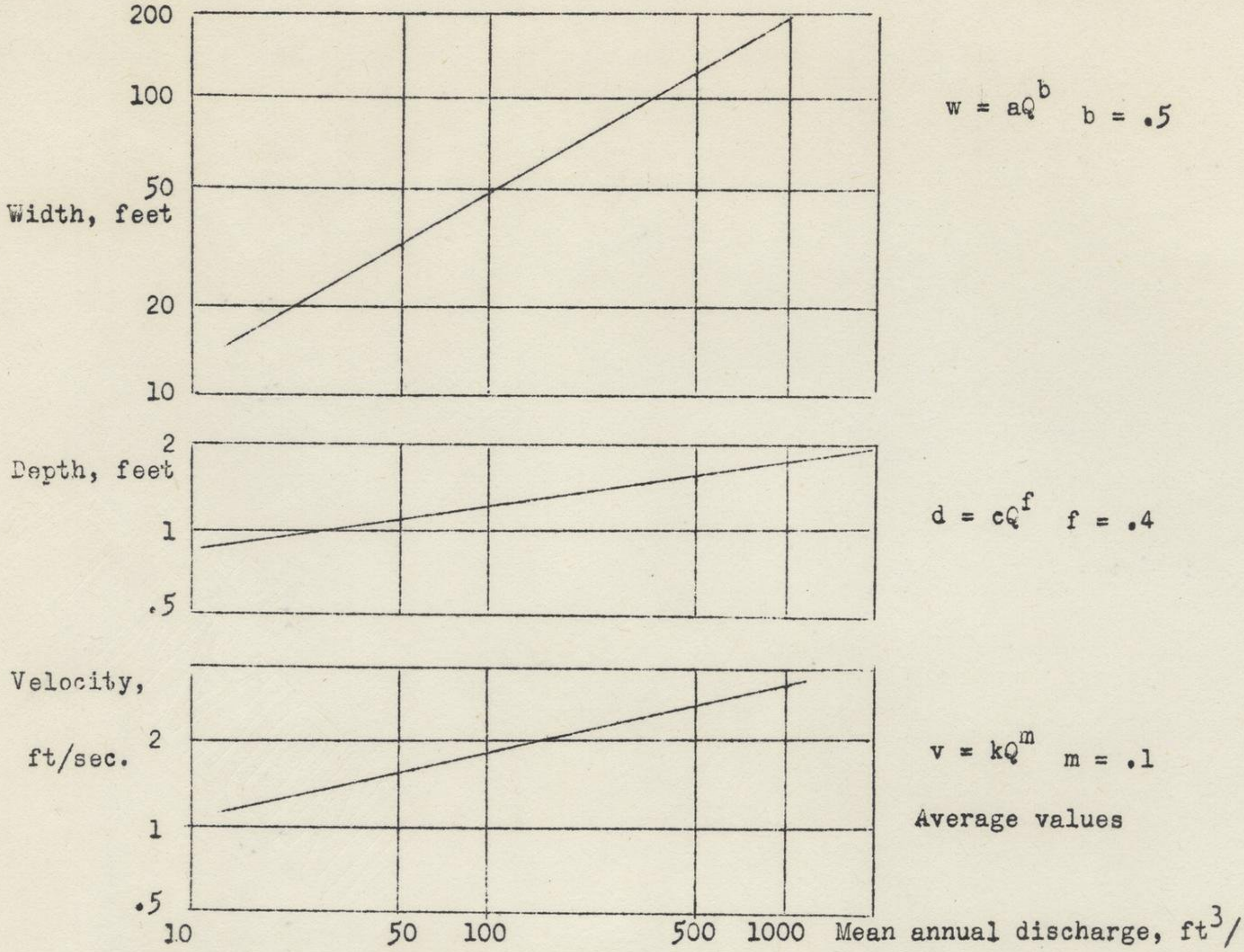
Fig. 3



Relations of width, depth and velocity to discharge as plotted on log-log paper, Scatter of points not shown.

<u>Downstream variations in channel shape</u>. In computing the relations of dimensions of stream channels in a downstream direction it is evident that all comparisons must be made for a specified discharge at every station. Most of the log-log plats were made for mean annual discharge which occurs or is exceeded on the average about one day in every four. In almost all rivers discharge increases downstream. Some were made for flows which occur less frequently.

Fig. 4



Downstream changes in dimensions of streams in Wyoming and Montana. Points not shown

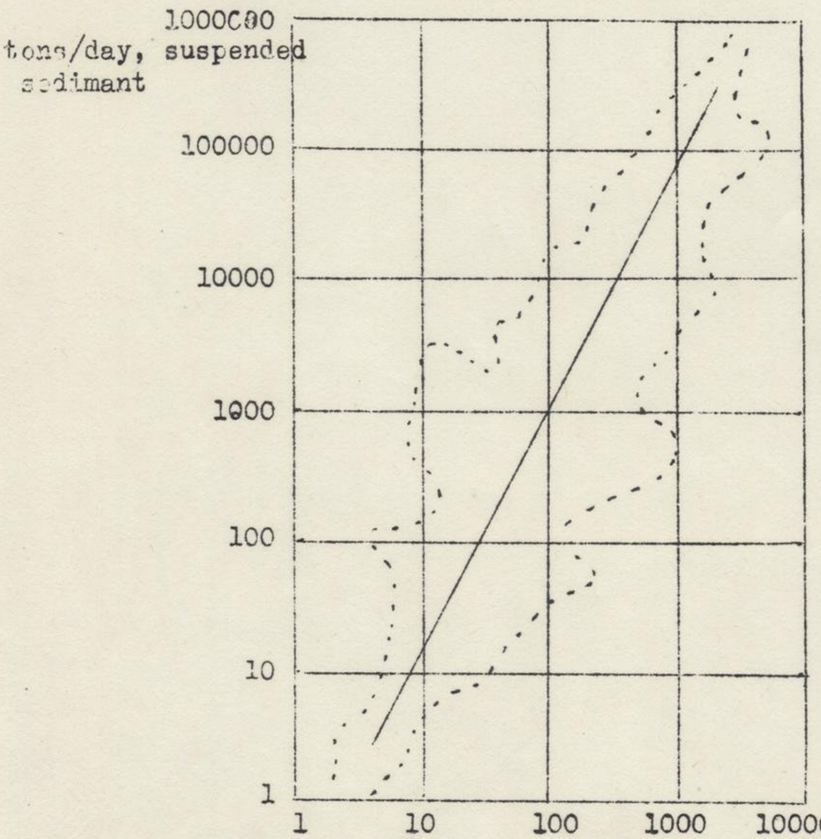
Despite the expectable "scatter" of points when platted, there is a remarkable agreement in results. Using the notation above, $w = aQ^b$, $d = cQ^f$, and $v = kQ^m$, the average values are $b = 0.5$, $f = 0.4$, and $m = 0.1$ This shows that for increase in discharge downstream all quantities including velocity increase. Increase in velocity is least and this quantity may be almost constant in some streams. Even in the headquaters however, the conclusion is demonstrable. It is contrary to what nearly everyone formerly thought and hence demands some explaination. To do this we will restate Mannings formula for velocity of a stream with turbulent flow: mean velocity (v) ft/sec $= \dfrac{1.5\ d^{2/3}\ S^{\frac{1}{2}}}{\text{roughness } (n)}$ (dimensions in feet) Note that for wide stream mean depth (d) replaces hydraulic radius (R or cross section area divided by width.) From this it may be seen that most geomorphologists have ignored both depth of water and roughness of the bed, Together these overcompensate for the fact seen in the field that slope of the water surface almost everywhere decreases downstream. Slope (s) in feet per foot $= 0.021Q^{-0.49}$ on the average.

<u>Sediment transport</u>. Streams carry sediment in two ways, (a) as bed load or bed-material, and (b) as in suspension or wash load. The two may change in proportion with alterations of the stream so that what is suspended at one time may be a portion of the bed and vice versa. The mathematical relations of the

two are only vaguely known for there is at present no accurate method of determining transport of material on a stream bottom. Any mechanical device to catch such load introduces changes in the currents which render the results valueless. Suspended load can be and is being measured at a number of localities. Possibly data on the filling of reservoirs may eventually supply some of the missing information. The following discussion is almost wholly on suspended load.

Suspended load. Platting of the weight of suspended load in given time against discharge of a stream shows at once (Fig. 5) that, despite scattering of points, the amount of sediment increases with discharge as a power function with with an exponent between 2 and 3, thus demonstrating an increase in more than direct proportion to discharge.

Fig. 5

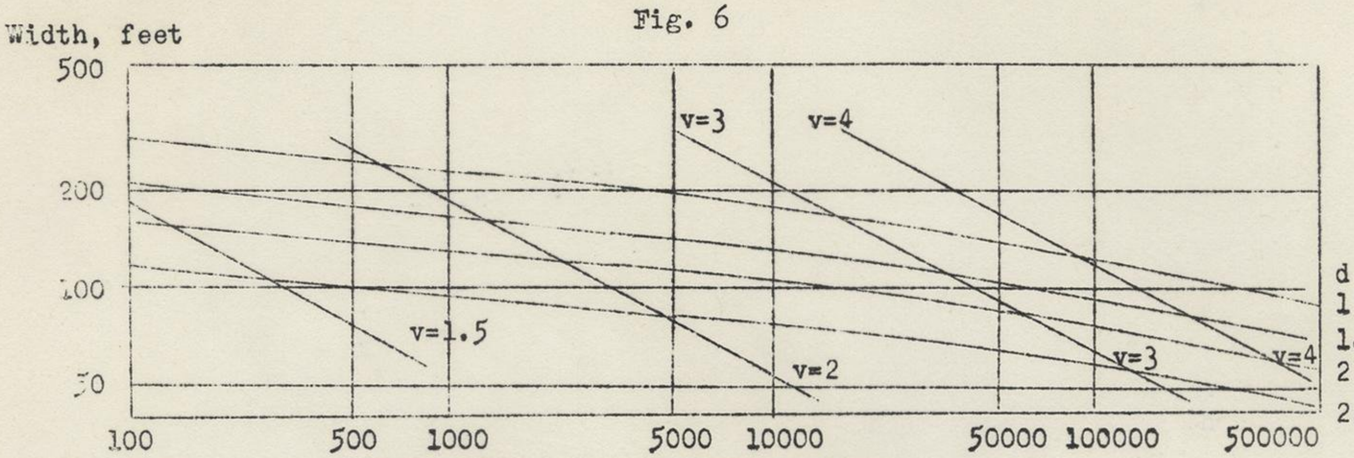


Relation of suspended sediment load to discharge, Powder River, Wyoming
Limit of scatter of determinations shown by dotted line

The cause of this rapid increase is known only in general terms. Factors are: (a) infiltration rate and storage of rain in puddles is greatest at start of a rain, (b) raindrop erosion increases with wetting of soil, (c) long duration of rainfall increases depth of and erosion by sheet wash, (d) increase in velocity of large streams enhances both scour of bottom and undercutting of banks, (e) changes in channel shape during a flood are caused by the suspended load, and (f) suspended sediment concentration may be considered as an independent variable on which both velocity and depth depend. Despite the known alteration of banks by floods, the conclusion of Leopold and Maddock is "that the observed increase in sediment concentration results primarily from erosion of the watershed rather than from scour of the bed of the main stream in the reach where the measurment is made." They found that there are not enough observations to permit of direct conclusions on changes in concentration downstream.

It appears to be slight so far as known for it is observed that increase in sediment with increase of drainage area is less for large basins than for small. It is possible to present a graph such as Fig. 6 showing the relations of width, depth and velocity to total suspended sediment load.

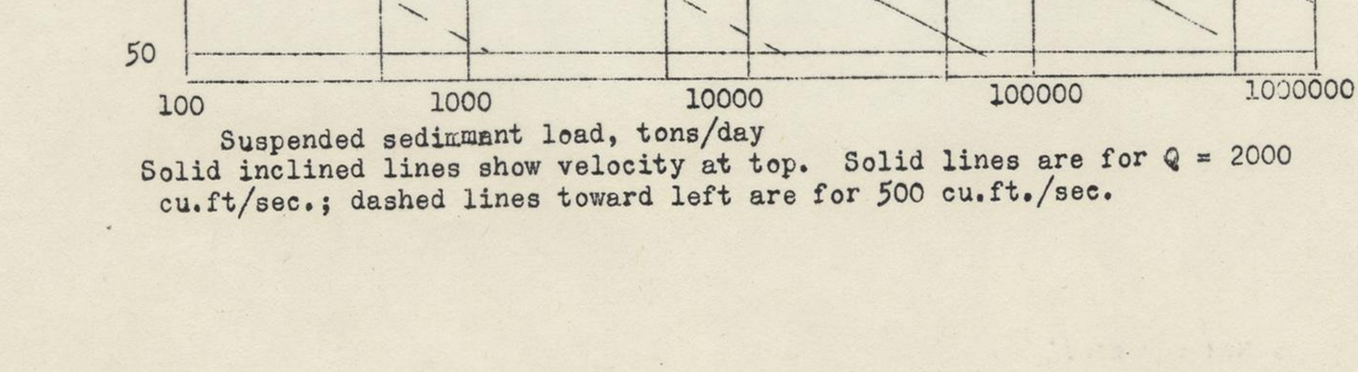
Fig. 6



Widths and depths in feet. Velocity in feet/second

General conclusions. (1) If discharge and width are constant increase in velocity means increase in total suspended sediment and a decrease in depth. (2) With velocity constant, increase in width decreases both the suspended sediment load and depth. (3) Both decreasing width with constant velocity and increasing velocity at constant width increase capacity for suspended load at constant discharge. (4) A wide river carries less suspended load than a narrow river with the same velocity and discharge. (5) Two rivers of equal width and discharge load of suspended solids is larger in that having the higher velocity. Suspended sediment transport with variable discharge. Due to fact that $Q = wdv$ the sum of the exponents $b + f + m$ must be unity as explained above. Hence if two of these exponents are known the third can be computed and from this fact some deductions may be made. First we draw Fig. 7 showing relation of suspended sediment to velocity, width, and discharge.

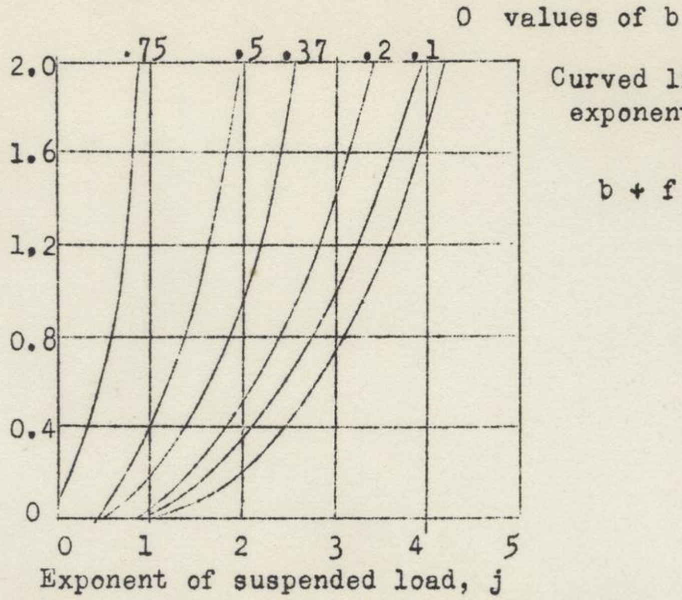
Fig. 7



Solid inclined lines show velocity at top. Solid lines are for Q = 2000 cu.ft/sec.; dashed lines toward left are for 500 cu.ft./sec.

Fig. 8

Ratio of exponents of veloity, m and for depth, f, m/f

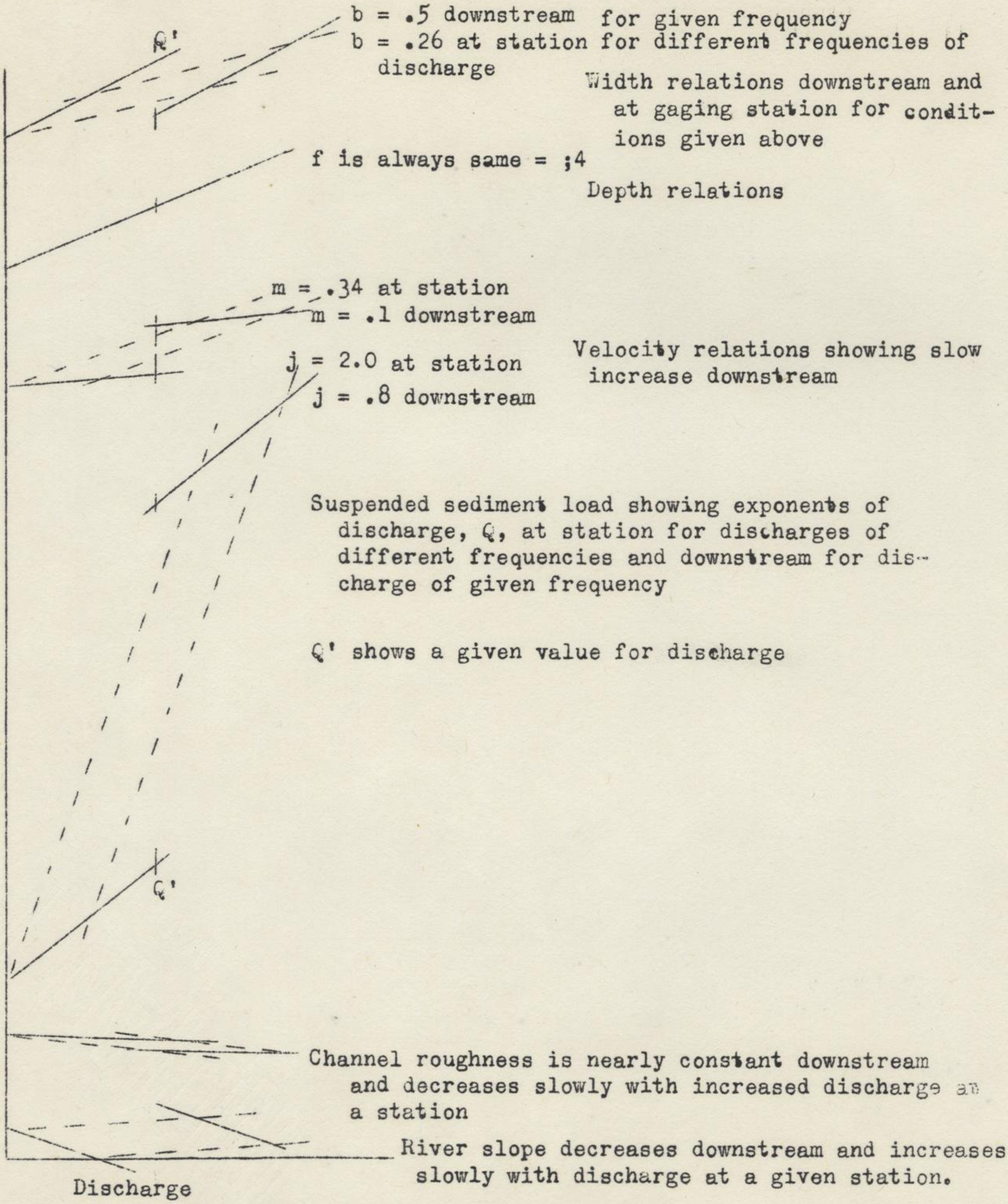


Curved lines are for the exponent b of width

$b + f + m = 1.0$

From this it is possible to draw curves showing values of j, the exponent of Q for suspended sediment, in terms of both b, and the ratio of m to f . The ratio between increase of velocity with discharge and increase of depth with discharge is, therefore, related to amount of suspended sediment. For the average cross section of a river m/f is 0.85, b = 0.26, and j = 2.3 This is in line with the statement that sediment concentration should decrease slightly downstream. (Fig. 8) Comparisons of different river cross sections indicate that: suspended sediment load varies: (1) directly with as a function of velocity, (2) directly as a function of depth, (3) inversely as a function of width, (4) as a large power of velocity, and (5) as small powers of depth and width.

Fig. 9

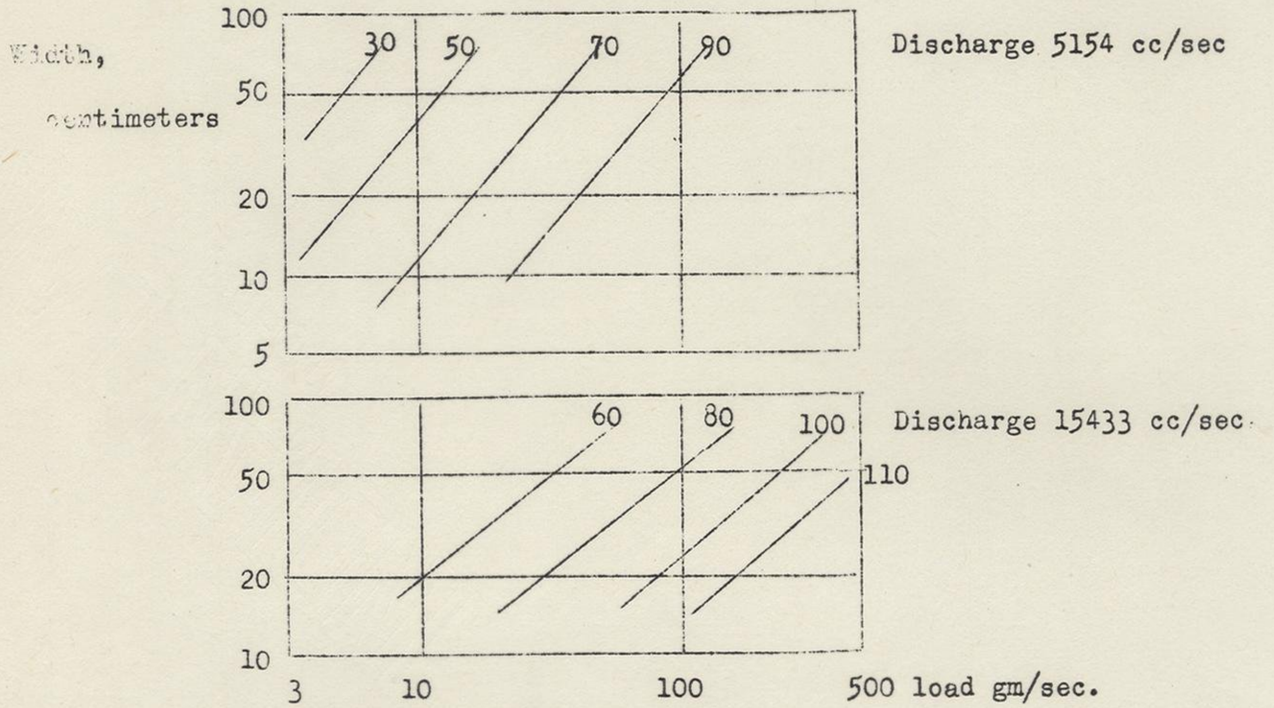


Solid lines for slope on exponent on log-log plat to show downstream change for Q of given frequency Dashed lines for changes at station.

Fig. 9 summarizes the information by showing the comparative changes at a station and downstream by giving the proper slopes of the lines which display the values of the exponents of discharge in log-log platting. We may say that for given width and discharge increase in suspended sediment requires increase in velocity and reduction in depth. The quantities involved are adjusted to the nature of the drainage basin so that they are independent of the channel system.

Bed load. Since there is little information of bed load transport in natural streams recourse must be had to the experiments of Gilbert in wooden troughs here restated in the C. G. S. system. Fig. 10 shows at once that the relation of the lines of equal velocity is exactly opposite to those of Fig. 6 for suspended sediment. Data are given for two different discharges both with same kind of sand. Tentative conclusions are! (1) with constant discharge and width increased velocity increases both bed load and suspended sediment, (2) with constant velocity and discharge increase of width decreases suspended load and increases bed load, (3) broad shallow channels are needed to transport a large bed load.

Fig. 10



Inclined lines show velocity in cm/sec.

Changes of channel form. At some gaging stations measurments have been made of changes in channel form during floods. Some places at the start of a flood, when concentration of suspended sediment is high, display a rise in level of the bottom. This is followed, when sediment decreases, by scour and lowering of the bed. Obviously the latter causes a lower velocity when less velocity is needed for transport. At other places erosion begins at once with the rise of discharge with high sediment concentration and later filling takes place during fall of water level. It has been noted that the spring floods of melted snow in western rivers lower river beds whereas later season floods due to rain result in fill. Filling often occurs during times of increasing velocity.

Roughness of channel. At constant width and discharge it is obvious that the product of v. d must be constant. Hence any increase in velocity requires a decrease in depth. From the usual velocity formula it is evident that for any increase in velocity and decrease in depth the factor $\left(\frac{S^{\frac{1}{2}}}{n}\right)$ must increase.

The two equations: $d = cQ^f$ and $v = kQ^m$ make it possible to set up another. $kQ^m = 1.5\ (cQ^f)^{2/3}\ S^{\frac{1}{2}}/n$ where the constants c and K vary. Hence $Q^m : Q^{2/3 \cdot f}\ (S^{\frac{1}{2}}/n)$ Where S and n are constant with discharge then $m = 2/3\ f$ or $m/f = 2/3$ From this it follows that if $S^{\frac{1}{2}}/n$ increases with discharge m/f is more than 2/3 and when this ratio decreases with discharge then m/f is less than 2/3. Now at a given station the average ratio of m/f is 0.85 whereas downstream this is only 0.25

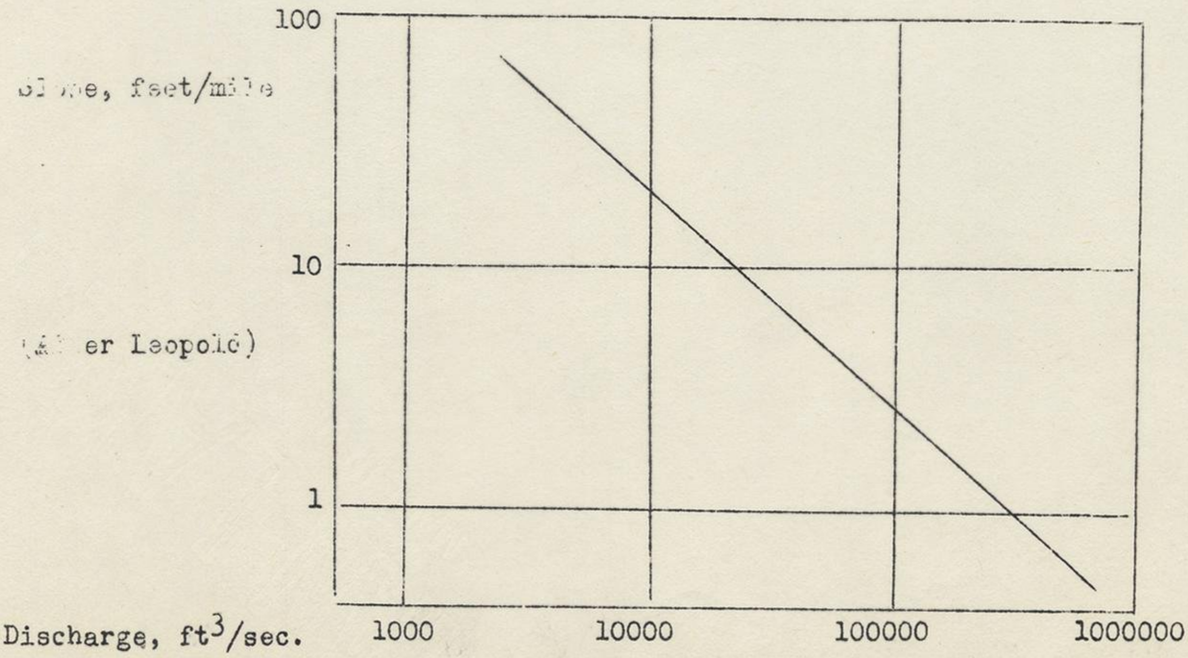
From this it appears that $S^{\frac{1}{2}}/n$ increases with discharge at a given station and decreases downstream. It has also been observed that in the downstream direction roughness (n) remains about constant so that slope must decrease to preserve the above relations. Observation has also disclosed that an increase in suspended load decreases channel resistance and hence increases velocity. Possibly this is really related to decreasing turbulence. Increased values of sediment concentration are associated with decreased values of n. At a given station, however, the slope does not change very much so that the alteration of n must be considerabl with change in concentration of sediment. Changes in velocity-depth relations might be attributed to change in sediment concentration where an increase diminishes the roughness, n, of the bottom. A check consists of the behavior of Colorado River after the completion of Boulder (Hoover) Dam which caught much of the sediment leaving clear water below. This is the same as a lake in the course of a river. Alterations below the dam consist of (1) increase in depth in spite of a lowering of surface elevation, (2) decrease in width due to reduction of flood volume, (3) decrease in mean velocity, (4) increase in roughness of bottom, apparently a result not of change in type of material but of decrease in suspended load, (5) reduction of bed load in the narrowed channel, (6) increase in capacity for suspended load due to change in velocity and discharge, (7) no appreciable change in slope.

Factors of channel roughness. Channel roughness is due to (1) particle size, (2) bed configuration, and (3) sediment load. It is commonly observed that the material of most stream beds diminishes in size of particles downstream although from this it does not necessarily follow that decrease in slope is directly attributable to this phenomenon. Waves and ripples on the stream bed are very important factors in roughness, although they are not permanent. Increased bed roughness decreases velocity in respect to depth hence affecting the capacity for load. These waves or ripples vary in nature with different kinds of sediment. They pass with increasing discharge from smooth bottom through successive forms into antidunes which travel upstream. For fixed slope and discharge decreased particle size tends to increase roughness. Bottom material is most important in the headwaters of streams where the bed consists of boulders, cobbles, and pebbles. Under this condition, downstream decrease in size of particles decreases roughness. The Powder River, Wyoming, has a value of n on gravel of .087 which falls to .017 on silt farther downstream. However, in other streams the value of n is about the same downstream despite marked differences in nature of bottom. There it must be that bottom configuration is dominant. In summary, it is clear that slope is the dependent factor which the stream is able to change. As noted above it is common to find at a given station that suspended load of streams increases rapidly with discharge. This requires a relatively rapid increase in velocity compared to depth, that is a high value of m/f. Such is accomplished primarily be an increase in the value of n which is related to increase in concentration of suspended load. However, in a downstream direction load does not keep pace with discharge and the concentration of suspended sediment decreases slightly. To do this depth must increase with discharge faster than does velocity so that the m/f ratio must be low. Hence $s^{\frac{1}{2}}/n$ must decrease downstream. With roughness about constant this can be done only by decreasing the slope.

Graded streams. By definition a graded stream can over a period of time just transport the amount of sediment furnished it. Engineers have constructed many irrigation canals which do exactly this, that is they neither erode nor silt up. Some rules were derived by experiment which used perimeter, P, instead of width and hydraulic radius, R, instead of mean depth. A sediment factor, F, is also introduced. The basic equations are: $P = 2.67\ Q^{\frac{1}{2}}$ and $V_{mean} = 1.15F^{\frac{1}{2}}R^{\frac{1}{2}}$. Note that in the studies of Leopold and Maddock they found that $w = aQ^{\frac{1}{2}}$ (downstream). By combining the relations $d = cQ^f$ and $v = kQ^m$ we find that $(d/c)^{1/f} = (v/k)^{1/m}$ or $v:d^{m/f}$. In natural streams this ratio of m to f downstream is only 1/4 whereas in the canals it was 1/2. But we must recall that canals for irrigation are not like streams because they loose discharge downstream as it is dispersed into laterals. They can have no change in suspended sediment concentration hence the value of j cannot be above 1.0. If b = .5 and j = 1.0 this means that m/f would be 0.43 or not far from that value already given. This suggests that j must in practice be less than 1. In summary, Maddock and Leopold conclude that with available data it is not possible to discriminate graded from ungraded sections of a river.

Longitudinal profile of rivers. It has long been assumed that the profile of a river bed is directly related to the maximum particle size of sediment in its bed. It has also been assumed that wear of the load results in a downstream reduction of size of particles. The latter can be checked in the field, although it is hard to distinguish material derived from tributaries and cut banks and not brought far downstream. Now if the velocity of flow really increases downstream how can competence of the current be the controlling factor of river profiles? Some have derived equations to substantiate this assumption but the issue is confused by several phenomena. (1) Decrease of particle size increases roughness by promoting ripples; (2) roughness is also related to concentration of suspended sediment and, (3) in practice roughness does not vary much downstream. Hence to preserve the required velocity-depth relations to transport the load the slope of a normal stream must decrease downstream. Leopold gives the empirical equation that slope, $S = 0.021\ Q^{-0.49}$, that is slope is approximately inverse to the square

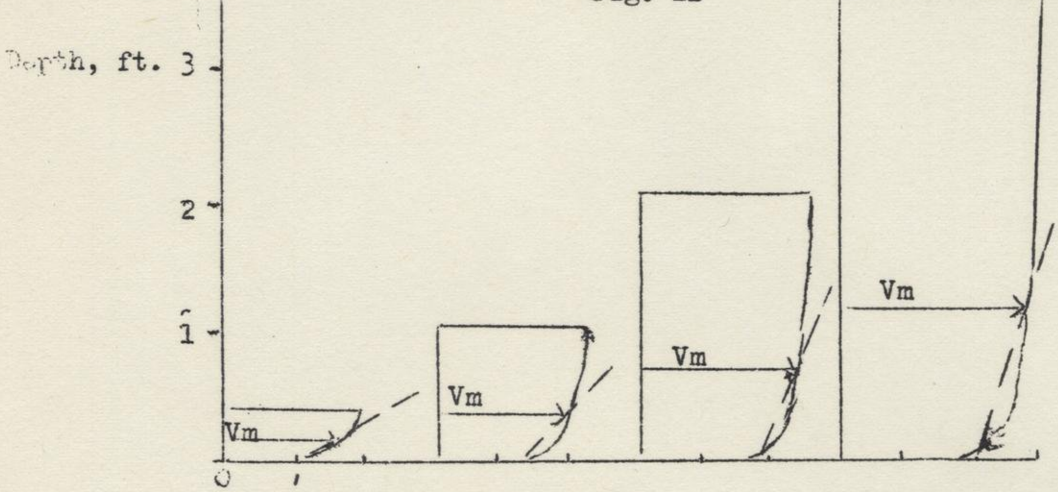
Fig. 11



root of discharge. We cannot, however, construct a longitudinal profile of a river from this without knowing how the discharge varies in a downstream direction. This is commonly in direct proportion to drainage area not to distance along the channel.

Vertical velocity distribution. It has long been known that in rivers which are relatively wide in proportion to depth, that is where the banks are readily erodable, the vertical distribution of velocity is approximately proportional to the logarithm of distance from the bottom, z. Such being the case the rate of increase of velocity with respect to distance from the bed is inverse (see any text book of Calculus). Now this rate of change in velocity upward from the bed determines the shear or rate of energy transfer from the stream to the bottom. Since in most streams depth increases downstream as a power function of discharge the slope of the line representing rate of velocity (dv/d2) change near to the bed must decrease with increase in total depth.

Fig. 12



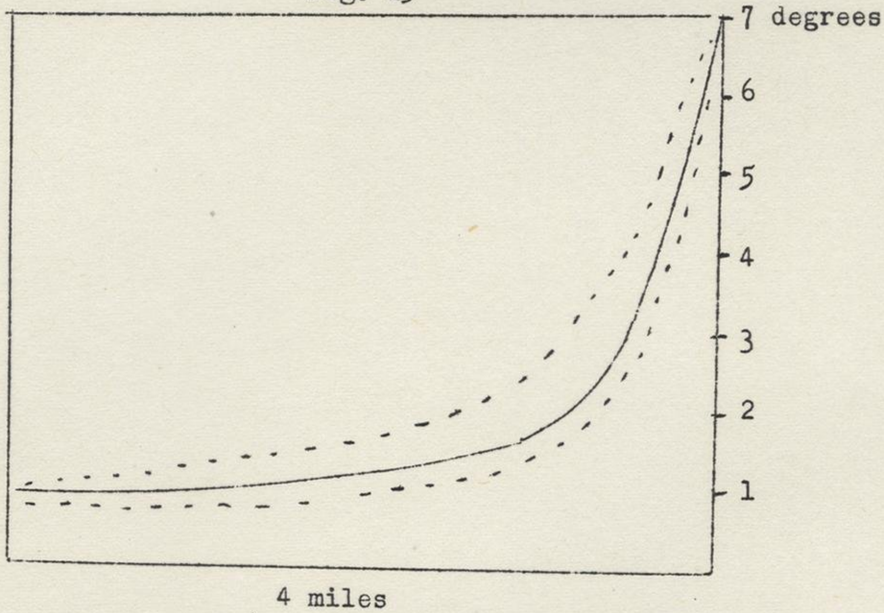
Velocity, ft/sec.
Inclined dashed lines show downstream decrease in shear

Another factor is that total force on the bed is proportional to depth times slope. As brought out above, depth increases on the average at the 4/10th power of discharge whereas slope decreases at approximately the square root of that quantity. Hence the product DS must decrease slowly downstream at about the minus 1/10th power of discharge.

Summary. Although the old idea that river slopes are directly related to velocity which decreases downstream thus decreasing competence must be abandoned, it is clear that there is a downstream decrease in competence. The details of just how this comes about are not simple. The vertical velocity profile and shear on the bed are interrelated and depend not only on mean velocity but also on depth, and on roughness of bottom. This shear also affects the intensity of turbulence which is necessary to keep material off the bed. Downstream decrease in roughness may diminish both shear and turbulence despite increase in mean velocity. Leopold lists the variables which enter into this problem: discharge, width, depth, velocity, slope, roughness, load, and size of particles in transit. These constitute eight simultaneous equations whose solution is at present impossible. Of them only the flow equation (Q=wdv) and Mannings formula for velocity are accepted by common use. The others comprise relation of load to nature of basin, rate of particle size change downstream, width-depth ration in relation to nature of the bed and banks, change in value of n, the roughness factor, with depth, material, discharge, and slope, and relation of n to sediment concentration. The interdependence of these factors is evident and it is clear that the stream is capable of adjusting its slope to fit the requirements of the others. The cross section of a stream is adjusted so as to equalize shear on both bed and banks. The form of the bed can be changed so as to alter roughness. All of these factors are much more complex than we were led to believe by the pioneer students of geomorphology who did not employ quanitative methods even if they were correct in general principle.

<u>Change of particle size downstream</u>. As explained above it is generally impracticable to measure the downstream reduction of size of particles transported by a river. On alluvial fans, however, all the debris is derived above the apex and reasonable success has been attained in comparing the maximum particle size with distance from the source. An article by Blissenbach based on fans in Arizona shows (Fig. 13) that despite considerable scatter a definite

Fig. 13



After Blissenbach Dotted lines show scatter of points

relationship does hold. From the known fact that diameter of pebbles is related to the square of velocity of transporting water it could then be concluded that the ratio of mean depth (or hydraulic radius) to bottom roughness must remain reasonably constant. On alluvial fans this might be expected for all the water is derived from the head so that the individual streams on the fan do not vary widely in size despite some loss by evaporation and perhaps by seepage. Roughness, which should decrease with smaller particles downward on the fan, could be maintained by more ripples in the bed on lower slopes. The log-log. platting (using slope as directly proportion to degrees measured) of the diagrams published show that slope is approximately inverse to the square root of horizontal distance from apex. Fall must, therefore (see integral calculus) be in proportion to the square root of distance from apex. The same paper also presents some data on relationship of maximum particle size to angle of slope (on steeper slope the degrees do not correspond directly to the technical definition of slope which is tangent of the angle) whic seem to confirm the determinations of Fair in South Africa. In the case of the Black Hills terrace gravels there is rough agreement of slope to logarithm of geometric mean size of stones. All of the above data is inconclusive for no attention has been paid to mean particle size of entire deposit and it is known that there is much finer material along with these maximum particles. On a table does the average or medium size control the coefficient of friction?

<u>References</u>

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GEOMORPHOLOGY

Supplement - Running water 1950

<u>Forces retarding stream flow</u>. As brought out in the text stream flow in any given portion of the course is retarded by forces whose sum is essentially equal to that of the current so that there is no acceleration. Following Cook, we may list:

(1) overcoming viscosity by internal shearing, for the most part raising temperature. In water this loss is apparently slight. Viscosity is neglected in formulas for velocity of turbulent flow.

(2) Kinetic energy of rotation due to turbulence - a considerable portion, although variable in amount; some of the loss is recovered whenever degree of turbulence is diminished.

(3) Transportation of debris, both suspended and bed load, the latter in part due to impact of particles on the bed loosening others. To maintain material in suspension demands a vertical component of motion at least equal to the settling velocity of particles concerned. Debris also increases viscosity.

(4) Erosion or changing of the shape of the bed. It is with this that the study of geomorphology is primarily concerned. Its energy demands must be considerable. This is in general treated simply as friction on the bed.

(5) Formation of surface waves - a negligable part of the total, except perhaps when the waves first start.

(6) Formation of ripples and ridges in loose material on the bottom. This increases turbulence and in laboratory experiments consumes a much higher amount of energy than in natural streams. At high velocity ripples disappear.

Now the relative amounts of available energy which can be devoted to these varies in a complex and, for the most part, unknown manner. The first two listed must always be met before any energy is available for the others.

<u>Power versus force</u>. A flowing stream develops a certain amount of power which is defined as force multiplied by velocity ($P = FV$) or as weight of water in a given time multiplied by slope ($P = WQS$), where W = unit weight of water and Q = volume passing in unit time. It is, therefore, possible to compute the amount of force which moving at the observed velocity would create the observed amount of power. It is then evident that this "fictitious" force is that component of the weight of unit column of water which is parallel to the bed of the stream. Thus we have the familiar DuBois equation, F = depth x slope ($F = WDS$). Although it is impossible with present knowledge exactly to evaluate all of the methods of energy dispersal, it is clear that no single one can be changed without affecting all the others. Cook concludes that, save possibly in exceptional circumstances, the addition of a suspended load decreases total energy available for debris movement. He comments that Rubey's attempt is too simple to take account of all factors for his formula considered only energy to keep this load off the bottom plus that consumed by " friction on the bottom."

Erodability of soils. Despite much work by those interested in soil conservation no very definite principles of wide application have been discovered which serve to compare the erodability of different soils. Most work has been done on relative amount of erosion with different types of crops which serves simply to reaffirm what has long been known, namely that erosion is at a minimum when the land is covered by either forest or grass. The importance of direct impact of raindrops on bare soil has also been proved.[2] Soils have been shown to vary widely in case of dispersion into suspension in water. Those which contain alkalic compounds are readily broken down. Other soils swell on wetting. The colloids in some rapidly fill up the voids when wet, increasing runoff. In this connection it is well to mention that mechanical analyses tell little on the subject of soil particles or granules because a very strong dispersing agent was first used before making the size separations. Many students of erosion lay much emphasis on the ratio of rainfall-intensity to infiltration-rate. Some experiments show definitely that under conditions tested percent of runoff ("water loss" as used by soil erosion students) increases with length of slope only when the rainfall-rate exceeds a certain value. With slow rains the opposite may occur. The same remarks apply also to soil loss measured in tons per acre. Few infiltration rates have been determined on undisturbed soils. Middleton considered as factors regulating erodibility of soils: amount of organic matter, silica-sequioxide ratio and total exchangable bases. The ratio of colloids to moisture equivalent or dispersion ratio is also given. This is the ratio of silt to clay in mechanical analyses expressed as percent. If this is less than 15 percent the soil might be classed as non-erosive.

Relative importance of sheet vs channel erosion. Most texts treat erosion of valleys as primarily the result of channel cutting, although the importance of slope or sheet erosion is recognized by some authors. Horton[3] presents evidence that erosion of the slopes by wash is much more important quanitatively than is channel or bank erosion. He concluded that since the total area of channels is a very small portion of the entire basin the relative importance must have about the same ratio. A sheet of water $\frac{1}{4}$ inch deep flowing at 1 ft./sec. will produce a surface runoff of 22,000 ft.^{3}/sec. from each linear mile of stream in a basin of 100 m^2 and a total stream length of 100 miles. If loaded to 10 percent by volume this would in 6 hours remove a depth of soil of nearly 1/3 inch. He cautions that some have failed to distinguish between erosion-rate and transporting capacity. Once capacity is attained erosion is governed by transporting rate alone. Both are in general expressed as a power of runoff-intensity. Letting amount of suspended matter at a given point by E, and distance from the divide by h, then E: h^n, where n is an exponent. This quantity, E, is equal to the sum or integral of all erosion down to that point. The rate of erosion at any given point is then (by differential calculus) proportional to $m \cdot h^{(n-1)}$ which is equivalent to dividing by h and multiplying by the exponent to obtain an average value or rate of increase per unit of horizontal distance. Experiment indicated that the value of n is greater than unity. Since the total depth of erosion at any given point is the product of rate times duration of flow, which is nearly constant on any given slope, the result is a curve which is concave upward. Horton carried his conclusions further pointing out that the average drainage basin is pear-shaped and that this closely approximates the shape of an excavation controlled by the above law which is cut into an inclined plane.

Knapp[4] lists factors controlling entrainment of sediment as: eroding forces: uniform boundary shear, local shear from eddies, fluid impact, particle impact, and increased buoyancy, and resisting forces as: gravity, support of adjacent particles, mechanical binding, protective surface coating, and resistance to internal or interstitial flow.

Straub[5] gives some data on silt load per square mile of drainage area of the Missouri River and some of its tributaries, as well as a comparison of the bed load and suspended load of some of them. The minimum is the James River at Scotland, S. D., 1 ton/m²; the maximum, Bad River, Pierre, S. D., with 1640 tons/m². Computing loads in parts per million of water by weight the bed load varies from a minimum of 3 in Osage River to a maximum of 2,120 in Niobrara River. Suspended load varies from 77 in Gasconade River to 21, 120 in White River. It must be noted that the contrast is between rivers whose basins are on fairly firm rocks and those which drain areas of slightly consolidated material. Rates of suspended to bed load vary from 0.26 to 1 in Niobrara River to 180.5 to 1 in Milk River.

Meandering of rivers, based on experiments by J. F. Friedkin at U. S. Waterways Experiment Station, Vicksburg:

The artificial stream used varied from 1 to 5 feet in width, 50 to 150 feet in length, 0.05 to 0.3 feet in depth, and had a discharge up to 0.15 ft^3/sec (second feet). The conclusions were (with a few additions by the writer):

Meandering is due to deflection of the current against a bank locally increasing turbulence and causing caving. Deflection is due to obstacles. No effect of the rotation of the earth was observed. The only requirement for meandering is bank erosion. Amount of material thus made available to the stream is governed by nature of the bank and the angle of attack of the current.

The stream channel is altered in an endeavor to carry off this locally-derived load and thus bring the channel into equilibrium with amount of sediment. Rate of bank erosion is not related directly to rate of downstream sand movement.

Sand is carried across the channel to next bar on the other side downstream, thus trading sediment from one bank to the other. At the bar velocity of current is at a minimum.

Rate of bank erosion is decreased by increase in slope or discharge, by increase in length of the stream, by straightening which reduces the angle of attack, and by shoaling. Thus rapid streams do not meander.

With relatively low velocity of the stream a bend begins to form, which by deflecting the current from side to side causes other bends below which are eroded into the bank. This is like a ball rolling down a trough when deflected to one side. Material thus derived forms bars which cross to the convex points of the other bank where sand is deposited. Thus a stream becomes a series of deeps and bars or <u>crossings</u>. Crossings are eroded at low water.

If the river banks are very easily eroded the channel becomes very wide and shallow thus making a braided stream. Braided streams maintain a relatively steep slope. Long straight reaches of rivers exhibit this character.

Rate of meandering depends upon the materials of the banks as does depth of the channel. Resistant material is associated with deep water and little, if any, meandering (or bank erosion). Slope is least in resistant materials. Scour of the deeper parts of the bed results from either decrease in amount of sediment or increase in velocity. Deposition is due to the converse.

The curves of the meanders are smooth in uniform material; non-uniformity of material causes irregularities.

Meanders migrate steadily downstream in uniform material and never form cutoffs through the narrow necks. Such are the result on non-uniform material slowing up part of one bend.

The radius of curvature of the meanders increases with both slope and discharge; thus only large rivers make large meanders.

Length of bends is directly proportional to discharge and slope but width of the meandering zone increases with both at less than direct proportion. Length of bends is inverse to angle of attack, but width increases at less than direct ratio; thus the radius of the bends is decreased. The angle of attack varies with velocity of flow.

Sinuosity (length of the stream compared with airline distance down valley) is directly increased by discharge but increases at less than direct ratio to slope.

Width of bends is limited by formation of <u>chutes</u> or short circuits across points (not to be confused with cutoffs).

The three variables, discharge combined with channel form, amount of moving sand, and rate of bank erosion are interrelated. Increase in slope is counteracted by increase in velocity so that a smaller channel with less hydraulic radius is required for the same discharge. Complete balance of these factors is never attained in nature. In the valley of the Mississippi below Cairo, the bank material becomes progressively finer and more resistant downstream. Thus slope decreases downstream along with rate of bank erosion. Meandering stops near New Orleans and depth of the channel increases downstream.

Rate of meandering is slowed in soft material by the wider and shallower stream channels. A natural river has a variable discharge and hence is continually changing the form of its bed.

Wood, Alan, The development of hillside slopes: Geol. Assoc., Proc. 53: 128-140, 1942

Wood applies his theory to reconciliation of the views of Davis vs Penck. Davis postulates uplift as sudden, Penck as gradual, then increasing in speed, last declining in rate. In this way Penck explained slopes as convex above and concave below.
Erosion of vertical faces. Wood explains formation of talus (scree) by upward growth on a receding rock base. He calls the exposed and later buried bed rock the FREE FACE and the talus the CONSTANT SLOPE.. Now if the talus is weathered it is washed out by rain and the finest material is carried farthest making the WANING SLOPE or what might be called a wash slope.
Erosion of valley walls. Streams tend to make valleys with vertical sides but weathering prevents this from being common. Now if downcutting is rapid weathering makes the valley walls convex. When downcutting is checked by baselevel talus slopes begin to form with waning slopes below. Talus will grow upward until all the free face is buried.
Formation of the upper convex or waxing slope. Normally the original convex valley sides will be destroyed by valley deepening. But when the valley is no longer being deepened rapidly then the tops of the walls are attacked from both sides and rounded off into a convex or WAXING slope. Rejuvination of the stream may deepen the valley and reexpose the free slope. Vegetation diminishes rate of erosion and the constant slope is preserved at original angle. But weathering may reduce angle of the talus and cause sliding. Increased rainfall might also cause slides. Constant slope may keep its angle as valley side is eroded back. Alluviation of a valley may bury the waning slope. Rise of sea level due to marine sedimentation would assist peneplaination. In application to actual conditions we must recognize a lag in adjustment.
Four independent factors: erosive activity of stream; intensity of transportation on slopes; material of valley sides; rate and products of weathering. Wood supports the idea of grade of streams. Such a stream is not in permanent equilibrium but has developed in relation to a baselevel, local or regional. Slope is that needed to carry available debris. Both constant and waning slopes are also "graded" . Stream beds are still degraded as the load becomes finer. Fits his ideas to soil erosion, gulleying, badlands. In case of pediments he says that deflation removes some of the talus and original free face wears back until entirely destroyed Water has distributed the talus material and after its removal the rock pediment may be eroded by sheet floods and deflation. Final result is lower slopes of pediments and formation of a desert peneplain. In tropical forest landsliding of the unstable soil mantle reexposes rock faces. A deep water table lowers baselevel of streams and makes for narrow razorbacked ridges with no waxing slopes. Chalk downs have no constant slope but convex-concave forms due to underground drainage. Solifluction is very efficient transportation in suitable climate. Result a very subdued topography.

Definitions of words peneplain and peneplane.

Wooster, p. 193
The peneplain, originally defined by Davis as almost a plain, -- designates the ultimate stage reached in a normal cycle of erosion. It represents a large land area that has been reduced nearly to baselevel by streams. In reality peneplains may not be "almost plains" but actual plains in the true topographic sense of the word. Some may approach the quality of a geometric plane, therefore, may be properly designated peneplanes (almost planes). The final process by which a land mass composed of rocks of varying structure and composition is reduced to a peneplain is planation brought about by the lateral erosion of streams--. As a rule, the surfaces of peneplains are not flat but gently rolling.

VonEngeln, p. 83
For an indefinitely long period is at the disposal of the normal degradational processes and agencies, namely weathering processes and streams flowing down to the sea, it is obvious that such activity will eventually bring about the reduction of the highest and broadest of uplifted regions to an ultimately lowest level.
As unchanged peneplains in situ are not available for observational study many of the characteristics of peneplains must be deductively inferred.

Lobeck, p. 634
It is admitted by most investigators that peneplanes may be formed subaerially by streams, or by marine planation, or by wind action under arid conditions. Some authorities restrict the term peneplane to surfaces developed only by stream action, but in this text it refers to an almost flat surface produced by destructive forces.

Cotton, p. 20
---the surface of very faint relief which the cycle theory requires shall eventually result from the prolonged action of normal erosion on a land surface without interruption by further uplift or other earth movements is a peneplain.

Salisbury, p. 153
It is doubtful whether any extensive land area was ever worn down to a perfect base-level; but great areas have been worn down almost to that level--- a region in this condition is called a peneplain (almost plain) (gives an illustration from Camp Douglas, Wis.)

Davis, Physical Geography, p. 152
It may be imagined that, at a very late stage of development, even the mesas and buttes of an old plateau may be worn away, the whole region being then reduced to a gently rolling lowland, a worn-down plain, or "plain of denudation" --a lowland of this kind may be called a "peneplain", because it is an "almost plain" surface.

Webster dictionary
Plain (noun) = level land or broad stretch of land having few irregularities of surface.
Plane (noun) = a surface, real or imaginary, in which if any two points

are taken, the straight line which joins them lies wholly in that surface; or a surface any section of which by a like surface is a straight line; a surface defined completely by any three points not colinear; or a surface more or less approximating a geometrical plane. (illustration, inclined plane).

Wooldridge and Morgan, p. 183
In the orthodox presentation of the cycle of erosion, the later stages are represented as largely concerned with the gradual lowering of the interfluves by atmospheric wasting. This process is regarded as continuing long after active valley deepening has ceased, so that it tends to the obliteration of the strong relief of maturity, producing in the limit, a rolling upland, on which rivers flowing with gentle gradients are separated by low swells of the surface. For such a surface W. M. Davis proposed the term "peneplain".

Johnson, D. W., Plains, planes, and peneplanes, Geogr. Rev. 1: 443-447, 1916
We must recognize (1) the perfectly plane surface of ultimate erosion and (2) the imperfect "almost plane" surface which characterizes the penultimate stages of the several erosion cycles.
(1) The level erosion surface produced in the ultimate stage of any cycle may be called a plane.
(2) The undulating erosion surface of moderate relief produced in the penultimate stage of any cycle may be called a peneplane. A low-relief region of horizontal rocks would be called a plain.

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GEOMORPHOLOGY

Supplement, 1949-50

Concavity of slopes

Introduction. Causes of convexity of hill and mountain summits have already been discussed. It is observed that most streams have a marked concave slope, although variation in discharge, load, and nature of underlying material commonly make mathematical analysis impossible. The lower slopes of fragmental volcanic cones, outwash terraces, alluvial fans, and pediments also display concavity. With those logarithmic plotting of profiles discloses that they obey definite laws. Two general groups occur: (a) most pediments, outwash plains, streams of essentially constant discharge with no tributaries within the part examined, all have an exponent, n, which lies between about 7/10 and 8/10, (b) the lower parts of volcanic cones and some pediments display a value of n which varies from 3/10 to 5/10, with rare examples of fall in proportion to logarithms of distance.

Explanations. Three possible explanations of the above facts are: (a) the slope is one of constant eroding and transporting force, that is a balance between erosion and transportation, (b) there is a difference in manner of flow of the water in different parts of its course, or (c) the slope is related to competence, that is to the largest particles available for transportation.

Slope of constant force. Little attempted to analyze slope erosion on the assumption that erosive force is proportional to V^2/D, explaining that this is the vertical gradient of kinetic energy of flowing water. However, this explanation is equivalent to saying that erosive force is proportional to the slope, for solution of the old Chezy formula (V^2:D.S.) for slope gives just this expression. Moreover, this view neglects the factor of total discharge. Although it happens that the exponents derived both by Little, and by modification of his method, agree rather well with those of many outwash plains and pediments the theory must be rejected. It assumed a discharge per unit width which increased in proportion to slope length and in the cases where there is approximate agreement it is obvious that the bulk of the discharge was acquired above the section under study. This is notably the case in front of a glacier and at the foot of mountains. Horton's analysis of slope erosion used the classic DuBoys depth-slope formula (F:D.S.) and also assumed that rainfall was gathered all the way down a slope. The reasoning here given (p.42) derives a much lower exponent for a slope of constant force that shown by any observations. It is clear that the slopes under consideration cannot be explained as those of constant force of erosion and transportation.

Change in manner of flow of water. Horton's experiments showed that in actual experiment the flow of water in thin sheets on the surface of the ground is neither laminar nor fully turbulent. The expression discharge in unit width, q = constant X depth 5/3 applies to true turbulent flow only. It should be 3 with laminar flow. In experiments the value of this exponent of depth varied from 1 to 2. Although if the bottom were truly smooth, laminar flow might occur, a little thought will show that it is incapable of important erosion and transportation. If any substantial depth is obtained turbulence must begin, a condition which greatly enhances both erosion and movement of material. It is likely that in certain spots where flow is concentrated until it attains a force considerably in excess of the resistance of the bed to erosion, there may be a transition from mixed to either fully turbulent flow or possibly from turbulent to shooting flow. Thus we may have gullies (dongas in South Africa) formed in the otherwise reasonably smooth surface of a pediment or outwash plain. It is possible that the phenomena of braiding may be in part related to changes in manner of flow of water where the sediment load is at its maximum value (so-called overloading). We may, therefore, conclude that nothing which is known proves the theory of change in manner of flow in descending a slope although such a change may explain gulleying.

Control by competence of water. Competence is defined as the ability of water to transport material measured by either diameter or weight of the largest particle. It has been shown that for particles over 1 mm, diameter, the diameter of the largest particle is proportional to the square of the water velocity and that for particles less than about 0.2 mm in diameter the ratio changes to the square root of the velocity Now with turbulent flow, V^2 is proportional to S (slope), other things being equal. With mixed flow V^2 is related to the 1.4 power of the slope. By substitution it is easy to demonstrate that the larger particle diameter is related directly to slope for turbulent flow and to the 1.4 power in the case of mixed flow. With small particles the law is different for more of the sediment is transported in suspension. By the above criteria diameter of particles should be proportioned to the $\frac{1}{4}$ power of slope with turbulent flow and to the 7/40 power of slope with mixed flow. Now it has been observed that along streams from the Black Hills average size of gravel stones is directly related to slope and it seems likely that such is generally the case.

Some have suggested that decreased slope downstream is due to "selective transportation", that is to progressive loss of the larger particles thus permitting a constant decrease in slope below that point. This could be explained by the fact that, as transporting ability of a stream varies with discharge, the larger pebbles are most apt to be left behind during the next flood, but in general the view seems hard to demonstrate.

Another suggestion is that wear of pebbles during transport decreases the need for velocity, so that slope is proportioned to size of largest material present at any given location. Sternberg proposed a law of pebble wear in which reduction in weight was assumed to be a function in which a constant is raised to a negative power in which distance is a factor. The usual form in which this is given is ratio of final weight to original weight = e^{-ah} where e is the well known constant, a is another constant depending on nature of the rock, and h is the distance. To compare diameters the constant a of the exponent is divided by 3. The above equation gives the percentage of weight lost in unit distance. It has been claimed that pebbles in the Rhine River lose 1 percent of weight per kilometer of travel. Unfortunately, neither the exact mechanics of pebble abrasion (impact and abrasion) are clear in all cases, nor is the source of a pebble readily determined. One wonders just how pebbles of local derivation were distinguished from those of distant origin! Another factor which seems to have been overlooked, is the tendency of some kinds of rock, like granite and sandstone, to pass abruptly from pebbles to much smaller particles. Weathering of pebbles during transit has also been ignored, Although the fact is recognized that pebbles are reduced in size and weight by transportation the validity of any universal mathematical law is doubtful and slope cannot be defined in terms of distance in this way. It seems likely that slopes with a very low exponent are all underlain by very fine materials like volcanic "ash" and clay.

An indirect approach to the problem of distance vs slope may be made by considering other hydraulic principles. It has already been shown that velocity of water is related not only to slope but also to depth (or hydraulic radius). In Mannings formula the latter quantity, D (or R) carries a higher exponent than does slope. In practice it is a larger number, for slope is in most instances of very small numerical value. In Horton's formula for slope wash discharge on unit width increases at a power of depth which is in general more than unity, and is much higher than that of slope. From this it may be concluded that increase in discharge of a stream, or in amount of slopewash in descending a slope, will in general need progressively less slope with increase of distance to take care of the discharge.

A somewhat more definite approach is by means of the Schoklitsch bed load formula which related the coarser material carried to total work of the stream in unit time by the familiar depth-slope relationship. As checked by experiment bed load per unit width = constant x slope to 1.25 to 2 power x excess of discharge per unit width after deducting discharge per unit width necessary for any movement to occur.

Using common notations: $G/p = C \cdot S^{(1.25 \text{ to } 2.0)} (q-q_o)$

Substituting V^2 for S for turbulent (where G = total bed load and p = wetted perimeter and C is a constant depending on nature of material) flow, using the average value of exponent of $S = (3/2)$, this becomes $G/p = C.V_m^3 (q-q_0)$ and substituting for $q = V_m.D$, $G/p = C.V_m^4 (D-D_0)$. It should be noted that all velocity formulas must have an exponent twice that using in a depth-slope expression. Now if we solve the equation using $S^{3/2}$ for slope it is evident that $S = \left(\frac{G/p}{C(q-q_o)}\right)^{2/3}$.

Now we may assume that bed load per unit width of channel (G/p) remains essentially constant below a zone near to the border of the belt of no erosion where erosion begins. Now if we assume, that unit width discharge, q, increases in direct proportion to h or horizontal distance away from the border of belt of no erosion this works out to S: $h^{-2/3}$ and to fall, f: $h^{1/3}$, since f is obtained by multiplying average slope by horizontal distance. In general streams do not increase at as high a rate measured in terms of average discharge and distance from one gauging station to the next. Reduction in rate of increase would then lessen the concavity by increasing the exponent. If, we assume a very low rate of increase, like that of streams fed from the mountains or from an ice sheet, for instance q: $h^{1/10}$ it follows that S: $h^{-1/15}$ and f: $h^{14/15}$. The last exponent, about .93, is not much higher than that observed in pediments and outwash plains. We would expect that with no increase in discharge per unit width a straight line unchanged slope would occur.

Another apporach might be along the line of capacity or limit of bed load, a feature not included in the above formula, which makes load directly proportional to discharge.

<u>Conclusion.</u> Although the foregoing discussion has failed to provide a conclusive universal answer to the problem of concavity of slope, the best solution seems to lie in the competence of water for transportation of the coarser debris conditioned by the size of material present. Save in exceptional conditions it is impossible to relate size of available material to distance downstream or downslope. Sternbergs supposed law is plainly inapplicable in most localities. It must also be realized that hydraulic principles alone tend toward concavity.

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Papers on Work of Running Water

Dury, G. H., Contribution to a general theory of meandering valleys: Am. Jour. Sci. 252: 193-224, 1954.

The paper by Dury of the University of London, England, takes up the long-disputed ideas on underfit or misfit rivers where there are two sizes of meanders, one of the valley and the other on a valley filling. One of the most important lines of the new approach was the study of the valley fills along a number of river in the south of England by means of auger borings to bed rock.

Older theories. Dury promptly discards the long-disproved hypothesis of a lessened volume by reason of headwater diversion, disappearance of glacial meltwater, diversion to underflow, erosion of the larger curves during flood stages, and influence of different formations of bed rock during downcutting. All of these are plainly inapplicable to the streams which he studied. He also briefly dismisses Wright's idea that the meanders of rock-bound streams grow much larger than those in alluvial deposits, as well as Bates' suggestion of a change of meander size due to aggradation of a valley.

Description. Meanders are carefully described. The "deeps" at the outside of bends are called swales, the shallows maintain the usual name of crossings, and the depositional features inside the curves are termed scrolls. It is admitted that meanders may be initiated on quite steep slopes and hence are not everywhere the result of a low stream velocity. The author's conclusion is that "if a straight channel becomes more sinuous, the hydraulic radius and mean velocity increase, while the wetted perimeter is reduced. Thus a deeper channel with more stable banks can be maintained and a more efficient runoff occur than with a straight course." The fact that meanders maintain themselves is thus explained despite the fact that in weak rocks the entrenched meanders which survive in firmer rocks are missing.

Formulas. Next, meanders are considered in relation to stream size. The variables considered include width, W, depth, D, wetted perimeter P, (all in feet) slope S, catchment area M, in m^2, width of meander belt M_b, and meander wave length M_l, (both in feet) velocity V, and discharge, Q (in ft3/sec). On the authority of others Dury presents several empirical formulas to relate M_b and M_l to W. M_b is from 14 to 17.38 times W and M_l is 6.06 W. The relation to discharge is $M_b = 84.7\ Q^{\frac{1}{2}}$; $P = 2.67\ Q^{\frac{1}{2}}$. W can also be approximated by the formula W = Beta $(CRM^{.6})$ where R is the annual runoff in inches, M, the catchment area is in m^2, and C the runoff coefficient. Beta varies from 0.3 to 0.375. Dury had not seen Leopold and Maddock's work on the mathematical relation of W, D, and V to Q. They found that W = constant $.Q^{.41}$. It may well be doubted that any of these equations give due weight to the nature of bottom and banks of a stream in regulating its width in relation to Q. No mention is made either of the inaccurate maps used by some of these investigators, or the fact that some were working with irrigation canals. Full data are tabulated for 6 streams and 9 localities. These include m, R. C, W. observed and computed, P observed and computed, width of filled channel W_f, its ratio to that of the present channel, Q at present from the formula $Q = (P/2.67)^2$ in ft^3/sec, and last rainfall intensity, i, necessary to fill the larger channel. The last is derived from a "rational formula" where $Q_{max} = 640$ CiM where i is rainfall in inches per hour and other quantities are given above. Hence $i = Q_{max}/640$ M. Dury concludes that an annual rainfall of 300 to 400 inches would be needed to fill the buried channels, with i equal to .20 to .33 inches per hour. The result could have been obtained if rainfall intensities, which now occur rather seldom, were once much more common. 300 inches per year could fall in 900 hours at a rate of .35 inches per hour.

<u>Discussion</u>. Dury concluded that the change in size of meanders is due primarily to a reduction in annual rainfall since the Pleistocene. He states that it is very difficult to compute radii of curviture which may account for neglect of the force directed against the outer bank due to a curved course. As all text books of physics demonstrate acceleration is proportional to V^2/r where r is the radius of curviture at the point under consideration. Since the formula is for acceleration, if force is desired the mass of water in the river which affects the outer bank must be considered. Since this is the location of greatest depth and highest turbulence, it is evident that the entire mass passing in unit time, Q, must be considered. However, it is clear that the force against the outer bank is only the lateral component of the total energy of the stream. In estimating this component rate of curvature must be considered. The formula gives this for it shows the portion of total kinetic energy which is necessary to keep the water flowing in a curved path. From the formula it is also evident that the inverse relation of force to radius is a factor which must limit the size of meanders at the point where resistance of the bank to erosion equals force directed against it. Another self-limiting factor is the obvious fact that meandering increases the length of a stream which at the same time decreases its slope. Now, other things being equal, velocity of a stream is related to square root of slope. Hence for V^2 we can substitute S and obtain the final result that force against outer bank is proportioned to slope divided by radius of the curve, multiplied by the mass of water passing unit length of bank in unit time. Although we have definitely shown that meanders themselves limit their size and that only big rivers can make big meanders we are met with an apparent contradiction. How is it that entrenched meanders which cut into bed rock are so much bigger than floodplain meanders in relatively soft material? Before we can answer this we must first consider three problems. First, what caused the deposition of the alluvial fill in a former rock-bound stream valley; second, what determines the wave length of meanders; and third, what effect does change in bank material have upon dimensions of the channel with the same discharge.

<u>Causes of valley filling.</u> Most text books ascribe the widening of a valley to lateral erosion of the stream when it has reached grade. If this were true, the thickness of the alluvial fill should be small and streams with entrenched meanders should develop wide flood plains. This condition does exist in some places; but in most parts of the world, valley filling is due to a change in level of the outlet of a stream. This change may arise (a) from change in sea level, (b) filling of an enclosed basin, (c) deposition of glacio-fluvial or other stream deposits at the outlet, or (d) lengthening of a stream by delta building . A change of climate is possible, as it also earth movement, both of which can affect the slope of a stream. In all of the cases outlined above, the slope of the stream is necessarily changed. The

Kickapoo River, Wisconsin, studied by Bates, had its outlet into Wisconsin River raised more than 150 feet by glacial outwash. Rivers of the Atlantic Coastal Plain and the British Isles all show drownding. Many rivers which flow into the Great Lakes show a similar feature due to tilting of the region which caused a rise in lake level. Thus, without any necessary change in runoff, the slope of the valley floor is changed. Streams aggrade, or degrade, their courses until a stable condition is reached in sediment transportation. Streams like the Kickapoo had a complex Pleistocene history. The aggraded to meet the ponding of the outlet by glacial outwash, then eroded when glacial meltwaters removed a part of this fill and are now aggrading apparently due to the increased supply of sediment since cultivation of the surrounding hills. Streams of the Great Plains which display terraces have undergone a complex combination of climatic changes, and tilting of the land. It is clear, therefore, that we must not ascribe all changes of stream slope to climatic change alone.

Wave length of meanders. We can consider meanders like the phenomena of a ball rolling down a flat-bottomed trough. If the ball is started straight, it may roll the entire length of the trough without ever striking the sides. The higher the velocity, the more likely this is to happen. But if the motion has a lateral component, collison is inevitable. On this happening the ball is reflected across, strikes the other side, is again reflected and so on. The angles of incidence and reflection alone affect the distance between collisions with the sides, the wave length. A stream behaves in much the same way except that it cannot be reflected as sharply. The wider the stream the harder it is to turn it. Another similie is the vibration set up in a hanging rope when struck which forms stationary waves. Hence it is easy to understand the effort of Nemenyi to liken meandering to some form of wave motion or rhythm. In nature, however, variation in material of the banks plus effect of tributaries make it difficult to determine a wave length. In the Vicksburg experiments Friedkin does not report it but used instead, length of bends from one shoal to the next and width of the bends, distance between line tangent to bends and parallel to axis of stream. The first comes closer to wave length.

Effect of bank material on channel dimensions. The quantitative results reported by Leopold and Maddock were derived from about 20 rivers in Western United States and hance do not represent all conditions. In Mississippi River it has long been noted that the finer and more compact the bank material the straighter and deeper the channel. On the other hand the Vicksburg experiments demonstrated that very soft erodible banks do not permit the formation of any meanders but result in a braided course. In Dury's examples there is everywhere a wide difference between observed channel dimensions and those computed from the formulas used. From this it is clear that none of the formulas can be relied upon except with the bank materials where they were derived. It is also evident that bank material has a profound influence upon meander size.

Vicksburg experiments. The experiments with model streams at the Vicksburg laboratory reported by Friedkin are almost the only ones with controlled conditions. Friedkin reports the following:

a) Length of bends is in direct proportion to discharge.

b) Width of bends increases at less than direct proportion with increase of discharge.

c) When slope was altered with discharge constant length of bends was almost in direct proportion to slope.

d) Under condition of (c) width of bends increases at less than direct proportion to slope.

e) The initial angle of attack, where the stream was deflected, is inverse to length of bends. This is exactly the same as with the ball rolling down an open trough.

f) Considering width of bends, the angle of attack is almost in direct proportion.

g) Turning to increase in length of the stream compared to original airline distance (sinuosity), the increase is almost in direct proportion to discharge.

h) Sinuosity increases at much less than direct proportion to slope.

i) In a meandering river shoaling or deepening takes place at any given spot depending on the relation of sand entering the area to the ability of the stream to carry such sediment.

j) The slope of a river is changed with change in level of its bed to bring about an adjustment between bank erosion and rate of sand movement.

k) Shape of the cross section of a channel is changed by the erodibility of the banks; the original form makes no difference to that fixed by flow, banks, slope, and alignment; slope is changed by cross section of channel.

l) There are three interrelated variables: discharge and channel form which regulate sand transport, amount of sand to be moved, and rate of bank erosion. No set formulas are possible. Stability involves a wide shallow stream which neither erodes its banks nor forms meanders.

m) The only reason an alluvial river does not erode it bed is the load of sand which it is carrying.

n) Although bank erosion causes the outside of a bend to be eroded back deposition builds up the inside of each curve thus reducing the channel area with sand eroded from the bend above. Width remains fairly constant.

o) What limits the size of bends is the formation of <u>chutes</u> across the points on the insides of the curves. Chutes form when a bend becomes too sharp and when the alignment upstream changes the direction of the current.

p) Variability of material on the floodplain, which is common in nature, disturbs regular growth of meanders producing dissimilar bends.

q) Meanders normally move downstream (<u>sweep</u>) and natural <u>cutoffs</u> across the neck only occur when a downstream meander is slowed up by variation of material.

r) Braided streams are often called "overloaded" and occur with steep slopes. The tests showed the primary cause is very soft bank material.

Three types of valleys can be distinguished: (1) resistant banks = deep narrow channel with low slope. (2) slowly eroding banks = meandering, fairly deep channel with fairly low slope. (3) easily eroded banks = stream with fairly steep slope and shallow meandering channel.. (4) extremely soft banks = braided stream with extremely high slope. Intermediate between (3) and (4) is a stream with any straight shallow wide reaches, islands and bars.

Applying the above to the Mississippi Valley one begins with the last alluviation probably associated with postglacial or lateglacial rise of sea level. Deposition took place until the stream was able to carry its load. Sediments decrease in particle size downstream. Subsequent development follows the above laws. A secondary reason for the downstream decrease in slope is that less velocity is required to transport the finer sediments (See Leopold and Maddock, "Dimension and competence of running water"). The easy erodibility of the sediments in the upper part of the valley is counteracted by the wide shallow bed of the river so that meandering is no more rapid than below. We must remember that natural rivers do not have constant discharge. Adjustment to bank conditions is never complete.

Entrenched meanders. With meandering valleys the problem arises as to whether the curves follow directly those acquired on a former floodplain or whether they have grown larger during downcutting. The latter are what Rich termed ingrown meanders. Leaving this question aside, it is evident that with meandering valleys we must in general have a stream bottom which is gravelly or sandy with abundant rock outcrops. Although such bends do migrate downstream more rapidly in soft than in resistant formations, it is clear that downward erosion occured so rapidly that no floodplain originated. Cutoffs and chutes could not be formed, although some cutoffs through caves are reported in the Ozarks. Very much elongated bends are common and it has long been observed that the ratio betweeen width of bends and width of channel is much higher than is the case on floodplains. The gravelly nature of much of the bottom means less easy erosion so that it may be presumed that the cross section of the streams during erosion was on the whole much deeper in proportion to width than is the case in soft sandy alluvium. It is also safe to assume that the slope during erosion was considerably more than on floodplains since more and coarser load was being transported.

Floodplain meanders. Floodplain meanders after the filling of a meandering valley with relatively soft alluvial deposits presented an entirely different problem to the stream. The slope was also decreased over much of its length. Following the laws discovered, above the bends should then be smaller and the stream channel wider and shallower, the latter counteracting to some extent the softer material of the banks. Shortening by chutes and cutoffs could occur readily. Shallow wide reaches should be more abundant.

Conclusion. With the above listed changes in controls other than a decrease in average discharge it is evident that Dury's conclusion of a climatic change should not be regarded as established beyond doubt.

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In a paper by Frey and Leonard, some of the practical problems of terrace correlation are discussed. Errors include mistaking a rock terrace for a depostional feature, mistaking a colluvial wash deposit from the valley wall for a terrace of material brought down the stream, concealment of a high terrace by loess, and confusion of terraces with flanking pediments. The effect of a resistant bedrock formation on the grade, both of stream and terraces, is also pointed out. Dissection of old terraces by stream valleys makes it difficult to discriminate between post-terrace erosion and original surface irregularities due to stream work. Miscorrelation of terraces along the stream valley is made possible by these chances of error.

Gilbert, G. K., The transportation of debris by running water: U. S. Geol. Survey Prof. Paper 86, 1914.

Introduction. Gilbert commenced his studies of movement of material by running water in a purely qualitative manner. When he began a study of the movement of the debris from hydraulic mining in California mines the need for quantitative knowledge became apparent and a study was begun at the University of California in Berkeley. The experiments, which almost all used straight wooden troughs, failled to discover a simple law but nevertheless are the most elaborate ever carried out. When sand is added to water flowing in a trough, it builds up the bottom until the slope is adjusted to that needed to transport the load. If the rate of feeding is not above a certain limit the bottom is stable, but if it exceeds that limit of capacity then the bottom is built up by the debris which cannot be carried forward. In experiments the several variables, slope, ratio of depth to width, discharge, nature of debris, and depth can be kept constant and hence the effects separated. These variables are not independent. We must recall two fundamental equations: discharge, Q = width times depth times velocity V, and Manning's formula where R = hydraulic radius and S = slope, $V = \text{constant}\ R^{2/3} S^{\frac{1}{2}}$, where the constant involves the nature of the bottom. Furthermore we should remember that a natural stream can adjust the form of its cross section to the discharge and debris in transport whereas the trough is fixed unless the sides are changed purposely. Flume transport is different in that no debris was allowed to accumulate on the bottom.

Natural streams. In natural streams those which are supplied with debris to less than their capacity erode their beds and bedrock is exposed in places and at certain times, these are corrading streams. When the supply of debris equals or exceeds capacity the stream bed is wholly composed of this debris although there may be some rock banks. These are rock-walled streams. Streams which have so much available debris that the entire bed is composed of loose material are termed alluvial. A given stream may have segments of its course in different catagories. Streams adjust their beds to meet the condition imposed by the local supply of debris. This process is termed gradation. Most alluvial streams are aggrading and have a flood plain. Gilbert regarded meandering streams as confined to low slopes which are underlain by fine material. Most river channels curve and on a bend the deepest and swiftest water is on the outside instead of in the middle of the channel. This develops an asymmetric cross section where the outside of the bend is eroding and the inside being built up thus preserving a nearly constant width. Shoals, called crossings, are built up diagonally across the channel from the inside of one bend to the inside of the next. These bars are built up at highwater and eroded down at low stages when the deep places are filled up. It is flood stages which perform a large part of modification of the channel to fit the river. With change in stage go changes in velocity and hence in the limits of competence. This is why the shoals come to be surfaced by large particles which cannot be moved at the lower stages. When the channel of a river is fully adjusted to discharge the same load will be transported in every section but the relative amounts in suspension and bed load may change with local conditions. On the whole the ratio of mean depth to width is much less in natural channels than in the optimum ratio for capacity found in the experiments. A complicating feature of natural streams is the nature of the debris supplied to them in reference to their competence. This load may bear no relation to capacity. Suspended load influences velocity in three ways: (1) its mass increases the mass of the stream and hence its energy, (2) suspended particles are always settling toward the bottom and work is required to keep them from doing this, (3) the load affects viscosity. An empirical formula derived from the experiments is that Mean velocity, $V_m = Q^{.25} S^{.3}$ times a constant, from which it concluded that addition of a suspended load increases velocity slowly with slope and inversely to the discharge. From quantitative comparison of the work needed to keep material off the bottom with its addition of energy of the stream it was concluded that the former is greater and hence the velocity of the stream is retarded by suspended load. Increased viscosity also retards the flow of the stream. Retardation by viscosity may reach 15%. Gilbert had difficulty in finding any retardation of velocity due to bed load. (tractional load). It is possible that such is related to the load, slope and velocity. He did find that a load influences the vertical distribution of velocity in a stream. Gilbert concluded that there is an automatic separation of suspended and traction loads. Were the Mississippi deprived of its bed load, he thought it would shoal the channel and reduce its slope until part of the load would be carried on the bottom. On the other hand he concluded that removal of the suspended load would increase velocity and lift some of the material now carried on the bottom. Checks are difficult because of the lack of measurements of bed load.

Application of laboratory results. Slope does not enter into computations at a given locality but applies to streams over a longer distance. In nature there is more variation than in the laboratory. Discharge must be measured to represent equal phases of stream work. The problem is complicated by simultaneous changes in nature of material carried, as well as by changes in velocity and competence. During low stages traction is confined to the tops of the crossings or bars. Turning to the ratio of depth to width, this is related to the resistance of the banks. In general bank resistance of a natural stream should be greater than that of the smooth wall of the laboratory trough. In general the difficulty in extending laboratory formulas to real streams is that they are empirical and not rational. Models do not have the same relations between dimensions that are present in the originals. A glaring example is the size of particles used for transportation. Gilbert simplified his experiments by using several size ranges of sand and fine gravel which were designated by letters. Some tests involved mixtures. Of his sizes the three smallest, .304 mm, .374 mm, and .508 mm average diameter are well below the point where competence of the current changes from the law that linear dimensions of particles are related to the square of velocity. It is not clear that this was recognized by Gilbert. For the sake of simplicity it was tried to set up equations for capacity which are power functions of some one of the variables. In each case a threshold value of that variable at which sand movement first is noticed must be subtracted before applying the exponent. On account of the change in competence with size of grains it is evident that this procedure would have to be changed at the critical point of about 1 mm diameter. When slope alone was used in the above manner, the exponent varied, from .93 to 2.37 and was found to be an inverse function of both discharge and coarseness of debris. When discharge was used the exponent was from .81 to 1.24 suggesting a nearly direct ratio, although the values are an inverse function of slope and size of debris. With fineness, the exponent varied from .5 to .62 suggesting a square root relationship. Values were inverse functions of slope and discharge. Capacity may be made to reach 0 either by making the stream very wide and shallow or very deep and narrow. The optimum ratio of depth to width varied from .5 to .04, inverse to slope, discharge, and fineness. Velocity, which many have thought of as the sole variable, could only be measured as mean velocity. With slope constant the average exponent was 3.2. With constant discharge it was 4.0, and with constant depth, 3.7, seemingly an inverse function of slope, discharge, and fineness. When a mixture of sizes was used the movement was more free. With addition of fine particles to coarse, the movement of the coarse was increased. In the case of changes in depth, results varied when other factors were held constant. With constant discharge velocity increases so that capacity is inverse to depth; However, when slope is held constant depth is related to discharge and capacity varies with depth. Depth is related to the .62 power of discharge. If velocity is held constant both direct and inverse relations were found. The average, considering sign, was -.54 suggesting an inverse square root relationship, but it is evident that depth is a dependent variable and cannot be used alone in a formula. The form ratio or ratio of depth to width has two zeros of capacity, one for a very high value, the other for a very low value.

Flume transportation. The conditions of a flume with no debris left on the bottom may occur in segments of natural streams. This condition leads to an increase of capacity because rolling is more important than jumping. With such motion, capacity is largest for coarse particles; whereas with leaping particles, the reverse is true. Capacity is reduced by roughness of the bottom.

<u>Criticism</u>. It seems clear from the above that any rational formula for capacity (1) must include several variables, possibly all of them, (2) must consider the change in competence with size of particles, (3) should include the relationship to both shearing force on the bed and to degree of turbulence resulting from that, (4) must include the form of the bed, rippled, smooth, or antidunes, all of which occur in succession with velocity increase. Some of these things have been discussed in other supplements and hence will not be here repeated. See especially "concavity of slopes" and "Dimensions and competence of running water."

Little, J. M., Erosional topography and erosion, a mathematical treatment, A. Carlisle and Company, San Francisco, 1940.

Little's book of 1940 appears to be one of the first, if not the first, attempts to find the mechanical features of erosion and the resulting topography. The primary approach of the author was to tie in erosional geomorphology with hydraulics and hydrology. Second to this, he desired to obtain an "erosional rating" for given slopes, soil and cover of vegetation which would be a basis for classification of lands for human use. Both flow in sheets and in channels was considered. He fully recognized the complexity of the problem and stated that some conclusions would have to wait for, or be modified by, the collection of more data. Fundamental assumptions included basing "erosive power of flow" on "some velocity to depth relationship that is exponential" (A power function as here defined).

<u>Types of flow and energy of flow</u>. Little dismisses laminar flow as having no erosive power and little or no coarse silt transporting power. He considers only turbulent flow and shooting (plunging) flow, which occurs when turbulence is excessive. Little concluded that "erosive power of turbulent flow is a function of velocity and depth and not of velocity alone," of which the exact nature under different conditions is undknon. He assumed that the relationship is V^2/D where the side walls of channel do not interfere. The exponent 2 was considered tentative. V = velocity and D = depth. He realized that any attempt to relate erosion to total force exerted on the stream bed in direction of flow is useless because turbulent flow is needed to raise material off the bed. In this he used the Schmidt computation of intensity of turbulence which related it to the total potential energy of a column (or prism) of water divided by the rate of change of velocity at the base. Since the rate of velocity change with depth is at a maximum near the bed of a stream, it is there that turbulence is greatest. "Since turbulence is proportional to kinetic energy of flow, $V^2/2g$, its intensity varies as V^2." Note that this expression is for kinetic energy of unit weight of water in British Engineering Units where mass is obtained by dividing by g, the acceleration of gravity. "The influence of D on turbulence in proximity to the bed is inverse." From this, the fact the relationship of erosive power, E, to V^2/D was deduced. As a check it was noted that this corresponds to loss of head in pipes but no mention was made that this is the same as slope, s, in an open channel. The expression was to apply to scouring and silt transportation by both suspension and bed movement. The equation $V^2/gD = 1$ was then set up with the value of unity expressing erosive power at critical flow, the passage from ordinary turbulent flow to plunging flow. It was concluded that in "the interaction of two materials, liquid and solid,

V^2/gD is a measure of the intensity-distribution of the internal forces which ultimately destroy turbulent flow--without regard to what actual values are assigned to V and D. Neglect of total pressure on the bed is accounted for because "loss of head and internal friction (related to turbulence) are independent of pressure, for the reason that viscosity and density of water are independent of pressure."

Turbulent flow in rectangular channels. In working out the laws of erosion in rectangular channels with water sides, i. e. vertical sections of a wide stream, substitution of Q (quantity in c.f.s.) = DV yields the conclusion that $D = Q^{2/3}/3.18$ and $V = 3.18\ Q^{1/3}$. Mannings formula for velocity is also substituted with some interesting results in solving, by various simple algebraic transformations, for the several quantities involved. Assuming that E (erosive power) = 1 and the roughness coefficient, n = .04, then $S = .03427\ Q^{-2/9}$ which may be compared with the empirical conclusion of Leopold, $S = .021\ Q^{-.49}$. Also, $Q = V^3/gE$.

Turbulent flow in trapezoidal channels. For channels with a flat bottom and sloping sides like many irrigation ditches, the development of formulas for critical flow when E = 1 simply modify results by the ratio of cross sectional area to area of a rectangle. Little then produces the result that for n =.04, $S = .4364\ Q^{-2/15}$ which departs widely from the observed result of Leopold.

Effect of roughness coefficient, n. By taking Mannings formula and substituting Q/D for V and solving for n, it is possible to give a formula by which n can be found experimentally by sprinkling a plot of land with fixed slope. It also appeared that E varies inversely as the 9/5 power of n.

Sheet runoff. It is with sheet runoff that the major relationship to geomorphology was found. Little suggested that the passage from sheet erosion to rill formation is a "point of breakdown or failure--analogous to the failure of any material in the testing laboratory." This he related to the attainment of unity as the value of E. Since he was concerned primarily with land use practices it was then necessary to introduce a "rainfall equation" to give quantity of rainfall. He chose R (rate in inches per hour)=8 T (duration in minutes)$^{-.25}$ A coefficient for relation of runoff to rainfall is needed. Q (runoff) = C (fixed coefficient) x area, A, x rainfall rate, R. Area is a direct proportion to horizontal distance from summit, h. C was placed at .7. By algebraic transformations using formulas given previously, Little arrived at $V = .3662\ h^{2/5} S^{3/10}\ t^{-.1}$ where t is in seconds. To relate observed concave slopes of hills to such erosion, Little concluded that purely convex profiles are typical of young topography which is the result of channel and not sheet erosion. Sheet erosion, in conjunction with removal of material at the bottoms of the slopes by streams, might result in uniform slopes. In older topography he observed that the common relationship is the compound reverse curve, convex above and concave below. Such regular curves develop only on homogeneous material and in nature there are always irregularities along any slope which can produce gulley erosion. He concluded that the work of man in cultivating the soil has increased this tendency and reduced the areas of uniform sheet wash.

Relation to coordinates of slope. In order to relate erosional power to coordinates of a slope, it was necessary to eliminate t and to give other quantities in terms of horizontal distance from divide, h, and to fall, f. Now to eliminate t, it was necessary to find two values for $t^{1/4}$ and then equate them. The first value was derived by equating two expressions for Q. $Q = V^3/gE$ is put as equal to $Q = Ch\ t^{-\frac{1}{4}}$ or $= .00035847h\ t^{-\frac{1}{4}}$ from 1 square foot, and this is solved for $t^{1/4}$ which equals 1404.5 $S^{9/2}$ h E^{-5}. Further algebraic transformations give V = .3662 $h^{2/5}$ $S^{3/10}$ $t^{-1/10}$. By setting V = dh/dt and multiplying both sides by dt/dh it appears that with S constant, $.3662\ S^{3/10} \int dt \cdot t^{1/10} = \int dh \cdot h^{-2/5}$. Integrating this and solving for $t^{\frac{1}{4}} = 1.479 h^{1/6} S^{-1/12}$. Now the two values of $t^{\frac{1}{4}}$ are equated eliminating that term. Solving for S, substituting df/dh for S, and multiplying both sides by dh yields df = .2229 $E^{12/11}$ $h^{-2/11}$. Integrating, it is clear that f = .2724 $E^{12/11}$ $h^{9/11}$. This equation represents fall in feet for horizontal distances in feet, a concave slope. H is supposed to be constant so does not enter into the result. Tables were presented for different values of E, some of them above unity. The basic idea is that slope wash forms the slope until the condition of constant E obtains. C is taken at .7 and n as .04, but the tables also show conditions for other values of these qualities.

Conclusions. Considerable discussion was devoted to the problem of the proper rainfall equation but none seems to be applied to an average over geologic time. No attention was given to the problem of erodibility of different sizes of particles or of mixtures of different sizes. Relation of fineness of particles to age of soil was also considered. Little stressed the idea that hill slopes are formed by sheet erosion only in the later stages of the erosion cycle starting by "gouging" at the bottoms of slopes next to streams. Old ridges should then be narrow and flanked with concave slopes. Little concluded that convex slopes must have a value of E which increases away from the divide whereas concave slopes have a constant E rating. When undisturbed by man a balance between erosion and soil resistance was approached but never quite attained. He desired to obtain E ratings of soils on different slopes by experiments with sprinkled plots rather than by physical tests of the soil. By the development of the concave slopes of fixed E rating the line of division between convex summits and concave lower slopes progresses uphill. It was recognized, however, that some convex divides survive in quite old topography. It was suggested that vegetation which retained rainfall on divides plus laminar flow there might account for this suggestion of Horton's "belt of no erosion." H, total horizontal distance from divide to stream channel, must remain constant during development of erosional topography; whereas F, the total fall, would decrease. A ratio of F to H might express maturity of development. Mass movement due to weight of water, swelling of soil and frost aiding gravity was recognized but not regarded as important. It was stated that "a prominent feature of top soil occurrence is its continuity and its proneness to maintain a uniform thickness on profiles." It was concluded that soil formation follows upon the development of slopes and is not important in forming them. The final conclusion "obviously implies that erosion has been, in general, the dominant process in geologic denundation."

Einstein, H. A., The bed-load function for sediment transportation in open channel flows, U. S. Dept. of Agriculture, Soil Conservation service, Tech. Bull. 1026, 1950.

Under the above title an attempt was made to reexamine an old problem, namely the rate of transportation of the bed load in streams. This problem is not only of scientific interest but is of great practical importance for at present it is difficult to predict just what changes in the bed of a stream will take place when one of the variables is altered by man. Such change upsets the equilibrium of nature which adjusted the size, shape, and slope to the amount of and variation of discharge. Einstein's solution is evidently intended to be a "new look" and he admits that it is not final. Although the title does not so indicate both suspended and bed loads are considered. He admits that no positive answer can now be given as to "what bed composition can be expected from a known sediment load in a known flow."

The general approach is highly mathematical which makes for very slow reading. However, the main points are explained in the text except where they were covered in previous publications. There are two pages of symbols and abbreviations.

From one of the earlier papers it is stated that velocities in the downstream direction vary with the logarithm of distance from the bed or the top of an inferred layer of laminar flow next to the bed. Hence a factor based on roughness of the bed in introduced. The importance of ripples on roughness is considered. Turbulence is considered under the idea of three components in the three primary directions. When Einstein states that all of these have a 0 time average, it is difficult to see how there could be any net downstream velocity! He states that velocity is variable and that a graph at a given level would show a very irregular line, although it would not reach 0 at any time.

Suspension. The primary idea of suspension as support of particles by water motion above their settling velocities is conventional. However, the discussion of distribution of concentration of solids in reference to depth is most unconvincing. His formulas are complex and involve the integral calculus and the fact seems to be ignored that turbulence must distribute suspended load fairly well. A hypothetical example is worked out which process takes three pages. Einstein's result is a tenth of what it should be because a decimal point was misplaced in computing the dry weight of sediment in a cubic foot of water, also his result is described as "per second foot," when it should have been per foot of stream width. The corrected answer is 3.29 pounds per foot width. However, if we compute by Mannings formula the velocity of his hypothetical stream as 3.5 feet per second, take the dry weight per foot as .0642 pounds, and multiply this by velocity, and by depth, 15 feet, to obtain total sediment, the result is 3.27 pounds per foot width. It is then obvious that we have only another example of a "hard way to do an easy thing."

Bed load. Einstein then turns to the particles which slide, roll or hop along the bed of a stream. He works out his theory on the basis of the probability of movement of particles of a given size with a bed load equation to show equilibrium between particles in motion and those at rest. He then evolves the ratio between lifting force and weight of particles. This involves a complex formula relating the submerged weight of a given particle size to the hydraulic radius, slope and square of the velocity at the bed. From this evolves a dimensionless figure for the intensity of transport of this given grain size. From this an actual example was worked out in 44 steps including an estimate of how the different sized particles of the real load affect the theory. Last, this bed load is combined with the suspended load to show the total sediment discharge of the stream. He does not tell how to check this result in the field!

Fisk, H. N., Geological Investigation of the Atchafalaya Basin and the problem of Mississippi River diversion, U. S. Army, Corps of Engineers, Mississippi River Commission, Waterways Experiment Station, 1952.

Introduction. It is a well-known fact that many rivers change their courses where they flow on an alluvial plain above a delta. Among the better known examples may be mentioned the Rhone, Po, Ganges, Yellow (of China) and Colorado. Although some of these events were quite well investigated and dated, in no case was the wealth of detail available that Fisk had for the study of the Atchafalaya distributary of the Mississippi. Thousands of logs of borings, many made especially for the study, detailed maps, air photographs, and measurements of stream discharge and sediment load were all provided. Although possibly the example may not be typical of streams which carry a more heavy sediment load than does the Mississippi it is thought that valuable lessons may be learned from it. The study was undertaken to ascertain if there is serious danger that the Atchfalaya will divert the Mississippi, leaving New Orleans without a major river.

History of the river. Fisk briefly reviews the recent history of the Mississippi River as worked out by him in his 1944 report. He notes the effect of eustatic change in sea level so that he discriminates the sediments of the last filling from those previously laid down, eroded and weathered when the Gulf was lower than it now is. These recent sediments are mainly of postglacial age. Fisk devised them into the sandy substratum which is overlain by a much finer-grained topstratum. River scour in many places extends through the topstratum undermining the more coherent material. The top stratum may be divided into: (a) deposits of natural levees, silt and fine sand; (b) point bar deposits on the insides of meander loops which are made of sand; (c) channel fillings in abandoned courses which are dominantly tight clay; (d) backswamp deposits also fine clay; and (e) deltaic deposits which are in part lacustrine and in part those of brackish water. Material derived from the Red River may be distinguished by its color from the deposits of the Mississippi. The different types of deltaic deposits are illustrated by mechanical analyses. No evidence of recent earth movement other than that due to compaction could be demonstrated.

Past changes in course., Several distinct courses of the Mississippi River, all occupied since the filling was essentially completed are described. The Teche-Mississippi was far to the west of the present course from a point well above Vicksburg and formed a delta slightly west of south of New Orleans. Traces of this former meandering course now form Teche Bayou. Next the LaFourch-Mississippi was diverted near the course of Red River to essentially the present river course as far south as Donaldsonville. Previously this was the course of the Yazoo. Here the river flowed more to the south forming its delta in essentially the same location as the Teche delta. The date of this change Fisk estimates at 300 to 400 A. D. Next followed another diversion near Vicksburg into the course of the Yazoo which joined the earlier phase of LaFourche Mississippi. This second diversion of the LaFourche route is thought to have occurred about 1000 to 1100 A. D. After this, the modern Mississippi route was formed by a diversion near Donaldsonville in 1100 to 1200 A. D. At first this route was more to the north than the present course below New Orleans and formed the St. Bernard delta well north of the present one. The present route below New Orleans Fisk concludes was formed about 1500 to 1600 A. D., probably not long before white men first saw the river. DeSoto discovered the river near Natchez in 1541 but the mouth was not used until later. It is now forming a delta closer to deep water than were any of the others.

Sediment load of Mississippi River. Fisk shows that the lower Mississippi is now anywhere nearly as highly loaded with sediment as many other rivers. At a discharge of 1,065,000 ft^3/sec. the Mississippi carries only about 871 parts per million of sediment. The Yellow River of China carries on the average about 50,000 p.p.m. at a discharge only a fifth that of the Mississippi, although this load may be quadrupled at times. The Colorado River carries 6,000 p.p.m. Sediment concentration is not wholly due to the hydraulic characteristics of the channel but mainly to the contributions of tributaries.

Cause of diversion. Diversion of a stream naturally occurs in order to secure a more favorable slope to the sea, which an more readily transport the load. In very heavily sediment-laden streams the bed may be built up so high that a break-through of a natural levee forms a permanent channel. Then diversion is sudden and the change of slope will in a short time affect the original channel above the point of diversion. Below this point the old channel is rapidly blocked by alluvial deposits. In the case of the Mississippi such break-throughs or crevasses have often been observed. The original channel is not high enough above the backswamp, which is mostly densely timbered, to make a permanent channel. Instead sand and silt are deposited in a braided channel which is abandoned when the flood subsides. Despite its nearness, the Mississippi has never succeeded in breaking through to Lake Ponchatrain above New Orleans. It is evident that here we must have a different process. Long ago the meandering Mississippi intersected the course of Red River. The northern arm of this loop was soon plugged with alluvial debris, but the south one was open. The Atchafaylaya was blocked with a "raft" of driftwood when whitemen first navigated the river. Absence of Indian mounds on the Atchafaylaya suggests it was not an important route prior to this. The southern arm is called Old River. The interconnection was both freed of driftwood and dredged by white men. Instead of being an outlet of the Red, together with its southward continuation of the Atchafaylaya, it is now a distributary. The diverted waters are steadily increasing in percentage of total Mississippi flow. Extrapolation of the curve indicates that by 1971 about 40% of the annual flow will go through the Atchafaylaya. Then there is grave danger of blocking of the course below with sediment leaving New Orleans without any river. The Atchafaylaya channel has also been much widened and deepened by the increased flow. The process has been probably accelerated by the building of artificial levees along the higher banks of the Atchafaylaya, but retarded by the building of delta in Grand Lakes farther down the course. Obviously some artificial control of the flow is urgently needed. to prevent the consumation of this type of gradual diversion.

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Notes on waves, reefs, etc. Supplement, 1952-53, part 4

Energy of a wave. Some features of wave work demand greater attention than in original manuscript. One of these is how the formulas for the energy of waves were derived. It is evident that wave motion involves both potential energy due to the position of the displaced water particles up or down from position of rest and kinetic energy due to the velocity of their motion, that is the work stored in displacing them. On general mechanical principles we would expect these to be equal. It may be urged, however, that the motion is not confined to the surface circles of rotation but dies out gradually below. Both derivations involve the integral calculus and are not "required." In each two separate steps are involved, first integration of the normal functions and, second, integration of the trigometric quantities which are brought in by roation. We will here perform the steps separately and then obtain the final result by multiplying the results.

Definitions: w = unit weight of water (dynes/cm^3 or lbs/ft^3); L = wave length, crest to crest; h = wave height, trough to crest; x = horizontal displacement of a particle; y = displacement of points on surface above still water level; z = vertical displacement of a particle; g = acceleration of gravity; which must be taken into account whichever system of units is used. Energy will be given for unit length along crest.

Potential energy due to displacement of surface of wave to position y = h/2 cos 2pix/L is E_p

$$E_p = w\int_{x=0}^{x=L}\int_{z=0}^{z=-y} z\,dz\,dx \quad = w/2 \int y^2\,dx$$

using above value for y this becomes: $wh^2/8 \int_{x=0}^{x=L} \cos^2 2pix/L\,dx$ which is $L/2 + \frac{1}{4}\sin 4pix/L$ Second term is 0 at both limits and multiplying by L/2 we get $E_p = wh^2L/16$

Kinetic energy, E_k, follows the formula of $\frac{1}{2}mV^2$. The double integral over a wave length L and wave height h is required. Now velocity in a circle of rotation is same, both vertical and horizontal and = 2 pi z/T the square of which is 4 pi^2 z^2/ T^2 Now T^2= 4 pi^2 z / g by laws of harmonic motion. hence by substitution this simplifies to z g.

Now the E_k for unit of length = $w/2g \int_{z=-y}^{z=h} g\,z\,dz$ g will cancel out and we have $E_k = w/4 \cdot z^2$ But $z^2 = h^2/4$ hence E_k (unit length) = $wh^2/16$

For entire length of wave, L we must integrate the horizontal and vertical components squared. This is the sum of $\sin^2 2\,pi x/L + \cos^2 2\,pi\,x/L$

Making the integration this becomes $x/2 + \frac{1}{4}\cos 4\,pi\,x/L + x/2 - \frac{1}{4}\cos 4\,pi\,x/L$ Sbustitution the limits of o and L, the trigonometic terms cancel and the final result is L. Multiplying result obtained above the entire result becomes $E_k = wh^2L/16$ confirming our assumption that it is equal to E_p.

Combining the two forms of energy total energy of a wave per unit of crest is $E = wH^2L / 8$ Note this is in terms of work and not of power.

Relation of wave height to time. Most discussions of waves do not consider length of time that the wind has been blowing. In regions of steady winds, such as the trades, this element is very important. It is obvious that when there are no waves, that is when the wind first starts to blow, the transfer of energy from wind to water will be at its maximum. As waves begin to roll with the wind this transfer must of necessity decrease. The diagrams given out in 1950 show the relations. Wave heights also decrease with increase of fetch for the same reason. Theories involve complex mathematical formulas, the practical importance of which is problematical.

Fundamental formulas are: $V = L/T$ or $V^2 = gL/2pi$ where V = velocity of travel, $L = 2 pi V^2/g$ or $gT^2/ 2pi$, $T^2 = 2 pi L/g$ or $2 pi V/g$ where T is time of period. Solving these to get results in seconds, feet, and land miles:

$V = 3.5 T$ or $V^2 = 2.23 L$, $L = .555 V^2$ or $5.12 T^2$ and $T^2 = .195 L$ or $T = .3$[illegible] V

Observation shown that wave height ratio to length is always less than 1/7. The following relationships appear well established. For a fetch of 11 miles or more maximum height is about 1.65 times the square root of the fetch in land miles. For a given wind speed wave speed increases with fetch. Maximum wave height is about .9 of the wind speed in land miles, or h = 0.0344 times the square of speed in land miles. Average maximum wave speed slightly exceeds wind speed up to about 29 m.p.h. wind speed and is less than wind speed above. Time required to develop maximum height increases with wind speed. High waves can be formed by strong winds in less than 12 hours. For a given fetch and wind speed wave speed increases rapidly with time. There is no well-established relation between wind speed and wave steepness, for the latter depends upon stage of development of the waves. During early stages of wave development waves are short and travel at less speed than the wind. Height of swell (old smooth waves with no wind) decreases as swell advances.

Roughly, waves lose a third of their height each time they travel a distance in miles equal to their height in feet. Period of swell may increase with distance of advance although this is not proved. When the speed of waves is equal to that of the wind, no energy is transferred.

A factor not considered by the students of waves is that they increase the roughness of the surface and hence raise the height of 0 velocity of wind as found by Bagnold. Although it is possible to treat many of the phenomena of waves by mathematics, it is well to remember that in practice variations in wind speed and direction introduce great irregularities.

Reef phenomena. In the last few years it has been discovered that some ancient buried reefs are very productive of petroleum and hence more attention has been given to recent reffs in order to understand them. Important papers are those by Cloud and Ladd. The fact has been brought out that reefs are mainly composed of clastic particles. Some desire the substitute term bioherm to be restricted to organic accumulations of doubtful form. The rigid framework of a reef may be both corals and algae and only make up a small part of the entire accumulation. On the outside of the reef growth is most rapid since the supply of food is largest. It took nearly 200,000 soundings to map the lagoon at Eniwetok for there are many terraces, depressions, and knobs of living coral. Igneous foundations of reefs have been reached by drilling at Eniwetok and Bermuda. At Eniwetok basalt was found in one hole at 4170 feet depth. There is only a few hundred feet of Pleistocene reef underlain by Tertiary limestone, dolomite, carbonaceous clay, and silt. The oldest sediment is Eocene. On Bermuda seismic work demonstrates that the boring is on the flank of a volcano. It disclosed Pleistocne limestone to depth 380 feet, Miocene 380 to 590, and Eocene 590 to 695, the top of the volcanics. The average depth to igneous rock is about 250 feet probably because of wave erosion with lower sea level during glaciation.

Ladd, Tracey and Lill gave a preliminary report on the boring on Bikini which failed to reach basement at 2556 feet. This showed Pleistocene reef to about 425 feet and Tertiary sediments below. The rock below about 1790 is Miocene. These borings all show shallow water calcareous sediments but demonstrate that subsidence has been going on throughout all of Tertiary and Quaternary time. It cannot be due wholly to glacial control of water level although the Pleistocne reef must have been affected by the process. Either the floor of the southwest Pacific has sunk, or the amount of water in the oceans has increased. Platforms on which reefs originated may be erosional, depositional, or local uplifts of the ocean floor. Ladd thinks that no reefs located on the rims of submerged volcanos have been discovered. Cloud uses the term <u>table reef</u> for small reefs without a true lagoon.

<u>Beach features</u>. Shepard has presented some new terms for beach features as shown below. He desires to restrict the term <u>bar</u> to submerged accumulations only.

Barrier beach = single elongate sand ridge parallel to mainland and separated by a lagoon.

Barrier island = multiple ridges together with dunes.

Barrier spit = a barrier tied to mainland at one end only.

Bay barrier = former bay bar extending across an inlet.

Barrier chain = series of barrier islands.

Longshore bar = submerged sand ridge or "low and ball".ridge, or subaqueous ridge.

Transverse bar = sand bar at right angles to shore line.

Reticulated bars = criss-cross pattern of bars inside barrier islands and in bays seen from air only.

Sandkey or sandcay = small island not parallet to shore.

Cuspate features = points 30 to 200 feet apart are caused by wave work. Larger ones have a ridge extending out to sea below water. These would include the famous cuspate capes of the Atlantic coast. Similar features occur inside lagoons on inside of barriers. Most of these have associated shoals or are opposite a cusp on the mainland with or without a connecting shoal.

Cuspate foreland = larger cuspate capes.

Cuspate bars and sandkeys = below-water features.

Lunate bars and sandkeys = crescentic bars off passes with tidal current channels in center or at ends.

Longshore bars (subaqueous bars). The Beach Erosion Board of the U. S. Army Engineers offers some ideas on Longshore bars and longshore troughs (subaqueous bars or "low and ball"). The theory of origin by Evans is supported. Repeated soundings along piers has shown that the positions and depths of both bars and troughs vary with intensity of wave work. The bars form where the waves break. After plunging on the bar, where observation by the writer showed water with considerable sand, the wave reforms and breaks again when a certain depth is reached. After several such breakings at each of which bars are formed, the wave reaches the beach. Troughs and bars become progressively smaller and shallower in approaching the shore, but their size and depth changes with height of waves. The ratio of depth of trough to depth on bar varies from 1.3 when mean sea level is used to 1.5 when "mean lower low water" is taken as datum, but otherwise no generally applicable relations were found. Although it is known that material is brought from both sides to build up a bar and that this building is due primarily to plunging breakers, the presence of longshore currents in the troughs is proved. These carry sand to breaks through which rip currents escape, spreading the sand on the sea bottom outside. On coasts where the depth increases rapidly offshore, waves break only on the beach and no ridges and troughs are formed. Keulegan and Krumbein showed mathematically that some seas could be so shallow that waves could not break anywhere and cite geological evidence of such conditions. Depth would increase at the 4/7 power of distance from shore.

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Definitions of technical terms, running water. Based upon usage by Lobeck, Vonengeln, Cotton, and others.

Abstraction= process of valley destruction due to development of larger valleys.
Accordant= applied to a stream which joins another without falls or rapids at junction, following Playfairs Law.
Accordant= applied to summit level of divides in hills or mountains
Aggradation= process of filling of valleys with water-deposited sediments
Alluvial dam= blocking of valley by alluvial fan of tributary
Alluvial fan= deposit made by a stream on land; due to change in capacity; form a segment of a cone.
Alluvial terrace= erosion remnant of valley filling.
Alluvium= deposit of a stream on land other than fan form, generally making a floodplain.
Air gap= wind gap; gap through a ridge not occupied by a stream.
Anaclinal stream= stream which flows in direction opposite to dip of rocks.
Anaclinal slope= slope opposite to direction of dip of rocks.
Anastomosing= applied to streams which branch and reunite in a braided pattern.
Anticonsequent= antecedent stream, which has survived earth movement without changing its course.
Antecedent drainage or stream= drainage which survived earth movement without change.
Anticlinal mountain= one controlled by an anticline of resistant rock but is not necessarily due to recent folding.
Anticlinorium= compund or complex anticlinal uplift.
Arroyo= valley or gulley of intermittent stream, generally in desert climate.
Arcuate delta= delta like that of Nile with many distributaries.
Autogenetic falls= falls which is working backward by erosion of plungepool.

Backslope= slope of faultblock opposite to the faulted side.
Badlands= areas of fine-textured drainage, steep valley walls, and narrow divides, generally little vegetation
Bahada (Bajadá)= slope due to adjoining or coalesing alluvial fans.
Baraboo= any area with a history and structure somewhat similar to Baraboo district, Wisconsin
Baselevel= level of termination of streams or that below which no stream erosion can occur.
Beheaded= applied to a stream whose headwaters have been diverted.
Betrunked= applied to stream disconnected from original main stream by drownding.
Birdsfoot delta= one with long natural levees where distributary channels project into body of water; one like that of Mississippi River.
Blunting of spurs= result on spur ends of successive erosion by meandering main stream.
Braided= applied to streams which branch and reunite repeatedly.
Butte= relatively small, steep-sided hill, generally an erosion remnant.

Canoe-shaped= applied to mountains or valleys whose form is controlled by pitching folds
Capture, see Piracy
Catoctin= linear monadnock; see Monadnock
Cigar-shaped= applied to mountains whose form is due to a pitching anticline
Canyon (Cañon)= narrow, steep-sided valley
Compound stream = stream which drains areas of different geologic structure, of different geomoprphic history
Composite stream= one which drains areas of different geologic structure.
Composite landscape= one due to rejuvination showing more than one cycle of erosion.
Cone-see alluvial fan.

Conoplain= form of pediment where slopes radiate from a point; see pediment.
Consequent stream= stream whose course was fixed by the orginal surface of a region; extended consequent= one whose course was extended as the shoreline progressed seaward during uplift of land.
Corrasion= scraping and wearing away of bedrock by mechanical abrasion, generally by stream action.
Cuesta= ridge with gentle slope on one side and a more abrupt declivity on the other; due to erosion directed by presence of a relatively resistant capping material with comparatively gentle dip.
Cutbank= slope where valley side has been undercut by lateral erosion of stream.
Cutoff meander= meander where narrow place has broken through
Cutoff spur= end of spur isolated by lateral stream erosion.
Cycle= system of development of results of a given process with increasing time of action; with running water cycle of erosion, generally in humid climate.

Defeated stream= stream whose course was changed by uplift of land.
Deferred junction= Yazoo type of stream junction where lower part of course of tributary is turned parallel to main stream by floodplain deposits.
Degradation= applied to erosion of land by running water; lowering of land surface.
Delta= deposit of a stream where it enters standing water (not another stream).
Dendritic= applied to pattern of streams developed in uniform material; resembles branching of common hardwood trees, oak, etc.
Denudation= wearing down of land or erosion.
Diastrophism= uplift of ear'ths crust by folding, faulting, etc.
Dip = inclination of rock units from horizontal; direction in which rock strata descend to lower level.
Dip slope= slope controlled by upper surface of an inclined layer of rock; the gentler side of a cuesta.
Dissected = cut up by erosion of valleys.
Distributary= channel of stream which leads water away from main stream.
Diverter= applied to stream which has captured part of another; see piracy.
Doab= ridge remaining between eroded valleys; part of interfluve. Donga
Drowned valley= valley which has been invaded by standing water as result of earth movement or obstruction of drainage.

Elbow of capture= abnormal angle in course of stream ascribed to capture and diversion.
Epeirogenetic= applied to uplift forming a continent
Epigenetic= drainage superimposed on a buried dome structure.
Endrumpf (German)= undisturbed peneplain.
Escarpment= slope which cuts across layers of rock or an abrupt declivity along a fault.
Exhumed topography= topography of land area once buried by younger material and then uncovered by erosion.
Extended, see Consequent

Fault= displacement in rocks along a plane of slip.
Fault scarp= escarpment due to relatively recent fauling where topography is due directly to the movement.
Fault valley= valley whose sides are due to relatively recent faults; a graben (German) between parallel faults.
Fault line scarp= escarpment along or near to a fault where the topography is due to the position of relatively resistant materials brought about by old movement; an indirect result of faulting which may not agree with the direction of movement of the fault.
Fault line valley= valley along a fault which is due primarily to erosion which was conditioned by faulting including fracturing of the rock.

Fault line scarp (or valley)= one due indirectly to faulting, i.e. to erosion controlled by the position of resistant material resulting from old faulting.

Felsenmeer (German)= area covered by loose rocks.

Fenster (German) ="window"or area eroded through overthrust sheet exposing rocks beneath a thrust fault.

Flatiron = triangular portion of a hogback ridge due to close spacing of valleys which cross it.

Flood-plain = channel of a stream which is covered by water during floods.

Flood-plain scrolls = deposits on inside of meander which fill the bend in order to maintain normal channel width.

Fossil surface = a surface once buried by younger deposits and later uncovered by erosion.

Foreset bedding = bedding on water-facing margin of delta formed by accumulation under water at angle of repose.

Geographic cycle, see Cycle of erosion.

Geomorphic cycle, see Cycle of erosion

Gorge = deep narrown valley or canyon.

Gravity slope = talus slope or Haldenhang (German)essentially.

Grade= profile or slope of a stream which is (on average) adjusted to transport available sediment with existing volume of water, the lowest point adjusted to baselevel = a profile of equilibrium.

Gulley = small young or new valley.

Haldenhang (German) = surface of bedrock beneath a graivty or constant angle talus slope.

Hanging valley = tributary valley without accordant junction to main valley, i.e. with falls or rapids at junction; also applied to valleys which end above level of a body of standing water.

Headward erosion = growth of a stream valley toward a divide or due to recession of a falls.

Helicoidal flow = assumed spiral course of line of maximum velocity in bends of a stream so as to stay near the outside banks.

Hogback = ridge formed by erosion on both sides of a resistant tabular body of rock with dip steeper than in a cuesta; may be interrupted by valleys or offset by faulting.

Homoclinal = monoclinal with dip in one direction only.

Horst = area between two normal faults which has either been raised or not lowered.

Incised meanders = meandering valley with rock bluffs.

Inface = side of a cuesta opposite to direction of dip = escarpment.

Ingrown meander = meander with rock bluffs which has grown larger during erosion.

Intermittent stream = one which carries running water only part of the time.

Intrenched meanders = meandering stream eroded into bedrock by downward extension only so that slopes on both sides of bends are about equal.

Inherited drainage = superimposed or epigenetic drainage derived from a cover over an older topography.

Inlier = area of older bedrock surrounded by younger formations.

Inner lowland = area at foot of a cusesta from which covering resistant formation has been removed; generally on the inland side of a cuesta of coastal plain with seaward dip.

Inselberge = island-like mountain with abrup sides rising from lowland, generally from a pediment; a form of monadnock.

Insequent stream = one whose course was fixed by obscure factors; often used for consequent.

Interfluve = surviving area of original surface left between eroded valleys.

Integrated drainage = streams joined as a result of growth in semi-arid climate, a result of filling of original depressions in surface.

Interscision, stream = result of lateral stream erosion reaching another stream or another portion of same stream.

Interrupted stream - one which is intermittent only in certain segments other than only at head; due to variation in capacity of underlying material to transport ground water flow.

Inverted drainage = change from synclinal to anticlinal stream courses.

Involution = complex folding of bedrock.

Kopje = butte. term mainly used in South Africa.

Krantz = escarpment, term used mainly in South Africa.

Lehmanns principle = shrinkage of a river by loss of flow to underground course in its own deposits resulting in decreased size of meanders.

Levee, see natural levee.

Louderback = fault block with backslope covered by lava; named from studies by Louderback on origin of mountain ranges of Great Basin.

Matched = terraces at same level on both sides of stream valley= paired terrace, due to erosion of valley filling.

Mature = applied to fully developed streams or drainage patterns; less commonly used for other geomorphic features.

Meander or meandering = regular symmetrical bends in course of stream.

Meander belt = width of area of meandering along a stream.

Meander core = end of a spur inside a meander which has been cutoff at the narrow neck of intrenched meander.

Meander cusp = point projecting into a valley between two meanders which have been eroded into the bank; see meander spur.

Meander scar = cutbank of curved outline formed where a meander undercuts side of valley; sometimes called meander scarp.

Meander spur = meander cusp, above.

Mendip = inlier

Mesa = hill with flat top due to survival of original surface or a resistant body of rock.

Misfit = applied to meanders of floodplain which are smaller than meanders of the valley walls, often called underfit.

Monadnock = unreduced elevation on peneplain presumably due to greater resistance of its bed rock than in surrounding area; German: härt linge, restberge, fernlinge.

Monoclinal = dip only in one direction; applied to monoclinal mountains.

Morvan = line of intersection of two peneplains due to tilting of land between their formation.

Natural levee= higher part of a floodplain along banks of the stream.

Neck = narrow part of a meander spur.

Nick, see nickpoint

Nickpoint = point of change in slope of a stream at upper termination of a new and steeper profile due to uplift of the land; German knickpunkt

Nip = cut bank of meander, a meander scar.

Obsequent = stream which flows opposite to direction of dip, a tributary of a subsequent stream developed on outcrop of a weak material.

Orogenic = applied to movements of earths crust forming mountains.

Orogency = mountain forming.

Overfit = term applied to a river which is too large for its valley.

Oxbow = applied to lake in cutoff section of meander.

Paired = applied to terraces at same level on two sides of valley due to erosion; see matched terrace.

Panplain = plane sloping seaward which was made by lateral stream planation.

Pediment = sloping plane underlain at shallow depth by bedrock and occuring at foot of a mountain range in semi-arid climate; ascribed to lateral stream erosion plus slope wash; may have a thin cover of stream sediments.

Peneplain (peneplane) = theoretical end-product of cycle of erosion where the stream divides have been greatly or entirely removed leaving a plain-like surface. German: fastebene, rumpfläche, endrumpf.

Piracy = diversion of part of another stream by headwater erosion of major stream; stream capture, drainage adjustment.

Pitch = inclination of fold along its axis.

Planation = erosion or degradation of the land.

Plateau = elevated region with fairly equal summit levels, generally nearly horizontal sedimentary rocks or lava flows.

Playfairs Law = fact that most streams join a larger one at same level without falls or rapids.

Plungepool = basin excavated by debris below a falls.

Pothole = hole in bedrock caused by swirling water in rapids; not to be confused with kettle hole.

Primärrumpf = slightly uplifted peneplain, an area always old because erosion kept pace with uplift (German).

Radial drainage = streams which lead away from a central point.

Raft lake = lake due to obstruction of stream by driftwood.

Rain-wash = slope wash

Regrading = erosion to a rough profile by removal of mantle rock.

Rejuvination = effect of uplift of land upon erosion.

Relief = difference in local elevations, say from valley bottom to divide.

Repose, angle of = angle at which loose granular material comes to rest; can also be applied to equilibrium in other slopes.

Resequent = stream flowing down-dip although not a consequent.

Resurrected = surface uncovered by erosion of later cover.

Rectangular = trellis drainage pattern.

Rock city = area of roughly rectangular blocks of resistant rock weatherd out along joints.

Rock-defended terrace = erosion remnant of terrace of stream deposits preserved on bed rock.

Rock fan = form of pediment superficially resembling alluvial fan.

Rock terrace = shelf on hillside due to presence of layer of resistant rock.

Saltation = movement of bed load of stream in series of jumps.

Scarp = escarpment due either to erosion or to faulting.

Scroll, meander = traces of old meanders, see meander scroll.

Slipoff slope = slope of end of spur inside bend of meandering valley, result of enlargement of meander radius during downward erosion; present with ingrown meanders.

Steppe = arid plain in interior of continent as in Siberia.

Strath = wide valley, generally applied to floor of a wide valley later trenched by erosion.

Strath terrace = strath which extends across divides of minor drainage, often called an incipient peneplain or graded erosion surface.

Strath valley = abandoned wide valley now higher than adjacent streams.

Suballuvial bench = Lawson's term for rock floor adjacent to mountains in arid climate.

Subsequent stream = one whose valley was formed subsequent to uplift by reason of relative weakness of underlying material; also subsequent valley.

Superimposed (superposed) = effect of stream courses of a now largely eroded cover wich once concealed underlying materials.

Suspension = method of transport of material in water where it is kept from settling by upward currents or turbulence.

Sweep = downstream component of motion of meanders due to greater force on banks which face the current.

Synclinal mountain = mountain in which rocks have synclinal structure.

Tepee butte = small hill with conical form; resistant material at top covers only small area.
Terrace = shelf on valley side, any origin.
Texture = density of streams per unit of area; may be computed either on length or number of streams; also termed drainage density or stream frequency.
Trellis drainage = rectangular stream pattern due to prsence of long parallel ridges of resistant rock whose abrupt sides cause large junction angle of tributaries.
Treppen(German) = steps or terraces forming a stairway; ascribed to sucessive uplifts.
Turbulence = mixing or rotational flow of water (also applies to other fluids and to gases).
Turf = sod.
Two-story valley = valley within a strath or a valley in a larger valley; called by some two-cycle valley.

Undercut = erosion at bottom of cut bank.
Underfit = river too small for valley; also applied to small meanders in bottom of a meandering valley.
Unaka = kind of monadnock or group of monadnocks.
Underflow = underground flow of a stream through the earth.
Uniclinal = shift of stream course down dip.

Vale = lowland between cuestas.
Volcanic neck = surviving igneous rock in conduit of volcano laid bare by erosion.

Wadi = dry river course rarely flooded (Arabic).
Water gap = valley of stream through a ridge of more resistant rock than elsewhere.
Wind gap = abandoned water gap or a low point in a divide, of any origin.
Wold = cuesta

Yazoo type of stream = one whose junction with main stream is diverted by natural levee of latter.

Personal (1)

GEOLOGY 109

GEOMORPHOLOGY

also Flow of ice 1949
wave height 2
Rainy winter 1950
endpoint of cycle 1951

Supplement, 1949-50

Concavity of slopes

Introduction. Causes of convexity of hill and mountain summits have already been discussed. It is observed that most streams have a marked concave slope, although variation in discharge, load, and nature of underlying material commonly make mathematical analysis impossible . The lower slopes of fragmental volcanic cones, outwash terraces, alluvial fans, and pediments also display concavity. With these logarithmic plotting of profiles discloses that they obey definite laws. Two general groups occur: (a) most pediments, outwash plains, streams of essentially constant discharge with no tributaries within the part examined, all have an exponent, n, which lies between about 7/10 and 8/10, (b) the lower parts of volcanic cones and some pediments display a value of n which varies from 3/10 to 5/10, with rare examples of fall in proportion to logarithms of distance.

Explanations. Three possible explanations of the above facts are: (a) the slope is one of constant eroding and transporting force, that is a balance between erosion and transportation, (b) there is a difference in manner of flow of the water in different parts of its course, or (c) the slope is related to competence, that is to the largest particles available for transportation.

Slope of constant force. Little attempted to analyze slope erosion on the assumption that erosive force is proportional to V^2/D, explaining that this is the vertical gradient of kinetic energy of flowing water. However, this explanation is equivalent to saying that erosive force is proportional to the slope, for solution of the old Chezy formula (V^2:D.S.) for slope gives just this expression. Moreover, this view neglects the factor of total discharge. Although it happens that the exponents derived both by Little, and by modification of his method, agree rather well with those of many outwash plains and pediments the theory must be rejected. It assumed a discharge per unit width which increased in proportion to slope length and in the cases where there is approximate agreement it is obvious that the bulk of the discharge was acquired above the section under study. This is notably the case in front of a glacier and at the foot of mountains. Horton's analysis of slope erosion used the classic DuBoys depth-slope formula (F:D.S.) and also assumed that rainfall was gathered all the way down a slope. The reasoning here given (p.42) derives a much lower exponent for a slope of constant force that shown by any observations. It is clear that the slopes under consideration cannot be explained as those of constant force of erosion and transportation.

Change in manner of flow of water. Horton's experiments showed that in actual experiment the flow of water in thin sheets on the surface of the ground is neither laminar nor fully turbulent. The expression discharge in unit width, q = constant X depth 5/3 applies to true turbulent flow only. It should be 3 with laminar flow. In experiments the value of this exponent of depth varied from 1 to 2. Although if the bottom were truly smooth, laminar flow might occur, a little thought will show that it is incapable of important erosion and transportation. If any substantial depth is obtained turbulence must begin, a condition which greatly enhances both erosion and movement of material. It is likely that in certain spots where flow is concentrated until it attains a force considerably in excess of the resistance of the bed to erosion, there may be a transition from mixed to either fully turbulent flow or possibly from turbulent to shooting flow. Thus we may have gullies (dongas in uth Africa) formed in the otherwise reasonably smooth surface of a pediment or wash plain. It is possible that the phenomena of braiding may be in part related changes in manner of flow of water where the sediment load is at its maximum lue (so-called overloading). We may, therefore, conclude that nothing which is own proves the theory of change in manner of flow in descending a slope although h a change may explain gulleying.

Control by competence of water. Competence is defined as the ability of water to transport material measured by either diameter or weight of the largest particle. It has been shown that for particles over 1 mm, diameter, the diameter of the largest particle is proportional to the square of the water velocity and that for particles less than about 0.2 mm in diameter the ratio changes to the square root of the velocity Now with turbulent flow, V^2 is proportional to S (slope), other things being equal. With mixed flow V^2 is related to the 1.4 power of the slope. By substitution it is easy to demonstrate that the larger particle diameter is related directly to slope for turbulent flow and to the 1.4 power in the case of mixed flow. With small particles the law is different for more of the sediment is transported in suspension. By the above criteria diameter of particles should be proportioned to the $\frac{1}{4}$ power of slope with turbulent flow and to the 7/40 power of slope with mixed flow. Now it has been observed that along streams from the Black Hills avg. average size of gravel stones is directly related to slope and it seems likely that such is generally the case.

Some have suggested that decreased slope downstream is due to "selective transportation", that is to progressive loss of the larger particles thus permitting a constant decrease in slope below that point. This could be explained by the fact that, as transporting ability of a stream varies with discharge, the larger pebbles are most apt to be left behind during the next flood, but in general the view seems hard to demonstrate.

Another suggestion is that wear of pebbles during transport decreases the need for velocity, so that slope is proportioned to size of largest material present at any given location. Sternberg proposed a law of pebble wear in which reduction in weight was assumed to be a function in which a constant is raised to a negative power in which distance is a factor. The usual form in which this is given is ratio of final weight to original weight = e^{-ah} where e is the well known constant, a is another constant depending on nature of the rock, and h is the distance. To compare diameters the constant a of the exponent is divided by 3. The above equation gives the percentage of weight lost in unit distance. It has been claimed that pebbles in the Rhine River lose 1 percent of weight per kilometer of travel. Unfortunately, neither the exact mechanics of pebble abrasion (impact and abrasion) are clear in all cases, nor is the source of a pebble readily determined. One wonders just how pebbles of local derivation were distinguished from those of distant origin! Another factor which seems to have been overlooked, is the tendency of some kinds of rock, like granite and sandstone, to pass abruptly from pebbles to much smaller particles. Weathering of pebbles during transit has also been ignored, Although the fact is recognized that pebbles are reduced in size and weight by transportation the validity of any universal mathematical law is doubtful and slope cannot be defined in terms of distance in this way. It seems likely that slopes with a very low exponent are all underlain by very fine materials like volcanic "ash" and clay.

An indirect approach to the problem of distance vs slope may be made by considering other hydraulic principles. It has already been shown that velocity of water is related not only to slope but also to depth (or hydraulic radius). In Mannings formula the latter quantity, D (or R) carries a higher exponent than does slope. In practice it is a larger number, for slope is in most instances of very small numerical value. In Horton's formula for slope wash discharge on unit width increases at a power of depth which is in general more than unity, and is much higher than that of slope. From this it may be concluded that increase in discharge of a stream, or in amount of slopewash in descending a slope, will in general need progressively less slope with increase of distance to take care of the discharge.

A somewhat more definite approach is by means of the Schoklitsch bed load formula which related the coarser material carried to total work of the stream in unit time by the familiar depth-slope relationship. As checked by experiment bed load per unit width = constant x slope to 1.25 to 2 power x excess of discharge per unit width after deducting discharge per unit width necessary for any movement to occur.

Using common notations: $G/p = C \cdot S^{(1.25 \text{ to } 2.0)}\ \overset{QQ}{(q-q_0)}$

Substituting V^2 for S for turbulent (where G = total bed load . and p = wetted perimeter and C is a constant depending on nature of material) flow, using the average value of exponent of S = (3/2), this becomes $G/p = C.V_m^3\ (q-q_0)$ and substituting for $q = V_m.D$, $G/p = C.V_m^4\ (D-D_0)$. It should be noted that all velocity formulas must have an exponent twice that using in a depth-slope expression. Now if we solve the equation using $S^{3/2}$ for slope it is evident that

$$S = \left(\frac{G/p}{C(q-q_0)}\right)^{2/3}.$$

Now we may assume that bed load per unit width of channel (G/p) remains essentially constant below a zone near to the border of the belt of no erosion where erosion b begins. Now if we assume, that unit width discharge, q, increases in direct proportion to h or horizontal distance away from the border of belt of no erosion this works out to S: $h^{-2/3}$ and to fall, f: $h^{1/3}$, since f is obtained by multiplying average slope by horizontal distance. In general streams do not increase at as high a rate measured in terms of average discharge and distance from one gauging station to the next. Reduction in rate of increase would then lessen the concavity by increasing the exponent. If, we assume a very low rate of increase, like that of streams fed from the mountains or from an ice sheet, for instance q: $h^{1/10}$ it follows that S: $h^{-1/15}$ and f: $h^{14/15}$. The last exponent, about .93, is not much higher than that observed in pediments and outwash plains. We would expect that with no increase in discharge per unit width a straight line unchanged slope would occur.

Another apporach might be along the line of capacity or limit of bed load, a feature not included in the above formula, which makes load directly proportional to discharge.

Conclusion. Although the foregoing discussion has failed to provide a conclusive universal answer to the problem of concavity of slope, the best solution seems to lie in the competence of water for transportation of the coarser debris conditioned by the size of material present. Save in exceptional conditions it is impossible to relate size of available material to distance downstream or downslope. Sternbergs supposed law is plainly inapplicable in most localities. It must also be realized that hydraulic principles alone tend toward concavity.

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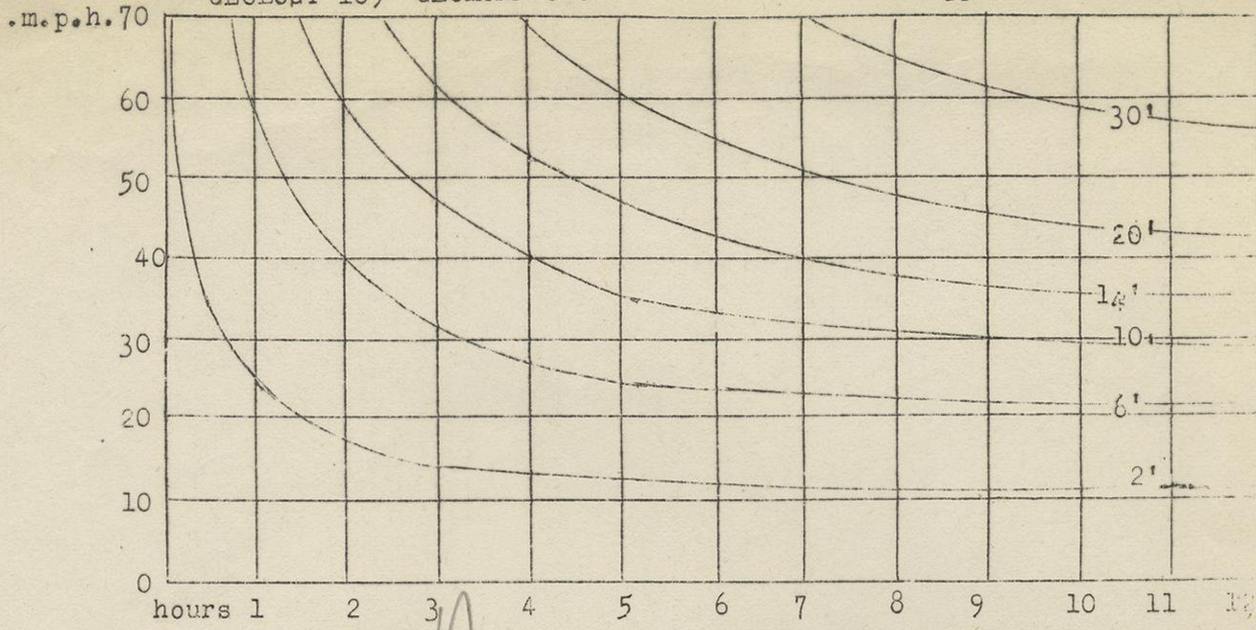
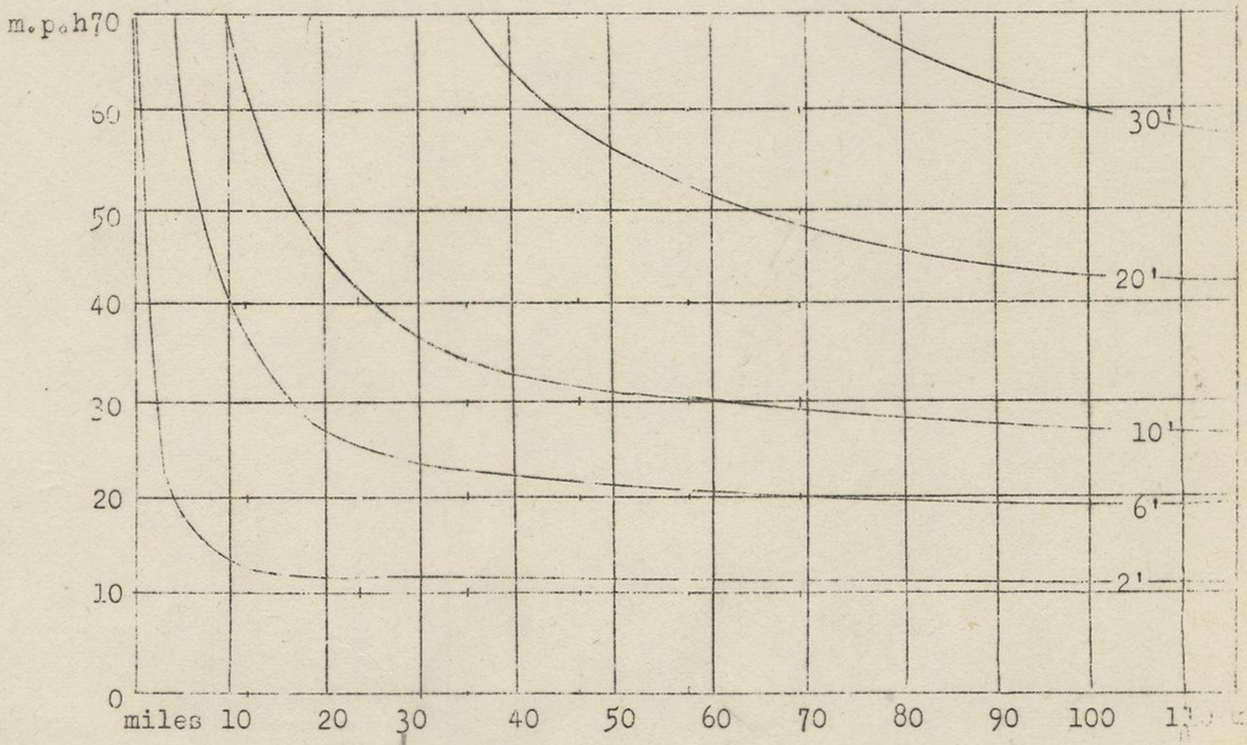


Wave heigth related to wind velocity and duration



Wave heigth related to wind velocity and fetch

Adapted from Scientific Monthly 69: 232, 1949 and Wind waves at sea, breakers and surf, U. S. Navy Dept., Hydrographic office, Pub. 602, 1947 Distances and velocities changed to land units.

Personal

(5)

GEOLOGY 109

GEOMORPHOLOGY

Supplement: Concepts of the endpoint of the cycle of erosion 1951

Introduction. The following is intended to clarify some of the material in the text. It includes neither all discussions nor the discrimination of now-dissected surfaces. The definitions of the word "peneplain" or "peneplane" given in another supplement show that the endpoint of the cycle of erosion has never been defined very clearly and that wide differences of opinion exist in just what is included. For instance Howard states: "The term peneplane is valuable, because it may be applied to any broad degradational surface of low relief even though its exact origin has not been determined.--------The term peneplanation signifies the reduction of a region to low relief without stating the process or processes involved." These views probably reflect the later opinions of Douglas Johnson for they coincide with those expressed by Lobeck. However, the majority of writers obviously do not use as broad a definition but stick much more closely to the idea that the peneplain is the final product of a cycle of erosion under a so-called "normal" climate. Such a view, which seems to agree with those of Davis, who first introduced the term peneplain, although not the idea of reduction to base-level, involves the corallaries that: (a) no earth movement intervened to interrupt the cycle during the long period necessary for even partial completion, (b) interstream areas were reduced to the approximate level of the streams mainly by mass movement of the mantle rock aided by slope wash, (c) static conditions prevailed for so long a time that resistance of the weathered mantle to erosion can be neglected, (d) the climate was humid and chemical decomposition was more rapid than the removal of the products of weathering by erosion, (e) the summits of the divides should become convex upward and, (f) the valley bottoms should become nearly flat by aggradation with excess debris which could not be removed. Some held that the great amount of sediment eroded would in time raise sea level and thus increase aggradation of the valleys as those of coastal regions are now being filled as a consequence of the increase of water in the oceans because of melting of glaciers.

Examples of Peneplains. The only areas in humid lands which really approach the theoretical peneplain are located either on soft clays and shales, or on limestone. Some enthusiasts used as examples for students Pleistocene outwash plains and lake beds, as well as floodplains of large rivers. Many maps purporting to show "old age" topography have now been shown to have been crudely surveyed with a very large contour interval. By no means should we either use such inaccurate maps or confuse depositional with erosional forms. Depositional topography simply does not belong in the erosion cycle. True undissected surfaces due to erosion beveling the geologic formations have been recorded only (a) in regions which have climates far different from that postulated by the pioneers and (b) beneath a cover of later sediments.

Objections to the peneplain hypothesis. In 1933 Crickmay ventured to suggest that "the fruitful conception of the 'cycle of erosion' carried with it into general currency certain ideas which rest on no real foundation.-------Geographic old age and peneplanation rest on nothing but pure deduction by a few---and a blind acquiescence by the rest of us--------only blindness could prevent us from seeing some serious inconsistencies---" Along with the total failure to find uneroded remnants of true peneplains he particularly noted the reported series or stairways of erosional levels in the same area, as well as the sharp outlines of many supposed monadnocks. In the last he might well have included the type Mount Monadnock, New Hampshire. Crickmay proposed that divides were removed primarily by lateral stream erosion and stated that "a normal floodplain is underlain everywhere at shallow depth by planed bed-rock. The alluvium is a mere carpet---" This endpoint of broad floodplains only slightly aggraded and separated by steep-sided remnants of divides he named a panplain. Absence of many modern panplains was explained by recent rise of the continents. Crickmay held that peneplanation or panplanation should start with the lower parts of rivers and grow laterally inland. Under his view stairways of erosional remnants are the expected results of successive uplifts. He recognized panfans and pediments in arid regions. Crickmay's view seem not to have received much recognition, possibly because the concept of pediments and pediplanation has since come to the front. King (1949) remarks "The development of a single true 'peneplain' involves of necessity the complete destruction of all its forerunners---" "instead of the 'normal cycle' concept, the idea of 'pediplanation', involving the extensive retreat of hill-slopes and the survival of planed remnants of earlier cycles upon upstanding topographic masses, has been advanced as the mode of continental planation." King concluded that the true peneplains occur only in a humid climate, whereas panplains could occur only in savanna and semi-arid climates. He also mentions the "etchplains" of tropical regions made by erosion of the decomposed rock along valleys leaving it in full thickness under the steep-sided divides. Pediplaned surfaces he ascribes to arid and steppe climates. They are associated with surface accumulations of silica, aluminum oxide, and iron oxide. This process involves the parallel retreat of slopes, plus removal of debris of weathering by sheet wash, and carries as a corallary the idea that the present distribution of climates has not always been present. King in fact ventured to suggest that most of the erosional bevels of the entire world are due to pediplanation, a theory tied in with the breakup of continents which has not been widely accepted in the northern hemisphere.

Development of the pediment concept. The following outline of the development of the pediment concept does not include all workers but aims simply to give the general growth of the idea. In 1880 Gilbert described the sloping rock surfaces around the Henry Mountains, Utah, and ascribed them to lateral erosion by streams. He did not explain the pediment terraces. The base level of these streams was fixed by local conditions. This same idea has been followed by many ever since. In 1897, however, McGee described the sheet floods of southwestern United States. Paige, in 1912, described rock pediments and ascribed them to a combination of lateral erosion and interstream degradation. Lawson's elaborate treatment of desert topography in 1915 recognized the retreat of mountain slopes whose inclination depends upon the size of the boulders in the talus. He concluded that with a rising base level due to filling of structural basins the rock surface beneath the alluvial cover should be convex upward. This surface was termed the "sub-alluvial bench". In 1919 Jutson described sheet floods in Australia but ascribed little erosive action tothem. In 1922 and 1925 Bryan elaborated on Lawson's ideas. In 1931 Blackwelder endorsed lateral erosion of weathered rock by streams as the major process of formation of pediments. Johnson in 1931 and 1932 followed up this view. In 1935 Rich revived the theory of slope wash working on a surface composed of weathered rock. The available water shaped this surface to a slope on which it could remove the debris formed by weathering. If, however, the material transported came from another source than local disintegration, only a thin veneer of transported debris was left. The degree of slope depends upon the size of particles which must be transported. Rich concluded that lateral erosion by definite streams "is not necessary for the production of rock fans and pediments though in many instances it contributes notably toward their formation." The origin of pediments along the foot of a retreating escarpment was also discussed. "Wasting and sheetwash, in conjunction with the blanketing effect of alluvial debris, are the essential factors, required for the production of rock fans and pediments. Lateral corrasion by streams is not necessary." "Drainage diversions and the dissection of abandoned fans and pediments are normal----They do not require----diastropic or climatic changes." Field, however, reaffirmed the lateral erosion theory. Davis argued in a paper published in 1938 that absence of cut-banks at the borders of the mountains adjacent to pediments favors sheet floods rather than lateral stream erosion. Howard in 1942 strongly supported lateral stream erosion stressing the work of small tributaries and correctly stating that desert streams are braided rather than meandering. A series of papers from 1948 to 1950 by King and Fair on phenomena in South Africa give important ideas. They were the first to employ Wood's classification of slopes and included much quanitative slope data. Fair noted that in the arid portion of Natal the slope of talus is related to average size of boulders and that the angle of slope decreases with weathering of the lower part of the talus making more or less of a transition to pediments below. The change is from 14 degrees to 4 degrees inclination. The pediments are eroded on shale and debris is moved by sheetwash. The major processes are weathering and removal of the mantle by sheetwash at a rate closely equal to that of its formation. Near the divides infiltration, aided by evaporation, prevents erosion suggesting Horton's "belt of no erosion". The slope difference at the top of the pediment was ascribed to change in the volume-load relation. Parallel retreat of slopes is present only where there is a hard cap rock. Sheetwash slopes are concave. No lateral erosion of streams was recognized.

King classified the different types of erosion bevels according to climate as outlined above, and concluded that pediplanation is more rapid than peneplanation as he interprets the term. Although it should lead to concave slopes the divides may be rounded by weathering into something more like the theoretical peneplain. In a paper of 1949 on the pediment landform King concluded that it is "adapted to and moulded by sheet-flow" and that the upper limit is fixed by the change from turbulent flow in ravines to laminar sheet-flow. The concavity of slope is classified as that of water erosion. He declares that pediments are "Possibly the most important of all land forms." "It is the fundamental form to which most, if not all, subaerial landscapes tend to be reduced, the world over." The retreat of escarpments is essential to the process and the pediment, on which erosion is at a minimum, is the result of thunderstorm type of rainfall. Erosion of escarpments is done by a multitude of gullies on their faces.

<u>Summary</u>. (1) Pediment formation can begin only where the original slope of the land is greater than that necessary for water to remove the products of weathering. Such slopes include normal valley sides, fault blocks and sides of folds (either anticlinal or monoclinal) (2) Original slopes in every case soon became of the constant or talus type, which means that in horizontal rocks there must be a resistant cap rock, (3) In order to maintain a constant retreat of the

talus the rock material must weather into particles which can be carried away by slopewash. Rocks which meet this condition include granite or other coarse-grained igneous and metamorphic rocks, sandstone, and some kinds of shale. In all of these the passage from large talus blocks to granular material is abrupt. (4) Retreat of the escarpment is maintained along a fairly definite line without much deep dissection by valleys by a combination of undermining by weathering, and erosion by minor ravines including streams which descend through the talus. The heads of the larger streams may in time retreat clear through the escarpment and coalesce with valleys on the other sides. Such cols are called pediment passes, and in many cases have a convex divide. The two sides may not accord in level, but join with a steep slope. Thus an elevated mass is transformed into a group of isolated remnants. (5) There is a relatively abrupt change in slope at the foot of the talus, to the gentler inclination of the pediment, but true cut banks appear to be rare. (6) The slope of the pediment is that on which the available water can remove the particles produced by weathering at the approximate rate of production by weathering of the talus blocks. It exceeds 10 degrees in many localities. (7) Ability of water to remove debris is facilitated both by sparse vegetation and by sudden downpours or cloudbursts, hence an arid or semi-arid climate is required for pediment formation. This checks with the surface accumulation of oxides and carbonates. (8) Obliteration of divides on the pediment is in part due to lateral erosion by braided rills, in part to erosion by sheet floods, both facilitated by deep mechanical decomposition of the bed rock, but sheet floods cannot form until other forces have prepared a smooth surface for them. Some observers report that shallow sheet floods have clear water and laminar (or mixed) flow and hence accomplish little erosion. (9) Most pediments have a cover of a few feet of debris in process of transit. A rising baselevel may increase the thickness to more than the depth to which floods disturb the material. A falling level minimizes the thickness of the mantle. Thickness may also be related to climatic changes which altered both the rate of supply and the amount of rainfall. (10) The angle of slope of a pediment must considerably exceed the theoretical slope of a peneplain where thorough chemical decay is believed to occur. (11) Filling of a structural basin below the pediment may proceed at the same time as erosion closer to the escarpment forming a pediplain (12) The concavity of slope of normal pediments is related to the laws of water transport. It is in part explicable by increase in depth and amount of flow and in part to decrease in mean diameter of the load by wear and weathering. The slope is probably not related to the size of the largest pebbles, for these required a only very small part of the energy of the water. Plumleys observation on the terrace gravels of the Black Hills was incorrectly stated in another supplement. It is a rough approximation to the "phi mean size" which is defined as the logarithm to base 2 of the geometric mean size. (Geometric mean of n quanities is the nth root of their product.) Observations on some outwash plains in Wisconsin showed a decrease in geometric mean size of the sampled portions at a rate less than the decrease in slope. A solution of this problem is yet to be reached.

(13) The end point of pediment development, which is not exhibited in the Basin and Range Province, involves disappearance of most of the residual highlands. Sites of former high places would be marked by convex domes, although the lower slopes should still be concave. Bed rock would be close the surface and not much altered by chemical weathering. If we postulate a former wide extent of arid climates many old erosion surfaces may be explained as pediments or pediplains. This would account for series of erosion terraces which cannot be ascribed to peneplanation.

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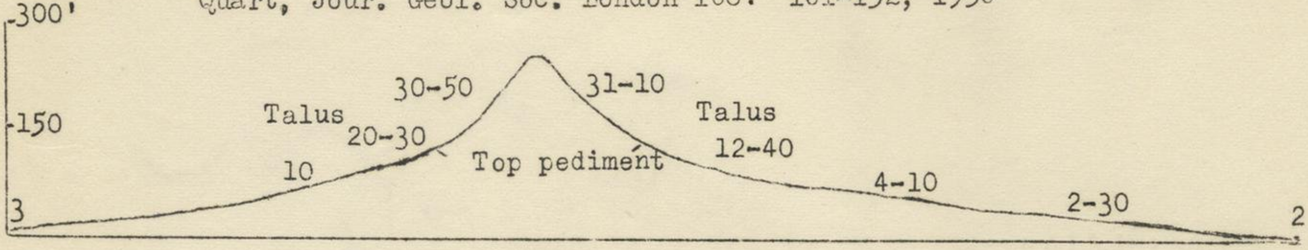
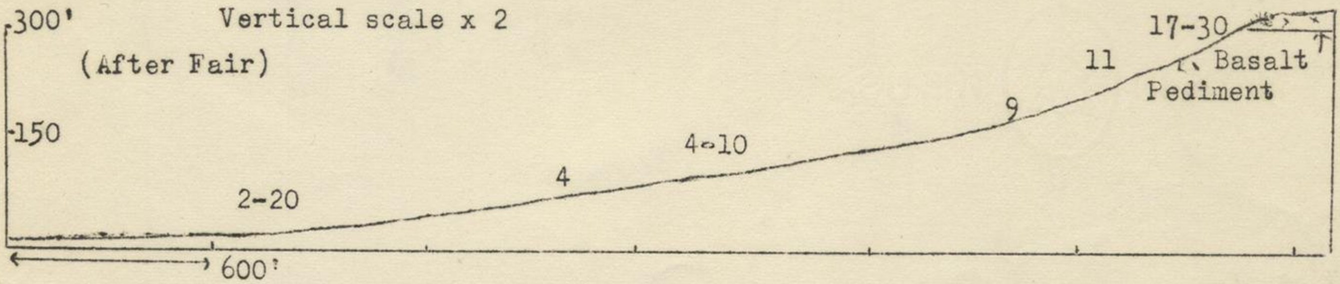
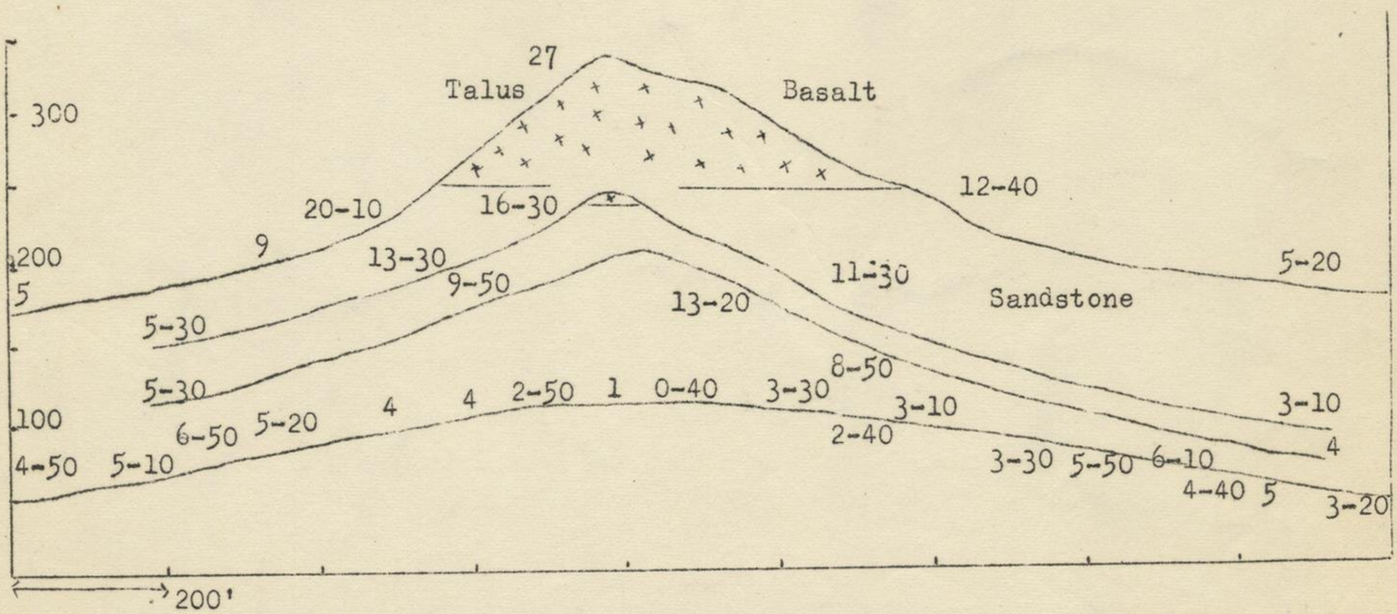


Formation of pediments on sides of divide capped by basalt. Slopes in degrees
Vertical scale x 2
(After Fair)



Hillside thinly capped with basalt and underlain by shale Slope in degrees
Vertical scale x 2



Hillslopes in interior of Natal (after Fair) showing stages in lowering of a divide of sandstone capped by basalt. Note development of convex summit in last stage. Elevations not relative to one another. Slopes in degrees, Vertical scale x 2

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GEOLOGY 109 GEOMORPHOLOGY Supplements, 1952, part 1

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1952 submarine canyons
2 great exam of erosion topog
- outline many water
3 soil mechanics

Carolina Bay Problem, 1952

The Carolina Bays could have been caused either by purely terrestrial forces, such as waves, currents of air or water, by solution by ground water, or by extra-terrestrial meteorite impact. Modern coverage with air photography demonstrates many things which were previously unknown or merely surmised.

Facts. (1) Typical "bays" occur only in the Coastal Plain and are absent both in the adjacent sea bottom and Piedmont. (2) Bays occur from Florida to New Jersey but are most abundant in the Carolinas. (3) The total number of well-formed elliptical bays is estimated at about a half million. (4) The rounded outline is less regular in regions underlain by limestone. (5) Most bays have a unstratified sand rim which is best developed around the southeast end. (6) Orientation of long axes of bays varies only slightly and changes are gradual between different regions. (7) Many bays overlap one another or have more than one rim. (8) Most bays are filled with peat which is thickest (15 to 30 feet) toward the SE end; this peat is underlain by lake silt. (9) Only a few springs occur in bays. (10) Bays are equally well developed in all regions suggesting the same age. (11) No similar basins occur anywhere else in the world.

The theory of Douglas Johnson puts its primary emphasis on solution by artesian springs. But there is no relation between bay distribution and either ground water circulation or presence of permeable or soluble rocks. The lack of any relation to a joint pattern is also evident. There is no relation of bays to rock structure and no noticeable difference of age of bays in different localities. The theory appears to exaggerate both the actual amount of rising ground water and its ability to dissolve material. The superimposed theory of wind and wave action in sinkholes to explain the sand rims fails to take into account the rim distribution which does not agree with the known SW direction of winds. The importance of water currents in such relatively small and shallow lakes also seems decidedly exaggerated. Compare the slight amount of such work in most lakes of glacial origin. Overlap of one bay on another is also hard to explain by Johnson's theory. The evidence of filling of a lake with peat does not agree with the idea of subsidence due to solution.

The suggestion of Grant that the springs were submarine and were frequented by great shoals of fish which swam around in circles seems entirely too far-fetched. One bay is 7 miles long. Besides it would not account for the sand rims.

Prouty's revised meteoritic hypothesis is based mainly upon the shock wave or compression cone which occurs with bodies moving through the air at supersonic speed. Impact of a vast shower of meteorites (a comet) would thus account for the formation of so many elongated craters in the sand of the Coastal Plain and not in adjacent firmer material. Many meteorites have been discovered in the Piedmont to the northwest. This idea also explains the striking perfection of outline, the marked parallel orientation, the overlaps, and the sand rims. Two checks have been presented. First the shapes of the bays agree with small craters formed in fine sand overlying clay by high-velocity rifle bullets fired at an angle of 30 to 35 degrees to the surface. Second, magnetometer work has disclosed many local "highs" nearly south of and distant from the rims by about the length of the short

Legrand, H. E Streamlining of the Carolina Bays
J G 61: 263-274, 1953

 Hager: Dorsey Crater mod (Meteor Crater) argon a geol. feature AAPG B 37 821-857, 1953

axis of the bay adjacent. Magnetic work is confused by linear magnetic highs due to basement(pre-Cretaceous)rocks, by the great number of bays in some districts, and the difficulty in making readings in swamps on unstable peat. The highs do not check either in strength or position with the idea of redeposition of iron oxide from solution of the bays. Limonite is not strongly magnetic. Final proof of the presence of metallic iron under the magnetic highs can only be established by test drilling. A survey by air-borne magnetometer might also be of value since it would not be affected by ground conditions.

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Technical terms suggested by Bryan (see next section on permafrost)

Pergelisol = permafrost or permanently frozen ground

Mollisol = surface zone thawed in summer, the "active layer"

Intergelisol = transition zone at bottom of mollisol, thawed at times

Tabetisol = unfrozen ground above, within, or below the pergelisol or frozen ground

Congelifraction = frost-splitting of rocks

Congeliturbation = frost action including frost-heaving and mass movement of the active layer or mollisol

Congeliturbate = disturbed material of active layer or mollisol

Cryopedology = science or study of intensive frost action, frozen ground, etc.

Cryoplanation=/process of leveling of topography under frozen ground conditions, similar to peneplantion is warmer climate

Pergelisol table = top of pergelisol

Subgelisol = unfrozen ground below the frozen zone

Supragelisol = zone above the pergelisol

Pergelation = process of forming permanently frozen ground at any time.

Introduction: Permafrost or perennially frozen ground is important to land forms in its effect on (a) weathering (b) erosion, (c) ground water circulation, and (d) formation of relief both by the freezing and the melting of ice. Although known for many years attention has been devoted to this problem recently because of its effects on the works of man. Criteria for its recognition from surface indications are, therefore, important. Attention has also been given to the former distribution of permafrost, which has left a record in land forms.

Origin--Sources of heat: The surface of the earth obtains heat by (a) direct solar radiation, (b) conduction from the air, (c) conduction from the interior heat of the globe, and (d) from latent heat set free by condensation of atmospheric moisture. Study of (c) has been based upon known rates of downward increase of temperature in drill holes and mines combined with laboratory determinations of the conductivity of the materials of the crust. It is generally believed that the rate of heat transmission from the interior of the earth is very slow, probably less than 0.2 calories/cm^2/day. Average conductivity for rocks is generally given as about 340 cal/cm^2/day/deg. C., for ice as 457, snow as about 43 and water as 112. In other words, the conductivity of solid ice is much above that of water or rock. Conductivity of mantle rock or permeable materials is greatly affected by the presence of either ice or water, particularly if the latter is moving. The low rate of heat escape from the earth is due to the prevailingly low temperature gradient. Although the conductivity of air is very low, about 1 percent of ice, its specific heat (.237) is relatively high, almost half that of ice (.502). This property makes it possible for warm winds and rain to contribute much heat to the earth by conduction. Direct solar radiation decreases with latitude because of the low angle of the sun's rays to the horizon but although the very long days somewhat compensate for the short northern summer. Direct radiation may contribute several hundred calories to a square centimeter per day. Formation of fog may contribute much heat.

Loss of heat: Heat is lost from the ground by (a) conduction into the air and (b) radiation. Loss of heat is impeded by both snow cover and a mat of vegetation. Both have very low conductivity and serve to keep out heat from the sun. But both are good radiators and promote loss of heat to the air when the ground is the warmer of the two. In the Arctic strong winds keep snow from accumulating to great depths by concentrating it into local drifts, whose summer melting is the source of streams of water. In most northern lands snowfall is not heavy. In order to have freezing of the ground it is necessary first to have moisture present, for dry materials cannot be frozen. Second, the temperature of the ground must be reduced to below the freezing point of water for some considerable time. High winds low air temperature, and good conditions for radiation from the earth all favor deep freezing. Whether or not bare ground freezes more quickly than areas with a matt of frozen mosses is debatable. The depth to which frost extends probably increases at less than direct ratio to the duration of low temperatures. If the summer thaw reaches only to a slight depth, the next winter, which is much longer than the summer, will add frost so that the congealed layer exteds deeper and deeper with time. Ice is a better conductor than dry rock or earth. Permafrost has been reported up to nearly 2000 feet deep. Ice occurs in irregular masses, sheets, wedges, and granules. Wedges narrow downward and may extend 30 feet below the surface. For ground temperature observations see figures 1 and 2, p. 7

Summer thaw: Depth to which thawing occurs in the short Arctic summers depends upon the (a) absorption of radiant heat from the sun, and (b) conductivity of the ground. The matt of tundra vegetation is a good insulator. The impermeability of the permafrost retains much water in the thawed layer. The prevailingly slight rainfall the the Arctic slows down melting. Summer ground conditions are then like those of more southerly latitudes during a dry spring when melting of the frost is very slow due to lack of warm rain.

Refreezing: When colder temperatures return in the fall a new frost layer forms on top of the melted zone. This thickens until it merges with the permafrost below. Locally ground water is trapped between two layers of frost. Fig. 3, p. 7

Distribution of permafrost: Permafrost is widespread in northern North America and in Siberia, as well as on some high mountains. Approaching more mild climates from either the Arctic or from the mountains, the permafrost becomes more and more patchy and shallower. Even in far northerly latitudes there is no permafrost under thick glaciers or adjacent to large rivers and the sea.

Topographic effects: The topgraphic effects of permafrost may be divided into (a) those due directly to the ice and (b) those caused by the thawed or "active" layer of summer. Frost heaving is believed to form hills ("pingos") up to 600 feet across and 230 feet high. These occur chiefly in fine-grained lake sediments. They contain radiating ice veins. Peat mounds with a core of ice do not exceed 25 feet high. One of the most prominent features which is readily seen in air photos, is the polygonal pattern due to melting of the tops of ice wedges. These display trenches up to 2 feet deep. The diameter of the polygons is for the most part only a few feet although a maximum of 600 feet is recorded. The shape varies widely and is rectangular on a slope. The best examples occur in fine-grained sediments and poorly drained areas. On rocky ground they grade into stone nets on flat areas and stone stripes on hillsides. The pattern of polygons is etched into the lower side of ice on ponds until that exceeds half a foot in thickness. In the centers the polygonal areas may be either depressed and covered with a mixture of ice and peat or elevated forming low mounds. This second condition is thought to be an older stage of development. When small streams form on an area of cracks they are angular in course interrupted by small pools due to ice melting in the centers of the polygons ("beaded streams"). The form of the cracks is thought to be similar to the contraction which formed columnar basalt or mudcracks. However, it is much larger units. The thawed or active layer is equivalent mechanically to thin mantle rock overlying impermeable massive bed rock. The confining of ground water to such a thin zone makes for extensive mass movement with folding, contortion, and considerable chemical weathering due to the abundant organic acids and high mineralization. Some regard the abundant silt of Alaska as a product of this weathering. Results of soil flow are also visible in leaning trees ("drunken forest"), and lobate waves on hillsides. The minimum slope showing them is above 5 degrees. Final result should be rounded convex hills covered by a mantle of frost-weathered rocks. Final production of a "peneplain" is problematical. Where thawing is deeper or complete "thermokarst" topography with sinkholes, dry valleys, cracks, and depressed areas is very prominent. "Cave-in lakes" are abundant in some areas underlain by silt and their outline in altered by wave work. South-facing slopes and the margins of sandy terraces show little frozen ground. Since most water comes from the south-facing slopes valleys are asymmetric. The sides of terraces next to the hills are eroded producing a slant down toward the margin of each terrace. Vegetation gives a clue to the depth to frost. Black spruce and tamarack may indicate as little as 2 feet, paper birch from 3 to 8 feet, poplar and balsam over 6 feet, willow and aspen 10 feet or more, white spruce 1 foot for each 10 feet of height. Much of the permafrost melts once the vegetation is removed suggesting it is a survival of past climate. See figure 3, p. 7

Past permafrost: Where the permafrost has melted completely along the south margins of the present areas and on the lower slopes of mountains some evidence of its former presence is left. Although the trenches may be filled with silt, sand, or gravel a cross section will display them. Some geologists have suggested that certain soil mounds, such as the Mima mounds of the outwash plains of the Puget Sound region, are relics of permafrost. In the Matanuska Valley of Alaska irregular hummocks are left when the frost melts as a result of cultivation of the ground.

Caution in interpretation of melted permafrost: Despite the fact that climate has indubitably changed in much of the world, certain cautions are necessary in the search for former permafrost phenomena. Many have assumed that of necessity the "periglacial" climate was much colder near to the margins of the continental glaciers than it now is. Granting that air drainage from the high continental ice caps might bring about many periods of cold winds, it is fair to recall that due to the large amount of heat needed to melt ice glaciers could (and still do) terminate in climates where growth is impossible. Moreover, descending winds are warmed by compression. To melt a glacier must require a warmer climate than required for permafrost. Such a conclusion may, however, apply only to the conditions which led to the final melting of a glacier, not to the climate during its growth. However, climate favorable to heavy snowfall is unfavorable to production of permafrost. Let us also recall that much mass movement perhaps forming cracks as well as folds of glacial drift undoubtedly took place because of the high water content when first deposited. Besides, mass movement of mantle rock undoubtedly takes place without the aid of underlying impermeable permafrost. Relics of ice wedges may also be confused with weathering along the courses of former tree roots, or cracks due to mass movement of mantle rock. Certainly neither all masses of crept material nor talus deposits demand the former presence of permafrost. The "stone rivers" around Baraboo, Wisconsin, are the result of present-day erosion by water from melting snow removing the finer material from the residual mantle above the impermeable quartzite. A few, which are higher in the center than at the margins, might, however, be true "rock glaciers". But when we find that the snow now lasts longer among the rocks than elsewhere does this idea demand a much different climate? Possibly the mantle of rocks and clay might have crept more rapidly when the climate was wetter than now, but the impervious quartzite bed rock could have taken the place of frozen ground. The hypothesis of permafrost origin of the mounds in Washington does not fit well with either the marine climate of today or their localization on well-drained sand and gravel outwash. Somewhat similar mounds in Oklahoma and other southern states have been explained by the hypothesis of former drying rather than of permafrost. Supposed "convulutions" in the surface of outwash plains in Illinois might be due to ground water work during soil formation. Some found in sand and gravel in northeastern Wisconsin are almost certainly due to shove by a glacier which left so little till that the soil-making processes have rendered it now unrecognizable. The "mettled ground" of the same region occurs in red glacial till which appears to overlie older end-moraines. The mottling is due to small knolls and ridges which show out in the air photography because the high spots photograph a lighter tone than do the damper hollows. It is probably a phenomenon of compaction of an irregular thickness of red glacial till with a high content of silt and clay. We must keep in mind the fact that the condition, which in permafrost regions lasts all summer, occurs almost every spring in lower latitudes and that in certain snowless winters frost penetrates to considerable depths.

See p. 2 for key to technical terms suggested by Bryan.

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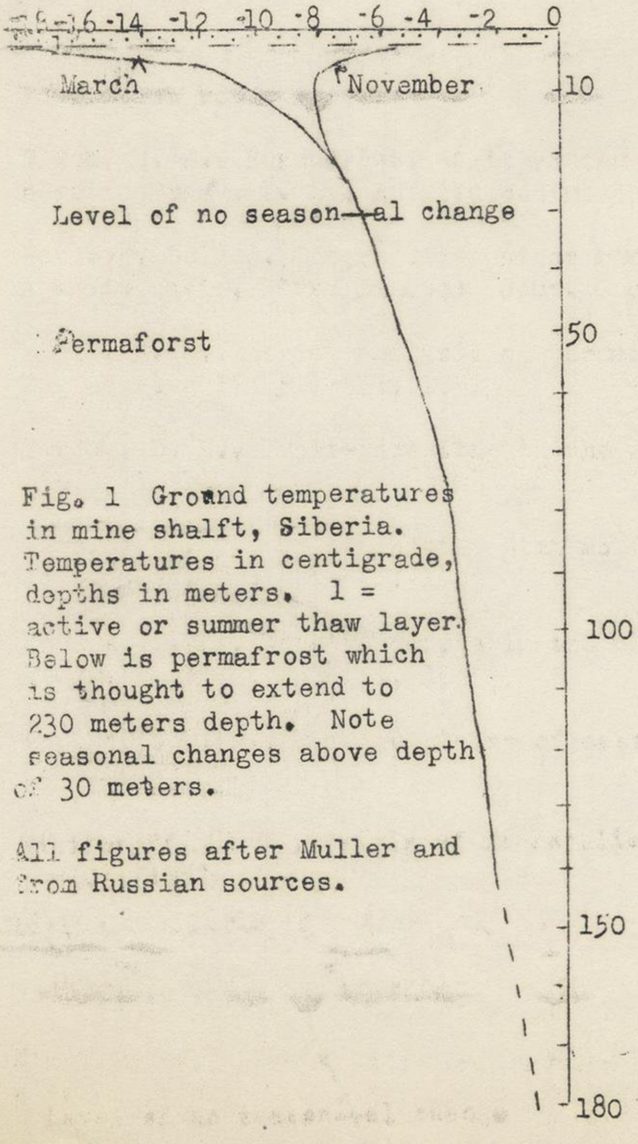
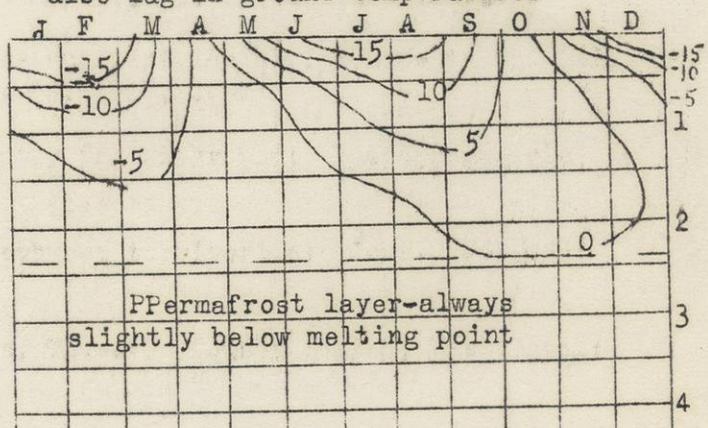


Fig. 1 Ground temperatures in mine shalft, Siberia. Temperatures in centigrade, depths in meters. 1 = active or summer thaw layer. Below is permafrost which is thought to extend to 230 meters depth. Note seasonal changes above depth of 30 meters.

All figures after Muller and from Russian sources.

Fig. 2 (below)
Seasonal temperatures of ground, Siberia. Months shown by initials. Temperatures in Centigrade. Note that melting temperature reaches top of permafrost for two months, also lag in ground temperatures



Depths in meters

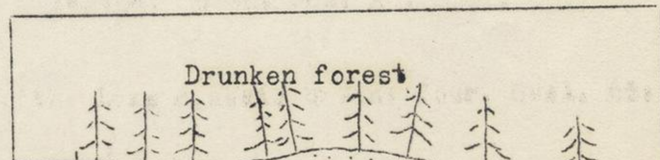


Fig. 3 Origin of pingo. 1= refrozen layer 2= water-bearing (artesian) layer 3= intrusion of water which later freezes and breaks through to surface in crack.

Since the original text was written there have been several important developments in the submarine canyon problem. First: funds have been available for much more deep water sounding both for navigation and for scientific interest. Second: methods of taking cores of unconsoldiated sea-bottom materials have been much improved. Third: movement of water due to the increase of density by sediment content have been observed and studied. Fourth: several deep drill holes have been put down on coral islands.

Soundings:

Recent sounding expeditions, using the acoustic method, have extended some of the submarine canyons scores of miles from where first observed into very deep water. For instance, the Hudson canyon has been traced about 200 miles out to sea, and according to newspaper reports, a recent voyage demonstrated a system of submerged valleys in the North Atlantic comparable in size to the Mississippi system. These valleys in the ocean bottom are by no means as spectacular as those in the continental slope and are possibly more like channels of rivers on a floodplain in being bordered by ridges which are larger and more massive than any natural levees on land. Nevertheless, they are distinct channels unquestionably due to some kind of flowing water. Besides the channels many flat-topped submarine mountains (guyots) have been discovered. There is little regularity in the depth of water above these.

Sediments:

The deposits off the Hudson canyon have been most fully described although scattered cores have been taken over a wide area. On both sides of the Hudson channel there is an extensive sand deposit with some associated layers of gray calcareous clay. In the channel itself there is gravel with pebbles up to 15 mm diameter. The sand is very well sorted but its otherwise much the same as the sand on the continental shelf. Interbedded with the sand are layers of normal red non-calcareous deep-sea clays. The pebbles can be matched with rocks which outcrop in the steep sides of the canyon. Opinion seems to now be trending to surf-zone erosion of the continental shelf and final desposition of clastic sediments at the foot of the continental slope.

Density currents:

Studies of sedimentation in reservoirs have demonstrated that the mud-laden water which enters at the head sinks to the bottom and flows along it to the dam. The rate of flow has been measured and seems to fit with a modified form of the formula for turbulent flow in open channels. The Chezy formula has been modified by introducing the excess of density of the water above unity and by changing the constant. Measured velocities in the low-gradient bed of Lake Mead, in the Colorado River Valley, strongly suggest the possibility that similar currents in the steeply-sloping submarine canyons could readily erode the bottom thus adding to their velocity. Trenches with marginal ridges due to sinking muddy water have long been known in Lake Geneva, Switzerland.

Evidence of subsidence of ocean bottom:

In recent years very deep drill holes were put down on several atolls of the southwest Pacific. On Bikini shallow water deposits occur to a depth of over 2500 feet. On Eniwetok volcanic rock is encountered below 4100 feet. These tests demonstrate that the ocean bottom has actually been subsiding but the age of the lower marine deposits runs back to early Tertiary . Whatever the cause, the change of sea level has been very slow (see section on coral reefs). The flat-topped peaks found in deep water, 3000 to 6000 feet, whose form suggests wave-eroded volcanoes, are in line with this conclusion.

Johnson's Theory:

Johnson's conclusion that submarine canyons are due to the recession of fresh water springs emerging from permiable formations of the Coastal Plain is now thoroughly discredited, because first, it disregards physical principles and second, the presence of canyons in impermeable rocks. Fresh water would rise directly to the suffface and could not make a canyon.

Density current theory:

The failure to demonstrate marked density or other currents in existing canyons of the continental slope is explained away by two suggestions. Daly originally proposed that during the moderately lower sea level of glacial times much more sediment was carried over the edge of the continental shelf than is now the case. This suggestion is supported by distribution of coarse sediments out to the margin of the shelf in many localities. On the California coast it has been suggested that alongshore transportation of wave-derived material is blocked at certain headlands. If , in such locations, the slope of the bottom is steep enough, a canyon is eroded by a descending density current. A further suggestion is that earthquakes loosened much sediment causing density currents down previously formed depressions. Under the density current theory no great change of sea level is required for the formation of submarine canyons. The sands and gravels now at great depths were thus transported from nearer the surface and interbedded with normal deep water deposits. Currents which spread out from the major channels deposited adjacent ridges.

Emergence theory:

Although the density current idea has gained much support in recent years, advocates of great emergence of the lands have not been idle. Landes proposed a theory that, with a periodically shrinking earth, the ocean bottoms of heavy sima should subside first. As a major cause of such sinking, he suggested the solidifications of basalt magma to solid rock with a very large volume decrease. He thought that the lands of lighter sial would not sink at once but that when they did horizontal compression would result. Such marked contraction is certainly not proved by present knowledge of the physical stae of the earth's interior, but, if such a process is possible, no distinct limit could be set on the depth to which the ocean level receeded without loss of water.

Compromise view:

Shepard offered a compromise view which is intended to avoid some of the above difficulties. He thought that the shoreward portions of the canyons, down to perhaps only 1000 feet depth, were eroded by streams during continental uplifts, not necessarily all at the same time. This is the only portion which has been examined by diving, photography, and detailed sounding. The lower extensions of canyons, which are extremely steep in slope and irregular in grade with few, if any, tributaries, are charageable to density currents. This applies with especial force to the very lowest parts which are not canyons but troughs. There is no definite lower limit to canyons with a series of deltas as there should be were land elevation alone the cause. Some submarine valleys could then lead into enclosed depressions in the ocean bottom. Some canyons might be very old in time of first erosion, then filled with sediments, and later reopened by slides of the soft material. It may be added that some might have originally have been tectonic in origin with only superficial alteration by erosion. Some of the Pacific coast canyons seem to be in part cut into very young sediments. Sedimentation and mass movement of deposits has been observed in the heads of some of these.

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Quantitative examination of erosion topography

Quantitative examination of land forms is relatively new, partly because no attention was paid in the past to such measurements, but also to the inaccurate topographic maps formerly available. One of the simpler problems is that considered by Smith, the quantitative reappraisal of some of Horton's work on stream or drainage density and stream frequency. ($Dd=\frac{\Sigma L}{A}$ and $Fs=\frac{N}{A}$ where A=area, L=length, N=number) It was recognized that these two quantities, length of streams per unit area (Dd) and number of streams per unit area (Fs) are interrelated. Natural regulatory factors must be: climate, including rainfall intensity and vegetation, resistance of underlying rock or soil, infiltration capacity, relief, and stage of development of the drainage system. Investigation was long hampered because the smaller or first order streams were omitted on most maps. Smith failed to find all small streams even on the more recent large scale maps and hence used the maximum number of crenulations shown in a contour which lies in the middle of a slope around an individual basin divided by the length of the divide in miles. (N/P where P= perimeter) He calls the result <u>stream texture</u>. He obtained a weighted mean value by using the area of the total region measured. This is the sum of all individual basins multiplied by area of each basin x total number of streams per mile of perimeter divided by total area.

$$T_m = \frac{\Sigma(A \cdot N/P)}{A}$$

Smith suggested that a value to T_m below 4.0 is "coarse;" a value between 4 and 10 is "medium" and one over 10.0 is "fine" texture.

He found in different areas and climates a range from 2.1 to 17.83. He compared the drainage density total (L/A) with texture (N/P) and found that within the natural limits of uncertainty for the 54 areas examined, the former is just about 5/3 the latter. ($Dd=1.657(\frac{N}{P})^{1.115}$)

Strahler made an extensive study of slopes and noted that the middle of most hillsides is a straight line, although the divides are rounded. Concavity at the base he found was only present where material had accumulated through either slides or slopewash. The general impression of concavity he declared was an optical illusion due to viewing spurs which do not trend in the direction of maximum declivity. He found that slope angle is essentially a constant under given relief and rock character. Slopes represent a steady rate of removal of debris in relation to rate of supply. It is related to grade of the channels which remove the material from the bottoms of the slopes. This is a condition of equilibrium in what is known as an "open system" in which there is a dissipation of energy. In this system potential energy of position is transformed into kinetic energy of moving water and debris. The supply of energy must diminish as the relief is lowered by long erosion so that the topographic forms must change with time. The slopes are actual where several processes of alteration are present, thus distinguished from theoretical slopes due to one method of formation only.

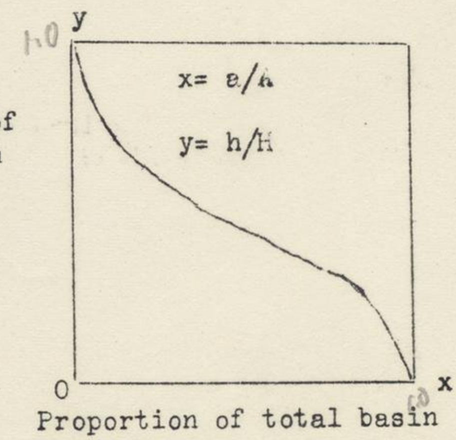
Profiles were measured directly down the steepest slopes from the divides. Only valley wall slopes which led down to a channel were measured. Divide slopes were not considered. Measurement was made with Abney level or Brunton compass. It was found that map measurements on the newer scales of 1:24000 or 1:25000 are satisfactory where drainage texture is suitable and that as few as 25 readings will give a satisfactory mean value. No maps gave good results with very fine-textured drainage. All results were averaged and considered from the statistical standpoint to find the expectable error. It was found that within a given area where the bed rock, soil, vegetation, climate, and stage of erosional development are similar the slopes are closely the same. The result is related to drainage density, relief, and slope-profile curvature, that is the relative area of convex compared to "constant slopes." Since length of slopes between adjacent channels is twice the drainage density it follows that, neglecting convex divides, the tangent of slope is approximately twice the relief multiplied by the drainage denstiy. tan S = 2VD$_d$, where V = relief, and D$_d$ = drainage density. He also found that there is a close relationship between slopes of valley sides and those of the adjacent stream channels. S$_c$ = channel slope and S$_g$ = ground slope, the ratio being S$_c$/S$_g$. Stream slopes were obtained from maps. The ratio was found by plotting on log-log paper and varies from about 0.2 for low slopes to about 0.5 with higher slopes, following the general equation; ground slope = 3.98 x channel slope 0.8 ($S_g = 3.98\ S_c\ 0.8$).

Strahler suggested the following tentative classifications: (1) High-cohesion slopes where the underlying material is dense clay or rocks such as granite, schist or gneiss. Slopes range from 40 to over 50 degrees. Mass movement of mantle rock is common in these areas.
(2) Repose slopes where the rock debris is loose fragments or talus (Wood's constant slope), sometimes called gravity slopes. Average slope is 30 to 35 degrees.
(3) Slopes reduced by wash and creep in areas where channel gradients are low. Here material lies at less than the angle of repose. These have been termed wash slopes and range downward from 20 degrees to about 1½ degrees.

Some studies included differences due to sompass direction of slope, which is reflected in effect on vegetation and soil. Although the appearance of the slopes differs considerably, actual measurment disclosed very nearly the same angle. The old question of parallel retreat versus decreasing angle was also considered. The correlation of stream slope with hillside slope expounded above definitely supports the latter concept. The only exception is there a slope has recently been undercut at the base thus maintaining a constant slope angle.

A much more elaborate system of topographic comparisons has been devised by Strahler consisting of area-altitude relations, the distribution of mass within a given drainage basin. A cumulative curve is prepared. It is evident that ratios are percentage or dimensionless numbers. In preparing this curve the datum point is take at the junction of the stream which drains the basin under study with another stream. Two ratios are computed: (1) , the area between a given contour and the divide to A, the total area of the basin (abscissa) and (2) ratio of height of contour to total height of basin (ordinate). These figures must vary from 1.0 to 0.0 and the resulting curve must pass through both the upper left hand and lower right hand corners of the diagram (x = 0, y = 1.0 and x = 1.0, y = 0.) After plotting and drawing in the curve the volume above base elevation is found by integration. Total volume = sume of segments (or slices) between successive contours times contour interval, or in mathematical terms, the intergral between base and summit elevations of a.dh. Both sides are then divided by the total area, A, and by the total difference of elevation, H.

Proportion of total basin height



Proportion of total basin area

Figure 1 (after Strahler)
a = area of slice A = toal basin area
h = elevation of slice above lowest part of basin
H = total difference of elevation in basin

Now in the diagram x = a/A and y = h/H. Dimensions are eliminated and the ratio of volume to product of H times A becomes the intergral of x/dy. Such an intergral cannot be determined by mathematical means but it is the ratio of area under the curve to area of entire square and can be measured on the diagram. Strahler, however, uses an approximate formula which he says conforms fairly well. Since this equation does not express any definite principle it is here omitted. It was used to draw a number of families of curves with different shapes and position of the point of inflection, the change in direction of curvature. This point does have morphological significance because it measures the position where the rate of decrease of mass with altitude changes from increasing to decreasing. The relation of the hypsometric curve to actual slope of the ground is not simple because it was built up on area, and not on lengths of contours at different elevations. A formula to take care of this problem involves measurment of length of longest contour and of the contour where slope is desired, a method obviously much more cumbersome than that of finding slopes directly from the spacing of the contours. Applications of Strahler's method to the geomorphic cycle appear to offer promise. In the early, youthful, or "inequilibrium" stage of development of valleys the curve is high showing a large percentage of material yet to be eroded (high value of the integral). Figure 2.

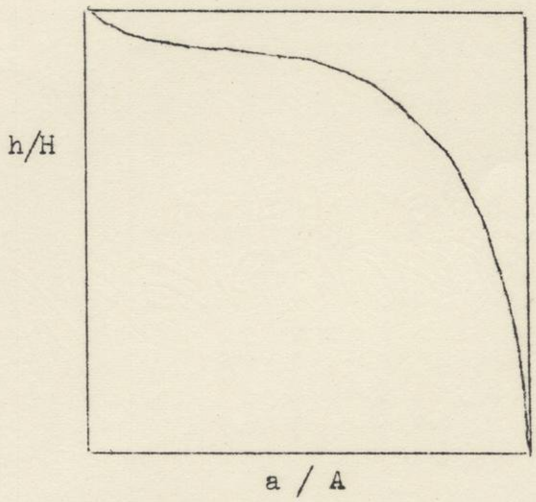


Figure 2 (after Strahler) Percentage curve of area and altitude for area in stage of youth or "inequilibrium" showing large amount of material not yet eroded.

Maturity or "equilibrium" shows roughly half (perhaps 40% to 60%) of the original still to be removed and the curve is S shaped. Figures 3 and 4, below.

Figure 3

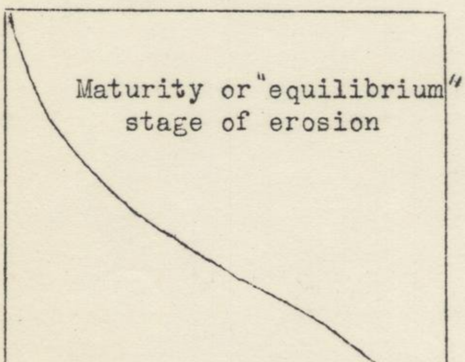
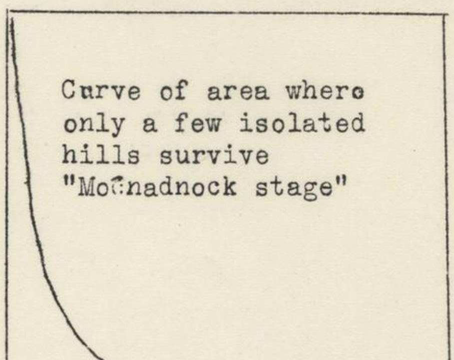


Figure 4 (both after Strahler)



No correlation between the hypsometric curves and either bifurcation or length ratios of streams could be distinguished. There is a marked relation between stream lengths and texture of drainage. Basins in nearly horizontal formations, whose resistance to erosion varies widely, have modified curves. Regions largely reduced to a plain have a strongly concave curve with a small amount of uneroded material.

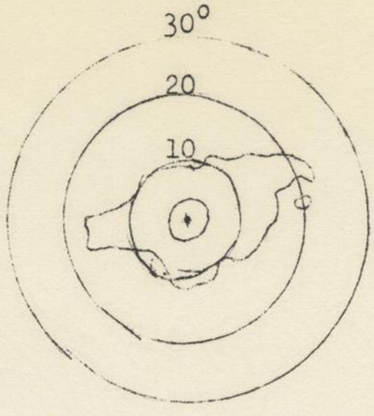


Figure 5 (after Chapman) Slope orientation diagram with contours of slope abundance at 2% and 10% only. Other disgrams show slope frequency of direction, average slopes, etc. North zt left in original diagram.

Strahler has modified Horton's system of giving stream orders. First order streams are unbranched, normally dry channels. Second order is only that part of a stream below the junction of two first order channels. A third order stream is below the junction of any two second order streams. This method removes the element of judgment and insures that a basin can have only one stream of the highest order.

Chapman presents a method for analysis of topography called the statistical slope orientation diagram. This diagram is similar to the petrofabric diagram showing orientation of mineral grains within a rock. Slopes measured in both amount and strike at many places on a map are plotted around a center or pole. This brings out the predominant direction of slopes and ignores altitude, at the same time eliminating the effect of minor or unimportant features of the landscape. Determinations may be made either along parallel traverses across the map or at corners of a rectangular grid. An ordinary slope scale is used to find the slopes from contour spacing. The diagram is a projection of a hemisphere. Since most points will fall within the 30 degree line plotting is relatively easy for angle is almost directly proportional to distance. A system of dots at varying compass directions and distances from the pole results. For contouring relative distribution of dots a hole or circle whose area is just one percent of the total area of the hemisphere is moved over the diagram and the numbers of dots witin counted at intersections of a définite grid pattern. Marking the number at each place contours are then drawn in the uaual manner unsing a contour interval expressed in percent. The resulting diagram is intended to bring out predominant slope angles and slope strikes of the area. The method is no more accurate than the map on which measurments were based plus inherent errors in doing this laborious task.

After some of the involved methods outline above it may be a relief to turn to an almost-forgotten attempt by Thompson to appraise evidences of cycles of erosion in a quanitative manner. The area he selected lies in Virginia and West Virginia. The evidences used by former students of the area included summit levels and ridge crests on which most place great importance. Long erosion in a region of diverse rocks ought to produce summit elevations at varying levels depending upon both rock character and distance from drainage lines. Summit maxima may occur a several different altitudes, hence any preponderance of summit levels does not prove past penepanation. Thompson prepared a curve with elevation horizontal and ordinates proportioned to area of summits at the several elevations. See Figure 6, next page

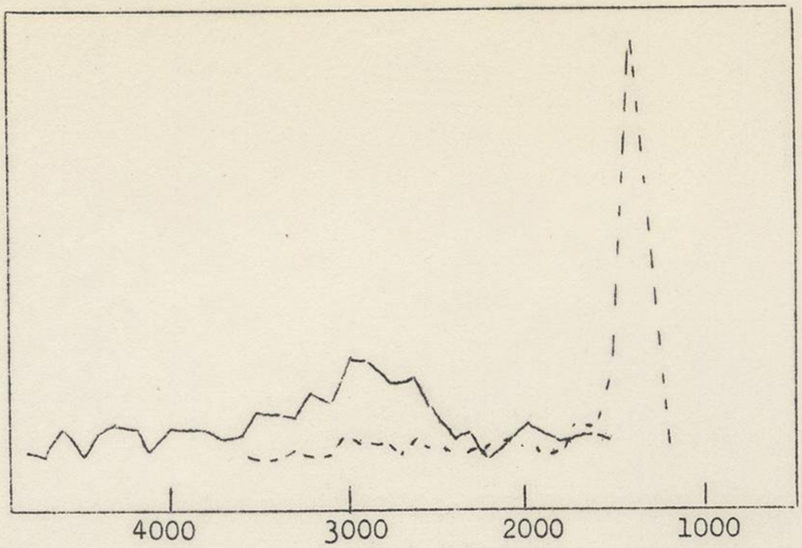


Figure 6 (after Thompson)
Number of summits at different elevations in two quadrangles

Although the older maps were not accurate this nevertheless gave a general picture which it checks with newer large scale maps. Tables with number at each level were also prepared showing the geological formations on which the summits occur. Since there is a maximum number and area of summits at intermediate altitudes the conclusion could be drawn that this is either an intermediate level of peneplanation or else a normal effect of prolonged erosion in diverse rocks. A similar study was directed to the ridge crests, finding that ridge tops occur at many levels. Possibly some of the highest major ridges are remnants of an east-sloping erosion surface, but, if so, the lower ridges have been varyingly reduced from that state. There is no relation between altitude of the ridge crests and notches. Some of the highest ridges are very knobby whereas some low ridges show even crests. Side slopes of the ridges were also studied but results indicate little. Elevations of ridge crests on the flanks of eroded anticlines suggest that erosion has been continuous since the resistant formation was breached. Hypsographic curves were also drawn where elevation is platted against area. Following Woolridge and Morgan it is stated that a region should show a concave curve if erosion has been constant. Old planation surfaces or platforms should cause convex irregularities.

The curves for the three quadrangles are similar in showing (Fig. 7) a

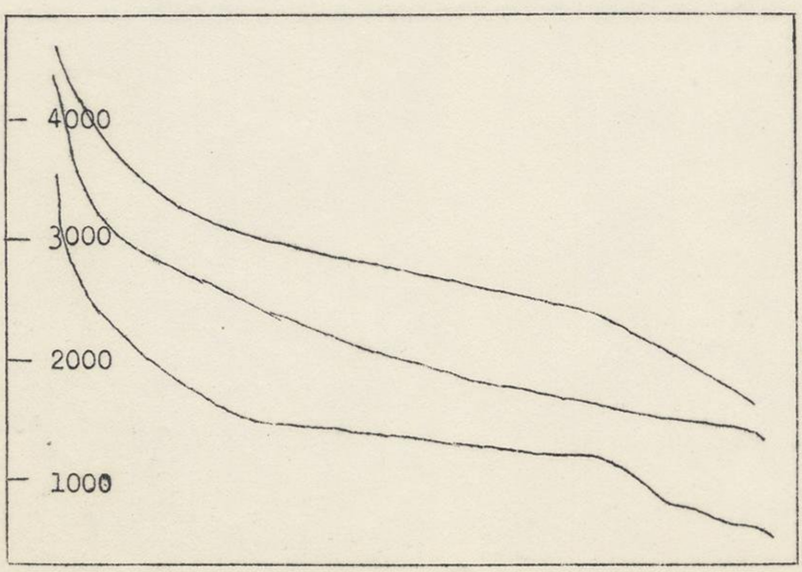


Figure 7 (after Thompson)
Area-altitude curves for three quadrangles. Compare with Strahler's curves which use no dimensions but are simply percentage distribution.
The preponderance of elevation at intermediate levels is shown by the "shelf" at level of the "Harrisburg" surface.

bulge at middle elevations which was thought to indicate an imperfect planation in the weak rocks of the valleys only, the "Harrisburg" surface. In conclusion it was decided that evidence for a summit peneplain is very weak but that for a halt in erosion at an intermediate level is good. The unreliability of many of the commonly-used criteria for past peneplanation is emphasized.

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Outline--Running Water

Mechanical principles
- Methods of flow-laminar (viscous or streamlined); turbulent; shooting; mixed.
- Expenditure of energy-resistance = force
  - Force exerted on bottom
  - Velocity formulas

Transportation of debris
- Force needed to move a particle-competence
- Methods of movement of debris
- Quantity of load-capacity
  - Competence and capacity formulas-experiments
- Solution

Erosion
- Slope, sheet wash, or overland flow-formulas-infiltration
- Channel formation, rills, ravines
- Sheet erosion-resistance to erosion-belt of no erosion
  - Profile of sheet erosion
- Channel erosion-cut banks, bottom erosion, profile and cross section, crossings, deeps, etc. kind of bottom.
- Distribution of turbulence-sand bars

Stream patterns
- Origin of drainage systems-rill modification, headward growth
  - Form of drainage basins
- Stream orders-law of numbers, lengths, slopes, spacing-texture or density growth of system-limits
- Horizontal form of stream channels-components of motion, forward, lateral, downward. Long profile of stream channels, floodplains, natural levees, deltas
  - Trading of debris across bends
  - Brading-meandering causes and controls
- Modification of stream channels
- Drainage development-terms applied to streams, consequent, etc.
  - Stream capture-subsequent courses. Controls.
  - Water gaps-wind gaps

Effects of earth movement on streams.
- Superposition
- Alluviation-outwash-effect on tributaries
- Uplift-nickpoints, terraces, intrenched meanders, misfit streams

Landforms due to stream work
- Relation of stream patterns to bed rocks
- Relation of valley sides to bed rock
- Summit levels-quantitative study

Cycle of erosion
- Concepts-parallel retreats vs. declining angle of slopes; speed of uplift vs. erosion
- Youth, maturity-old age
- Peneplain-processes
- Pediplain-processes
  - Evidences of former surfaces

Some principles of soils mechanics important to geology and geomorphology.

Introduction. Soil mechanics is a branch of engineering which has to do with those physical properties of unconsolidated materials which are important in engineering operations. The term "soil" is here employed in the sense of all mantle rock regardless of depth or origin. "Mechanics" refers to the resistance of these materials to either fracture or compaction (settling or consolidation.) Fracture or other movement of the material is termed "failure". It is evident that the properties are related to several geomorphic processes, for instance the slumping of wet glacial drift, and land forms due to mass movement of unconsolidated material. Moreover, the engineering determinations are a valuable tool in the description, correlation, and history of the surficial materials of the earth. Since geologists are frequently consulted in relation to engineering problems in subsidence, excavating, and mining it is very important to understand these relatively new tests and physical measurements.

Geological descriptions. In the past geologists have to a large extent ignored physical properties of unconsolidated materials. Their descriptions have been almost entirely origin, particle-size distribution, mass chemical analysis, and to some extent mineralogy. It is evident that origin is too general to furnish much help in most problems. The second is known as mechanical analysis and consists in screen separation of the particles down to a diameter of about 0.07 mm. The smaller diameters after dissociation by use of a strong alkalie are placed in suspension in water. Use is made of the known rates of settling and the density of the mixture to find relative proportions of different grades. Results of such analyses are presented in various kinds of diagrams. Prior to the development of X-ray examination and the electron microscope, mass chemical analysis was the only possible tool for examination of the sub-microscopic particles. Attempts to apportion the elements reported by the chemist into minerals were most uncertain. Now the mineralogy, shapes, and arrangement of the small particles is much better known. Their diameters are often expressed in microns or thousandths of a millimeter. The shapes of those smaller than about 2 microns cannot be seen with the ordinary microscope. Most of the small particles are flaky and are lumped together as clay minerals. Particles smaller than 0.1 micron are termed colloids and possesses peculiar properties which, together with those of other small particles, influence the physical nature of the entire mass to an extent out of proportion to their quantity. One of these properties of colloids is a negative electric charge which attracts the hydrogen of water molecules. The resulting layers of adsorbed water contain the ions of electrolytes. These products of dissociation of molecules react with one another causing the phenomenon of base exchange. Much of the void space between small particles is filled with adsorbed substances. Both cohesion and plasticity are properties due to colloids and the physical arrangement of the small minerals varies widely with the state of consolidation due to pressure.

Soils mechanics determinations. It is evident that the ordinary geological description of a mantle rock which contains a large proportion of fine particles leaves much to be desired in knowledge of its physical nature. For this reason engineers have used a wide variety of other determinations. Those most commonly measured comprise: (a) bulk density or unit weight (in gm/cm^3 or lbs./ft^3);

(b) voids in percent of volume of solids; (c) water content in percent of dry weight; (d) <u>Atterburg limits</u> which consist of <u>plastic limit</u> or percent of water of dry weight, at which crumbling ceases, <u>liquid limit</u> or percent of water of dry weight at which flow begins under specified conditions, and <u>plasticity index</u> or difference of these two; <u>permeability</u> or rate of water movement through the material under specified conditions as a coefficient; <u>shear strength</u> under standard conditions (gm/cm^2); <u>unconfined compressive strength</u> similar to the measurement on firmer material (gm/cm^2); <u>cohesion</u> determined from compression with sides under pressure (gm/cm^2); <u>compression rate</u> as tested with force applied to one end of a cylinder; and <u>precompression limits</u> an estimate of apparent compression (kg/cm^2 or tons/in.^{2}) of the material earlier in its history, or prior to being brought to the surface. (In reports of these tests it is important to note that kg/cm^2 is almost exactly equal to short tons/in.^{2})

<u>Plasticity</u>. We do not need here to detail the arbitrary standards which have been set up to make plasticity measurements but their relation to the origin of the clays is important to geology. When liquid limit is plotted against plasticity index on ordinary coordinates all results on the same kind of clay from the standpoint of origin fall either on or close to a straight line. The slope of lines for different clays does not vary much (Fig. 1). Clays which

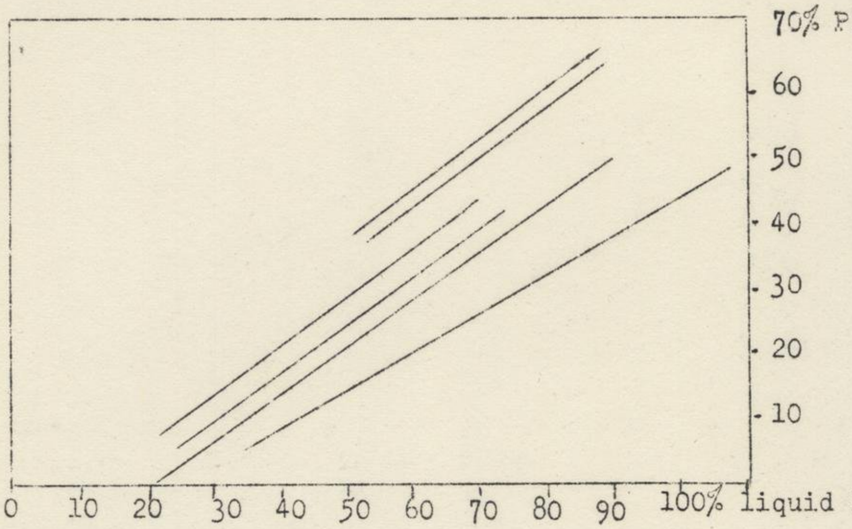

Figure 1 Relation of liquid limit to plasticity index. Redrawn from Casagrande Quanities in percent water of dry weight.

contain sodium require much more water to become plastic than those with minerals containing calcium or hydrogen. It is also to be noted that in liquid limits we have an approximation to the point at which clays become similar to liquids. Considerable difficulty is found in duplicating these tests and different laboratories do not always agree.

<u>Strength tests</u>. Long ago the strength of unconsolidated material was expressed by <u>Coulombs equation</u>; shearing force = cohesion plus force times the tangent of the "angle of internal friction". S = c+ p tan phi. In the case of a sand which is dry and shows no cohesion the angle phi is the <u>angle of repose</u> at which the material will rest. This angle is about 34 degrees in dry sand with angular grains and slightly less when the sand is below water. As in a talus, the sand is held together by internal friction. When a finer material than sand is below water the value of p is reduced by the amount of

pressure of the water. Unconfined compressive strength is readily measured on a cylinder of cohesive material. It ranges in clays from .25 to about 4.0 kg/cm^2. The values of cohesion and of phi are not so easily determined. Some tests have been made by finding the force needed to break a cylinder of undisturbed material by sliding one half of a containing box over the other. Another method is to enclose the specimen in water-tight flexible cover. It is then immersed in a liquid which can be put under pressure before force is applied to one end. Pore water may or may not be allowed to escape from the container. Pressure is applied until the specimen fails. This type of test is known as <u>triaxial</u>. The confining force or pressure is plotted on the horizontal line of Fig. 2. The value of the force at failure then lies to the

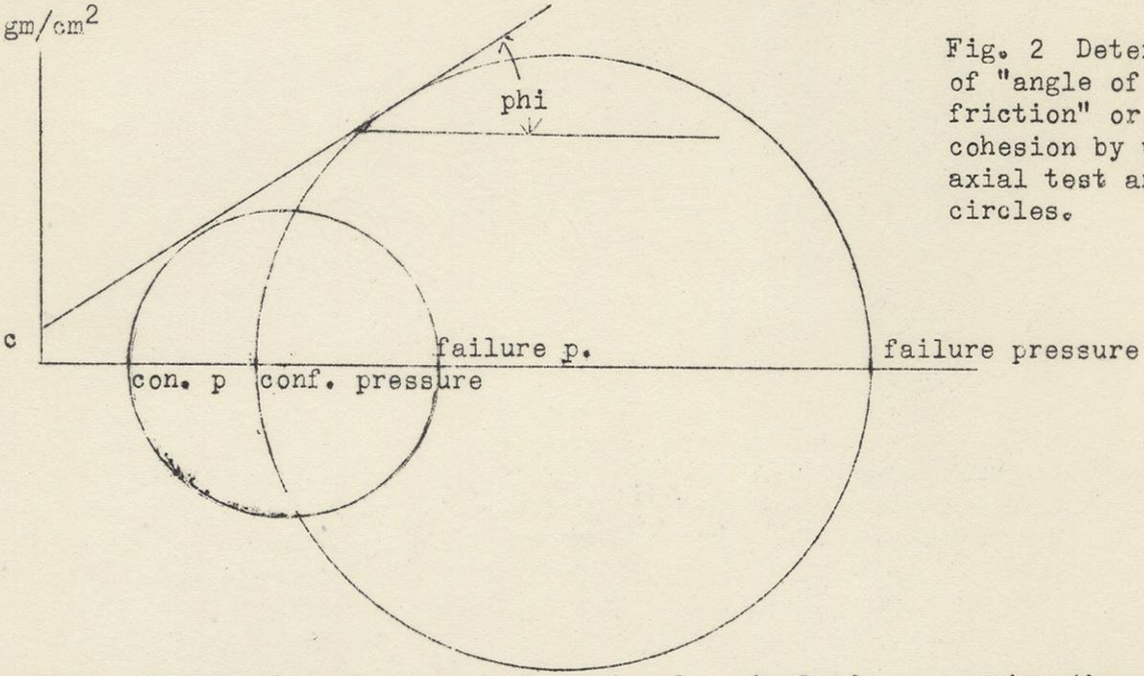

Fig. 2 Determination of "angle of internal friction" or phi and cohesion by use of triaxial test and Mohr circles.

right. The distance between is halved and a circle drawn passing through both points. This is known as the <u>Mohr circle of stress</u>. If the proceedure is repeated with another specimen of same material and larger stresses a second circle can be drawn. Then a line is drawn tangent to both circles. Its slope from the horizontal then determines the value of phi and the distance of its intersection with a vertical line through the origin at left measures the value of cohesion. However, a commonly used value of shearing strength for soft wet clay is half the unconfined compressive strength. Specimens must be from cores, not cuttings.

<u>Compression</u>. The phenomena of compression are measured by placing a short section of undisturbed core in a circular ring. Opportunity for escape of water is provided at the bottom and pressure is applied at the top. At first the rate of dimension change is rapid, then it slows up and, if the test is carried on far enough, would eventually cease. (Fig. 3). However, it is customary to plot percentage of voids against logarithm of pressure as in Fig. 4. This enlargement of the horizontal scale for small forces changes the curve so that the first part has a gentle slope which on increasing pressure changes to a straight line. Under this condition, rate of change of voids is inverse to pressure. This line may be extended upward in the diagram until it intersects the horizontal line representing the estimated original void ratio.

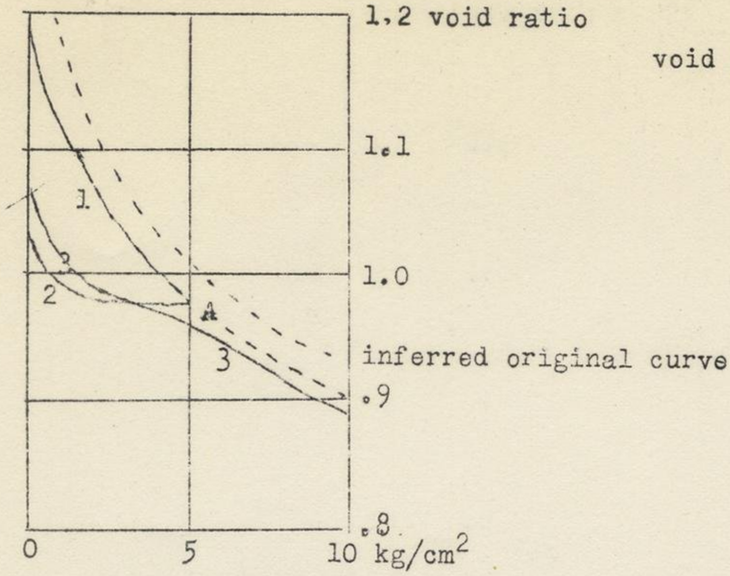

Figure 3 Compression test with recompression. Arithmetical scale Recovery shown by curve 2 void ratio = ratio volume voids to volume of solids.

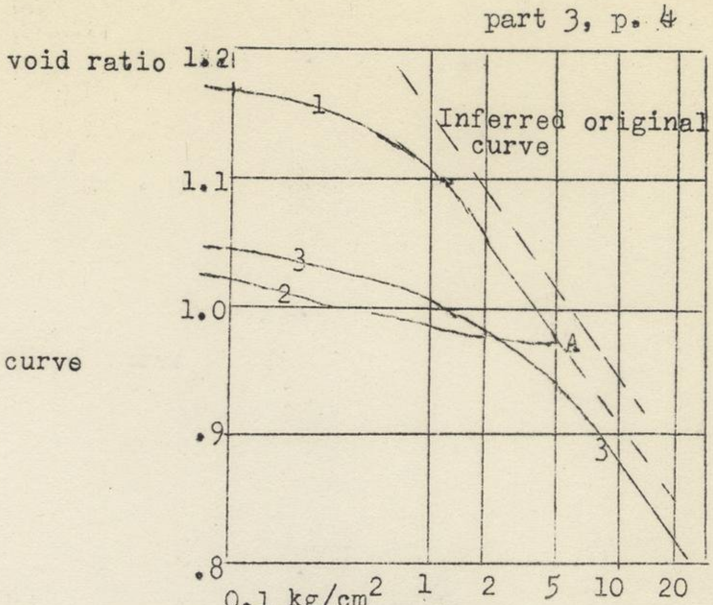

Figure 4. Same data as Fig. 3 with pressure to log. scale. Curves show first compression to A, recovery, recompression

Recompression. If after reaching the straight line portion of the graph pressure is gradually reduced the sample expands, although the original pore space ratio is not attained. Recompression then gives a curve (3) which has been displaced to the left when the straight portion is reached (Fig. 4). It has been claimed that the characteristics of this cur e enable the discovery of the point at which pressure was reduced in the first experiment. The procedure for finding an earlier stress is to first draw a tangent to the compression (or recompression) curve at the point of minimum radius as estimated by eye. A horizontal line is then drawn through the point of tangency. (Fig. 5) The angle between these two lines is bisected and the line from point of tangency

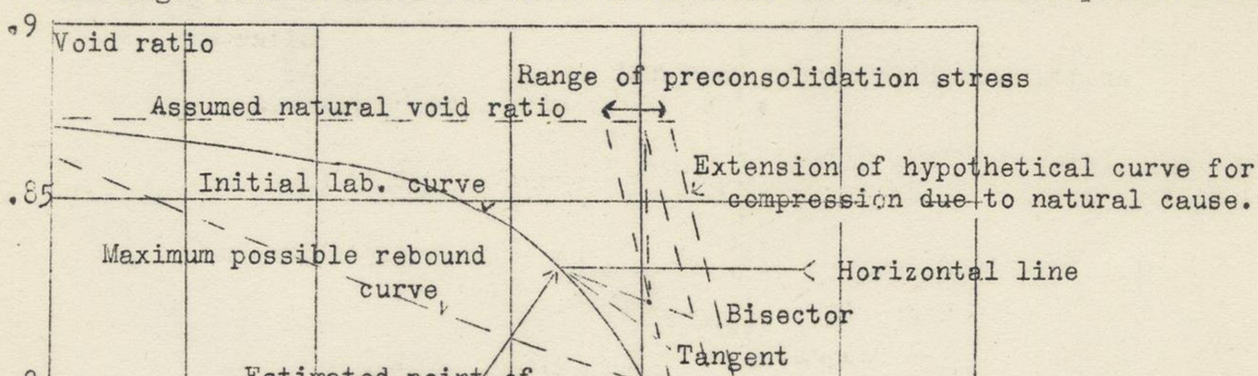

Figure 5 Casagrande's method of finding preconsolidation stress. After Rominger and Rutledge The middle on the three lines at right is often called the "Most probable preconsolidation stress". Note the displacement to left of curve in successive times of consolidation by pressure.

extended until it intersects the extension of the straight portion of the curve. A straight vertical line through the point is supposed to be the "original " pressure or stress, also termed the preconsolidation stress.

Precompression stress. Some investigators have used the above method to estimate the amount of pressure that a clay once sustained prior to either erosion of overlying material or melting of glacial ice. A load of water has no effect on precompression of a clay which it enters. If the method were always reliable it would afford a valuable tool to the geologist. Unfortunately, a very similar effect results from drying of a clay. Note that in the figure the straight portion of the final compression curve is extended upward until it meets a horizontal line drawn at the level of an assumed "original void ratio". Another line parallel to the final curve is also drawn which is supposed to be the maximum possible position of a curve if the specimen had been compressed when in its original condition prior to the deposition of any overburden. This also is extended until it intersects the line or original void ratio. The difference of pressure read on this line between this and the actual recompression (or compression) curve is then recorded as the "range of precompression stress". A mark is often placed to indicate the "probable value", as found in Fig. 5. It has been stated that this method often gives too small a thickness of eroded material. In samples taken from test holes or pits it may be checked with the load which rested on the specimen before it was brought to the surface. A marked consistent departure of the values of precompression stress from actual load is nevertheless a proof of either erosion of overlying material or former drying. The value of this range is in a sense a measure of the amount of compaction which the material has undergone. However, it seems as if it is based on too many estimates to ever be an exact determination.

Failure of slopes. One of the ever-present problems of engineers is how high and how steep is it safe to leave the side of an excavation in unconsolidated material. Geologists are interested in this problem in considering the natural reduction of slope of valley sides and the attainment of equilibrium in slopes. We must recognize at the outset that the physical conditions within a bank of "soil" may vary greatly by reason not only of its original chemical and mechanical make-up but also because of subsequent changes, for instance by weathering or percolation of water. Engineers use a number of different assumptions as to the strength of materials and the amount of pressure which tends to collapse a slope. One of these is that the shearing strength of coherent material in a bank is half the unconfined compressive strength. Fig. 6 shows

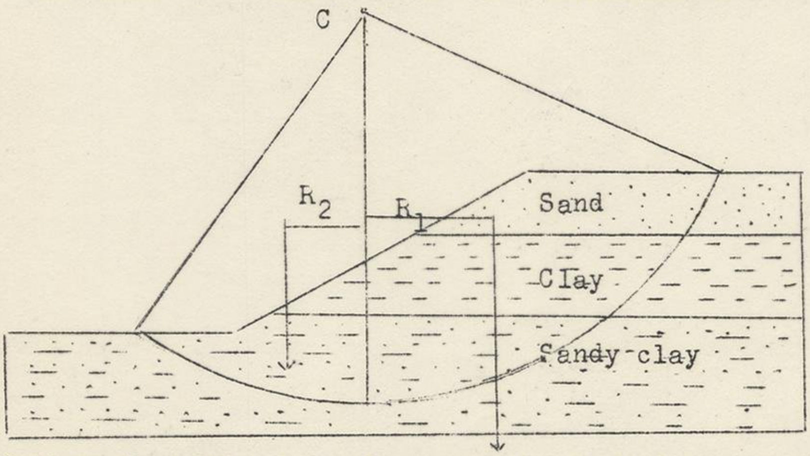

Figure 6. Adapted from Kaye. Center, C, and radius of circle must be assumed. Total force tending to move the segment to right of vertical line is its weight acting on arm R_1. Resting force is weight of other segment on arm R_2 plus shear resistance on the arc within the bank. Unit weight and shear strength of each material must be found, also length of circle in each.

Factor of safety is resting moment divided by disturbing moment. This allows a comparison of results with different centers and radii. The method affords little help in study of natural slopes because of the assumptions. Physical conditions within the bank may change with time and amount of water.

the computation by which the strength of a bank is determined. Here both the radius of a circle and location of its center are assumed. Then several circles are drawn with various changes in both these conditions. The moments of force due to weight of material which causes failure and that which resists it are readily computed from mass density (unit weight) in a section of unit width. The shear strength along the assumed circle of sliding is then computed for the different types of material cut by that circle. The total resisting force is then compared with that which might cause failure. The ratio of the two is the "factor of safety" and the structure is designed to keep this as great as is economical. It is evident that such analysis is not of much value to the geologist. It ignores all natural planes of weakness such as shrinkage cracks in clay. Fig. 7 presents a somewhat different analysis of the forces in a vertical slice of unconsolidated material of uniform physical state. Total

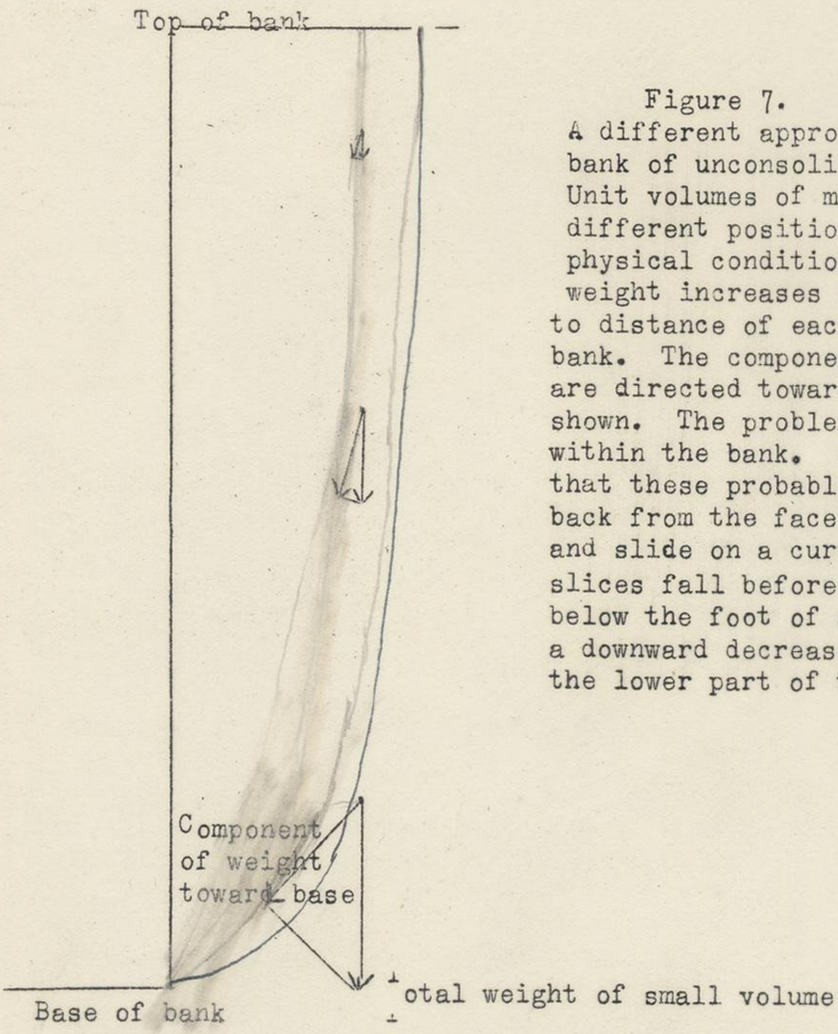

Figure 7.
A different approach to stability of a vertical bank of unconsolidated material or "soil". Unit volumes of material are shown in three different positions. It is evident that if physical conditions are uniform the total weight increases downward in direct proportion to distance of each point from the top of the bank. The components of total weight whish are directed toward the foot of the bank are shown. The problem is complicated by cracks within the bank. In the field it is evident that these probably determine the distance back from the face at which a slice will shear and slide on a curve. Often a number of such slices fall before stability is reached. Motion below the foot of a bank very likely indicates a downward decrease in shear strength so that the lower part of the slope behaves like a fluid.

weight increases directly with the height but only that component which is directed to the foot of the slope is important. As long as this does not exceed the shearing strength on a curved surface the bank is safe. Since the angle of the force is inverse to height of the bank a curve of distance back from the face proportioned to the logarithm of distance below top of bank should result. The unknown feature of this analysis is the factors which control the distance back from the top of the bank at which breaking starts. Possibly this is related to drying and to shrinkage cracks in clay. It is here assumed that shearing incoherent material has no relation to the angle phi and can take place in any direction. An involved derivation used in most text books of soil mechanics arrives at the conclusion that the safe height of a bank is four times the shear strength divided by the unit weight. Taking an unconfined compressive strength at 1000 gm/cm^2 and a unit weight of 2 gm/cm^3 this figures out at 4 x 500 /2 or 1000 cm (10 meters) as the safe height of a vertical face. Under the view taken above, the pressure on a square cm at depth of 1000 cm would be 2 kg/cm^2 (horizontal) and the component on a surface inclined about 45 degrees would be half this per square centimeter which is twice the assumed shear strength. As a matter of actual field conditions the problem in many cases defies analysis for the presence of water in the pores of a clay may greatly reduce its strength and the amount of such water may vary widely. Besides this, the above analysis neglects the fact that the shear strength is not surpassed throughout half of the probable surface of failure.

Conclusion. The subject of soil mechanics offers an important field for the advancement of knowledge of the nature of unconsolidated materials but considerable study, especially from the geological standpoint, is still required. The existing state of knowledge of "pre-compressed" clays leaves much to be desired. Fig. 8 is some data on an actual test hole where foundations settling had been excessive with geological interpretations. (Next page)

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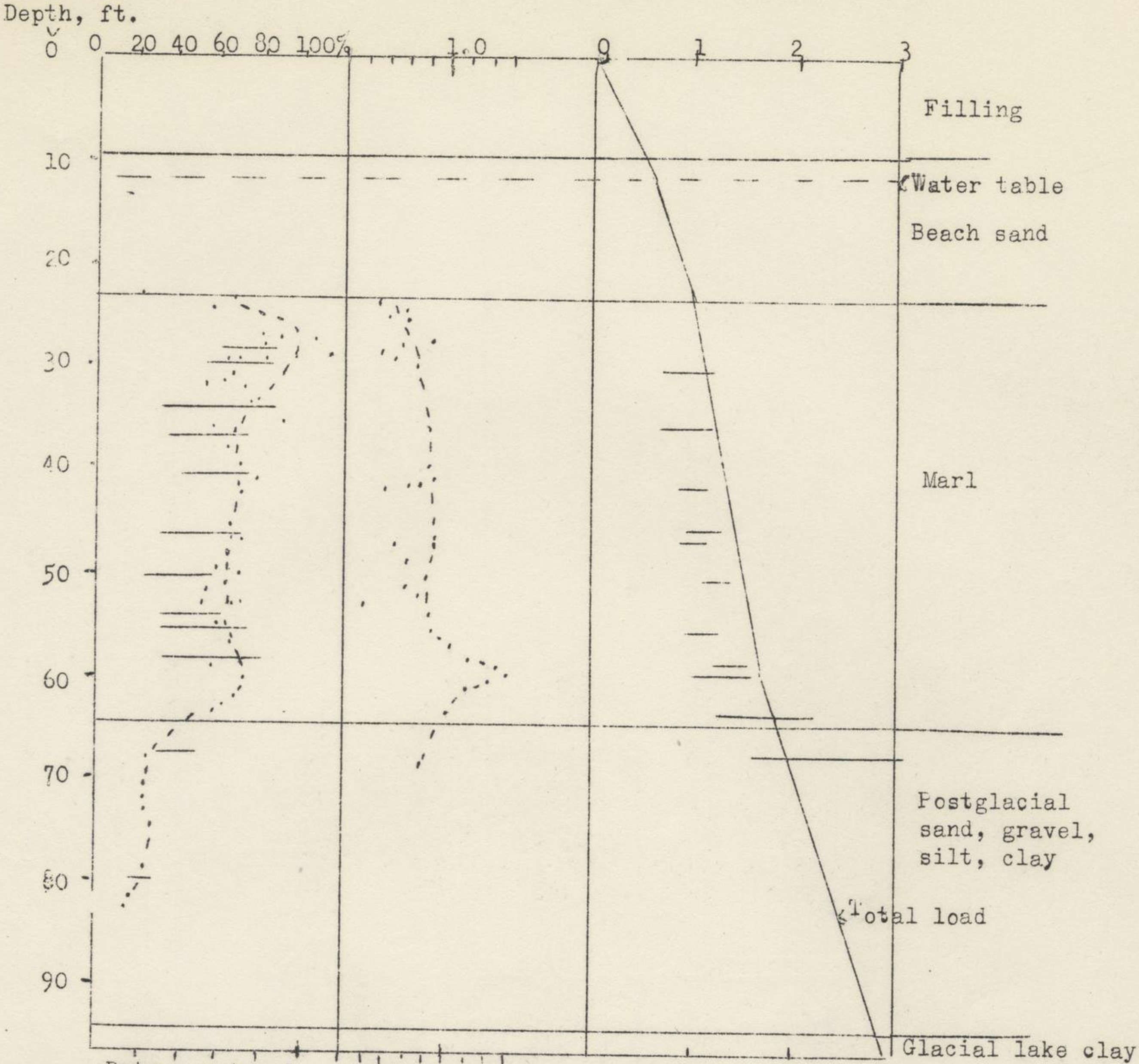
Figure 8 Soil mechanics determinations in test hole on Jones Island, Milwaukee, Wis. Data furnished by Klug and Smith, Engineers.



Dots = natural water content; Lines = plastic and liquid limits Broken line = average water content All % dry weight

Unconfined compressive strength, tons/ ft^2 Broken line= average

Range of preconsolidation stress = horizontal lines. Total load = solid line All in tons /ft^2

(3) Jeremy

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Notes on waves, reefs, etc. Supplement, 1952-53, part 4

Energy of a wave. Some features of wave work demand greater attention than in original manuscript. One of these is how the formulas for the energy of waves were derived. It is evident that wave motion involves both potential energy due to the position of the displaced water particles up or down from position of rest and kinetic energy due to the velocity of their motion, that is the work stored in displacing them. On general mechanical principles we would expect these to be equal. It may be urged, however, that the motion is not confined to the surface circles of rotation but dies out gradually below. Both derivations involve the integral calculus and are not "required." In each two separate steps are involved, first integration of the normal functions and, second, integration of the trigometric quantities which are brought in by rotation. We will here perform the steps separately and then obtain the final result by multiplying the results.

Definitions: w = unit weight of water (dynes/cm^3 or lbs/ft^3); L = wave length, crest to crest; h = wave height, trough to crest; x = horizontal displacement of a particle; y = displacement of points on surface above still water level; z = vertical displacement of a particle; g = acceleration of gravity; which must be taken into account whichever system of units is used. Energy will be given for unit length along crest.

Potential energy due to displacement of surface of wave to position y = h/2 cos 2pix/L is E_p

$$E_p = w\int_{x=0}^{x=L}\int_{z=0}^{z=-y} z\,dz\,dx \quad = w/2\int y^2\,dx$$ using above value for y this becomes: $wh^2/8\int_{x=0}^{x=1} \cos^2 2pix/L\,dx$ which is $L/2 + \frac{1}{4}\sin 4pix/L$ Second term is 0 at both limits and multiplying by L/2 we get $E_p = wh^2L/16$

Kinetic energy E_k, follows the formula of $\frac{1}{2}mV^2$. The double integral over a wave length L and wave height h is required. Now velocity in a circle of rotation is same, both vertical and horizontal and = 2 pi z/T the square of which is 4 $pi^2\ z^2 / T^2$ Now $T^2 = 4\ pi^2\ z / g$ by laws of harmonic motion. hence by substitution this simplifies to z g.

Now the E_k for unit of length = $w/2g\int_{z=-y}^{z=h} g\,z\,dz$ g will cancel out and we have $E_k = w/4 \cdot z^2$ But $z^2 = h^2/4$ hence E_k (unit length) = $wh^2/16$

For entire length of wave, L we must integrate the horizontal and vertical components squared. This is the sum of $\sin^2 2\ pix/L + \cos^2 2\ pi\ x/L$

Making the integration this becomes $x/2 + \frac{1}{4}\cos 4\ pi\ x/L + x/2 - \frac{1}{4}\cos 4\ pi\ x/L$ Sbustitution the limits of o and L, the trigonometic terms cancel and the final result is L. Multiplying result obtained above the entire result becomes $E_k = wh^2L/16$ confirming our assumption that it is equal to E_p.

Combining the two forms of energy total energy of a wave per unit of crest is $E = wH^2L / 8$ Note this is in terms of work and not of power.

Relation of wave height to time. Most discussions of waves do not consider length of time that the wind has been blowing. In regions of steady winds, such as the trades, this element is very important. It is obvious that when there are no waves, that is when the wind first starts to blow, the transfer of energy from wind to water will be at its maximum. As waves begin to roll with the wind this transfer must of necessity decrease. The diagrams given out in 1950 show the relations. Wave heights also decrease with increase of fetch for the same reason. Theories involve complex mathematical formulas, the practical importance of which is problematical.

Fundamental formulas are: $V = L/T$ or $V^2 = gL/2pi$ where V = velocity of travel, $L = 2\ pi\ V^2/g$ or $gT^2/\ 2pi$, $T^2 = 2\ pi\ L/g$ or $2\ pi\ V/g$ where T is time of period. Solving these to get results in seconds, feet, and land miles:

$V = 3.5\ T$ or $V^2 = 2.23\ L$, $L = .555\ V^2$ or $5.12\ T^2$ and $T^2 = .195\ L$ or $T = .346\ V$. Observation shown that wave height ratio to length is always less than 1/7.. The following relationships appear well established. For a fetch of 11 miles or more maximum height is about 1.65 times the square root of the fetch in land miles. For a given wind speed wave speed increases with fetch. Maximum wave height is about .9 of the wind speed in land miles, or h = 0.0344 times the square of speed in land miles. Average maximum wave speed slightly exceeds wind speed up to about 29 m.p.h. wind speed and is less than wind speed above. Time required to develop maximum height increases with wind speed. High waves can be formed by strong winds in less than 12 hours. For a given fetch and wind speed wave speed increases rapidly with time. There is no well-established relation between wind speed and wave steepness, for the latter depends upon stage of development of the waves. During early stages of wave development waves are short and travel at less speed than the wind. Height of swell (old smooth waves with no wind) decreases as swell advances.

Roughly, waves lose a third of their height each time they travel a distance in miles equal to their height in feet. Period of swell may increase with distance of advance although this is not proved. When the speed of waves is equal to that of the wind, no energy is transferred.

A factor not considered by the students of waves is that they increase the roughness of the surface and hence raise the height of 0 velocity of wind as found by Bagnold. Although it is possible to treat many of the phenomena of waves by mathematics, it is well to remember that in practice variations in wind speed and direction introduce great irregularities.

Reef phenomena. In the last few years it has been discovered that some ancient buried reefs are very productive of petroleum and hence more attention has been given to recent reffs in order to understand them. Important papers are those by Cloud and Ladd. The fact has been brought out that reefs are mainly composed of clastic particles. Some desire the substitute term bioherm to be restricted to organic accumulations of doubtful form. The rigid framework of a reef may be both corals and algae and only make up a small part of the entire accumulation. On the outside of the reef growth is most rapid since the supply of food is largest. It took nearly 200,000 soundings to map the lagoon at Eniwetok for there are many terraces, depressions, and knobs of living coral. Igneous foundations of reefs have been reached by drilling at Eniwetok and Bermuda. At Eniwetok basalt was found in one hole at 4170 feet depth. There is only a few hundred feet of Pleistocene reef underlain by Tertiary limestone, dolomite, carbonaceous clay, and silt. The oldest sediment is Eocene. On Bermuda seismic work demonstrates that the boring is on the flank of a volcano. It disclosed Pleistocne limestone to depth 380 feet, Miocene 380 to 590, and Eocene 590 to 695, the top of the volcanics. The average depth to igneous rock is about 250 feet probably because of wave erosion with lower sea level during glaciation.

Ladd, Tracey and Lill gave a preliminary report on the boring on Bikini which failed to reach basement at 2556 feet. This showed Pleistocene reef to about 425 feet and Tertiary sediments below. The rock below about 1790 is Miocene. These borings all show shallow water calcareous sediments but demonstrate that subsidence has been going on throughout all of Tertiary and Quaternary time. It cannot be due wholly to glacial control of water level although the Pleistocne reef must have been affected by the process. Either the floor of the southwest Pacific has sunk, or the amount of water in the oceans has increased. Platforms on which reefs originated may be erosional, depositional, or local uplifts of the ocean floor. Ladd thinks that no reefs located on the rims of submerged volcanos have been discovered. Cloud uses the term <u>table reef</u> for small reefs without a true lagoon.

<u>Beach features</u>. Shepard has presented some new terms for beach features as shown below. He desires to restrict the term <u>bar</u> to submerged accumulations only.

Barrier beach = single elongate sand ridge parallel to mainland and separated by a lagoon.

Barrier island = multiple ridges together with dunes.

Barrier spit = a barrier tied to mainland at one end only.

Bay barrier = former bay bar extending across an inlet.

Barrier chain = series of barrier islands.

Longshore bar = submerged sand ridge or "low and ball".ridge, or subaqueous ridge.

Transverse bar = sand bar at right angles to shore line.

Reticulated bars = criss-cross pattern of bars inside barrier islands and in bays seen from air only.

Sandkey or sandcay = small island not parallel to shore.

Cuspate features = points 30 to 200 feet apart are caused by wave work. Larger ones have a ridge extending out to sea below water. These would include the famous cuspate capes of the Atlantic coast. Similar features occur inside lagoons on inside of barriers. Most of these have associated shoals or are opposite a cusp on the mainland with or without a connecting shoal.

Cuspate foreland = larger cuspate capes.

Cuspate bars and sandkeys = below-water features.

Lunate bars and sandkeys = crescentic bars off passes with tidal current channels in center or at ends.

Longshore bars (subaqueous bars). The Beach Erosion Board of the U. S. Army Engineers offers some ideas on Longshore bars and longshore troughs (subaqueous bars or "low and ball"). The theory of origin by Evans is supported. Repeated soundings along piers has shown that the positions and depths of both bars and troughs vary with intensity of wave work. The bars form where the waves break. After plunging on the bar, where observation by the writer showed water with considerable sand, the wave reforms and breaks again when a certain depth is reached. After several such breakings at each of which bars are formed, the wave reaches the beach. Troughs and bars become progressively smaller and shallower in approaching the shore, but their size and depth changes with height of waves. The ratio of depth of trough to depth on bar varies from 1.3 when mean sea level is used to 1.5 when "mean lower low water" is taken as datum, but otherwise no generally applicable relations were found. Although it is known that material is brought from both sides to build up a bar and that this building is due primarily to plunging breakers, the presence of longshore currents in the troughs is proved. These carry sand to breaks through which rip currents escape, spreading the sand on the sea bottom outside. On coasts where the depth increases rapidly offshore, waves break only on the beach and no ridges and troughs are formed. Keulegan and Krumbein showed mathematically that some seas could be so shallow that waves could not break anywhere and cite geological evidence of such conditions. Depth would increase at the 4/7 power of distance from shore.

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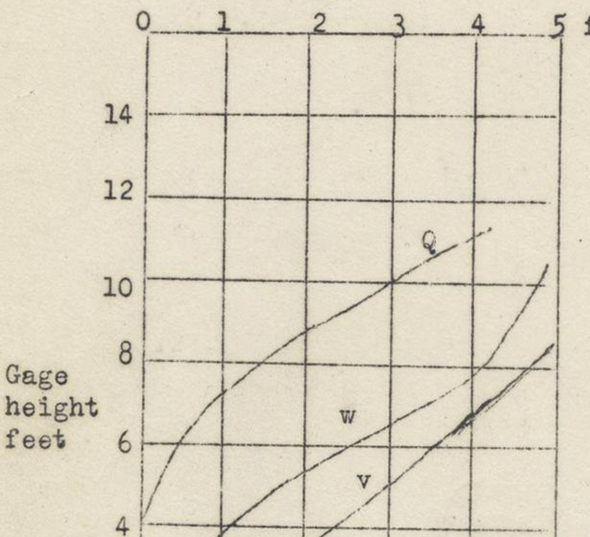
Dimensions and competence of running water. Supplement I, 1953, p. 1

Introduction. Three papers have appeared on the subject of running water which appear to show marked progress in understanding of some problems. Two of these not only clarify some of the basic points of the physics of streams but also point the way to solution of many important problems of sediment transport. The third, deals with particle size distribution on an alluvial fan.

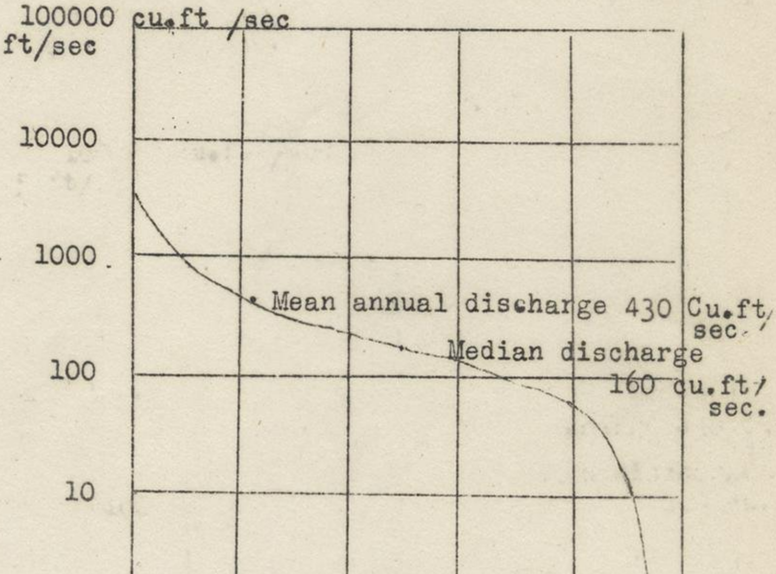
Discharge of streams. The fundmental quantity measured by hydraulic engineers is the discharge of streams. To find this figure they first discover a suitable cross section of the channel. This is subdivided into segments of known dimensions, then the average velocity of flow is found in each segment giving its discharge and the final sum of the segments is the Discharge (Q) = average width of channel (w) X average depth (d), X average velocity, (v) or Q = w.d.v. British engineering units are employed, cubic feet per second, and feet. Since the discharge of all rivers varies constantly it is necessary to connect each actual measurment to the gage reading of water level in the river at that time. Most discharge determinations are read from a curve (Fig. 1) which indicates this relationship. Next a curve (Fig. 2) must be prepared which shows the percent of days that any given discharge is equalled or exceeded. The mean discharge is also computed as the aritmetical average of all recorded daily discharges. This quantity is generally larger than the median discharge which is equalled or exceeded exactly 50% of the time.

Fig. 1



Relation of quanities to gage reading in feet.
0 200 400 600 800 ft. width
0 10000 20000 30000 40000 cu.ft./sec.

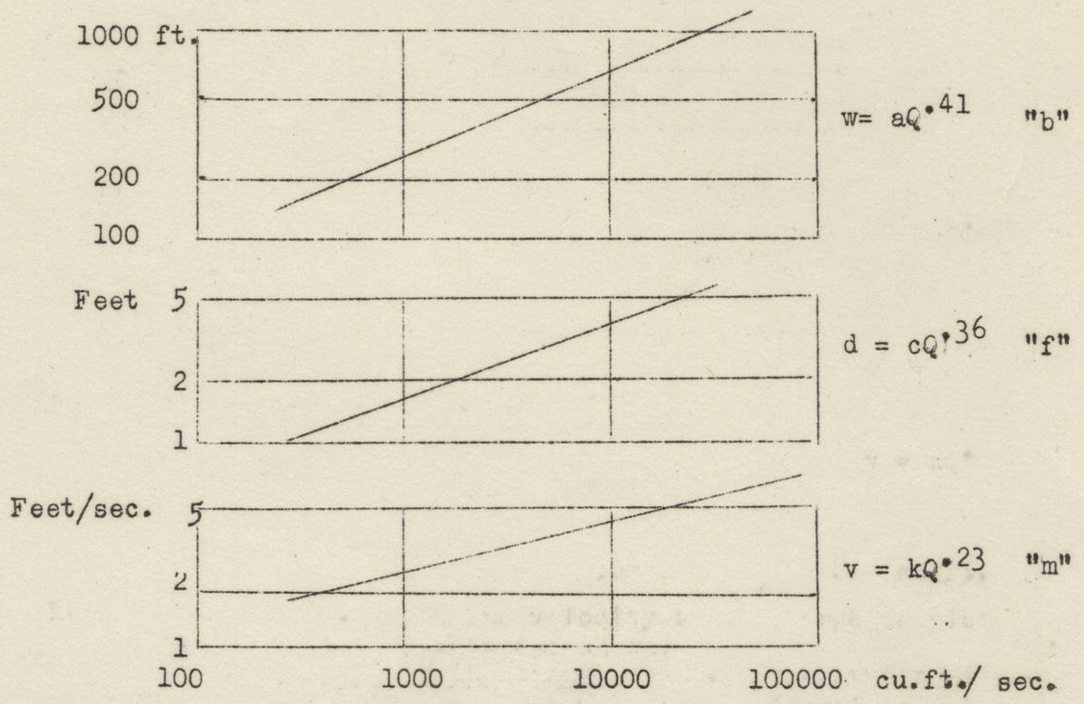
Fig. 2



Typical frequency or flow-duration curve of a river

<u>Inter-relations</u> of quanities. Platting on log-log paper demonstrates, as shown in Fig. 3, that w, d, and v are simple <u>power functions</u> of Q, the primary determination. In mathematical expressions $Q = wdv = aQ^b \times cQ^f \times kQ^m = ackQ^{(b+f+m)}$ From this it is evident that the sum of the exponents of Q must be unity and the product of the numberical constants must be the same. An average of 20 river sections studied gave b = 0.26, f = 0.40, m = 0.34 but the values of the constants varies much more widely than do the exponents. Evidently the values are related to the materials of the stream beds and possible to other factors. The limits of variation are unknown. Depth increases with discharge faster than does width.

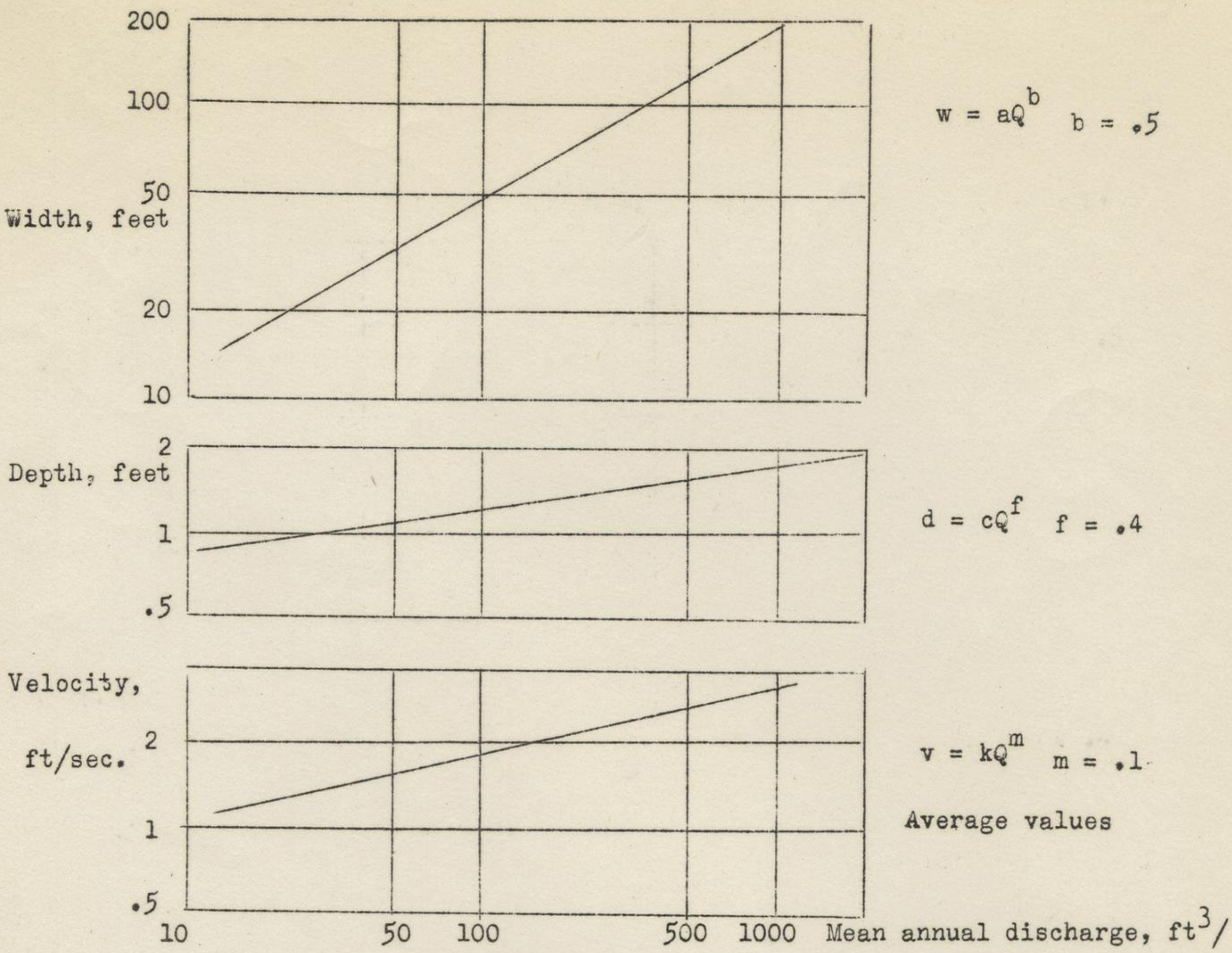
Fig. 3



Relations of width, depth and velocity to discharge as plotted on log-log paper. Scatter of points not shown.

<u>Downstream variations in channel shape</u>. In computing the relations of dimensions of stream channels in a downstream direction it is evident that all comparisons must be made for a specified discharge at every station. Most of the log-log plats were made for mean annual discharge which occurs or is exceeded on the average about one day in every four. In almost all rivers discharge increases downstream. Some were made for flows which occur less frequently.

Fig. 4



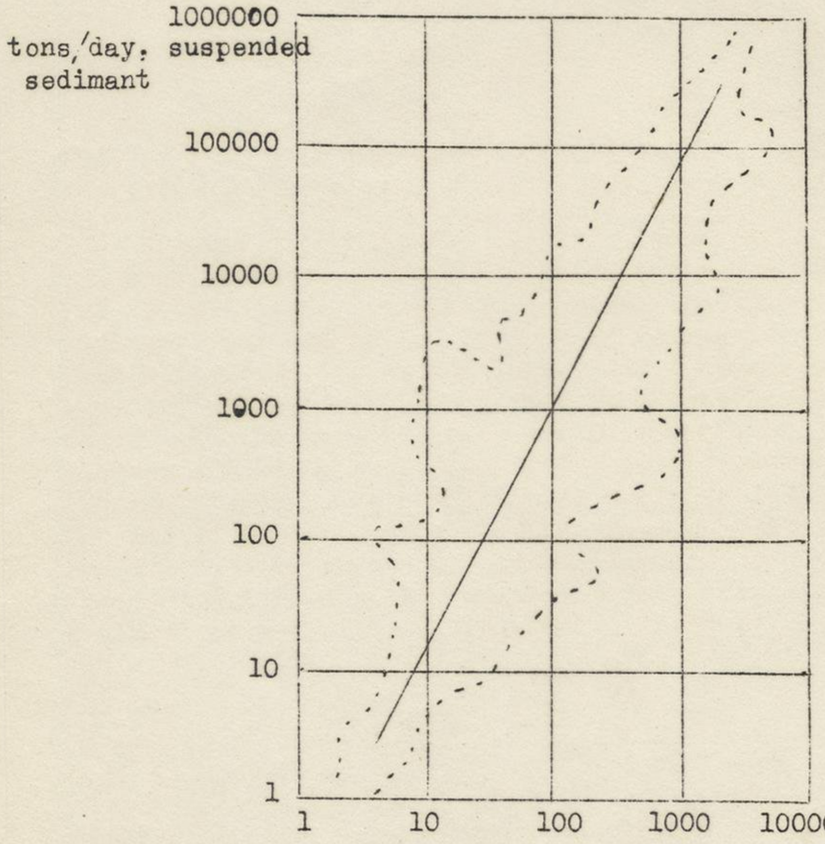
Downstream changes in dimensions of streams in Wyoming and Montana. Points not shown

Despite the expectable "scatter" of points when platted, there is a remarkable agreement in results. Using the notation above, $w = aQ^b$, $d = cQ^f$, and $v = kQ^m$, the average values are $b = 0.5$, $f = 0.4$, and $m = 0.1$ This shows that for increase in discharge downstream all quantities including velocity increase. Increase in velocity is least and this quantity may be almost constant in some streams. Even in the headquaters however, the conclusion is demonstrable. It is contrary to what nearly everyone formerly thought and hence demands some explaination. To do this we will restate Mannings formula for velocity of a stream with turbulent flow: mean velocity (v) ft/sec = $\frac{1.5 \ d^{2/3} \ S^{\frac{1}{2}}}{\text{roughness } (n)}$ (dimensions in feet) Note that for wide stream mean depth (d) replaces hydraulic radius (R or cross section area divided by width.) From this it may be seen that most geomorphologists have ignored both depth of water and roughness of the bed. Together these overcompensate for the fact seen in the field that slope of the water surface almost everywhere decreases downstream. Slope (s) in feet per foot = $0.021Q^{-0.49}$ on the average.

Sediment transport. Streams carry sediment in two ways, (a) as bed load or bed-material, and (b) as in suspension or wash load. The two may change in proportion with alterations of the stream so that what is suspended at one time may be a portion of the bed and vice versa. The mathematical relations of the two are only vaguely known for there is at present no accurate method of determining transport of material on a stream bottom. Any mechanical device to catch such load introduces changes in the currents which render the results valueless. Suspended load can be and is being measured at a number of localities. Possibly data on the filling of reservoirs may eventually supply some of the missing information. The following discussion is almost wholly on suspended load.

Suspended load. Platting of the weight of suspended load in given time against discharge of a stream shows at once (Fig. 5) that, despite scattering of points, the amount of sediment increases with discharge as a power function with with an exponent between 2 and 3, thus demonstrating an increase in more than direct proportion to discharge.

Fig. 5

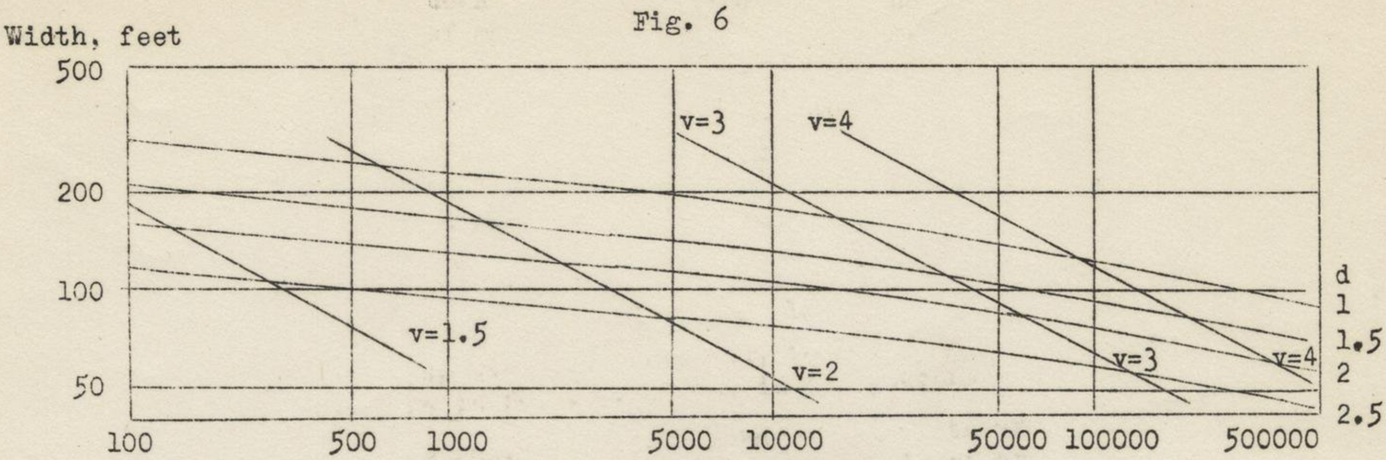


Relation of suspended sediment load to discharge, Powder River, Wyoming
Limit of scatter of determinations shown by dotted line

The cause of this rapid increase is known only in general terms. Factors are: (a) infiltration rate and storage of rain in puddles is greatest at start of a rain, (b) raindrop erosion increases with wetting of soil, (c) long duration of rainfall increases depth of and erosion by sheet wash, (d) increase in velocity of large streams enhances both scour of bottom and undercutting of banks, (e) changes in channel shape during a flood are caused by the suspended load, and (f) suspended sediment concentration may be considered as an independent variable on which both velocity and depth depend. Despite the known alteration of banks by floods, the conclusion of Leopold and Maddock is "that the observed increase in sediment concentration results primarily from erosion of the watershed rather than from scour of the bed of the main stream in the reach where the measurment is made." They found that there are not enough observations to permit of direct conclusions on changes in concentration downstream.

It appears to be slight so far as known for it is observed that increase in sediment with increase of drainage area is less for large basins than for small. It is possible to present a graph such as Fig. 6 showing the relations of width, depth and velocity to total suspended sediment load.

Fig. 6

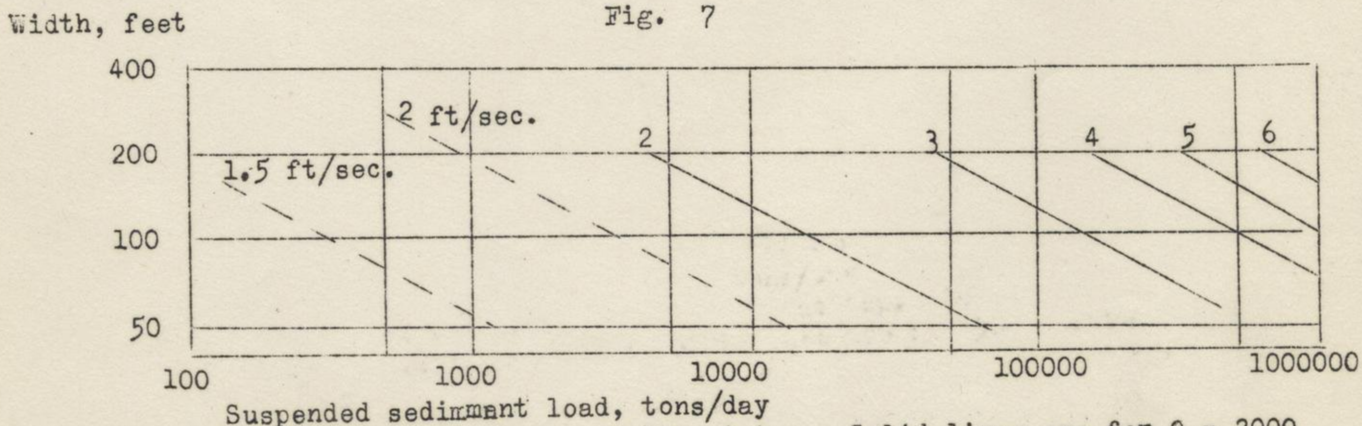


Widths and depths in feet. Velocity in feet/second

General conclusions. (1) If discharge and width are constant increase in velocity means increase in total suspended sediment and a decrease in depth. (2) With velocity constant, increase in width decreases both the suspended sediment load and depth. (3) Both decreasing width with constant velocity and increasing velocity at constant width increase capacity for suspended load at constant discharge. (4) A wide river carries less suspended load than a narrow river with the same velocity and discharge. (5) Two rivers of equal width and discharge load of suspended solids is larger in that having the higher velocity.

Suspended sediment transport with variable discharge. Due to fact that Q = wdv the sum of the exponents b + f + m must be unity as explained above. Hence if two of these exponents are known the third can be computed and from this fact some deductions may be made. First we draw Fig. 7 showing relation of suspended sediment to velocity, width, and discharge.

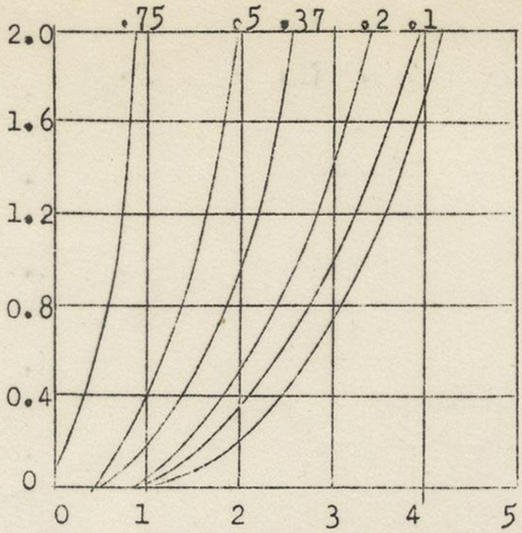
Fig. 7



Solid inclined lines show velocity at top. Solid lines are for Q = 2000 cu.ft/sec.; dashed lines toward left are for 500 cu.ft./sec.

0 values of b

Ratio of exponents of veloity, m and for depth, f, m/f



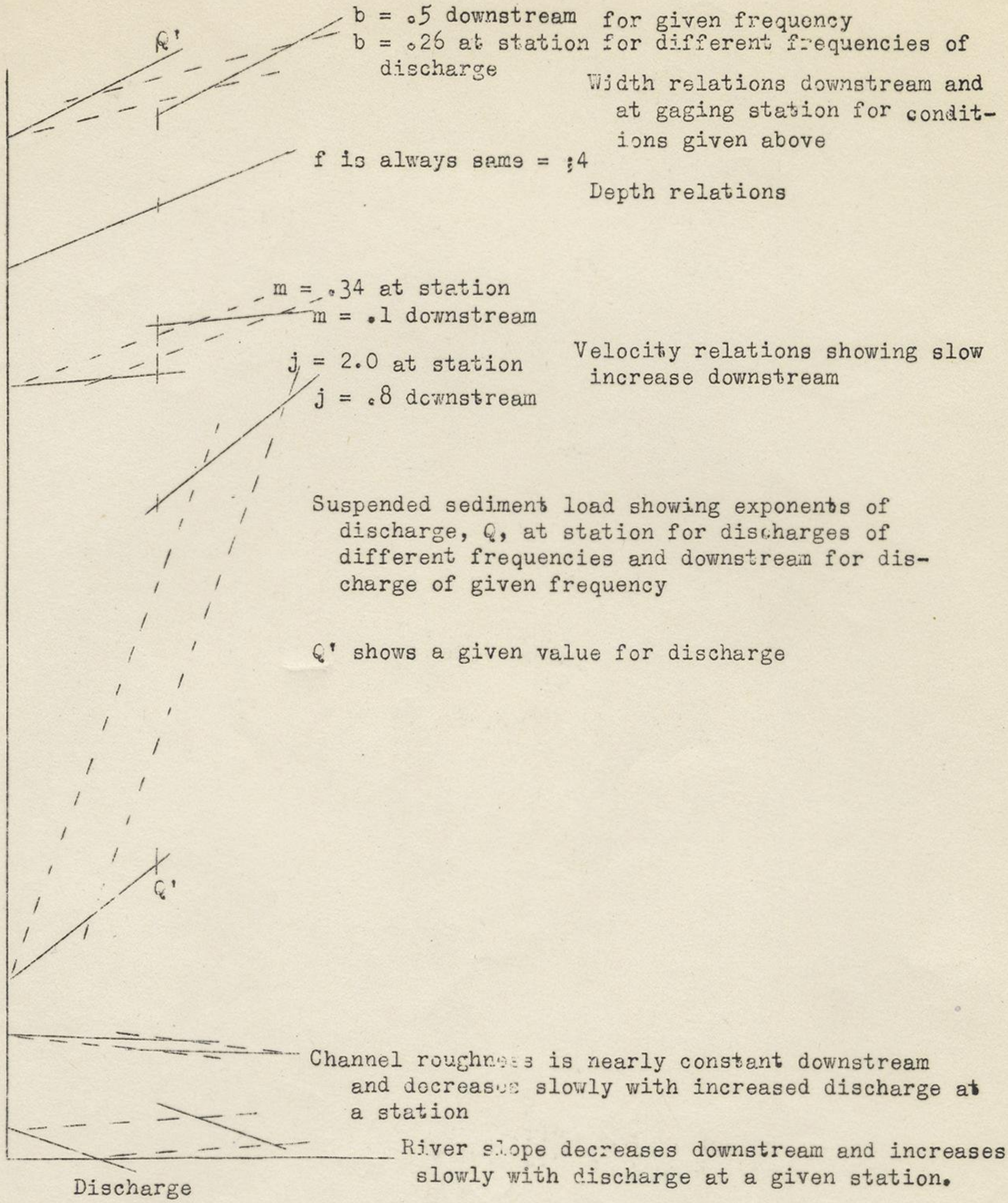
Curved lines are for the exponent b of width

b + f + m = 1.0

Exponent of suspended load, j

From this it is possible to draw curves showing values of j, the exponent of Q for suspended sediment, in terms of both b, and the ratio of m to f. The ratio between increase of velocity with discharge and increase of depth with discharge is, therefore, related to amount of suspended sediment. For the average cross section of a river m/f is 0.85, b = 0.26, and j = 2.3 This is in line with the statement that sediment concentration should decrease slightly downstream. (Fig. 8) Comparisons of different river cross sections indicate that: suspended sediment load varies: (1) directly ~~with~~ as a function of velocity, (2) directly as a function of depth, (3) inversely as a function of width, (4) as a large power of velocity, and (5) as small powers of depth and width.

Fig. 9

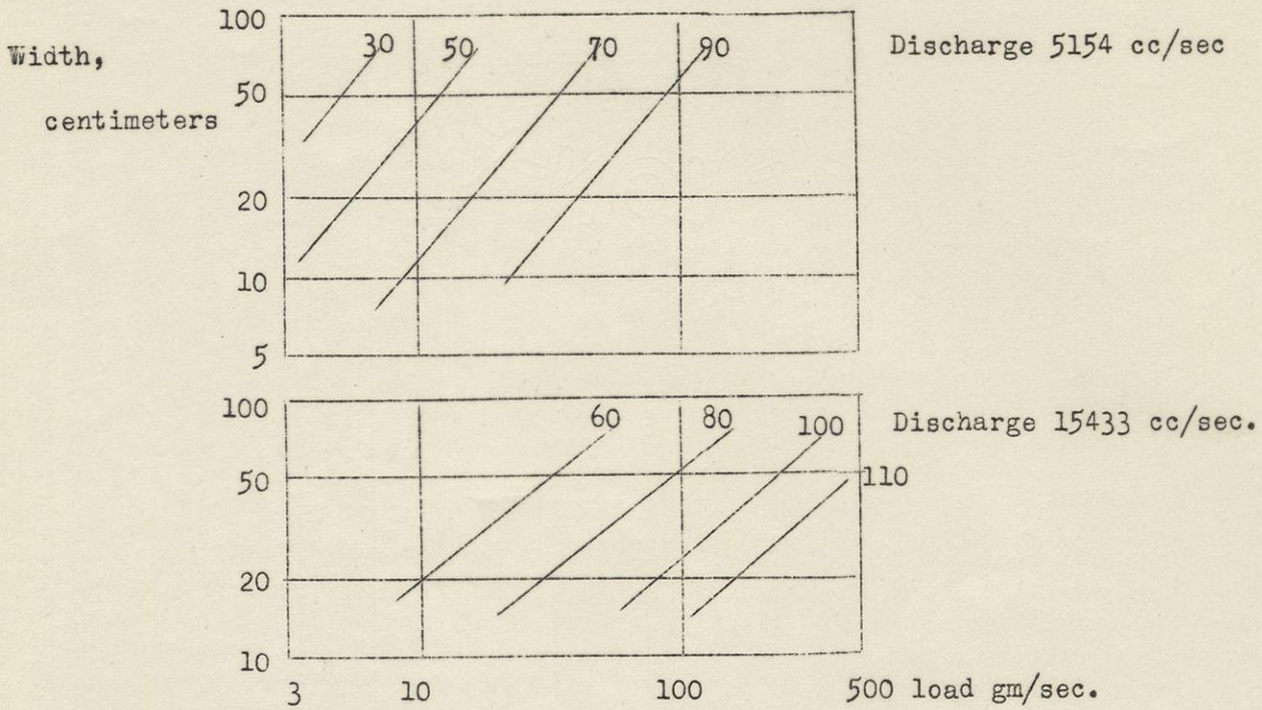


Solid lines for slope on exponent on log-log plat to show downstream change for Q of given frequency Dashed lines for changes at station.

Fig. 9 summarizes the information by showing the comparative changes at a station and downstream by giving the proper slopes of the lines which display the values of the exponents of discharge in log-log platting. We may say that for given width and discharge increase in suspended sediment requires increase in velocity and reduction in depth. The quantities involved are adjusted to the nature of the drainage basin so that they are independent of the channel system.

<u>Bed load</u>. Since there is little information of bed load transport in natural streams recourse must be had to the experiments of Gilbert in wooden troughs here restated in the C. G. S. system. Fig. 10 shows at once that the relation of the lines of equal velocity is exactly opposite to those of Fig. 6 for suspended sediment. Data are given for two different discharges both with same kind of sand. Tentative conclusions are: (1) with constant discharge and width increased velocity increases both bed load and suspended sediment, (2) with constant velocity and discharge increase of width decreases suspended load and increases bed load, (3) broad shallow channels are needed to transport a large bed load.

Fig. 10



Inclined lines show velocity in cm/sec.

<u>Changes of channel form</u>. At some gaging stations measurments have been made of changes in channel form during floods. Some places at the start of a flood, when concentration of suspended sediment is high, display a rise in level of the bottom. This is followed, when sediment decreases, by scour and lowering of the bed. Obviously the latter causes a lower velocity when less velocity is needed for transport. At other places erosion begins at once with the rise of discharge with high sediment concentration and later filling takes place during fall of water level. It has been noted that the spring floods of melted snow in western rivers lower river beds whereas later season floods due to rain result in fill. Filling often occurs during times of increasing velocity.

<u>Roughness of channel</u>. At constant width and discharge it is obvious that the product of v. d must be constant. Hence any increase in velocity requires a decrease in depth. From the usual velocity formula it is evident that for any increase in velocity and decrease in depth the factor $\left(\frac{S^{\frac{1}{2}}}{n}\right)$ must increase.

The two equations: $d = cQ^f$ and $v = kQ^m$ make it possible to set up another. $kQ^m = 1.5\ (cQ^f)^{2/3}\ S^{\frac{1}{2}}/n$ where the constants c and K vary. Hence $Q^m : Q^{2/3 \cdot f}\ (S^{\frac{1}{2}}/n)$ Where S and n are constant with discharge then $m = 2/3\ f$ or $m/f = 2/3$ From this it follows that if $S^{\frac{1}{2}}/n$ increases with discharge m/f is more than 2/3 and when this ratio decreases with discharge then m/f is less than 2/3. Now at a given station the average ratio of m/f is 0.85 whereas downstream this is only 0.25

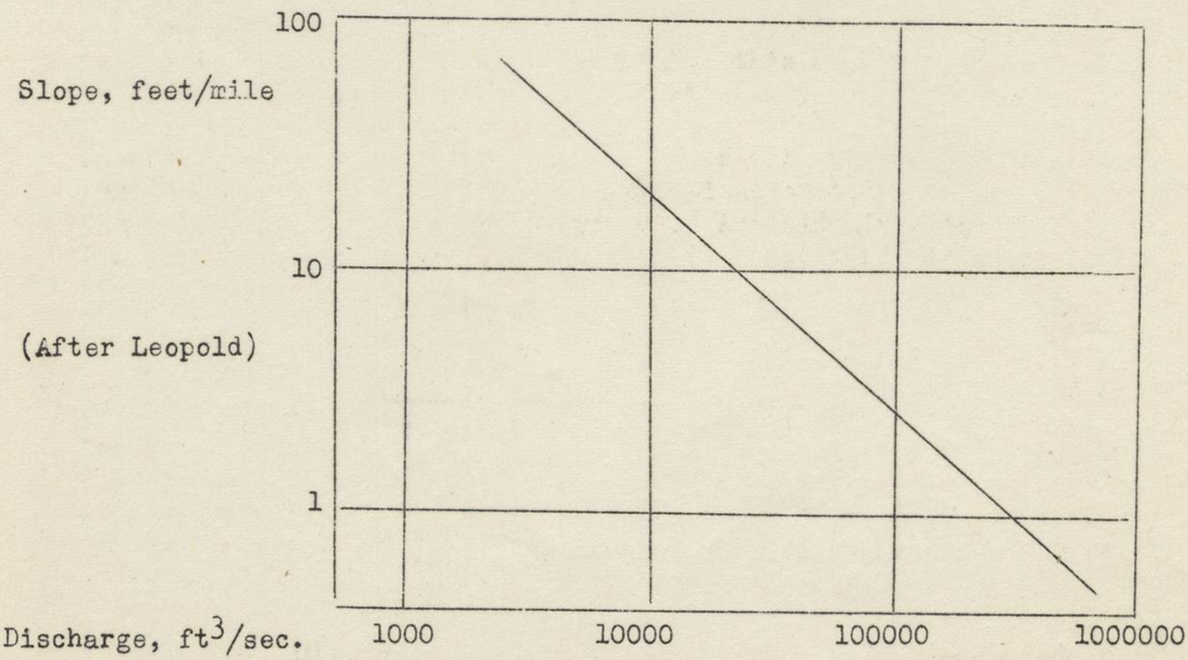
From this it appears that $S^{\frac{1}{2}}/n$ increases with discharge at a given station and decreases downstream. It has also been observed that in the downstream direction roughness (n) remains about constant so that slope must decrease to preserve the above relations. Observation has also disclosed that an increase in suspended load decreases channel resistance and hence increases velocity. Possibly this is really related to decreasing turbulence. Increased values of sediment concentration are associated with decreased values of n. At a given station, however, the slope does not change very much so that the alteration of n must be considerabl with change in concentration of sediment. Changes in velocity-depth relations might be attributed to change in sediment concentration where an increase diminishes the roughness, n, of the bottom. A check consists of the behavior of Colorado River after the completion of Boulder (Hoover) Dam which caught much of the sediment leaving clear water below. This is the same as a lake in the course of a river. Alterations below the dam consist of (1) increase in depth in spite of a lowering of surface elevation, (2) decrease in width due to reduction of flood volume, (3) decrease in mean velocity, (4) increase in roughness of bottom, apparently a result not of change in type of material but of decrease in suspended load, (5) reduction of bed load in the narrowed channel, (6) increase in capacity for suspended load due to change in velocity and discharge, (7) no appreciable change in slope.

Factors of channel roughness. Channel roughness is due to (1) particle size, (2) bed configuration, and (3) sediment load. It is commonly observed that the material of most stream beds diminishes in size of particles downstream although from this it does not necessarily follow that decrease in slope is directly attributable to this phenomenon. Waves and ripples on the stream bed are very important factors in roughness, although they are not permanent. Increased bed roughness decreases velocity in respect to depth hence affecting the capacity for load. These waves or ripples vary in nature with different kinds of sediment. They pass with increasing discharge from smooth bottom through successive forms into antidunes which travel upstream. For fixed slope and discharge decreased particle size tends to increase roughness. Bottom material is most important in the headwaters of streams where the bed consists of boulders, cobbles, and pebbles. Under this condition, downstream decrease in size of particles decreases roughness. The Powder River, Wyoming, has a value of n on gravel of .087 which falls to .017 on silt farther downstream. However, in other streams the value of n is about the same downstream despite marked differences in nature of bottom. There it must be that bottom configuration is dominant. In summary, it is clear that slope is the dependent factor which the stream is able to change. As noted above it is common to find at a given station that suspended load of streams increases rapidly with discharge. This requires a relatively rapid increase in velocity compared to depth, that is a high value of m/f. Such is accomplished primarily be an increase in the value of n which is related to increase in concentration of suspended load. However, in a downstream direction load does not keep pace with discharge and the concentration of suspended sediment decreases slightly. To do this depth must increase with discharge faster than does velocity so that the m/f ratio must be low. Hence $s^{\frac{1}{2}}/n$ must decrease downstream. With roughness about constant this can be done only by decreasing the slope.

Graded streams. By definition a graded stream can over a period of time just transport the amount of sediment furnished it. Engineers have constructed many irrigation canals which do exactly this, that is they neither erode nor silt up. Some rules were derived by experiment which used perimeter, P, instead of width and hydraulic radius, R, instead of mean depth. A sediment factor, F, is also introduced. The basic equations are: $P = 2.67\ Q^{\frac{1}{2}}$ and $V_{mean_1} = 1.15F^{\frac{1}{2}}R^{\frac{1}{2}}$. Note that in the studies of Leopold and Maddock they found that $w = aQ^{\frac{1}{2}}$ (downstream). By combining the relations $d = cQ^f$ and $v = kQ^m$ we find that $(d/c)^{1/f} = (v/k)^{1/m}$ or $v:d^{m/f}$. In natural streams this ratio of m to f downstream is only 1/4 whereas in the canals it was 1/2. But we must recall that canals for irrigation are not like streams because they loose discharge downstream as it is dispersed into laterals. They can have no change in suspended sediment concentration hence the value of j cannot be above 1.0. If b = .5 and j = 1.0 this means that m/f would be 0.43 or not far from that value already given. This suggests that j must in practice be less than 1. In summary, Maddock and Leopold conclude that with available data it is not possible to discriminate graded from ungraded sections of a river.

Longitudinal profile of rivers. It has long been assumed that the profile of a river bed is directly related to the maximum particle size of sediment in its bed. It has also been assumed that wear of the load results in a downstream reduction of size of particles. The latter can be checked in the field, although it is hard to distinguish material derived from tributaries and cut banks and not brought far downstream. Now if the velocity of flow really increases downstream how can competence of the current be the controlling factor of river profiles? Some have derived equations to substantiate this assumption but the issue is confused by several phenomena. (1) Decrease of particle size increases roughness by promoting ripples; (2) roughness is also related to concentration of suspended sediment and, (3) in practice roughness does not vary much downstream. Hence to preserve the required velocity-depth relations to transport the load the slope of a normal stream must decrease downstream. Leopold gives the empirical equation that slope, $S = 0.021\ Q^{-0.49}$, that is slope is approximately inverse to the square

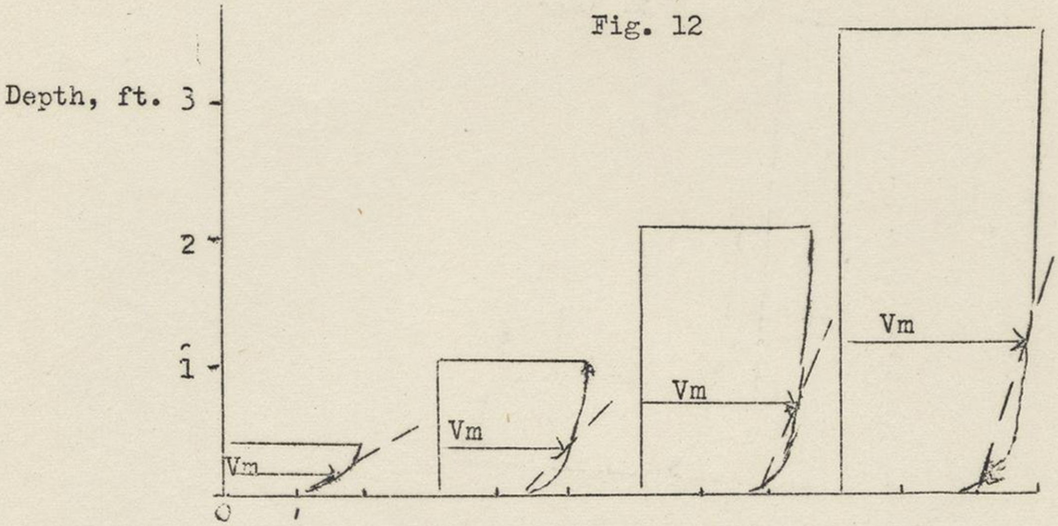
Fig. 11



root of discharge. We cannot, however, construct a longitudinal profile of a river from this without knowing how the discharge varies in a downstream direction. This is commonly in direct proportion to drainage area not to distance along the channel.

Vertical velocity distribution. It has long been known that in rivers which are relatively wide in proportion to depth, that is where the banks are readily erodable, the vertical distribution of velocity is approximately proportional to the logarithm of distance from the bottom, z. Such being the case the rate of increase of velocity with respect to distance from the bed is inverse (see any text book of Calculus). Now this rate of change in velocity upward from the bed determines the shear or rate of energy transfer from the stream to the bottom. Since in most streams depth increases downstream as a power function of discharge the slope of the line representing rate of velocity (dv/d2) change near to the bed must decrease with increase in total depth.

Fig. 12



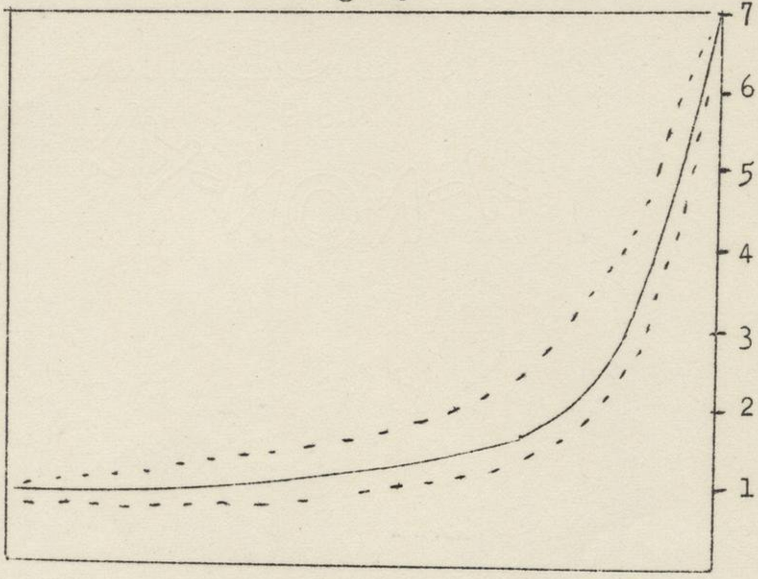
Velocity, ft/sec.
Inclined dashed lines show downstream decrease in shear

Another factor is that total force on the bed is proportional to depth times slope. As brought out above, depth increases on the average at the 4/10th power of discharge whereas slope decreases at approximately the square root of that quantity. Hence the product DS must decrease slowly downstream at about the minus 1/10th power of discharge.

Summary. Although the old idea that river slopes are directly related to velocity which decreases downstream thus decreasing competence must be abandoned, it is clear that there is a downstream decrease in competence. The details of just how this comes about are not simple. The vertical velocity profile and shear on the bed are interrelated and depend not only on mean velocity but also on depth, and on roughness of bottom. This shear also affects the intensity of turbulence which is necessary to keep material off the bed. Downstream decrease in roughness may diminish both shear and turbulence despite increase in mean velocity. Leopold lists the variables which enter into this problem: discharge, width, depth, velocity, slope, roughness, load, and size of particles in transit. These constitute eight simultaneous equations whose solution is at present impossible. Of them only the flow equation (Q=wdv) and Mannings formula for velocity are accepted by common use. The others comprise relation of load to nature of basin, rate of particle size change downstream, width-depth ration in relation to nature of the bed and banks, change in value of n, the roughness factor, with depth, material, discharge, and slope, and relation of n to sediment concentration. The interdependence of these factors is evident and it is clear that the stream is capable of adjusting its slope to fit the requirements of the others. The cross section of a stream is adjusted so as to equalize shear on both bed and banks. The form of the bed can be changed so as to alter roughness. All of these factors are much more complex than we were led to believe by the pioneer students of geomorphology who did not employ quanitative methods even if they were correct in general principle.

Change of particle size downstream. As explained above it is generally impracticable to measure the downstream reduction of size of particles transported by a river. On alluvial fans, however, all the debris is derived above the apex and reasonable success has been attained in comparing the maximum particle size with distance from the source. An article by Blissenbach based on fans in Arizona shows (Fig. 13) that despite considerable scatter a definite

Fig. 13



After Blissenbach Dotted lines show scatter of points

relationship does hold. From the known fact that diameter of pebbles is related to the square of velocity of transporting water it could then be concluded that the ratio of mean depth (or hydraulic radius) to bottom roughness must remain reasonably constant. On alluvial fans this might be expected for all the water is derived from the head so that the individual streams on the fan do not vary widely in size despite some loss by evaporation and perhaps by seepage. Roughness, which should decrease with smaller particles downward on the fan, could be maintained by more ripples in the bed on lower slopes. The log-log. platting (using slope as directly proportion to degrees measured) of the diagrams published show that slope is approximately inverse to the square root of horizontal distance from apex. Fall must, therefore (see integral calculus) be in proportion to the square root of distance from apex. The same paper also presents some data on relationship of maximum particle size to angle of slope (on steeper slope the degrees do not correspond directly to the technical definition of slope which is tangent of the angle) whic seem to confirm the determinations of Fair in South Africa. In the case of the Black Hills terrace gravels there is rough agreement of slope to logarithm of geometric mean size of stones. All of the above data is inconclusive for no attention has been paid to mean particle size of entire deposit and it is known that there is much finer material along with these maximum particles. On a ~~table~~ talus does the average or medium size control the coefficient of friction?

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(7) personal

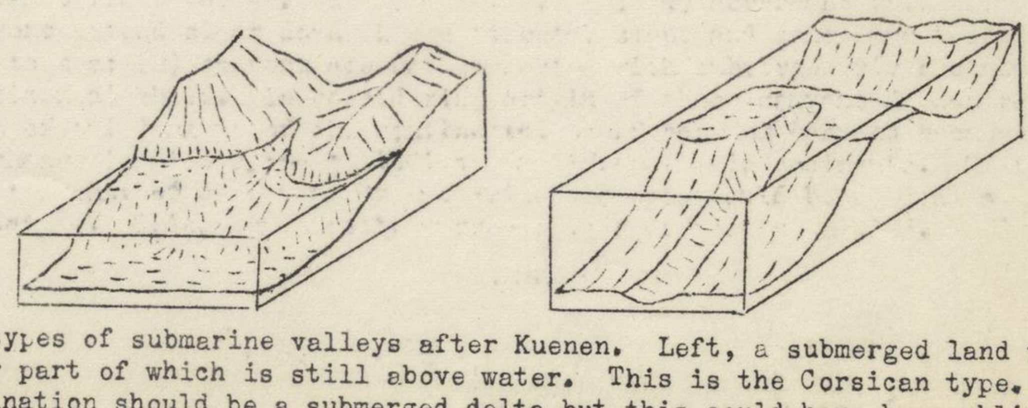
Submarine canyon problem, cont., 1952 Supplement III, 1953, p.1

Introduction. Since the supplement on the problem of submarine canyons appeared some important papers have been published and it is desirable to add some material which was omitted previously. Progress has been mainly along two lines: (a) suggestions that canyons are of more than one origin, and (b) tracing of canyons into the deeper parts of the ocean.

Hypotheses. Kuenen has suggested that submarine canyons may be divided into two great classes: (a) true drowned valleys, the Corsican type, and (b) troughs due to density currents, the New England type. In addition, it is recognized that drowned valleys may have been clogged with marine sediment and then reexcavated and/or extended by density currents. This history can be regarded as in a way a transition between the two major divisions. The author presents charts which show the difference between the submarine extensions of land valleys on coasts where there has been relatively recent orogeny. Although theoretically the drowned valleys should terminate in submerged deltas that fact is hard to prove, and it is possible that slides and density currents may have obliterated or altered them beyond recognition as suggested by Shepard. Shepard also suggested that the continental shelf between valleys may have been built up with sediments during the time that the valleys were kept open by slides and density currents. We must not lose sight of the strong probability that there are tectonic depressions on the continental shelf in regions of mountain building. And last we must always give due weight to the "personal equation" in the drawing of submarine contours, as well as to the limitations of acoustic sounding.

New England type of canyons. Kuenen lists the major characteristic of canyons of the New England type which cannot possibly be regarded as the submerged extensions of land valleys. In brief these are: (a) V cross-section with sides sloping at about 22%; (b) straight course down the continental slope; (c) course locally different inside the edge of the shelf; (d) widely rounded curves; (e) continuous seaward slope of bottom; (f) steepest grades near head with decrease outward; (g) with rare exception, no break in slope at the continental terrace border; (h) no abrupt falls; (i) accordant tributaries; (j) all canyons extend clear down the continental slope and some have been traced far out to sea; (k) canyons are not connected with submerged river channels on the continental shelf. In considering origin of these canyons Kuenen rejects Shepard's idea of building up of the continental shelf between canyons because there is no change in side slope to indicate a difference in sediments. He thinks a very great amount of deepening by submarine currents would be needed to eliminate such a feature. Origin by density currents is therefore concluded. (Figure 1).

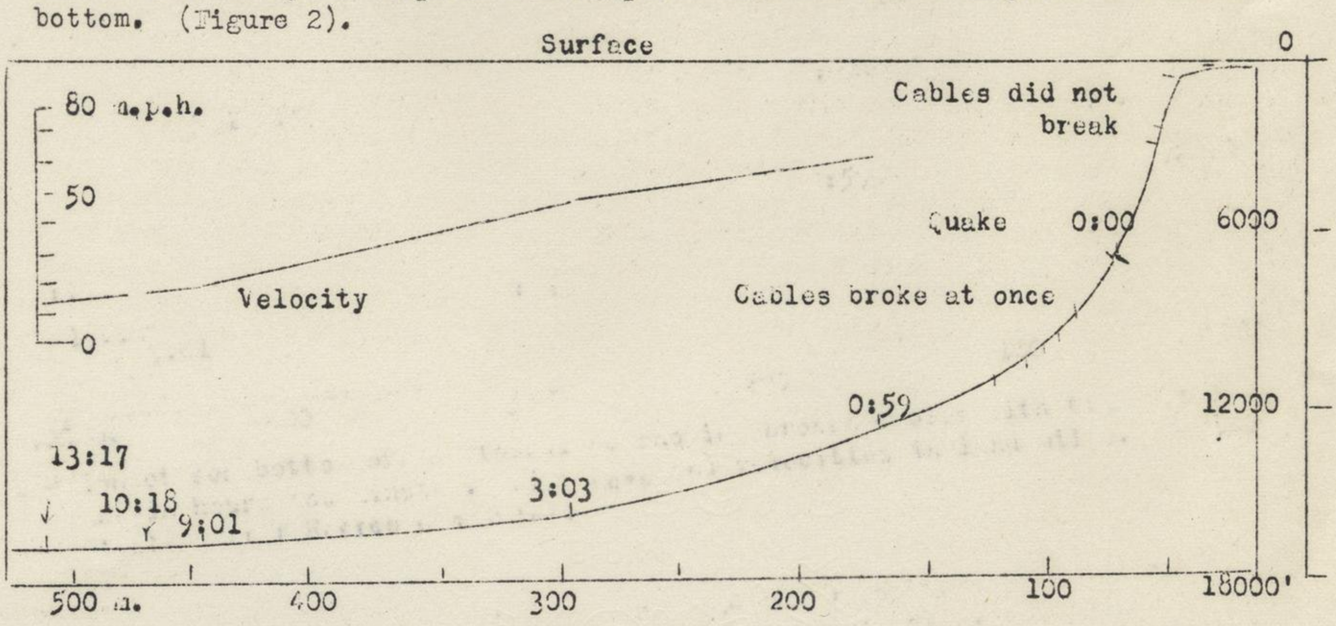
FIG. 1



Two types of submarine valleys after Kuenen. Left, a submerged land valley, the upper part of which is still above water. This is the Corsican type. Lower termination should be a submerged delta but this could have been obliterated by slides and density current erosion. Right, conditions on Georges Bank, off New England. Here the top of the bank is a submerged cuesta but the valley on its seaward slope has nothing to do with land valleys. It was eroded by density currents when the level of the sea was lower than it now is.

Proof of density currents. As noted in the former supplement one of the weakest points of the density current hypothesis, originally proposed by Daly, is that it is extremely difficult to find such penomena actually at work. Daly concluded that their maximum activity is a thing of the past because low glacial sea levels furnished much more sediment which flowed down the continental slope than is now the case. It is well to note that this is entirely in line with recent theories of the origin of the continental shelf which shows indubitable evidence of a lower sea level. It is suggested that the observed density currents in freshwater lakes and reservoirs are not a fair comparison because of the gentle grades and the presence of concurrent sedimentation from water which, in the case of glacial meltwaters, floated on top of the lake by reason of the temperature difference. The channels on delta fronts are more of the levee type and are not true canyons. Sliding may have taken place, for instance, on the delta of the Mississippi.

The Grand Banks earthquake. It is to phenomena which followed upon the earthquake on the Grand Banks of Newfoundland in 1929 that advocates of density currently mainly turn for evidence. The quake occurred on 18 Nov., 1929, at 2032 hours G.C.T. Instantly six cables in water from 900 to 10800 feet deep broke, but for 13 hours 17 minutes thereafter there was an orderly sequence of breaks of other cables at progressively greater distances from the epicenter, to about 375 miles. The velocity of the force which brought about these delayed breaks can easily be computed and compared with the known slope of the ocean bottom. (Figure 2).



Section of sea bottom off Newfoundland showing broken cables with time after quake in hours and minutes. Distances and velocities in land miles. Depths in feet. After Heezen and Ewing.

The affected area broadened with distance and the velocity decreased from 63 miles per hour to about 14 miles per hour at the last cable. Every cable broke in at least two places 100 miles or more apart. The cable between the breaks was in all cases either buried or carried away so far that it could not be recovered. Although most geologists at first considered the cause of breaking to be faulting, the opinion of Heezen and Ewing is that it was the transformation of a landslide to a turbidity current. In this connection we may note that the breaking strength of a new submarine cable is about 12 short tons and its weight under water is about 1.3 short tons per mile. The necessity of assuming a more powerful force than lack of support due to erosion of the bottom is evident. In making repairs to the cables "sharp sand and small pebbles" were dredged up in about 16800 feet of water. Kuenen showed that with existing formulas the size and velocity of the inferred density current are credible.

Midocean canyons. A recent publication by Ewing and his associates gives much more data on canyons in deep water than was available only a few years ago. They state "In recent years exploration has revealed that the canyons do not end at the base of the continental slope but continue across the continental rise to the abyssal plains of the ocean floor. Studies of the sediments from the floors, walls and seaward extremities of these canyons---have proved that powerful turbidity currents have repeatedly carried large volumes of sediment through the canyons and deposited them in well-sorted beds on the abyssal plains". One of the canyons has been definitely traced for nearly 1400 land miles and it may extend for more than 2800 miles. Steep sides and flat floors are indicated by the cross sections with a depth below the adjacent ocean bottom of 60 to 600 feet. The longitudinal slope is from 2.5 to 5 feet per mile. Maximum recorded depth to the bottom of the channel is about 16500 feet. The mid-Atlantic channel crosses the Southeast Newfoundland Ridge in a narrow gap. It is not certain that its end has yet been reached. "In general the cores indicated that the turbidity currents depositing sand and silt in the canyon feathered out on the banks. The burial of these sands and silts by over a meter of clay and silty clay would indicate that the last major turbidity current probably occured in Wisconsin time." "Faulting offers no explanations of the sediment relations or the stream-like longitudinal profile so easily explained by turbidity currents." The evidence of these channels and their associated sediments seems to present a very much more formidable case for the reality of turbidity currents than was even dreamed of when the theory was first advanced. Preservation of topographic forms with little alteration in the normally quiet realm of the ocean depths can readily be understood.

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GEOLOGY 109
GEOMORPHOLOGY

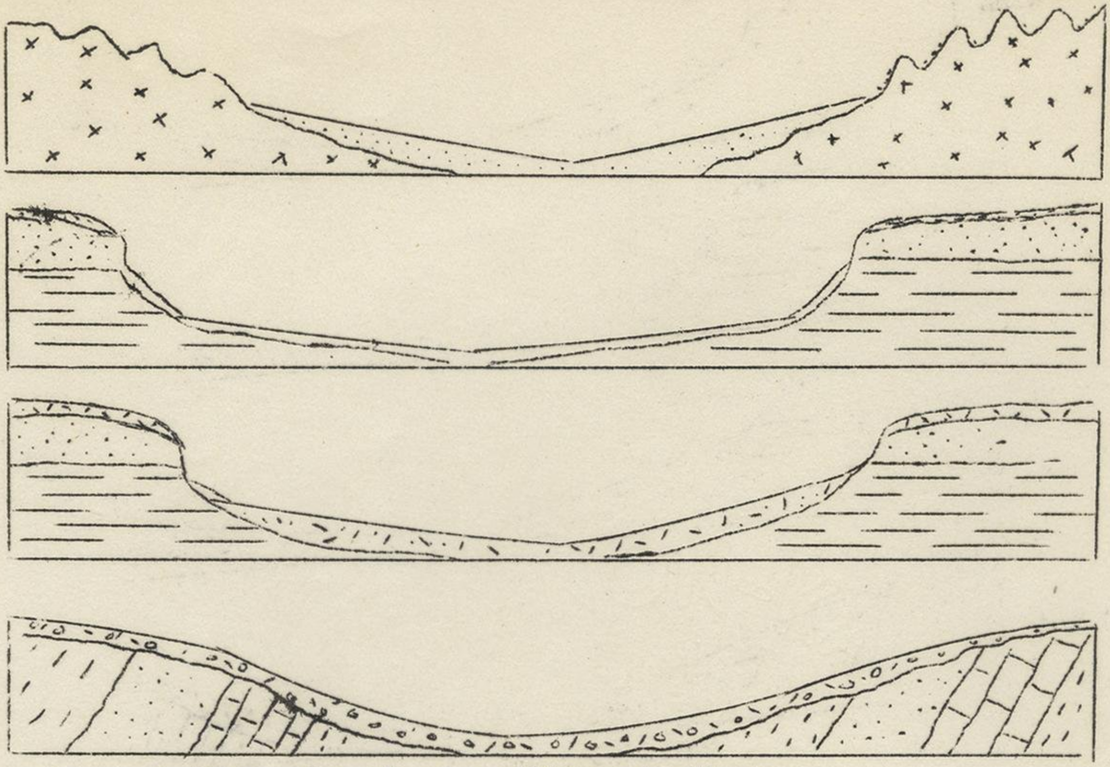
Supplements, 1953, part II

Pediplanation vs peneplanation.

Introduction. Although the subject of final stages in denudation by running water has been covered in previous supplements, data which has appeared in the past year offer further food for though on this extremely important problem. Before beginning a discussion, however, it is well to repeat that the definition of the word peneplain (peneplane of Johnson) is far from uniform among students of geomorphology. This makes it extremely difficult to argue about either processes or end results. Let us here return to the original ideas and neglect later attempts to change the definition to one which is so broad as to be almost meaningless.

"Normal climate" One of the often unwritten but necessary conditions for the origin of a peneplain (under the original meaning) is the so-called "normal climate", in other words a climate similar to that of northeastern North America and northwestern Europe where temperatures are moderate, rainfall well distributed seasonally, vegetation abundant, and chemical decomposition of the material of the earth's surface well developed. True, this climate is that in which a very large part of the civilized inhabitants of the world dwell, but from the areal standpoint it is certainly not that of the main portion of the present lands. We must look at a globe and not at a Mercator projection map to form an intelligent opinion on this point. Besides this fact, we must recognize the strong possibility that the present distribution of climates was not a permanent feature during the history of the earth. Evidence to prove this is not easy to obtain and rests largely upon inference. Soil profiles are not much help for many are not more than a few thousand years old. Marine deposits offer even less aid except insofar as they demonstrate wind and current directions. Hence we must turn to continental deposits and evaporites. With them the influence of now-eroded mountain chains must be evaluated. Besides this, many geologists offer the time-honored excuse of movement of either or both poles and continents. Whatever might be the correct conclusion on this debatable subject for the older geological periods, considerable evidence has been presented to demonstrate that the hypothesis of changes in latitude must be rejected for the Tertiary and Quaternary. Distribution of plants and of glaciation substantiate this. The occurence of glaciation alone proves that climatic changes of the first magnitude took place at that time. The later Tertiary is notable for the immense alluvial deposits of Western United States which must have been laid down under a decidely different climate than now prevails in the same place. It has often been suggested with considerable assurance than the present day wind and climatic belts still show the effects of the Pleistocene glaciation because of surviving icecaps. Such being the case it is best to forget about such a thing as a "normal" climate and to realize that much more of the globe may have once been semi-arid. We should then reject the idea that either aridity or semi-aridity is a "climatic accident".

Climatic control of debris removal. Fig. 1 shows cross sections of slopes in arid, semi-arid, humid, and sub-arctic climates. All but the last have in common the presence of enough rain to remove more or less completely the debris formed by weathering. In the truly arid environment weathering is almost wholly mechanical. When it does rain the water is not enough in amount or duration of flow to remove the debris of weathering from the area but instead it accumulates in alluvial fans and filling of enclosed basins. Both chemical changes and restraint by vegetation are at a minimum. Resistant crusts of chemical origin are formed. In a semi-arid region some chemical weathering is pre-



Cross sections of valleys in different climates

FIG. 1

sent but vegetation is not important. Enough rainfall occurs to keep the debris shed from steep slopes moving toward the sea or other base level. Much debris is water-born, the only proviso being that the particle size distribution be within the competence of running water. In a humid land, however, chemical alteration of the bed rock is very important. Although the average particle size is thus reduced, the presence of vegetation slows down removal. Mass movement is, however, very important. Slow erosion is especially conspicious where grass is present for all experiments demonstrate that it is by all means the most effective of all vegetation in restraining erosion. In a region of perpetually frozen ground the net result is to make all bed rocks and mantle rock alike into a solid, massive material. The seasonally thawed or "active" layer of the hills is moved in large part by mass movement to the streams.

Changes in climate. Due to the indubitable fact that climates change at any given locality it is expectable that we should find the characteristic climatic landscapes superimposed one upon another. Many believe that adjacent to the Pleistocene ice sheets vast areas were once frozen. Consideration of the heat requirements for melting of ice show that such could have been possible only during the advancing stages of the glaciers, if indeed it ever affected areas of marine climate. However, changes in amount of rainfall and vegetation can be and have been detected. Pluvial periods with more rain than at present have been postulated by many geologists in areas which are now semi-arid. Students of soils have also noted past climatic changes, particularly near to major lines of division due to climatic control. In this discussion, however, we will mainly concern ourselves with the later stages of erosion, the production of surfaces of low relief late in the progress of erosion.

<u>Davis' idea of the peneplain</u>. Fig. 2 shows two contrasted theories of the retreat of slopes. W. M. Davis held that the slopes on the sides of a stream

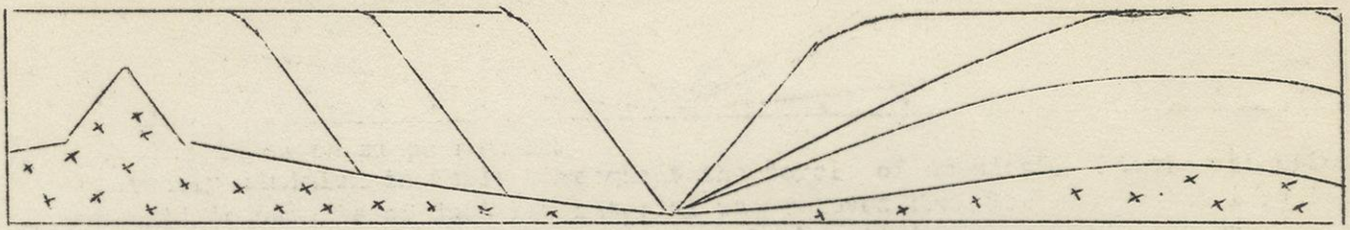


FIG. 2 Two ideas on slope retreat

constantly diminish in angle throughout the "cycle of erosion". Little attention was paid to details of just how material was removed from low slopes and less to the conclusion that a balance must ultimately be attained between the force available to remove material and the resistance of that material to erosion. Fig. 3 shows the original concept of the peneplain where it was concluded that

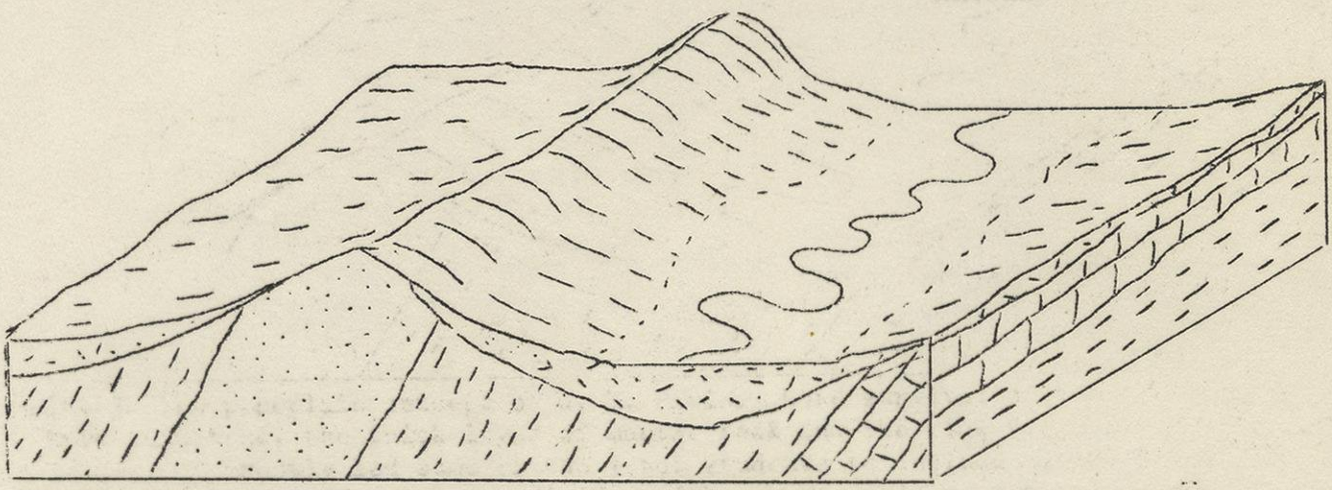


FIG. 3 The peneplain concept of W. M. Davis. Note survival of the ridge on hard sandstone, the thick layer of mantle rock and the wide floodplain. The last is what apparently led some of the later students to include depositional areas with peneplains. Note convex divides with any possible concave slopes buried under floodplain deposits.

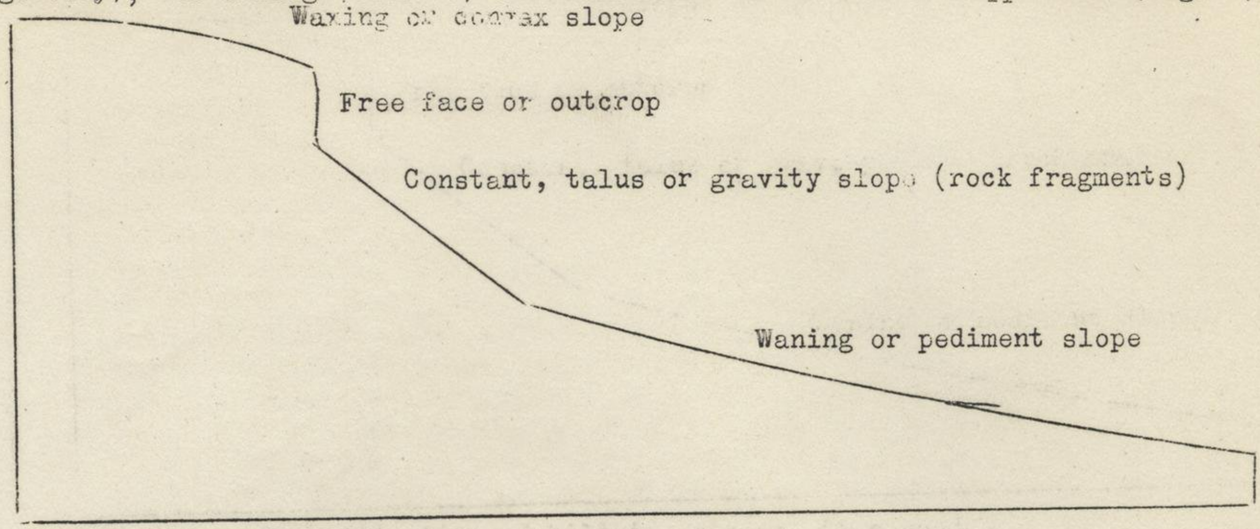
the streams would no longer be able to remove the debris of weathering as fast as it formed and would hence form extensive floodplains. A deep mantle of disintegrated rock was assumed to be present all over the area and residual elevations or monadnocks were left only where the bed rock was particularly obdurate to weathering and erosion. Elsewhere rounded convex divides should merge into the falts of the floodplains. The pre-Cambrian surface of Canada and north-central United States appears to fit fairly well with this concept, although we must recognize that it has been buried by marine sediments and later exhumed. There monadnocks are confined to extremely resistant materials, quartzite, hard iron formation, and fine-grained igneous rocks. Between these, slopes are in many places very low and divides are inconspicious. Bed rock is disintegrated to considerable depths not only in exposed areas but also where the cover of later rocks still persists.

Objections to the peneplain hypothesis. Other than buried and resurrected surfaces such as mentioned above, very subdued erosion topography is observable only on soft shales and on limestones where it is greatly aided by solution. The whole idea that a thick mantle of weathered material would form on surfaces of low relief ignores the head necessary to force water far below the surface. Absence of a deep residual mantle on the pre-Cambrian is generally ascribed to glaciation and it is true that in the lightly glaciated or unglaciated pre-Cambrian of central Wisconsin the mantle rock locally exceeds 140 feet in thickness in schist. The problem remains, however, to what extent was this due to chemical reaction by ground water while still buried. For that matter, how much erosion was caused by the waves and currents of the sea which transgressed this surface long ago! Other objections are of a more theoretical nature, Just how could debris be removed on very low slopes? Monadnocks should be rounded and grade into the adjacent landscape save perhaps where difference in bed rock geology is abrupt. A very serious objection lies in the apparent presence of old subdued surfaces near together and separated by a steep escarpments. Are these all explicable by differences in bed rock geology? Or is there something radically wrong in the hypothesis of origin of subdued erosion surfaces? Why did not the process that made the younger surface obliterate all those of older levels? Horton held that under his hydrophysical approach there must be "a definite end point for both stream and valley development." This point would be reached when the area between the streams is all within the belt of no erosion. Indeed Horton held that "most of the observed gradation of divides takes place before the streams which are separated by the given divide are developed--in other words, the terrain where the divide is located is graded in advance at a time when sheet erosion is taking place along or across the line which subsequently becomes the divide." He rejected entirely the idea that divides are graded down indefinitely. Horton also stated "The ultimate surface of erosion within a main basin boundary is neither 'almost a plane, as the prefix 'pene' implies, nor is it usually as close to being a plane as was the original surface area from which it has been derived. It seems better to call it a 'base surface' generally concave upward except along divides". Horton appears to have assumed soft material to considerable depths.

Parallel retreat of slopes. The theory that slopes do not lessen with time but retreat parallel to themselves after the initial formation was first presented by Penck and is shown on the left side of Fig. 2. This view requires the formation of a gently sloping surface between the foot of the steep slope and the channel of the adjacent stream. Material derived from the wearing back of the steeper slopes must be transported across this area by running water. This was the concept of the pediment, an idea also put forward by Gilbert from his observations in the semi-arid western part of this country. Davis did at one time write a paper on rock floors in which something of this theory was recognized although he rejected the idea of parallel retreat of slopes.

Strahler's equilibrium theory. Strahler used a statistical analysis of certain measurements in California and concluded that slopes lessen to a point where the adjacent streams can just remove the debris shed by weathering and fed into them by slopewash and mass movement. He found that these slopes have the same angle from top to bottom. It is apparent, however, that the area in the Coast Range probably represents a very early stage in the cycle of erosion, possibly prior to stabilization of slopes in relation to kinds of rock debris, each of which probably has a distinctive particle size distribution.

Wood's classification of slopes. Allan Wood's discrimination of types of hillside slopes into the waxing (convex), free face (outcrop), constant (talus or gravity), and waning (concave) was summarized in an earlier supplement (Fig. 4).



FIG° 4 Classification of hillside slopes after Wood.

Examples of each are found in almost all climates, although some may be absent at any given locality. We will first consider the methods by which each is formed and altered.

Convex or waxing slope. Formation of a rounded edge or convex surface on hill tops is not due to one process alone. It implies a removal of material toward lower ground at a rate which increases downshope. As pointed out by Davis long ago a sharp angle between original surface and hillside, such as is formed early in the cycle of erosion, is vulnerable since it is attacked by the agents of weathering from two sides. Once weathered, removal may occur either by slopewash or mass movement. Variation in intensity of rainfall causes the boundary of Horton's "belt of no erosion" to fluctuate in position. This should result in rounding off the corner. King has a similar idea for he states: "as the volume of water increases with distance from the crest of the slope and its speed downhill increases with the steepening declivity, there comes a stage where modification of the surface under the action of running water exceeds the modification due to soil creep. This is the end of the waxing slope." Soil creep is favored by this rounding off of the corner, by rock which weathers into a mantle which has low viscosity when wet, and by the presence of a restraining cover of sod or other vegetation which minimizes sheet wash. In the White River Badlands of South Dakota it has long been noted that convex divides occur only on the weaker layers.

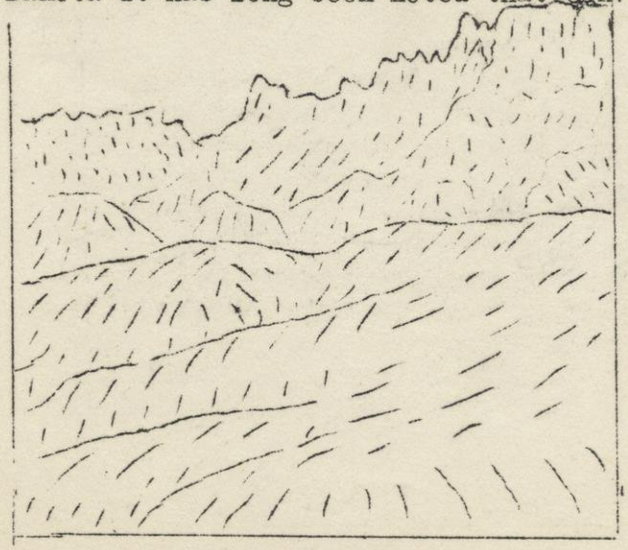


FIG. 5 Convex divides in White River Badlands of South Dakota from photograph by F. T. Thwaites. Note that these are confined to a certain soft stratum whereas the harder beds above make the craggy divides in the background. Note also the very steep sides below the convex crests which slope down to beds of ravines and in other places to true pediments. Small residual masses of the soft clay resemble haystacks.

Wherever firm material is present the divides are jagged and narrow. Convex divides are, then, best developed in humid lands with weak bed rock and abundant protecting vegetation. Rock exposures, other than large boulders moved from their original position, are rare in true convex slopes. However, convex slopes are not of universal occurrence.

Free face or outcrop. In the zone of the free face or rock outcrop it is evident that the debris of weathering derived from above must be moving with much greater speed than it does near the hill summit. Acutal outcrops can occur only where the bed rock is fairly resistant to weathering and are best developed in regions of horizontal strata, particularly where the resistance of different layers varies considerably. In the latter case there may be more than one such line of exposure. The actual type of rock forming outcrops varies with climate. In semi-arid regions we even find that gypsum, which is water soluble, is exposed because of its mechanical resistance. In very humid regions sandstone, quartzite, or fine-grained igneous rocks are common ledge-makers. Where slope development is reaching its endpoint, due either to a long time or to the weakness of the underlying material to both weathering and erosion the free face may be absent. Obviously this is most common where relief is low.

Talus, debris slope, or constant slope. Since the free face or outcrop is exposed to the elements it sheds fragments of rock. The size distribution of these depends upon bedding and jointing which is in turn an inherent feature of the type of rock. These fragments roll, slide, or fall into the slope below which is varyingly described as talus, scree, debris slope, or constant slope. The mechanics of this zone, which in many localities has a constant declivity, have been previously discussed. However, the fact that with most rocks and in most climates talus fragments disintegrate through weathering. The resulting finer material may be retained between the larger rocks for a time because of their protection and the restraint of vegetation. If there is enough moisture and clay has been formed, mass movement of the talus is possible. Landslides may then reveal the sloping surface of only slightly weathered bed rock which is the underlying basement of these slopes. This may reduce the slope of the lower part of the talus. If removal of material both thus and by rill erosion is not fast enough the free face above will be buried and talus formation will cease. Rill erosion is more probable than unconfined slope wash because the steep slope promotes high turbulence with associated channel erosion. It is the view of King that in South Africa such erosion is enough to cause retreat of the face of a hill so that the burial of the outcrop is postponed and the entire slope retreats at a constant angle, that determined by the size of rock fragments. Some talus slopes are interrupted by ledges where resistant formations have not been buried and by projecting butresses of bed rock which is more resistant than adjacent material. Rock outcrops may, therefore, be found in some places within this zone. Material which is removed from the talus only when its particle size is within that which can be transported by water on the available gradient, but it is evident that running water will be unable to decrease the angle of the entire slope because of the protection afforded by the larger rock fragments. To wear back a talus slope to a significant distance must involve weathering and erosion of its bed rock floor.

X

Waning or pediment slope. In many localities valley filling has obscured and buried everything below the talus slope. This is the case throughout the Driftless Area of the Upper Mississippi Valley and the cause is valley filling consequent upon nearby glaciation. In the Coastal Plain a recent rise of sea level has had the same effect and in much of the western part of the United

States climatic change has interferred with the normal development of hill slopes. In many places slopes are undercut by streams of considerable size which also prevents the formation of a concave lower slope. It is in semi-arid regions of sparse vegetation that these slopes are best observed and many of them were at first confused with the somewhat similar form of coalescing alluvial fans. Where typically developed these slopes are underlain by rocks which readily disintegrate to particles within the range of water movement. They have a thin veneer, locally absent at the top, of water-transported detritus which rests upon relatively fresh bed rock, The surface is scarred with rill marks which grade into less abundant ravines (dongas of South Africa). It is the problem of just how these smooth surfaces developed which is not yet solved to the satisfaction of everyone. Suggestions include (a) lateral erosion by streams which are at local baselevel fixed by a balance between erosion and deposition, for many grade into depositional slopes downhill; (b) erosion by many rills similar to those described from the talus slopes; and (c) erosion by sheet or slope wash including the sheet floods of McGee. King has gone out onto such slopes during rains to observe what actually happens. Higgins has dug trenches across little pediments and filled them with a different sand to check on rills vs. sheet wash. In rains of moderate intensity King found only clear water in the sheet flood close to the upper limit of the slopes. This disclosed laminar flow by having a depressed surface above obstacles. Just how such flow, which was not eroding or transporting material,could shape the pediment was a problem. Material eroded in the talus above must in this case have been deposited temporarily at or near its lower border. However, later studies showed that farther downslope and in heavier rains turbulent sediment-transporting flow is present, although deep floods like those described by McGee were not observed. It is obvious that to have sheet flow there must first be a smooth surface on which the water can spread out. King explains this by the multitude of small rivulets which descend the talus. He rejects the idea of lateral stream rosion largely because the great escarpments of South Africa are parallel to the coast, do not extend far up rivers. He thinks of them as originally as great monoclines which erosion has worn back parallel to themselves through several geologic periods at a rate of one foot in 150 to 300 years. He also rejects the stream erosion hypothesis because of the comparatively straight and level bases of the escarpments. However, this view does not seem to meet all observed conditions. The lateral extension of pedimented surfaces joining into a pediplain with only small residual, steep-sided hills rising above it implies recession of valley sides. In other areas it is evident that pediments have formed along fault scarps. Movcover, some form of channel erosion would seem a prerequisite for preparing the ground for widespread sheet floods. Possibly Horton's theory of rill grading

FIG. 6 After King Steep-sided residuals of granite rising from smooth pediment which has a thin grass cover. Slopes of hills are talus blocks. Similar residuals are common in The Great Plains. From photograph. East of Pietersburg, Transvaal

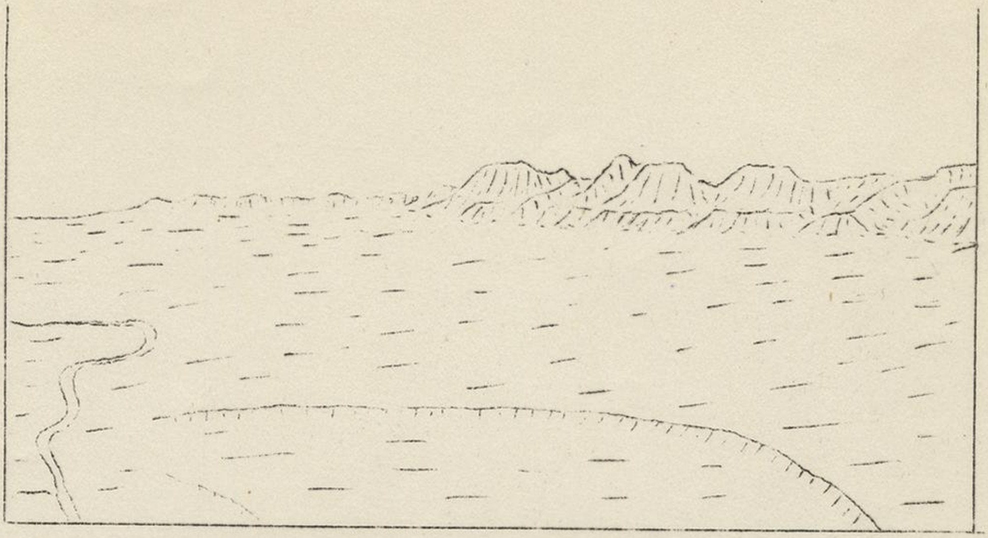


FIG. 7 After photograph by Fair published by King. The river has no real flood plain but pediments rise gently to the steep-sided residuals of andstone and dolerite (basalt). No convex hilltops can be distinguished but the concavity of the pediment is plainly shown. Vegetation appears to be scanty brush, possibly some thin grass. The Karroo, South Aftrica.

is the key. But when all is said and done the reality of these rock-cut slopes must be admitted. They do not fit in with the old concept of peneplains. They could explain preservation of remnants of more than one erosion cycle in adjacent hills for on top of the remnants erosion is very slow. They do not require for formation a very arid climate and might occur in somewhat modified form in humid regions unless deeply buried by crept mantle rock. They explain the apparent youthfulness of the mountains of the Basin and Range province despite the width of valleys, a fact which puzzled early students of that area.

Form of pediment cross section. Pediments have a characteristic concave cross section leading down from the more or less level, abrupt upper limit either to streams or to an alluvial fill in the center of the adjacent valley. Wide stream spacing may be a factor in pediment formation. The sharpness of the upper contact is best developed in hard rocks. In weak rocks this contact is a gradational curve. The various causes of the concavity due to running water have been explained in a previous supplement. The matter is not simple and is unlike conditions on alluvial fans for rain falls all across the pediment slope giving increased depth down slope with consequent decrease in shearing force. Indeed, it has been declared that pediment slopes are formed in order to facilitate disposition of sudden heavy downpours which are common in semi-arid regions. As pediments join at divides the divide is commonly abrupt and angular rather than rounded, although both forms may occur apparently depending upon the resistance of the bed rock. The best-developed pediment profiles occur where the bed rock is granite rather than soft sediments such as shale or limestone. Residual elevations within a pediment or pediplain (area of coalescing pediments) characteristically have concave sides. Mount Monadnock, New Hampshire, rises in this fashion from adjacent uplands of the same kind of rock. However, this area was glaciated and a basal mantle of decomposed rock might have been eroded by the ice or the base might have been eroded by waves. A feature of pediment slopes is that gullies (dongas of South Africa) occur entirely on them rather than on higher slopes locally extending to the upper border. Some change to low alluvial fans below. It is thought that these ravines are due to local concentrations of the sheet flow which set up more turbulent flow which causes erosion. Some are certainly due to disturbance of the ground by farming. Rock outcrops occur in the walls of such gullies, at the head of the slope of pediments, and in small isolated "islands" or residuals. The only cause of convex profiles in pedimented areas is erosion at an accelerating rate due to later uplift, or to climatic change toward greater humidity. In this connection we may ask if erosion surfaces which bevel the bed rock and yet show deep weathering are (a) pediments developed in humid climates or (b) pediments which have been altered by a change of climate. Since the theory of pedimentation can explain the occurence of several different levels in the same region it opens up many new possibilities in interpretation. Could it be that the Piedmont Plateau of southeastern United States is a pediment whose surface was later eroded by a more humid climate possibly associated with uplift? Such a view would explain the anomaly of stream capture along the youthful divide of the Blue Ridge to the northwest, features which seem impossible under the peneplain hypothesis. The convex divides of the Piedmont together with deep disintegration of the bed rock would be more recent than the original bevel. Widespread gravels of late Tertiary age in the Coastal Plain seemingly support this view. The Harrisburg terrace, which is so conspicious throughout the entire Appalachian region, would then be correlated with the Piedmont and possibly also the Highland Rim surface west of the high plateaus. Many will object to this suggestion because it seems to imply a marked climatic change, but just how much of a change is debatable. Perhaps only enough to affect the vegetation cover to a moderate extent. Turning to the Rockies, it is obvious that the upland surfaces are true pediments correlated with alluvial filling of adjacent lowlands. Climatic change, possibly associated with, or due to, uplift, has removed much of the fill but a remnant persists in the Gang Plank west of Cheyenne, Wyoming. Surely, it is inappropriate to call the upland surface a peneplain if we stick to the original meaning of that word. Some in the Uinta Mountains have in part been described as pediments. Throughout the Great Plains many of the residual hills have steep concave sides which appear to demonstrate pedimentation.

<u>Summary</u>. The following table, adapted from King shows the differences between what may be inferred as characteristics of peneplains (under the original Davis view) and those of pediplains. We must note that peneplains are inferences, whereas pediplains may be actually observed in the field. Moreover, it seems doubtful that there can be any sharp line of division on the basis of either climate or kind of rock. King declares that the peneplain,as originally defined, is an "imaginary landform", so that it may be that debate is futile. The exact method of formation of the theoretical peneplain is only vaguely described in the literature and is not backed by actual observation. A factor in comparison, which King suggests, is that the mantle of grass which so effectively restrains erosion and makes for convex divides was not present prior to the middle Tertiary. Indeed, others have suggested that vegetation on the lands was absent in the earlier geologic periods, and that erosion was then everywhere like that of semi-arid

regions of today.

Peneplain (Davis, theoretical)	Pediplain (observational)
Broad flood plains.	Narrow flood plains.
Convex or subdued divides with much creep of a deep mantle.	Divides sharp with concave slopes on both sides, locally convex over a narrow width.
Residuals gentle and convex.	Residuals sharp with concave sides except where top is very weak rock.
Lower slopes only, concave.	Dominantly concave slopes, except on very weak rock.
Origin by slope flattening.	Origin by scarp retreat and pedimentation by running water.
Origin destroyed all older surfaces.	Several levels may be present in one locality.
Mantle rock due only to weathering and creep.	Mantle rock thin, and water-transported.
Bed rock deeply weathered(?)	Bed rock fresh.

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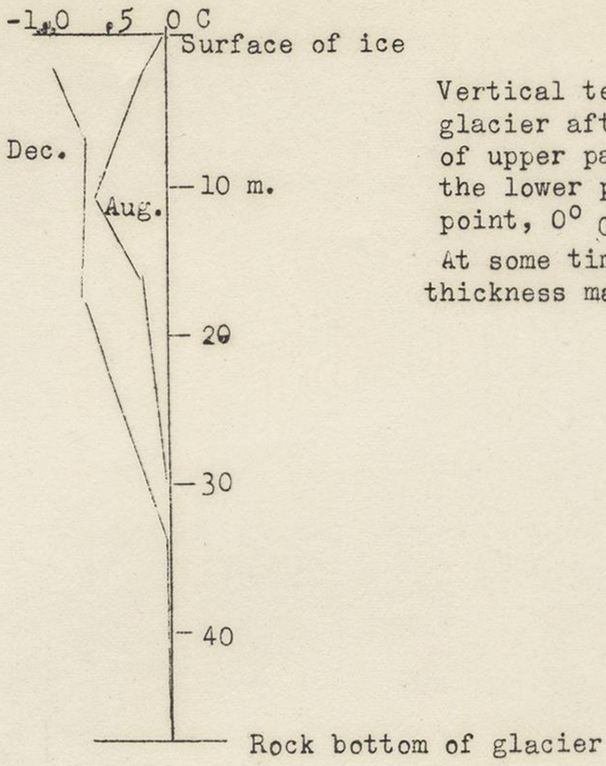
<u>Introduction</u>. The subject of the physics of the flow of glaciers has received much attention in recent years. A large portion of the results are scattered in publications which are not readily accessible to American students and many are in other languages than English. However, a series of papers on phases of the subject appeared in the <u>Journal of Geology</u> of 1951 and in all numbers of the British <u>Journal of Glaciology</u>. Results of these researches are not all harmonious and the following is an attempt to offer suggestions on the meaning of these differences. Field investigation has been carried on chiefly in Switzerland but also in Alaska, Baffin Land, Norway, etc. We cannot here summarize all this highly technical data, but a number of problems can now be outlined. Chief of these on which there is much difference of opinion, we may list: (a) is extrusion flow as postulated by Demorest a fact, (b) to what extent is melting and refreezing a factor both in motion and in glacial erosion, (c) does ice flow like a plastic substance where the "coefficient of viscosity" is a function of applied force, (d) how is the crystal structure of ice related to flow, and (e) is the flow of ice due to just one process or to a combination in varying proportions depending upon conditions which prevail at certain places and certain times in varying relations.

<u>Extrusion flow</u>. The hypothesis of <u>extrusion flow</u> of ice at depth was promulgated by the late Max Demorest and adopted by a few other American students. It seems to have met with considerable skepticism abroad, particularly with British glacialists. Here we must note that <u>glaciology</u>, the study of the physics of glaciers, is not identical with <u>glacial geology</u> which is primarily the study of glacial and glacioaqueous sedimentation and the resulting land forms. One of the basic concepts of the hypothesis of extrusive flow is that the viscosity of ice decreases with load so that the basal portion of a thick glacier is more fluid than the upper portion. This upper rigid portion obeys the laws of solid rather than those of fluid mechanics. Its thickness would then measure the <u>threshold stress</u> where the change takes place. Now the existance of such a transition is abundantly supported by experiments with both metals and the softer rocks. The disagreement seems to rest mainly on the problem of whether or not the velocity of motion of thick ice increases downward or upward. It is well recognized then in mountain or alpine glaciers (also termed valley glaciers) the increase is upward, although it is concluded that the surficial rigid ice which breaks into crevasses rides along on the more mobile bottom ice. In deriving the formulas for velocity based upon analogy to laminar flow of a fluid integration is carried on from bottom up with a maximum toward the top of the ice just below the rigid zone. However, in the case of extrusion flow it would be necessary to reverse the computation and integrate downward from this point thus placing the maximum velocity near to or at the bottom of the ice. Apparently it is this reversal of rate of increase plus difficulty in seeing why the surficial rigid ice does not travel along on top of the more fluid zone which has led to doubts. In the very nature of things extrusion flow is an invisible phenomenon. Obviously the best practical check is to drill a deep hole in a glacier and find out from repeated surveys where velocity is at a maximum. Such holes have been made in relatively thin alpine glaciers but the completion of one in a thick polar glacier has been delayed by drilling difficulties in very cold ice. It has long been recognized that the upper portion of a moving glacier yields by breaking into crevasses. The problem is whether or not crevassed areas are the only parts of continental glaciers

which move. This opinion is still widely held but it is asserted that it would be difficult to account for the observed discharge of ice from Greenland glaciers and some other thick ice masses without a deep-seated flow beneath unbroken ice. It may also be argued that glacial erosion of the deep basins of the Great Lakes and Finger Lakes of New York would be hard to explain if motion of the continental glacier were confined to a marginal zone. A similar remark could also be made in respect to the long distances of transport of some glacial erratics. Demorest explained the failure of the top central zone of a continental glacier to move by his idea of obstructed flow. Under this theory the greater resistance of the rigid ice, which descends to the surface at the margin of the ice sheet, would restrain any tendency of the surificial ice and firn to move out of the central area. It is not due to cohesion for ice yields readily to tension. It seems to the present writer that this is perhaps the greatest objection to the idea of extrusion flow which is otherwise a valid working hypothesis.

Freezing and thawing due to pressure. Many criticisms have been urged against pressure-controlled melting followed by refreezing. The process would seem a prerequisite to extensive glacial plucking of bed rock, which very slight consideration of energy requirements shows must indubitably have been much more important than grinding from the quanitative standpoint in the glacial erosion of hard rocks. Fortunately for the argument pressure melting and refreezing are easily demonstrated in ice which is at the melting point. One can press two ice cubes together in the hand and they stay frozen upon release. In the time-honored experiment of the passing of a weighted wire through a block of ice, it is evident that the energy of the falling weights furnishes the requisite 80 calories of heat absorbed in melting a gram of ice. Above the wire the opening freezes shut and each gram should set free the same amount of heat which was required to melt it. With the slow motion of the wire this heat is dissipated into the adjacent ice and to the atmosphere thus representing the final disposition of the available energy. The change in pressure is just sufficient to permit this heat exchange. Now within a thick glacier which is in motion we must give thought to the fact that the net pressure on any given point is not only the weight of the column of ice and firn above but is in fact the resultant of this force and that of motion. Even if the weight does not change, a very slight alteration in rate of motion must indubitably affect the net pressure at a given point. Turning to the hypothetical temperature curve of a thick glacier whose top is in a cold climate, Fig. 1, 1949--50, it seems impossible to think that a very large portion of the deeper parts of a continental glacier can be at any other temperature than the pressure-controlled melting point. Any possible extension of the known portion of the near-surface curve must intersect the melting curve because ice cannot exist above that temperature. The writer ventures to suggest that pressure melting and refreezing is a vital process not only in glacial plucking but also in glacial motion. May it not well be that the thickness of the cold ice near the surface is an important control in extrusion flow and hence in stagnation of a decaying ice sheet which thinned to the point that there was no thick basal zone at the melting point? Such a case could occur where the climate is cold. On the other hand the lower southern margins of a continental glacier must have behaved like temperate glaciers which are often at the melting point throughout. May not these be the location of a different type of flow, perhaps similar to Demorest's gravity flow of valley

Fig. 1 shows temperature curve of a temperate glacier (next page)



Vertical temperature curve of a Swiss glacier after McCall. Temperatures of upper part vary with the season but the lower part remains at the melting point, 0° C. Depths in meters At some time of the year the entire thickness may reach that point.

glaciers? Could not the difference between extrusion and gravity flow be thus explained? Before leaving the subject of glacial plucking, attention may be directed to the necessity of assuming that crevices are present in the bed rock. It is true that most water wells in granite are dry if no crevices are found within about 200 feet of the surface. But dry crevices have been found at much greater depths and there seems no necessary limit to the depth at which shearing may take place along faults. If holes may be drilled to depths of over 20,000 feet in sedimentary rocks we certainly cannot set a downward limit to open cracks in hard crystallines. Another factor which must not be neglected is that erosion lightens the load on creviced rock. If the erosion merely replaces rock with density about 2.6 with ice at density below 0.9 we have still furnished a reason for cracks to open up and admit plastic ice or water which freezes into ice.

Viscosity of ice. Almost all physicists now admit that the coefficient of viscosity of ice is not a constant as it is in a true liquid. In fact some desire that the term be abandoned in favor of another. Most measurements agree that velocity of strain is a power function of stress, with an exponent variously given from 2 to slightly over 4. It is suggested that the value of the exponent is a function of both temperature and nature of the ice. Glen conducted some carefully controlled experiments by stretching ice crystals at mean temperature of 1.5 c and arrived at the expression:

strain rate (in years) = 0.0074 stress $(\text{bars})^4$, 1 bar = 10^6 dynes/cm^2. Threshold stress is about 1 bar corresponding to a thickness of 11 meters of ice, much less than the postulated value given before. Observations in tunnels have also

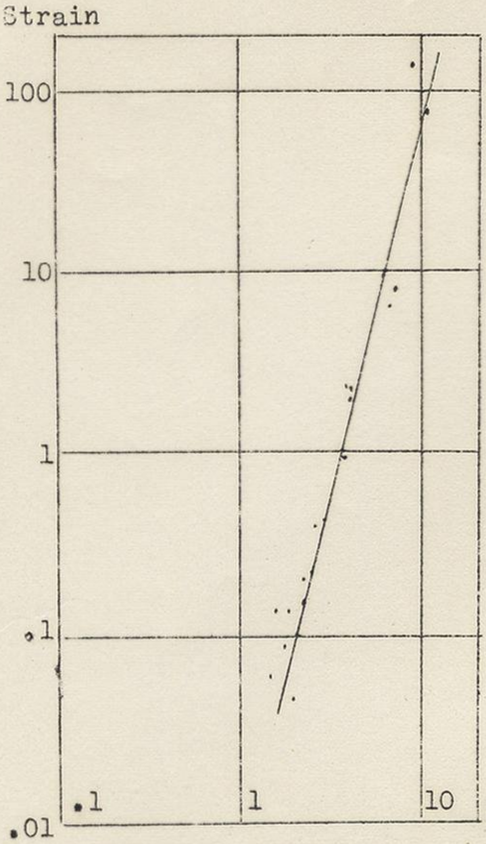


Fig. 2 (left) Strain rate in cm/yr vs stress in bars after Glen Log-log plat showing value of exponent, n.

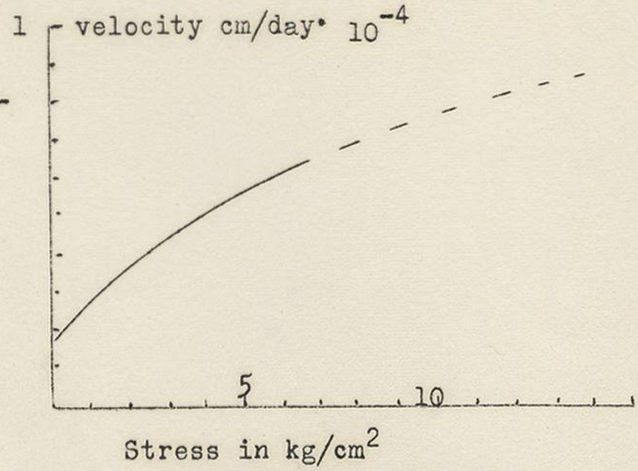


Fig. 3 Strain-stress relations in drill hole through a Swiss glacier after Perutz Although top 50 m of ice did not behave as a fluid value of the threshold stress is low.

been made by measuring the rate of contraction of the sides. Very similar results with a threshold value of slightly over 10 meters of ice are reported by McCall. Haefeli also describes measurements on rate of tunnel contraction as well as by direct determinations of the rate at which a metal ball was intruded into the ice. Calculations based on the former gave results for viscosity ranging from 8 to 3.7 $X10^{14}$ poise, decreasing toward the interior of the ice. He remarks "The tests with the sinking ball confirm the importance of pressure melting in glacier flow. Owing to local pressure concentrations the fluidity and thus the deformation of the ice is locally increased." Nye gives the formula for force F at depth d on slope S with unit weight of ice **w**, **was:**

$F = w\ d\ S$ (on unit area)

By integration the difference of **top** velocity (U_o) and bottom velocity (U_b) of a glacier with total thickness D is:

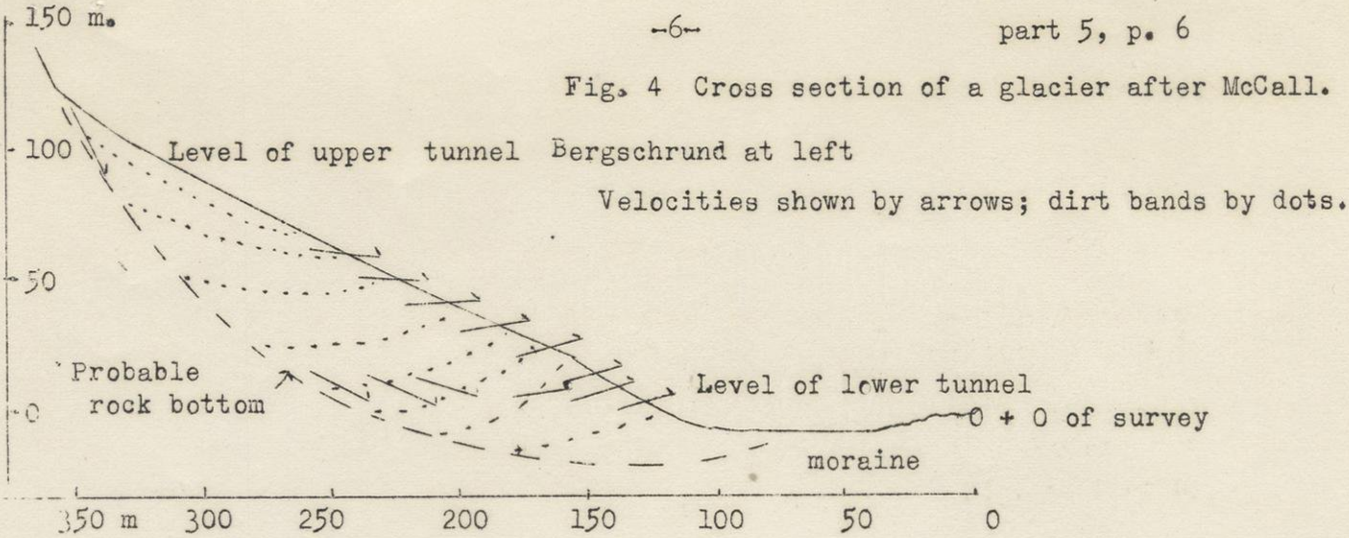
$$U_o - U_b = \frac{W^n}{B\,(n+1)}\ S^n \cdot D^{n+1}$$

B is a constant. Note that W/B = reciprocal of the coefficient of viscosity. With strain rate cm/year and force in bars (within the range of 0.8 to 5.5 bars) B = 1.62 and n = 4.1 but the internal physical state of the ice and the temperature cause changes in these values. It is worthwhile to compare the above formula for velocity with that derived for laminar flow of a true liquid. Bader suggests that viscosity is controlled by temperature, stress, grain size, grain shape and grain orientation. It is to be noted that some authorities now admit a distinct bottom velocity in an alpine glacier, a fact of great importance to glacial erosion whatever the process by which that goes on.

Crystal structure of ice. Much work has been done on the crystal structure of ice both in the field and in the laboratory. We can mention here only the papers by Bader, Rigsby, and Bjerrum, the last of which apparently made use of the X-ray. The large size of some of the individual crystals is remarkable for they measure many inches across. Bader remarks: "it is difficult to visualize shearing along tortuous grain boundaries as an important factor in grain deformation". Nevertheless there is a marked preferred orientation or schistose structure although it is poorly developed compared to that of a mica schist. Dead ice, which is no longer moving, has larger crystals than living, moving ice which contains bubbles of air and water. Bjerrum ascribes plasticity wholly to gliding on planes perpendicular to the optic axes of the crystals. Rate of gliding is related directly to temperature. Threshold stress was estimated at about 5 kg/cm^2 which corresponds roughly to about 55 meters of ice. He also calculates the heat of sublimation of ice at 675 cal./gm. In conclusion it seems that nothing has been discovered to contradict former conclusions on ice flow by recrystallization or crystal gliding except that the threshold stresses are decidedly less than that postulated by some older students of the subject.

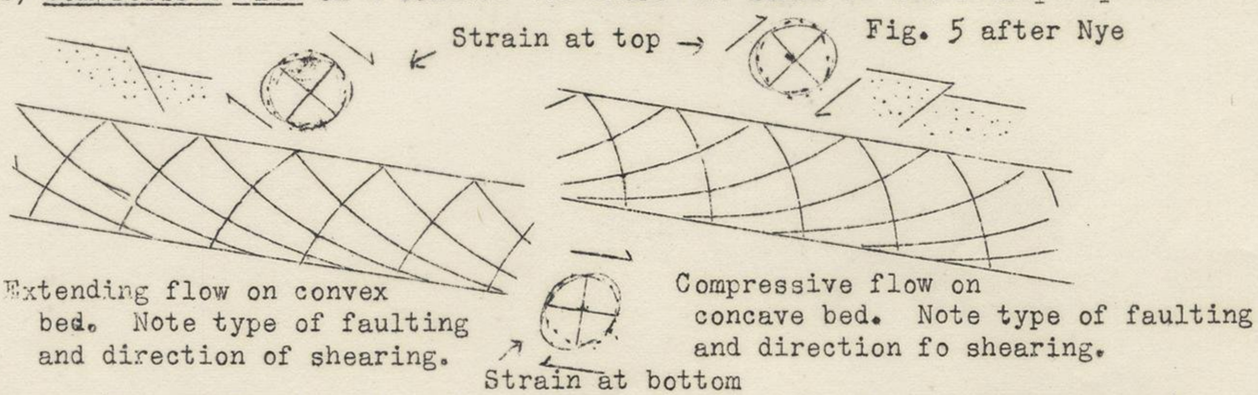
Conclusion. In conclusion of the subject of glacial flow it appears to be well established that a number of processes must occur either at once or under somewhat different physical conditions. We may summarize that motion can be by: (a) mechanical shear, (b) pressure melting and refreezing, and (c) recrystallization or gliding. The dominance of any one process is certainly related to temperature, total stress, rate of stress change, and physical makeup of the ice in amount of air bubbles and possibly the amount of water-soluble salts in water bubbles. Nothing which has been discovered prohibits the occurrence of extrusion flow at depth, although it must be admitted that such a theory offers certain mechanical difficulties. It is in fact possible to set up a theory of motion of both valley and continental glaciers by repeated gravity slicing along planes which dip outward from the center or source. (See section on soil mechanics) Lewis suggested such an explanation in cirque glaciers by a method of sliding which strongly resembles the hypothetical analysis used by engineers to establish the safety of the side of an artificial excavation. Such a theory of rotational sliding has been combined with yield by flow rather than by fracture. Some positive evidence to support its reality has been collected from motion of the ice in the sides of tunnels and from the dip of apparent shear planes in the ice which are difficult to explain as bedding or accumulation planes. However, McCall found that (a) ice movement near to terminus of a cirque glacier has a marked upward component, (b) motion of basal layers is not always in the same vertical or horizontal direction as that near the surface, (c) there is no effective over-thrusting as demanded by the theory of rotational slip, (d) ice is quasi-viscous and (e) there is no evidence that debris is being raised into the ice from the ice bottom. See Fig. 4 on next page.

Fig. 4 Cross section of a glacier after McCall.

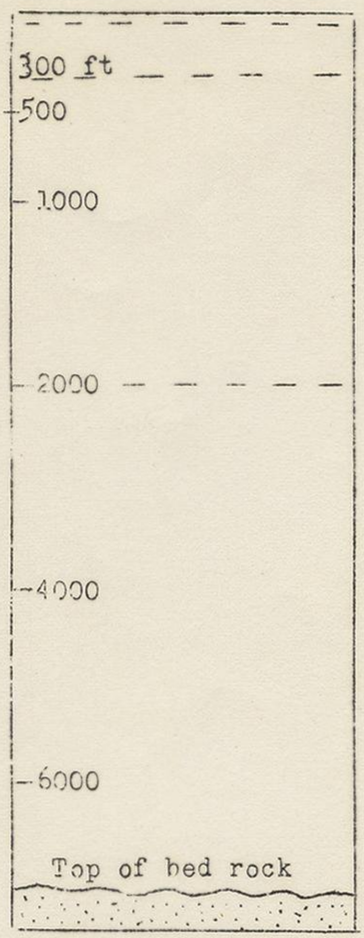


Nye made a theoretical analysis of the changes from normal laminar flow such as postulated by Demorest as unobstructed gravity flow into two phases: (a) extending flow on a convex bed where there is a marked slope of lines of shear downhill at a greater angle than the slope of either the bed or the top of the ice, and (b) compressive flow on a concave bed where the lines of shear slope upward

Fig. 5 after Nye



at a greater angle than the slope of the glacier. The latter is Demorests' obstructed gravity flow, a conclusion checked by Nye's statement that the former is characteristic of the accumulation area and the latter of the terminus of a glacier. The figure here given explains the phenomena by the aid of the strain ellipsoid and Nye's mathematical analysis may better be omitted. But the observations in ice tunnels seems to show very definitely that ice flow is predominantly due to shear within crystals and to a considerable extent to pressure melting and refreezing in all areas where the ice is just at the melting point. In considering the threshold stresses outlined above it is well to recall that such were given in meters of ice with density of 0.9 thus giving a much lower figure than the actual overburden of snow and firn. In this connection it is also clear that actual measurements of the depth of crevasses in the brittle surface ice have almost everywhere shown that previous estimates of depth were far too great. The difficulty in finding the bodies of persons who, like Demorest, perished by falling into crevasses may be due to very soft snow in the bottom.



Surface Thickness down to solid ice unknown in accumulation area. Ice mainly too cold to flow

No differential flow noted in Malispina Glacier above 300 ft.

Ice always below melting point to estimated 2000 feet in a true polar glacier. Flow by gliding possible Fluidity increases with depth.

Depth of this line depends upon mean air temperature of top, and conductivity of ice. Gradient measured only near top in Greenland.

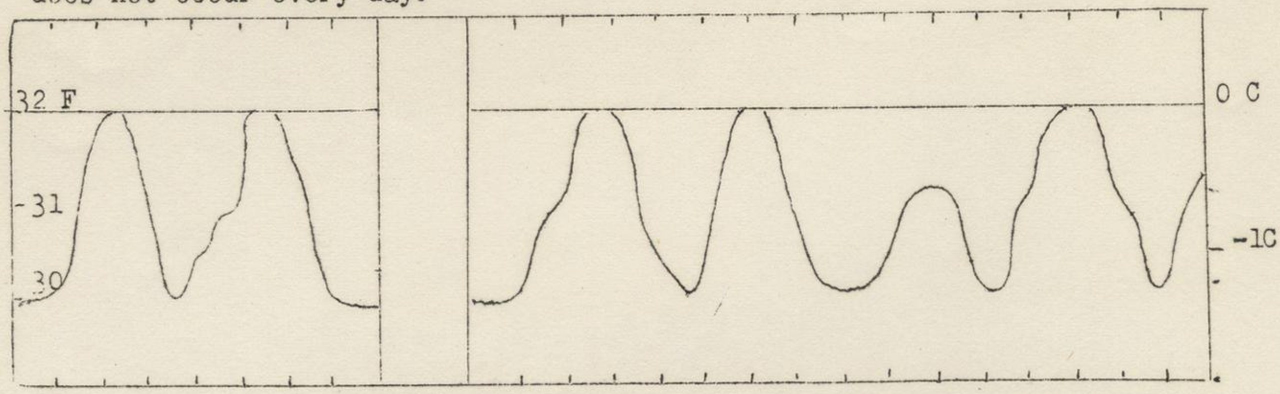
Zone where ice is at pressure-controlled melting point. Flow by gliding important but local pressure melting and refreezing can occur. This is zone of maximum fluidity (minimum coefficient of viscosity). Demorest held that the zones above cannot move and that hence velocity must increase downward giving extrusion flow. No observations display this conclusion and the major difficulty with the view is how the surface zones could remain immobile. That fluidity increases with depth is not important.

Figure 6 (top part not to scale)

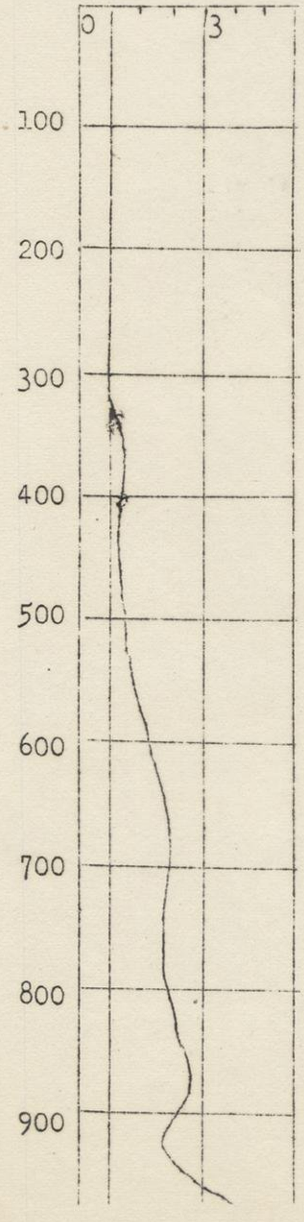
Fig. 6 is an attempt to define the zones of a polar glacier in which the different processes of flow should be dominant. In concluding this controversial subject we may recall that science thrives by the introduction of new ideas and hypotheses. Some of them may be false, but nevertheless they compell us to overhaul our knowlege and to complete our proofs. Unfortunately, many students of science, who ought to have known better and admitted ignorance, evade the areas where their personal knowledge was lacking. Witness the many prophecies of the imminent exhaustion of all petroleum resources, the condemnation of certain little-known areas before testing, the conclusion that a machine heavier than air could not fly, that atomic energy could not be released, and that holes could only be drilled to a modest depth until filled by rock flow. Let us beware of falling into the attitude that what we do not observe or do not know about does not exist.

Temperature in the bergschrund. An long argument has been waged as to the presence of freeze and thaw within the bergschrund due to actual air temperature change or to the entry of meltwater from above. Recently some actual observations show, as might well be expected from the presence of so much ice, that air

Fig. 7 after Battle and Lewis
Temperatures measured in a bergschrund Note that the melting point is never exceeded although no very low freezing temperatures were observed. Melting does not occur every day.



temperature rarely if ever attains more than 0 C. (32°F.) Higher temperatures during the days outside are evidently due to winds which do not penetrate underground. Thus the melt water hypothesis of Lewis appears to be well substantiated. Fig. 7 above



Postscript. Just as the above was being completed Sharp published the results of a 1000 foot hole in the Malispina Glacier, Alaska. This test was drilled by the thermal method and the hole was lost at about half the estimated depth of the glacier by reason of failure of the apparatus. Absolute geographic position of the hole cannot be found because of the distance to fixed landmarks, hence the deformation shown in a year is only the net change of inclination which does not exceed 6 feet. No change was found above depth 300 feet. No evidence of extrusion flow was found. It is difficult to see what would keep the surface ice from moving. This is a piedmont and not a true continental glacier and the hole is not in an area of accumulation.

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See also: Gerrard, J. A. F., Perutz, M. F., and Roche, Andre, Measurment of the velocity distribution along a vertical line through a glacier: Royal Soc. Proc.: A 213: 546-558, 1952

Fig. 8 after Sharp. Deformation of a test hole in Malispina Glacier, Alaska. Distances in feet

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Personal

GEOLOGY 109

GEOMORPHOLOGY

Penck, Walther, Morphological Analysis of Land Forms, MacMillan and Co., London, 1953.

Supplements, 1954, Part I

Introduction: The above work by Walther Penck has at last been translated into English. Its fame rests upon the challenge it presented to the hypotheses of W. M. Davis in presenting a different version of the relation between uplift and erosion, including application of this to the explanation of different forms of slopes, and by that the history of uplifted areas. On account of the difficulty of reading from the original German, Penck's views appear to have been in many cases misinterpreted. The text is still difficult to read even in the excellent translation of Hella Czech and Katherine C. Boswell. The following is an attempt to relate the conclusions of this controversial work, published after the death of the author, to some more recent ideas.

Primary Approach: Penck's primary approach is the relation in time between crustal uplift and erosion. Davis and most of his followers presented the cycle of erosion as alteration of a surface rather suddenly upraised and and then subjected to erosion under climatic conditions similar to those of the better populated portions of the globe. Such a view Penck held to be inadequate. He stated that the surface form of the earth is the net result of the conflict between forces which cause elevation and those which bring about degradation. These he felt could not be separated in time and the topography resulting from from erosion must be explained on the basis of the ratio between these opposed processes. He also called attention to the importance of study of the eroded materials which were deposited concurrently with the erosion. "The physical character of the morphological problem comes out clearly. The task before us is to find out not only the kind of formative processes, but also the ratios of their intensities with respect to one another. None of the usual geological or specifically morphological methods is sufficient for the solution of this problem. As is now self-evident, it requires the application of the methods of physics." (p. 3) Penck's application of physics involves what he terms the differential method or study of the progress of a process by considering small increments, each developed in a time unit. He states that this is like the methods of differential calculus (p. 13) but it should not be assumed that solutions were attained by use of higher mathematics. Instead, many drawings were used which are rather difficult to comprehend and almost impossible to reproduce by mimeograph. Penck's discussion of processes of weathering is, for the most part, in accordance with present day knowledge. He stressed the increase in mobility of the weathered material due to the formation of clay minerals and colloids. This refers to mass movement and not to ease of water erosion. Penck's renewal of exposure means the downward growth of the zone of weathering as material is removed above.

Mass-movement. Removal of material from slopes was regarded by Penck as due almost wholly to mass movement, the material being thus fed into the water courses. He uses the term denudation for lowering of slopes between such channels by removal of weathered debris, and erosion for the work of running water in them. Intensity of denudation is the term given to quantity of broken up material removed in unit time. Penck did not accept the explanation of "Block Seas" or boulder trains as products of a periglacial climate, pointing out that he had found the same phenomenon in Uruguay and Brazil. He related them to the kind of rock and a gradient of 15 to 30 degrees. Resistance to denudation comprises cohesion, friction, and the roots of plants. Methods of movement are listed as: Free, including dry rubble on slopes of only 2 or 3 degrees; Slumping, which may occur on forest or grass-covered slopes, and Soliflucition which is flow of material above frozen ground. Penck regarded the stone rivers of the Falkland Islands as similar to those of the Baraboo District, Wisconsin: concentration of stones due to erosion of the finer material between them. Bound-down movement could occur even beneath a cover of vegetation. Corrasion is the term applied to wearing of the substratum by mass movement of debris. Penck ascribed the smooth upper parts of valleys which lack definite water courses to this process. The present writer has observed that on fairly steep slopes there is almost everywhere a sharp contact between the mantle rock and the bed rock instead of the transition or gradation described in most text books. But nowhere has he observed any sign of marked wear of the bed rock surface, although in many instances there is a bending down of the layers of the weakened bed rock. Penck nowhere mentions the conditions necessary for removal of debris which is produced uniformly over an entire slope. Gilbert long ago deduced that with a reasonably uniform thickness of mantle rock this necessitates a downslope increase of gradient thus leading to convexity. Convexity of this type may be observed in many localities of this country, for instance all over the Driftless Area of the Upper Mississippi Valley and the Piedmont Plateau. Penck differs with others in minimizing the effect of climate on land forms. Despite his travels in Asia Minor and in Argentina he concluded that there is no fundamental difference in land forms in humid and arid areas. He thought that corrasion by moving mantle rock is more important where vegetation, aridity, or a high rate of infiltration of rainfall impeded erosion by running water.

Formation of slopes. Perhaps nothing in Penck's work has been more consistently misunderstood than his treatment of slope formation. One of his cardinal principles is that slopes decrease in gradient with time unless renewed by erosion by running water. He might well have included wave work and earth movement but these do not seem to be mentioned. He classified slopes into straight, convex, and concave and set out to use the association of these types in certain regions to decipher their geomorphic history. He also mentioned that slopes of a certain inclination appear to be associated with definite regions.

For his first analysis Penck chose a valley wall where the stream at the base is neither deepening nor filling its bed because it can exactly take care of the debris furnished to it. A diagram then demonstrates that, as the cliff or rock face is weathered away, a talus accumulates below. This talus protects the underlying rock so that as the outcrop is weathered back there develops a sub-talus rock slope termed Halenhang. The final result is the formation of a slope at a less angle than the cliff above. The face of the cliff keeps the same inclination as it retreats until the talus slope finially meets that of the next adjoining valley forming a sharp divide not rounded to any material extent. The next step in formation is that the older basal part of this talus slope weathers. By this weathering it acquires mobility and begins mass movement which reduces its slope. In time this lower slope extends farther and farther back reducing the slope of the valley side. Thus was propounded the law that slopes flatten from below upward until the final result would be such low slopes that mass movement would not be able to persist. Eventually this would lead to what Penck terms the end-peneplain (Endrumpf), the same as the peneplain of Davis, but Penck deduced concave slopes and sharp divides. To the present writer it is difficult to see just how Penck allowed either for the increase in volume of the talus over that of the rock in place or for the volume changes arising from weathering. In one place he states that the thickness of mantle rock diminishes uphill, which would increase the rate of denudation of the divides. It must be realized that this retreat of cliffs or free-face of Wood at constant slope has been confused by several authors as the retreat of talus or "gravity" slopes. This is a fundamental error, as was the claim by some that the lower slopes are pediments where running water is important. No such statements can be found in the translation.

The second case considered was when the intensity of erosion was increasing or waxing and the river therefore deepening its bed. Similar analysis disclosed that this leads to a convex slope. It was concluded that such began at river level and worked upward because of the higher rate of denudation of the steeper slope. The higher, and gentler slopes were then left undisturbed and are independent of the base level of erosion. Conversely, concave slopes were regarded as the result of decreasing or waning activity of erosion, although they can be formed with a stationary baselevel. To the present writer the major weakness of this painstaking analysis is that it ignores slope wash and rill erosion, putting all slope formation the result of talus formation and mass movement.

Effect of variation in rock resistance. Penck described the scarplands of Germany where there are alternating permeable resistant formations with intervening impermeable soft strata. The result is a series of terraces. Penck ascribed the flattish areas on the soft strata as "peneplains" developed by the control of the local baselevels due to the outcropping resistant formations. He specifically states that such "peneplains" are simply areas of low slopes and are not peneplains in the sense that the term was used by Davis.

Inselberg landscapes. Penck held that since the lower slopes of isolated hills or inselbergs are concave they are the product of waning erosion with a constant base level. He denies that they are characteristic of any particular climate or type of rock. They are not monadnocks of very resistant rock masses. He concedes that in arid climates the basal slopes are more concave than elsewhere. It is not clear how he explained removal of material at the bottoms of the steep sides to maintain their slope.

Piedmont benches. A point on which many differ with Penck is his explanation of the benches which form a stairway below certain higher mountains. These he called piedmont benches (Treppen) and ascribed to erosion on the border of a constantly uplifted area. The surfaces were regarded as peneplains which grew horizontally, leaving the higher benches or primärrumpf almost unaltered. But such surfaces were not thought of as end-peneplains of the Davisian variety. In Davis' criticism of the theory it was stated that Penck postulated uplift at an accelerating rate but the translation does not confirm this. Each "peneplain" extends along the valley floors which are dissecting the next higher and older level. Zones of convex ivides separate the levels. The baselevel for erosion of each higher level is the surface or "peneplain" next below, or, rather, the nickpoint or break in grade of main streams. Waning development then starts on each "peneplain" level and spreads upslope from it. Inselbergs occur on the "peneplains". It is not at all clear just how Penck distinguished these steps of a stairway from true peneplains of the original definition, the end of a cycle of erosion in a humid climate.

For an example in North America he mentions the levels of the South Atlantic slope. There the high mountains are bordered by an imperfect level which rises above the top of the Blue Ridge and the high level plateau of New England. This second surface, the Blue Ridge crest, was at that time thought to continue under the Cretaceous formations of the Coastal Plain. The now-dissected Piedmont Plateau was the third in this series of benches. Penck fully recognized that, if these surfaces were each end-peneplains, "the step-like repetition of peneplains represented insuperable difficulties of interpretation, and it remained a complete enigma." For this reason a fault along the Blue Ridge had been suggested by some geologists, although not proved. Attention was directed by Penck to the distribution of the Coastal Plain sediments to the southwest to disprove the fault hypothesis. Penck concluded that the Piedmont is a "zone of oscillation" between the higher Appalachians which have continuously undergone erosion and the Atlantic Ocean where deposition has been uninterrupted. He thought that unconformities and submarine valleys demonstrate that the Coastal Plain has shared in this oscillation of level. To the present writer Penck's explanation cannot explain the origin of the Blue Ridge escarpment. It was not intended to suggest that the Blue Ridge is an ancient sea cliff. Other important objections to the piedmont bench hypothesis have been made by others considering downward erosion versus reduction of divides. It is not clear how the streams could transport the debris from a zone of youthful erosion across one of subdued slopes. Certainly alternation of belts of youthful and very advanced erosion is peculiar. Some have thought that these phenomena reduce the entire peneplain concept as originally proposed to an absurdity. Penck evidently felt that his idea of concurrent and possible accelerating earth movement in a series of waves of uplift causing erosion which advanced laterally was sufficient to answer the objections. This hypothesis ignores King's theory of successive pediplains are scarp retreat with the debris carried away by running water. The fact that King's idea entails in much of the world a different climate than that of today has deterred many from accepting it. We can here neither go farther into this problem nor summarize Penck's interpretation of the geomorphic history of a number of mountain chains.

Summary. Penck's hypothesis of earth movement at the same time as erosion is undoubtedly a distinct advance over the original simple hypothesis of Davis. However, it does not answer all problems. Steep slopes cannot be maintained unless the debris shed and washed from them is constantly carried away from their bases. Penck ignored the phenomena now so well demonstrated by the students of soil erosion. He also ignored climate and climatic changes. His main theme is that past erosional history in relation to earth movement may be deciphered by the single criterion of the present distribution of convex and concave slopes. This ignores other ways in which such phenomena can be accounted for. Certainly Penck does not describe anything like pediments nor advocate the parallel retreat of anything but rock cliffs and the "risers" of peneplain "stairways". His theories have been misrepresented especially by the school of geomorphologists who attach great importance to running water outside definite stream beds, a condition which is indubitably important where scanty vegetation furnishes little resistance to erosion. Penck was not a leader of a new school of thought but an advocate of one theory only, his own.

F. T. Thwaites, October, 1954

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GEOLOGY 109
GEOMORPHOLOGY

Supplements, 1954, part II

Introduction

Geomorphology is a term which seems to have replaced the older name physiography for the study of the origin and classification of land forms. The surface of the earth is the end result of a conflict between two major classes of forces: (a) those that build that surface up and (b) those which wear it down. The land forms we observe are the products of this conflict.

Methods of approach. Originally the classification and nomenclature of land forms was considered a phase of geography and the problem of process by which each originated was stated only in very broad terms, if at all. Through the years a vast number of technical terms have been introduced, so many in fact that the student is readily confused. Von Engeln remarks: "The competence of the geomorphologist. . . depends very much on how well he is informed in regard to the established and accepted nomenclature." To the present writer, however, there is grave danger in this method of approach, although it may have been necessary in the past. Penck remarks, "The physical character of the morphological problem comes out clearly. The task before us is to find out not only the kind of formative processes, but also the development of the ratio of their intensities with respect to one another. . . . As is now self-evident, it requires the application of the methods of physics." Rich, in 1938, voiced something of this criticism of the early approach by stating: "Comparatively few papers dealing with the basic principles of geomorphology have been published in the United States in the past 20 years. Only a small percentage cover the nature and origin of slopes, the details of differential erosion, the relative roles of rain-wash and creep, or the exact way in which peneplained surfaces are formed or destroyed. . . . Much of modern physiography, therefore, might almost be called a science without a foundation." It is true that since the above was published much attention has been devoted to problems of soil erosion and a considerable amount of experimental work has been accomplished along this line. However, much of this is not entirely applicable to the study of erosion where unaffected by the works of man. Nevertheless, Rich's criticism still remains sound in respect to geomorphology as taught in most universities. Teaching of this subject in secondary schools has fallen almost to nothing and has been replaced by purely descriptive geography. Perhaps the most serious objection, however, is that use of technical term commits the user to a certain hypothesis of origin.

Aim of present approach. The present method of approach is an attempt to supply some of the foundation knowledge by an anlaysis of the physical processes which alter the face of the earth. Results are subject to the limitations imposed by present knowledge and by the elementary fact that in most localities more than one process operates simultaneously to produce the observed end result. Mathematical analysis can show whether or not a given result follows a definite physical law, but it cannot in all cases discover the relative importance of the different processes involved. The different processes may be so related to one another or they may not be so related. In some cases the interrelation is so complicated that no one has evolved a rational explanation. In many subjects it will be found that the methods of solving problems used by engineers have progressed far beyond those used by geologists and deserve careful consideration. In other cases, however, the engineering study presents only the "factor of safety" of construction and hence is inapplicable to an explanatory approach.

Mathematical analysis. In order to discover physical laws it is necessary to resort to mathematical analysis. Mathematics are exceedingly useful but can, like many other good things, be misused or abused. A humorist is said to have remarked, "Figures don't lie, but liars can figure." We must always be on our guard both to avoid pitfalls and not to think that because we have evolved a mathematical expression for a process or land form, that it is the end result of our search. It is to no avail if it does not point to the process which alone (or mainly) has led to the observed result. Most mathematical studies in geomorphology do not lead us much beyond simple algebra; only a few require calculus.

Deducing a mathematical expression. In most cases the data offered for analysis is a curve or the figures necessary to plot one. The question presented is: Does this curve represent the result of some definite process or law? All of us are familiar with the preparation of curves or graphs on ordinary coordiates where the quantity represented is directly proportional to the distance from the 0 point or point of origin. Since the quantities plotted on the two axes are for the most part not the same quantities, the scales may be chosen with any desired value. For instance we may plot a cross section or profile of a land form; in this case we are comparing distance in the horizontal direction with differences in elevation. We can take as point of origin either the top or the bottom of the slope and we can use, if desired, a different scale for each. Since both could be measured in the same units such an alteration of scale is called exaggeration. Its use is necessary in some cases but may lead to gross misrepresentation and consequent error in others. A curve may be either regular and smooth or irregular and complex.

Logarithmic scale. A logarithm is the power to which a certain number is is raised to obtain another number. In most tables the base number is 10 but some mathematicians use e which is approximately 2.718. Such logarithms are designated as $\log_e$, Natural logarithm, or Ln. $\log_{10} = 2.31 \log_e$. Example: $10^1 = 10$, $10^2 = 100$, $10^3 = 1000$ and so on; $10^0 = 1$, $10^{-1} = 0.1$, $10^{-2} = 0.01$ and so on. Note that negative exponents are the same as reciprocals; i.e., $10^{-2} = 1/100$. Logarithms are most commonly used to simplify multiplication and division which then become addition and subtraction. Powers need not be whole numbers; the square root is the 1/2 power. If we plot quantities on a logarithmic scale, note that there is no zero. Such a scale is, say, .01, .1, 1, 10, 100, 1000 correspondeng to logarithms -2, -1, 0, 1, 2, 3. Many physical measurements are shown better on such a scale than on the more common one of direct proportion to the numbers involved. Another use of powers of 10 is to show very large numbers, as for instance 1,500,000 as 1.5×10^6. This saves space, confusion, and possible error in placing the decimal point in the computations.

Power Functions. A function is some mathematical expression whose value upon the power of a given quantity, for instance $y=x^2$. If any similar expression is plotted on logarithmic scales for both variables, the result is a straight line. Here we have shown $\log y = 2 \log x$. Evidently the line will slope at the rate of two vertical normal units for each horizontal unit provided x is shown on the horizontal axis. This fact enables one to find: First, that a series of observations is a power function of a variable; and second, obtain

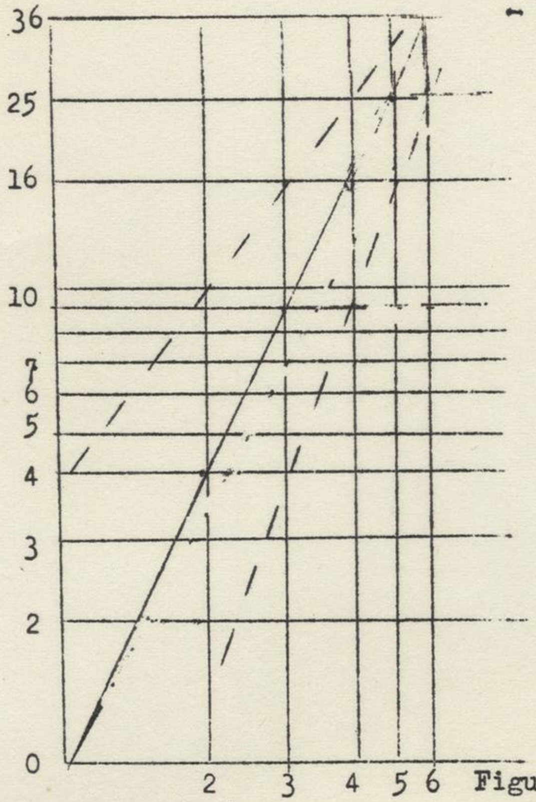


Figure 1

Logarithmic plot of equation
$y = x^2$
Correct result is shown by solid line. If the 0 point had been wrong we would have obtained one of the broken lines. Correct exponent is 2. The left line would have given 1.3 and the right broken line 2.6 Note how slope of line shows the exponent of a power function.

the value of the exponent involved. This exponent may be positive or negative and is not always a whole number. If the expression is x = constant · x^2, the straight line will cross the vertical line for x = 1 at the reading for the logarithm of the constant for log x = log c + 2 log y. Constants do not affect the slope of the line. In proctice with observations of natural quantities, we must not expect all points to fall exacctly on a line. This failure is called scatter and may be so great as to cast doubt on the interpretation as a power function. We may also find that there is a change in slope of the line which shows that some other law has come into play or that we have passed to observations which have a different point of origin. (See section on shape of volcanic cones.) The dotted lines of Figure 1 show the effect of eross in the point of origin. The points plotted still fall on nearly straight lines, but the deduced values of the exponent are altered. In natural phenomena it is often hard to find the origin of a power function.

Exponential Functions. A very common relationship which is exhibited by many natural phenomena is represented by the exponential function. This is a power relationship in which the exponent is one of the variables. An example is y = constant ^{x}. If we take the constant as 10 and increase the exponent from 0 to 4, values of y will be 1, 10, 100, 1000, 10000. Mathematicians commonly prefer e as a base. If the exponent is negative, a decreasing or inverse relationship to x is shown. For instance, with exponents decreasing from 0 to -3, and 10 as the base, the values are 1, .1, .01, .001. It is easily seen in Figure 2 that, if we plot the x axis or horizontal axis with normal proportion and the vertical or y axis as logarithms, we obtain a straight line. The same relationship holds for a geometric progression or series. Given such a series with the ratio of 2 the values are 2, 4, 8, 16, etc. This will also plot as a straight line. Such series are common natural phenomena. In the case of an exponential function, $y = mb^{ax}$, the letters m, b, and a represent constants. b, by common usage, is taken as the quantity e described above. m may be found by taking x = 0; this gives the value of y at the

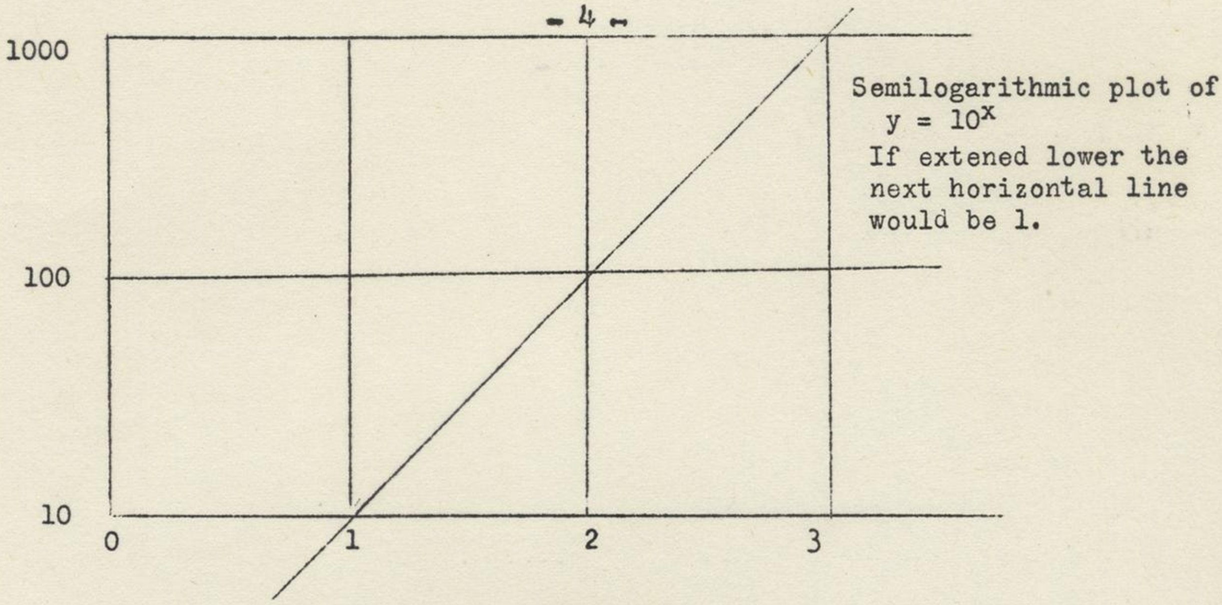


Figure 2

starting point or origin of the curve, which is often called y_0. a can be found in several ways, one of the commonest being to put the value of y at half that of y_0. Then $y/y_0 = \frac{1}{2}$. From this it follows that $\frac{1}{2} = e^{-ax_{\frac{1}{2}}}$ or $-\log_e\frac{1}{2} = ax_{\frac{1}{2}}$ or $-\log_e\frac{1}{2} = +\log_e 2 = .693$. Hence $a = .693/x_{\frac{1}{2}}$. See Krumbein, W. C., Sediments and exponential curves: Jour. Geol. 45: 577-601, 1937.

Rate of change. In many studies it is important to know the rate of change of a variable. For instance, in a profile across a land form we speak of the slope at any given point. Slope is the rate of change of elevation or fall depending upon where we take the origin of the coordinates. Mathematicians call the rate of change the differential. For instance, in the equation $y = x^2$ the rate at which y is changing with respect to x is expressed as dy/dx. The d's expressigg change in a small interval. It is demonstrated in works on the differential calculus that with such power functions the differential is found by reducing the exponent by one and multiplying by the original exponent. For $y = x^2$, $dy/dx = 2x$. For $\log_e x$ it is $1/x$ or x^{-1}.

Integration. It often happens that it is necessary to put the taking of a differential into reverse. For instance, if we have the rate at which slope is changing, we can compute the difference of elevation. This process of finding the mathematical expression from which the differential was derived is called integration. It is denoted by the sign and denotes a totaling up process. If the limits between which this summation is to be made are given, it is the same thing as finding the area enclosed under a curve. For power functions it is necessary to increase the exponent by one and divide by this increased exponent. An exception is where the exponent is -1; then the integral is $\log_e$. Integrals can be found in tables in books on the calculus.

Trigonometry. The diagram of Figure 3 shows the essentials of trigonometry in respect to right triangles. Note that for small angles sine and tangent are nearly the same. Moreover for small angles tangent of x degrees approximates x times tangent of 1 degree.

Limitations of mathematical analysis. If we have any expression where one quantity is dependent for value on another, it is easy to express the results in a curve or graph. If, however, there are two or more independent variables, the expression cannot be thus shown. Where several processes are at work at the same time, the mathematical relations become very complex and a series of simultaneous equations expressing the interrelations are needed. If some of these are unknown, mathematical analysis becomes impracticable. Attempts to fit mathematical curves to natural phenomena simply because they somewhat resemble the facts are in many cases preposterous.

Physical definitions. The following are some of the definitions of physical quantities which are met with in the studies of geomorphic processes:

Mass, m = an inherent property of matter independent of Weight, w, which varies with changes in attraction of the earth (g) in different localities.

Neither space or distance, s, needs any definition, nor does time, t.

Force, F, is that which causes a mass to alter its condition of motion.

Velocity, v, is rate of motion in respect to time = s/t.

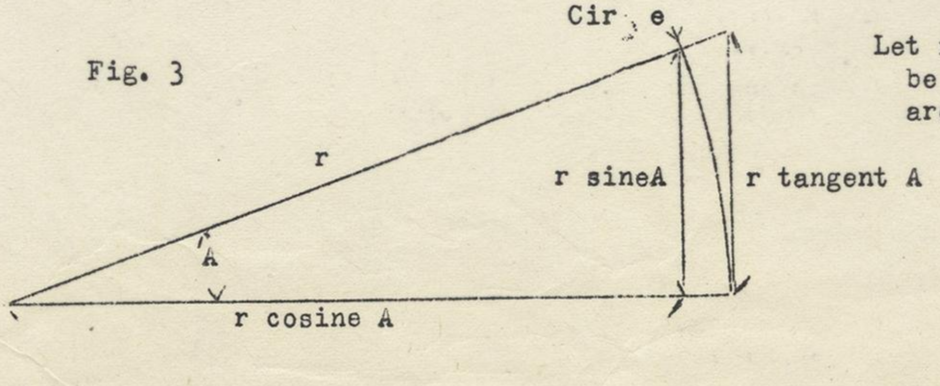
Newton's First and Second Laws of Motion hold that rate of motion(or velocity) is changed by application of force at a rate called acceleration, a. Acceleration is then the time rate of change of velocity = v/t and force, F = ma (mass times acceleration).

Work, W, is force exerted over a certain distance, W. = Fs.

Power, P, is the time rate of work. P = Fs/t or, by substitution, of s = Vt. P = FV.

Kinetic energy, E, is work done in bringing a body of mass, m, from rest to velocity, v, in time, t, over distance, s. Knowing that F = ma, and that s = t times average v, which is v/2, and that a = v/t. Substitution gives E = (mvvt)/2t. Cancellation then yields the result $E = \frac{1}{2}mv^2$ If the motion is not along a line but around an axis (rotational like a flywheel) the same formula applies to every particle. If the revolving body is a cylinder or disk then the calculus shows that it is half the product of the total mass times the square of the radius times square of angular velocity in radians per second (a radian is the angle subtended by the radius of a circle). Newton's Third Law of Motion, that action and reaction are equal and opposite, is an invaluable concept in explaining equilibrium of forces which is widely demonstrated in natural phenomena.

Fig. 3



Let radius of the circle be r Other quantities are found in tables

Personal

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GEOMORPHOLOGY

Further discussion of the problem of "impact craters". Supplements, 1954, pt. III

Several papers have appeared in regard to the controversy over the origin of supposed impact craters. Chief of these are Crater Mound, Arizona, and the Carolina Bays.

Crater Mound. Dorsey Hager, a consulting petroleum geologist, has contributed the latest data on the famous Crater Mound of Arizona. In his paper of 1953 a great amount of information on the surrounding geology and the results of test drilling is presented. Although the author admits that meteor fragments are abundant in and near the crater, he concludes that its relation to a structural nose in the bed rock precludes impact origin. The meteor shower is declared to be much younger than the depression and its location a pure coincidence. However, the log of the deep test hole on the southern rim showed nickle almost to the bottom where the tools were lost in boulders. The accompanying section is compiled by Hager.

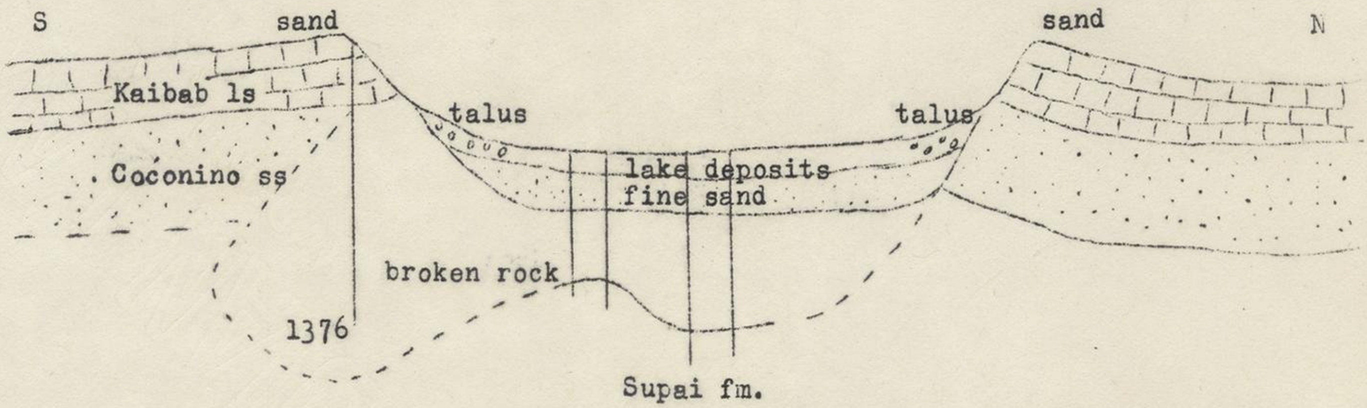

Figure 1. North-south section of Crater Mound, Arizona, after Hager
Scale about 1" = 1000' Vertical lines = drill holes

The crux of the argument is the presence of very fine sand and silica glass. Proponents of the impact hypothesis ascribe this to a violent concussion and heat. Hager, pointing out the paucity of meteorite fragments in this material, suggests that it is either volcanic ash or is due to fracturing of sand during deformation. Although volcanic rocks are known only 25 miles away, the gas (or steam) explosion theory is summarily rejected because of the lack of evidence of an explosion. The sinkhole idea is rejected for obvious reasons. Hager suggests a former salt dome due to intrusion of salt from an inferred evaporite series below which has been discovered in some wildcat wells for oil 35 miles away. The solution of the salt later formed the crater. An igneous plug is disproved by results of test drilling and the negative results of magnetic and electrical surveys. Hager leaves the problem open although stating that the salt dome hypothesis "satisfied most of the geologic data."

Ninenger, who has done much work searching for meteorites in and around the crater replied with an article on the inferred evidence of heat and violent fracturing in the formation of the silica glass and fine sand. To this Hager replied that the fine slivers of quartz show no evidence of heat referring their origin to weathering and folding. He suggests that no final answer to the problem is possible unless someone drills "a series of drill holes deep enough to penetrate a datum bed within the crater, on the rim, and outside the crater." To this Ninenger replied again stating "the following known facts. No sinkhole or graben in this area can be remotely compared with the crater as to (1) its general appearance, (2) the symmetry and uplifted form of its walls, (3) the blanket of broken-up rubble derived from the three formations of the area, (4) the presence in its pit and among the rubble on the rim of millions of tons of rock flour --------, (5) the scattered deposit of large boulders, from the normally buried sediments, found on various sides of the pit out to distances of more than a mile in locations remote from any visible trace of water courses, (6) the surrounding deposit of some 40,000 recovered meteorite fragments, (7) the presence of meteoritic material mingled with the fill and in the rubble of the rim, (8) the surrounding deposit of several thousand tons of nickeliferous particles in the form of metallic spheroids, (9) the presence in the rubble on the rim of millions of lumps of fused country rock in the form of slaggy bombs and droplets recently described by me as impactite." He also stresses "the following facts that have been overlooked: (1) A radiate distribution of meteoritic fragments, (2) a zonal distribution of meteoritic fragments as to size, (3) a zonal distribution of meteoritic fragments as to heat effects, (4) the presence among the fragments of a small percentage which differ from the majority in both structure and composition, (5) the presence around the crater of an enormous tonnage of metallic spheroids composed of meteoritic metal, but exceptionally rich in Ni, Co, and Pt, (6) the presence of a considerable tonnage of slug-like particles comparable in size with the spheroids, but having the same composition as the well-known Canyon Diablo irons, (7) an abundance of impactite slag."

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The Carolina Bays

Livingston, D. A., On the orientation of lake basins, Am. Jour. Sci. 252: 547-554, 1954

The similarity of the form of the Carolina Bays to the elliptical lakes of the Alaskan Coastal Plain has long been noted. Black and Barksdale concluded that the latter are thaw lakes in the permafrost which have grown downwind in the direction of prevailing winds by reason of the stronger waves on the exposed or lee shore. The long axes of the these basins trend slightly west of north but modern knowledge of the prevailing winds above 15 m.p.h. shows that they blow almost exactly at right angles to the long axes of the lakes! The author then tries to show how wave action can explain this apparent contradiction. He points out that removal of material from the shores depends more upon rip currents than upon direct wave action where the eroded material is removed directly out into the lake. The wave action on the lee shore tends to move water toward the sides of the lake. There it piles up and the pressure is relieved by rip currents which transport the eroded material to the deeper waters. A mathem ical analysis is given which attempts to show this fact, but since it is admittedly only a first approximation, it will not be repeated. Erosion of the shores of Lake Michigan north and south of Two Rivers, Wisconsin is readily observed to be at a maximum in the bays and not on the headland between them. Evidently it is not due so much to direct wave action as to the same rip currents. Hence the author's point seems well taken.

Turning to the Carolina Bays, the present winds are too variable to explain any such action. Odum, however, suggested that the Pleistocene winds were parallel to the long axes of the Bays. This is admittedly based upon assumptions as to the effect of the continental ice sheet on wind directions. Granting the correctness of these assumptions the contradiction is obvious. Livingstone explains this however, in another way. The Carolina Bays date from a time of drier climate when they were blowouts due to wind erosion of the sand.

Reference

Black, R. F., and Barksdale, W. L., Oriented lakes of northern Alaska: Jour. Geol. 57: 105-118, 1949

Odum, H. T., The Carolina Bays and a Pleistocene weather map: Am. Jour. Sci. 250: 263-270, 1952.

Odum did some work on the Bays and observed the wave patterns in those which have open water. From this he deduced that the elongation and streamlining could be explained by winds of Pleistocene time when the pressure gradient south of the glacier led to stronger winds than those of the present. He noted the wave refraction in the shallow lakes, a feature apparently overlooked by Livingston. Odum attempted the construction of a map showing air pressures at 10,000 feet elevation as governed by isotherms on the ice and to the south as they could have been during a Pleistocene glaciation. The -20 C isotherm is drawn approximately along the south border of the ice sheet and the 0 C isotherm would then be approximately along the Gulf coast and Atlantic coast south of Cape Hatteras. These are the January positions shown on the present weather charts for this altitude. From this data the isobars are shown which trend much more northeast-southwest across the Atlantic coast.

The sea level pressures would then show a more marked parallelism to the coast. Prevailing winds are then drawn at an angle of about 40 degrees to the isobars. The postulated wind directions agree very well with the longer axes of the Bays. Such winds might very well be dry and cause blowouts although neither the perfect streamlining nor the size of the Bays is in harmony with wind-excavated hollows of today.

LeGrand, H. E., Streamlining of the Carolina Bays: Jour. Geol. 61: 263-274: 1953

In the paper cited above we have an attempt to confirm the hypothesis of Douglas Johnson in respect to the origin of the Carolina Bays by action of ascending ground waters. The author has been engaged in ground water studies and hence had available much more sub-surface information than did Johnson, or anyone else. His theory must, therefore, command respect. Some of the results of this ground water study at once point to a relationship to the Bays. Bays are confined to the area where limestone is present at relatively shallow depths, not over 150 feet, and beneath a clay cover. This limestone is water-bearing and must once have possessed an artesian head. Bays are absent on areas where there is nothing but sand. The sand rims, of which so much has been made, are far from universal and are confined to an area where there is a thin Pleistocene sand above the clay. Eolian origin is conceded for these rims. Overlapping Bays, a thorn in the flesh to some hypotheses, are confined to areas where there is more than one limestone formation. The limestone does not extend to the outcrop and solution is the probable cause of its absence. LeGrand's theory is that solution was localized in the limestone with formation of ordinary sinks. The overlying clay subsided into these sinks. Escaping artesian waters or water bypassing a clay plug formed a streamlined flow pattern and thus affected the surface outline of the depressions called Bays. Alignment of the long axes of Bays agrees with dip of the aquifers. Admittedly, the perfection of the streamlined or oval outline of the Bays is the chief stumbling block of this hypothesis but the conclusion of clay near the surface may not be favorable to the meteoritic hypothesis.

Wolfe, P. E., Periglacial frost-thaw basins in New Jersey: Jour. Geol. 61: 133-141, 1953

The paper on supposed frost-thaw basins in New Jersey bears not only on the problem of past permafrost but also on that of the origin of the Carolina Bays. Wolfe describes shallow basins which dot the sandy coastal plain of New Jersey. Although not as perfectly streamlined as those of the Carolinas the resemblance is nevertheless striking. The meteoritic hypothesis is rejected because of much lower magnetic anomalies than are known to the south. An origin due to solution is also rejected because auger borings showed that underlying marl is not affected. Instead an origin different from that of the Bays is formulated. Backed by observations of irregularities in gravel pits in the zone of weathering, Wolfe concluded that New Jersey once had permafrost. The depressions would then be relics of thaw lakes like those of the Alaskan coastal plain.

Several objections may be raised to such a theory. First, the supposed frost wedges and convulutions of an "active layer" are in gravels and not in fine-grained materials like those of Alaska. Second, New Jersey was always close to the ocean and its climate is molified by that body of water. Third, New Jersey is near the southern margin of the continental ice sheet and presumably there was always a mean temperature high enough to melt ice at the surface. Presence of continental ice is not of itself sufficient proof of an Arctic climate nearby. Just how did the frost-thaw lakes serve to remove material? If the material did not sink doen and was not removed by explosion or impact, then wind action is the only hypothesis left. Wolfe admits that the basins have been altered by wind action and illustrates wind-worn shores.

Cooke, C. W., Carolina Bays and the shapes of eddies: U. S. Geol. Survey Prof. Paper 254, pp. 195-206, 1954

In a recent report Cooke returns to his explanation of the Carolina Bays since he regards Prouty's meteoritic hypothesis as "implausible". The occurence of bays higher than areas of artesian flow and in areas where there is no artesian head in the surficial deposits is urged against acceptance of Johnson's theory. Cooke modifies his earlier view of wind-driven eddies by adopting a tidal origin. This involves the fact that the bays on the higher marine terraces are decidedly older than those near the ocean. The basic idea is that tidal currents set up rotary eddies which have the properties of a gyroscope. A gyroscope is like a child's top. Rotated at high speed its inertia keeps the axis of rotation in a fixed position in space. The rotation of the earth was once proved in this way. Cook conceived of the tidal eddies as fast moving enough to develop this property although made of a substance, water, which yields to the force of gravity. However, a spinning gyroscope can be affected by a force which tends to tilt the axis. A top if thus touched has a secondary motion of the axis which starts at right angles to the direction of the disturbing force. This secondary revolution of the axis is called <u>precession</u>. "The eddy, like a spinning flywheel, may react in either of two ways: the plane of the impulse to rotate may precess about the east-west diameter of the eddy or the plane may set itself perpendicular to the axis of the earth, about which it is forced to revolve." However, the actual form of the eddy will be level and thus a circular shape would be altered to an ellipse. The horizontal components of motion are thus changed from a circle to an ellipse, the projection of a tilted circle. In a precessing eddy the ratio of shorter to longer axes is equal to the cosign of the latitude. In a fixed eddy the ratio equals the sine of the latitude. In a precessing eddy there are two simultaneous motions: revolution about the axis of the earth and precession about its east-west diameter. The resultant of this is rotation about an inclined axis between the axis of the earth and an east-west horizontal line. Then cosine latitude = 1/tangent of declination from north. A fixed eddy has an apparent motion like that of the sun from east to west and in Northern Hemisphere another from north to south. The two rotations are one quarter phase apart. This tilts the long axis of the eddy 45 degrees toward the northwest. Cooke admits that these conclusions are purely deductive and without either experimental or mathematical proof. Nevertheless on the many air photographs he presents there is a striking approximation to a N-45 and N. 51-41 W strike for the long axes of the Bays. The ratio of diameters is not stated in many examples. He states that his manuscript has been checked by engineers who are familiar with gyroscopes.

<u>Criticism</u>. Cooke's explanations are very far from clear and repeated reading of them and references on the gyroscope did not serve to elucidate them satisfactorily. Just why are there eddies of two different types? One may well question if the velocity of tidal eddies is enough to show true gyroscopic motion. Rotational inertia = $\frac{1}{2}$ mass radius^2 in a disk. Another question is what effect the two reversals of tidal flow per day should have on the phenomena. Moreover, only two examples of modern curved bars which resemble those of Bays are presented. Besides this why are the Bays on the older high marine terraces so well formed if they are so much older than those near the seacoast as the theory demands? Despite the endorsement of the U. S. Geological Survey one is left with a sense that the problem is simply more confused than it was before.

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GEOMORPHOLOGY

Classification of shorelines and miscellaneous. Supplements, 1954, part V

Introduction. The placing of shorelines in a suitable genetic classification has been a vexed problem for a long time, and it is generally admitted that no perfect system has been evolved. The time-honored classification into "submergent", "emergent", with Johnson's contribution of "neutral" and "compound" has held the adherence of the great majority of geomorphologists. Shepard's rival classification of "primary" and"secondary" coasts has not met with many supporters. A third system divides shorelines by motion in a horizontal direction, that is shore which are being worn back and those which are being built outward into the body of water. This is really the same as submergent and emergent. The problem has also been confused by attemptes to introduce a cyclic classification similar to that used so successfully by Davis on landforms due to stream erosion and weathering.

Criticism. The greatest difficulty in the application of any system is that everywhere and at all times there is a contest between the forces of the land and those of the waves and currents. For instance, the east coast of Florida, which is exposed to the full force of the Atlantic, shows dominantly wave-built features. Not far to the west, the Mississippi River is building a delta into the Gulf of Mexico with only moderate alteration by the waves. The exposed coast of New Jersey and Maryland is almost wholly wave-built, whereas the shore lines of Delaware and Chesapeake bays are obviously due to a submerged landscape which was molded by stream erosion. Again a shoreline may be backed by many abandoned shorelines due to a higher stand of the waves (or lower level of the land), yet between the formation of two of the strand lines the waters may have stood at a much lower level. An example is the coast of northeastern Wisconsin which is lined with beautiful examples of raised beaches, some of them tilted by later earth movements. Yet just prior to the formation of the lowest major abandoned beach there is now known to have been a time when the waters of Lake Michigan were 350 feet lower than they now are. Should we term this an emergent coast because of the old beach levels or a submergent one because of the rise in water level after the low stage? Much of the west coast of the United States shows abandoned beach lines up to over a thousand feet elevation alongside obvious drowned valleys.

And in respect to cyclical development it is clear that the relative speed of alteration of the horizontal form of a coast depends upon both material and exposure to waves. Almost all coasts are demonstrably rather young in years. It is also clear that depth of water adjacent to a coast line is an important factor in rate of erosion. The writer has observed that there may, in many places, be much more coast erosion in bays with deep water than on adjacent headlands. This is explained by the piling up of water in the bays with removal of material by rip currents. On the headlands the only removal process is transportation in the zone of breakers. In this connection it is well to reexamine the old idea that barrier beaches are a temporary youthful feature later removed by the waves to form a cliffed coast. The only evidence that the writer can find in the literature is the occasional discovery of peat on the beach which was taken to indicate retreat of a barrier into the former lagoon. It could be, however, that this peat was formed during a former low stand of the water level and that it does not demonstrate retreat of the shoreline.

Cotton's criticism. Cotton has been one of the chief critics of the Shepard system of shoreline classification. He particularly objects to the omission of "emergent" coasts as a group, saying that Shepard "failed to realize the significance of important features of emergence, such as raised beaches and marine terraces..." In Cotton's own classification new classes within Johnson's neutral group comprise coasts produced by: "(a) volcanic accumulation; (b) faulting..; and (c) glacial over-deepening below sea level..". In considering genetic classification he mentions: (a) tempo of development, that is the effect of differences in material and exposure, (2) convergence of development, (3) progradational outbuilding of a coast which may occur during any part of the cycle, due primarily to accumulation of material furnished by rivers. Another suggestion is: (a) eroded or retrograded coasts, (b) coasts prograded by alluvial (delta) deposits, and (c) coasts prograded by beach development.

Conclusion. In the opinion of the present writer, the usual classification of coast lines into emergent and submergent is a good illustration of an undesirable nomenclature whose use commits one to a definite interpretation of geologic history. A radical change in water level may have occurred without leaving any definite evidence which may be observed at the surface. A good instance lies in the postglacial beaches of Lakes Michigan and Huron mentioned above. It is true that careful study does indicate that the Nipissing beach does bridge older erosion valleys but until subsurface information from drilling and lake bottom coring became available this fact was almost entirely ignored. The post Nipissing beaches are truly emergent, however, because they were due to progressive erosion of the lake outlet. The effects of glacial eustatic changes in ocean level, however, make almost all, if not all, marine coasts really submergent. In conclusion, it is desirable to emphasize that no map studies of shorelines should be undertaken without (a) chart showing water depths, (b) information on prevailing winds and their strength, (c) information on subsurface geology, and (d) information on currents. Much information can be gleaned from "The Coast Pilot", published for mariners.

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Permafrost phenomena

Black, Robert, Permafrost: a review: Geol. Soc. Am. Bull. 65: 839-856, 1954.

Flow of ice, etc.

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Sharp, R. P., Glacier flow: a review: Geol. Soc. Am. Bull., 65: 821-838, 1954.

Wentworth, C. K., The physical behavior of basaltic lava flows: Jour. Geology, vol. 62: 425-438, 1954.

Wentworth writes in respect to the courses taken by flows from the Hawaiian volcanoes over preexisting topography. He concludes that lava is rarely a perfect fluid because temperature differences, escape of gas, and crystallization bring about great differences in viscosity in small distances. The heat which causes fluidity originates within the vent so that heat is lost all along the flow both at the bottom and the top. Flows often develop a crust on which persons can walk while motion continues at considerable velocity beneath. With streams of water, the bottom only causes important frictional resistance, and the maximum velocity is obtained when a semicircular cross section is present. In a lava flow loss of heat to the atmosphere is very important; hence one might well presume that a circular cross section should afford minimum friction. This agrees with the form of lava tunnels. In fact, the "Devils Sewer", a pipe of about two feet diameter at Craters of the Moon, Idaho, seems to have had this origin in a thread of lava. Another factor in lava flow is the floating of blocks of solidified lava and fragmental material. Although flows tend to build up barriers of such blocks along the sides renewed flow may float away masses of rock. Wentworth notes velocities of flow up to 15 m.p.h. in channels up to 50 feet wide with slopes of 500 to 5000 feet per mile. However, the viscosity of lava is such that turbulent flow is never present, all flow being lamminar. Viscosity is affected greatly by escape of gas. Average values of the coefficient are 10^2 to 10^5 poise. Density is less than 3. Level surfaces on lava pools are rare and the idea that the flows behave like water is erroneous. Slopes of solidified flows as low as 200 feet per mile (about 2 degrees) are rare in Hawaii; much more abundant are slopes of 500 to 1000 feet per mile. Inspection of the topographic map of the Craters of the Moon shows that i is very rare to find the slope of a flow as low as 50 feet per mile, and that slopes normally exceed this, reaching in short distances over two hundred feet per mile. It is fair to conclude that the apparent levelness of basalt flows as viewed without instruments is misleading.

Evidently the filling up of a lava plateau area is not due to low initial slopes of flows so much as to successive flows in different directions plus some water deposits in low places. Lava flows behave much like aggrading streams. Wentworth mentions temperatures of basalt up to 1050 deg. C. and that movement ceases at around 760 deg. C. One flow lost only 100 deg. in 10 miles. Lava cannot penetrate into fissures of previously cooled rock because it loses heat too rapidly.

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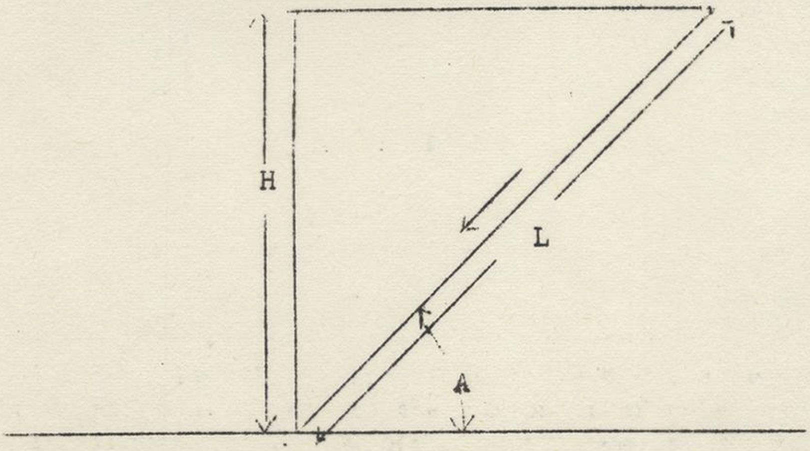
Nichols, R. L., Viscosity of lava: Jour. Geol. 47: 290-302, 1939.
Wentworth, C. K., Carson, M. H., and Finch, R. H., Discussion on the viscosity of lava, Jour. Geol. 53: 94-104, 1945.

Derivation of limit of height of a vertical bank at four times the shearing strength.

In the supplement on soil mechanics no proof was offered for the above conclusion. Most text books endeavor to derive this law by involved and difficult processes. The following drivation aims to reduce the matter to its simplest terms. A bank of height H is cut by a trial plane which is equivalent to the surface of a cylinder of infinite radius. The length of the inclined plane along which failure may occur is L and its angle with horizontal = A (see diagram).



The weight of this triangular slice of unit width = its area times unit weight y. Hence weight = $\frac{1}{2}$yLHcosA. Force is the component of this weight parallel to the slope which is $\frac{1}{2}$yLHcosAsinA. Now books on trignometry demonstrate that sinAcosA = $\frac{1}{2}$sin2A hence force = $\frac{1}{4}$yLHsin2A. Opposing this force is the shear strength of unit area, C. Hence total opposing force = CL. Failure occurs when the force down exceeds this resistance. Then $\frac{1}{4}$yLH sin2A = CL. L occurs on both sides and can be cancelled. Solving for H, H = 4Csin2 A/y. A little consideration shows that the maximum value of the right hand side of this equation is reached if A = 45° for then sin 2A = sin 90° or 1. The catch in the whole solution is how to determine C accurately, especially when water is present. Otherwise it is evident that failure will in general be along a plane of 45° inclination. The effect of vertical joints is also ignored.