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Climatic Controls

Introduction. Many of the materials of the crust of the earth are not in equilibrium with their environment at its surface for they were either cooled from a molten state or deposited under water. In contact with water and air they undergo changes in both physical and chemical states. The control of the nature and rapidity of these adjustments lies in the climate. We will here discuss the climates of the earth from this standpoint only, considering precipitation, together with seasonal distribution and its disposal on reaching the ground, winds, and temperatures. We must, however, not regard climate as fixed. Through geologic time changes have been very marked.

Transmission of moisture. All rain and snow which falls on the earth ultimately comes from evaporation of bodies of standing water. However, its journey through the atmosphere in the form of vapor may be made in two or more stages separated by one or more intervals during which it was precipitated and then evaporated from the ground. The ultimate cause of precipitation is reduction in the vapor-carrying capacity of the air by lowering its temperature. It is a coincidence that the vapor pressure of water in millimeters of mercury at a given temperature is, within the range ordinarily met with, almost exactly the same as the maximum possible number of grams of water in a cubic meter of air. There is no necessary connection between these two quanities. Ordinarily air does not contain all the moisture it can hold at that temperature and the percentage of what it does contain of the maximum possible is termed relative humidity. Now, when air is cooled its relative humidity increases until the amount is 100 percent. The temperature at which this occurs is called the dew point. If cooled below this temperature precipitation occurs. If the temperature is above freezing rain or fog results. If below freezing snow crystallizes directly from the vapor.

Causes of precipitation. Cooling of air which contains moisture is due primarily to ascent for temperature decreases with elevation. Rising of air is due to (a) local heating, (b) winds encountering a mountain range, and (c) winds rising over a mass of colder heavier air. The first process accounts for both local afternoon thunder showers and the tropical rainbelt which is located in the latitude of maximum solar radiation moving with the seasonal change in position of vertical sun rays. Precipitation on mountains is common throughout the world and requires no further comment. At the present the polar regions are perpetually frozen and cold. Cold air which flows south (north in the southern hemisphere) encounters the warmer air of equatorial regions. The contact of northern and southern air is sometimes called the polar front. It is irregular in outline, associated with gigantic swirls termed cyclones. Within these warm, moist air rises above the cold northern air giving rise to cyclonic precipitation, now spoken of as storms along fronts or contacts between air of different source, humidity and temperature. These areas have low atmosphere pressure and are often termed lows. If, as seems probable, in the geologic past the poles were warmer than now the width of the area through which this belt of storms migrates with the seasons was once smaller and the vigor of weather changes within it was much less than now.

Winds. The belt of tropical rainfall is marked by rising currents of air, low barometric pressure, and by absence of surface winds. The air which flows into this belt of calms from farther north and south and, deflected by the rotation of the earth, forms the very constant northeast and southeast trade winds. North and south of the trades descending currents occupy the tropics producing another belt of calms, here associated with relatively high atmospheric pressure. At present about half the area of each hemisphere is occupied by a belt of variable winds along the polar front whose prevailing direction is from the west. This is the belt of westerlies. The larger continents in the belt of west winds afford an exception in that the great seasonal range of temperature in the interiors causes winds to blow inward in summer and outward in winter. These temperature-controlled winds are termed monsoons.

Seasonal distribution of precipitation. In few types of climate is precipitation uniform in amount from month to month of the year. The belt of tropical rains follows the apparent movement of the sun. In equatorial regions this means two rainy seasons in every 12 months, but farther north and south there is only one rainy season and the dry period is much longer and more pronounced. This is the Savanna belt. Seasonal migration of the subtropical calms brings dry weather to regions which at other times are either in the westerlies or the trades. It even causes aridity in the Mississippi Valley whenever the movement is unusually far north. In past geologic times it doubtless caused deserts like the Sahara to extend much farther north than they do now. Relative temperatures of sea and land also have an effect on seasonal changes in rainfall. In many places most rain falls when the land is cold. In regions far from the sea outflowing winter winds plus reduced evaporation spell a winter minimum.

Disposition of precipitation. Rain which falls on the surface of the ground is disposed of by (a) surface runoff into streams, (b) percolation into the ground which may or may not emerge later from springs to join the surface runoff, (c) direct evaporation from soil or free water surfaces, (d) transpiration from vegetation, (e) chemical combination in vegetation and minerals. The relative amounts of each disposition is difficult to ascertain although estimates of the first two are not particularly difficult. Exact determination of total precipitation on a watershed is inexact because of the spotty distribution of individual storms.

Percolation. Snow which melts on frozen ground must nearly all join the surface runoff. The proportion of rain falling on unfrozen ground which percolates into the soil is frequently estimated as a simple percentage of total precipitation. Because water which enters the soil must displace air between the mineral grains and causes physical changes in the soil it is obvious that this is a very inexact method. Experiments by Horton indicated that the initial rate of infiltration decreases rapidly at the beginning of a rain but attains a constant rate after from half an hour to three hours. Fig. 2. He arrived at the empirical formula:

$$f = f_c + (f_o - f_c)\, e^{Ct}$$

where f = rate of infiltration at any time, inches per hour, f_o = initial rate, f_c = constant rate, e = 2.718, C = a constant factor, and t = time of rainfall duration in hours. Physical properties of the soil which control entrance of rain comprise (a) texture, (b) structure, (c) vegetation, (d) biologic structures such as burrows, (e) moisture content, and (f) condition of soil if cultivated, sun-cracked, etc. The value of f_c is attained only during heavy and long-continued rains thus increasing surface runoff to cause both floods and surface washing. Infiltration capacity is greatest in loose sands which display a very low proportion of surface runoff. When soil is entirely saturated with water the rate of movement is termed by Horton "transmission capacity". His conclusion was that such saturation is actually attained only in the heaviest clay soils. The relation of infiltration capacity to rainfall is vitally important to geomorphology and more exact data are needed for quanitative study of drainage.

Runoff in relation to rate of rainfall. It is evident that whenever rain falls at a rate less than that of infiltration no runoff is possible. From this it follows that no storm may ever reach the point where there is a surplus above infiltration on the divides. This accounts for the paucity of streams in areas of sandy soil. Rainfall rate generally is highest in brief storms. Various empirical formulae show this. Little used the formula:

$$r = \frac{8}{t^{\frac{1}{4}}}$$

where r = inches per hour and t is in minutes. Other expressions do not involve exponents.

Return of percolation to streams. Some of the water which percolated into the soil is used by plants. Much is evaporated or transpired by plants and returned to the vapor of the atmosphere. Evaporation losses in eastern U. S. range from about 18 to over 38 inches per year. Correlation with summer temperature is approximate. In the region of low rainfall the lines of equal loss cross lines of temperature at right angles because there is not enough rainfall to supply potential evaporation. Some enters into chemical combinations with minerals. The remainder becomes ground water. Much of the ground water reaches the streams either through definite springs or by seepage into streams.

Actual stream flow. The discharge of a stream is determined by measurement of its cross section and mean velocity at several different water levels. A curve is then drawn to show relation between water level and discharge. When discharges are plotted in respect to time a very irregular curve is obtained of which the crests each correspond to a particular storm. The discharge between storms is essentially all ground water runoff provided there is no storage in lakes and swamps. From a study of surface and ground water runoff in the United States it appears that: (a) The total runoff is greatest with the highest precipitation. (b) Ground water runoff is largest in regions of porous bed rock where there is large subsurface storage. (c) Losses due to evaporation are more nearly constant than are other quantities.

Summary. Climatic control of geomorphic forms is concerned chiefly with causes for variation in amount of runoff, intensity of rainfall, frequency of heavy rainfall, frequency of freezing and thawing, direction and intensity of winds, duration of frozen ground, rather than with the information given on conventional climatic maps. Some of these features will be described more fully in later sections where their bearing is more fully explained. In this section precipitation alone is considered. Water which falls on the ground is disposed of by evaporation, including water transpired by plants, by soaking into the ground beyond the reach of subsequent evaporation, and by direct runoff. Much of the water which enters the earth returns via springs and seepage to form the ground water runoff. Ground water runoff may be determined from the discharge of streams between storms. Its quanitative ratio to surface runoff depends in part on climate but to a large extent of the geology of the watershed. Evaporation loss is related both to temperature and to total precipitation. In arid regions it disposes of almost all the precipitation.

Section 2.
Materials of the earth's surface.

Introduction. In describing the materials of the earth's surface for the purpose of accounting for the present topography the method of approach must of necessity be different from that employed in other branches of geology. We must distinguish between consolidated and unconsolidated materials. The former are rocks and owe their firm condition to either (a) the irregular shapes of constituent particles (commonly minerals) or (b) the presence of a cementing compound between the particles. The shapes of particles as well as their physical character determines the mechanical strength of the rock. The chemical composition and size of the particles determine the reaction of the rock to the chemical effects of the atmosphere and water. From the standpoint of geomorphology the conventional division of rocks into igneous, sedimentary, and metamorphic is almost meaningless. What is important is the relative resistance or durability of rocks to those forces which act upon them when they are at or near the surface. For this reason it is to reports upon building stones that we must turn for information. The ordinary geologic map ignores many of the factors of durability, such as grain size (texture), porosity, permeability, and structure, all of which by controlling both breaking strength and entrance of water are of profound influence on durability. Chemical composition, if shown on a map, is only one factor. Maps which indicate only geologic age are almost worthless for geomorphic studies. Because solid rock is commonly found beneath unconsolidated material it is often referred to as bed rock. Rocks are often divided into two great classed: (a) hard rocks of igneous and metamorphic origin mainly crystalline and (b) soft rocks, mainly sedimentary, although including some kinds of igneous rocks.

Texture. The term texture refers to the size, or range in sizes, of individual particles (commonly minerals) of a rock. It also includes their shape and arrangement. Division of igneous rocks into coarse and fine texture is generally only qualitative and no definite standards have been set up. Fragmental sedimentary rocks are classified by texture. Texture has a marked influence not only on mechanical strength but also on chemical resistance to alteration. Where the constituent particles are large it is evident that failure of a single one either by breaking or chemical change is much more important than in the case of a small particle of a fine-grained rock. Where the constituent minerals have good cleavage, as do the micas and feldspars for instance, parting along those planes of weakness extends farther in a coarse-grained than in a fine grained rock. Such failure of a rock then allows more water to penetrate causing chemical alteration. Available data also appear to indicate that fine-grained massive igneous rocks have a higher crushing strength than do coarse-grained rocks of similar composition. In general the crystalline rocks with interlocking crystals possesses much higher mechanical strength than do fragmental rocks which have been cemented together. In this connection it is well to realize that some sedimentary rocks, such as dolomite, are distinctly crystalline and hence have high crushing strength. The commoner cements are silica, calcite, dolomite, and iron oxides. In the case of cemented rocks the degree of cementation is of the first importance because it affects not only crushing strength but also the entrance of water. Sandstone, as shown on a geologic map, may vary from very well cemented with low porosity and permeability to a rock little more resistant than is loose sand. Quartzite, or sandstone cemented by quartz into a rock so hard that it breaks through the original quartz grains, has very high crushing strength as well as low porosity and permeability. Shales also vary greatly in durability as well as in mechanical strength. Some have been thoroughly compressed or have a cementing substance. Others have a high porosity and low crushing strength. All have low permeability because of the small size of individual openings and are made of minerals which are in large part resistant to chemical alteration. Although much data is available in the literature on porosities of rocks there is little on permeability except in connection with studies of underground water and petroleum, most of which have little bearing upon conditions which exist at the surface. In reference to chemical composition we must be sure to discriminate between true limestones made of calcite and dolomites or magnesian limestones because of their difference in reaction with water.

Structure. Structure refers to the larger features of rocks, the partings which divide them, including the attitude of such planes of division. Sedimentary and volcanic rocks display bedding planes which are the result of interruptions in deposition. Igneous flows are finer-grained at both top and bottom than they are in the middle where cooling was slowest. Gas bubbles are common near the top of a flow and make the rock much weaker than is the rest of the flow. Many metamorphic rocks have the crystals of readily cleavable minerals such as mica and hornblend arranged parallel producing schistosity. Others have bands of different chemical composition producing foliation. Both these factors result in weakness of the rock along definite planes. There are no definite standards of comparison in regard to the distance apart of bedding planes or other planes of weakness. The terms thick-bedded or thin-bedded are very indefinite and many geologic descriptions ignore such information. All rocks are more or less broken by planes which are due to earth movement. In some localities these planes or joints follow a more or less definite pattern in response to the forces which produced them. In other localities they are irregular in direction, inclination, spacing and continuity because caused by settling or cooling. Many lava flows, however, show regular hexagonal columns due to contraction. Standards of comparison between closely spaced jointing and widely spaced joints are wanting and many geologic reports ignore this point. In regions of disturbed sedimentary rocks the inclination of bedding planes, the position and direction of folds and faults is delineated. In making geologic maps the effect of such earth movements on the arrangement of relatively resistant bodies of rock with consequent shaping of the topography is an immense aid.

Unconsolidated deposits or mantle rock. Over most of the earth's surface there is a variable thickness of unconsolidated material above the solid bed rock. For the most part this surficial mantle rock is due to action of the atmosphere (weathering) on the underlying rock. (See Sec. 3) In other localities, such as some glaciated districts and the Coastal Plain of southeastern United States sedimentary deposits have not yet become consolidated. In these localities bed rock lies hundreds or even thousands of feet below the surface. The mantle rock in many places contains fragments of consolidated rocks which range from small granules to boulders of large size. Mantle rocks may be mixtures with a wide range in size of particles or be assorted to a narrow range of grain size; they may be massive (unstratified) or arranged in layers either of the same or of different composition and texture. In some places certain layers have been consolidated into rock. In many localities unconsolidated or semi-consolidated materials are firmest close to the surface which is exposed to the atmosphere. This phenomenon is due to evaporation of ground water leaving a cement and is known as case-hardening. It is of great importance in geomorphology.

<u>Units for description</u>. The lack of quantitative standards of comparison by which spacing of joints and bedding planes and grain size of igneous rocks may be compared has been noted. Porosity is expressed in per cent of voids. It is determinable from the difference between the density of a substance when dry and when fully penetrated by water. Permeability is given in many different units. In the petroleum industry the commonest is the <u>darcy</u>, which is measured in cubic centimeters of water at a given temperature which are forced through a section one centimeter square and one centimeter long by a pressure difference of one atmosphere in one second. Crushing strength is given in either pounds per square inch or in kilograms per square centimeter. <u>Density</u> is for the metric system synonymous with <u>specific gravity</u>, namely a comparison of weight of a specimen in air with that submerged in water where it loses the weight of its volume of water. The following table presents some data which are of interest. Reports on building stones also contain information on results of freezing and high temperatures upon specimens of different kinds of rocks.

Tables of data bearing on durability of materials

Densities and porosities (Birch)

	Porosity, percent	Density	
		Dry	Wet
Uncondclidated Gumbo soil	54.1	1.19	1.73
Clay	40.0-50.0	1.30-1.60	1.80-2.00
Sandy Soil	53.2	1.25	1.73
Loess	20.0-69.4	0.8 -1.6	1.4-1.93
Silt	49.9	1.36	1.86
Sand	30.0-48.0	1.37-1.81	1.85-2.14
Gravel	20.0-37.0	1.36-2.05	1.65-2.39
Soft Rocks			
Sandstone	0.9-38.0	1.60-2.68	1.99-2.73
Shale	1.5-44.8	1.56-3.17	1.92-3.21
Limestone	.9-37.6	1.74-2.72	2.43-2.77
Hard rocks			
Granite	slight	2.667	
Gabbro	"	2.976	
Diabase	"	2.965	
Ultrabasic	"	3.370	

Crushing strengths, kg/cm^2 (Birch)

	Average	- Range +
Hard Rocks		
Granite	1480	1110-2310
Gabbro, basalt	1800	1340-2900
Gneiss	1560	750-1710
Quartzite	2020	1760-1180
Slate	1230	780-1650
(Buckley)		
Rhyolite, Berlin, Wis.	3210	
Granite, fine-grained, Montello, Wis.	3080	
Granite, coarse-grained, Pike R., Wis.	1615	

Soft Rocks	Average		- Range +
Limestone	960		900 -2640
Sandstone	740		330-1700
Tuff	310		210-210
Marble	1020		710-1600
Niagara dolomite up to	2800	(Buckley)	
Arkosic sandstone, Wisconsin	340	"	
Quartz-cemented sandstone, Wisconsin	830	"	

Grain sizes of sediments (actual deposits show considerable mixture of sizes)

Materials	U.S. Bureau of Soils mm	Wentworth mm
Boulders		over 256
Cobbles		64-256
Pebbles (gravel)	over 1	4-64
Granules		2-4
Sand	0.5-1	0.0625-2.0
Silt	0.005-.5	0.0039-0.0625
Clay	below 0.005	below 0.0039

Summary. In the study of rocks and other materials of the crust of the earth in relation to topographic forms the qualities considered are those which affect durability at the surface. Some of these properties have been outlined in the tables above but their relation to methods of alteration by weathering is taken up in the following section.

Section 3. Weathering

Introduction. The processes of weathering are all directed toward placing the physical and mineral properties of the earth's surface into harmony with their environment. Conditions at the surface are much different from those under which most of the materials originated. Outstanding results of weathering are (a) breaking up of solid rocks into small fragments, (b) chemical alteration, mainly in the direction of reduction of density, (c) chemical combination with water including solution, and (d) the formation of prevailingly simpler compounds which resist further alteration. The processes of weathering include those which are purely mechanical, those which involve chemical change, and those due to the presence of organisms.

Mechanical weathering. Mechanical weathering consists of reduction in size of particles of material without the aid of chemical change. Breaking up of rocks and minerals into small particles is also an accompanyment of chemical alteration. A very striking feature of breaking up of materials is the enormous increase of surface area which results. Areas are proportioned to the cube of linear dimensions. Thus if we break up a single particle of a given diameter into similar shaped particles of a tenth the linear dimension the surface is increased a thousand fold. This rapid rate of increase in surface prepares the way for the agents of chemical weathering. One of the most potent of all purely physical processes which results in breaking up of rocks and other materials is frost. Water which enters into pores, bedding planes, joints, gas bubbles, and other openings near the surface is frozen. In many regions freezing and thawing take place many times during a year. On many mountains it occurs almost every day. Expansion of water when changed into ice is estimated to give a pressure of about 150 tons per square foot or over a ton to the square inch. It is true that this pressure is well below the crushing strength of many rocks but frost does not crush rocks. Instead it breaks them by setting up tension. Tension tests are not included in tables of physical properties although some shearing tests are made. The increase of volume by freezing is about 9 per cent and that of linear dimensions about 3 per cent. Although this seems small the effect is cumulative because when ice melts the water again fills the opening completely. The depth below the surface at which freezing occurs varies widely. In southern latitudes freezing is rare except in mountains. Going poleward the depth of winter frost increases until a normal of several feet is attained in middle latitudes. In the far north as in Siberia and Alaska the ground is permanently frozen to a great depth, locally several hundred feet, and only the surface thaws in summer. This frozen ground probably dates from a time of colder climate probably associated with continental glaciation. A second mechanical process on which most text books lay great stress is expansion of rocks from diurnal or seasonal increase in temperature. The following table gives some data on this subject. The figures are for linear expansion which is very near to a third of the volumetric change. It is well to recall that crystals vary in rate of expansion according to the internal arrangement of the atoms.

Expansion in per cent, from 20 C to 100 C

Quartz	.08 to .14	according to direction in crystal	Volume .36
Hornblende	.05 to .06	" " " " "	.16
Calcite	.17 to -.05	" " " " "	.08
Orthoclase	.00 to .12	" " " " "	.12
Steel	.09 (for comparison)		

Ratio of linear expansion of rocks to temperature change

Granites and rhyolites	$8 \pm 3 \times 10^{-6}$
Andesites and diorites	7 ± 2
Basalt, gabbro, diabase	5.4 ± 1
Sandstones	10 ± 2
Quartzite	11
Limestone, marble	7 ± 4
Slates	$9 \pm 1 \times 10^{-6}$

Thermal conductivity of common rocks in watts per centimeter per degree C
(multiply by .239 to obtain calories /sec /cm /deg.)

Granites	16 to 35×10^{-3}
Diabase, basalt	14 to 3[illegible] "
Gabbro	20 to 30 "

Limestone, marble	14 to 34 x 10^{-3}
Quartzite	39 to 65 "
Sandstone	8 to 42 "
Slate	18 to 28 "
Shale	10 to 17 "
Sand, dry	2.6 (wet up to 23) x 10^{-3}
Clay, dry	2.4 (wet 9 to 16) x 10^{-3}
Snow	2.1
Ice	22.2 x 10^{-3}
Water	5.5 x 10^{-3}
Steel (for comparison)	460 x 10^{-3}

Although the rates of expansion of rocks are less than those of many common metals there are several weak points in the argument that temperature changes do not break rocks. First, rock temperature is often much higher than adjacent air temperatures and is not recorded at weather stations. Second, the low conductivity of rock causes a much moré rapid decrease in temperature (steeper temperature gradient) in rocks than in many other materials thus bringing about marked shearing stress not far below the surface. Third, the expansion of rocks is best shown at joints and other openings; if these are far apart a considerable total expansion is caused. Fourth, the differences in coefficient of expansion in different directions in crystals causes marked shearing stresses in them. Fifth, daily repetitions of temperature-induced shear may readily cause failure through fatigue. Experiments on small laboratory pieces of rock are inconclusive because of the limited total expansion and temperature gradient. It may be true, however, that temperature changes do not break up relatively small rocks. Expansion is naturally most potent in regions of large diurnal temperature variation, that is on mountains and in deserts.

<u>Chemical weathering</u>. Chemical weathering is defined as the work of any agent which causes changes in the composition of the molecules; it is certainly incorrect to limit the agencies concerned to purely inorganic processes. Details of the subject of chemical alteration by the atmosphere and by water with associated substances in solution are far too complex for discussion in this connection. What concerns geomorphology is mainly the alteration in physical state brought about by chemical changes. As with the work of temperature changes, including freezing, these agents cause an immense increase in surface area of particles. This results in speeding up the attack of chemical agents. Among the most active and abundant of chemical agents we may list water, oxygen, carbon dioxide, as well as acids derived either from organisms or from the alteration of sulphides. A large part of the chemical reactions of weathering result in minerals which are simpler in chemical compositions than they were before, less in density, and consequently in many cases larger in volume.

Exfoliation. A result of weathering which is of much importance in geomorphology is the breaking off of concentric shells of rock, a process called exfoliation. Once regarded as due simply to temperature changes, perhaps aided by surficial chemical alteration, it is now known that the concentric fractures which are best developed in massive crystalline rocks like granite, extend too far below the surface for such an explanation. Distance between the partings increases with depth as has been noted in many granite quarries. It is treu, however, that the rounding of boulders of crystalline rocks is in part due to chemical attack from both sides of an angular projection. It is now believed that a large part of exfoliation is due to relief of load on the rock because of erosion of overlying material. It also thought the hydration of feldspar, once regarded as due to weathering, is brought about during the crystallization of the rock by the water which is then present. This probably leaves the rock under stress so that fracturing occurs upon lessening of the overlying load. Massive igneous rocks form rounded summits on account of exfoliation. World-famous exfoliation domes are Sugar Loaf in the harbor of Rio de Janeiro, Brazil, Stone Mountain, Georgia, and Half Dome in Yosemite Valley, Calfiornia. The last named has had one side removed by glacial actiom.

Other chemical changes. In general the igneous and metamorphic rocks are more susceptible to chemical alteration than are sedimentary rocks formed from the products of previous chemical weathering. An example is shale, which, although mechanically weak, is made of clay minerals resulting from chemical alteration. Exceptions to this rule are limestones, gypsum, and salt formed from material which was dissolved in water and are therefore relatively soluble under weathering conditions. The last two rarely reach the surface in humid climates. Susceptibility of silicates to weathering increases from quartz, through muscovite mica, orthoclase feldspar, biotite mica, alkalie plagioclase, hornblends, augite calcium plagioclase, to olivine.

Soil formation. The word soil has been used in different ways. Students of soils (pedology) confine its application to the surficial layer, in few places much more than a foot deep, in which plants grow and other organisms thrive. Many of the older geologists, however, applied, and some still apply, the word to the entire unconsolidated material or mantle rock which overlies solid bed rock. They spoke of the transported soils of glaciated districts whereas if we mean only the surface layer we must recognize that its origin is essentially the same as in non-glaciated districts, namely alteration in situ of broken up rocks. It is only on floodplains and dunes that we find material which was made into true soil and then moved to another locality. Soil making involves not only the inorganic processes mentioned above but also the work of organisms. Bacteria, moulds, fungi, etc. are very abundant in soils. Among the minerals formed are many which aid in plant growth by the capacity of exchanging bases, calcium for sodium for instance. Bacteria and other organisms do not normally extend far below the surface because of adverse temperature, lack of oxygen, lack of food, and the presence of products made by other organisms which are poisonous to them. Plants possess the power to synthesize new chemical compounds taking the requisite materials from the air, water, and minerals already in the soil. On their death the decay of organic substances produces many chemical reagents which promote further mineral changes. Minerals like quartz which are extremely resistant to alteration form very unfavorable soil for the growth of plants.

Soil profiles. Provided that erosion by wind or water does not remove soil as fast as it forms we find a definite order of layers or horizons beneath the surface. The horizons differ in chemical and physical nature. At the surface alteration from original material is so marked that many of the older geologists thought that they were dealing with transported materials. The succession of layers is known as a soil profile and was first discriminated by Russian scientists. The idea was widely disseminated at first that the surface soil is determined by climate and vegetation rather than by the rock from which it was ultimately derived. Such a view may be correct in some localities but it must be realized that a sandstone which consisted wholly of quartz grains could form nothing but a sandy soil regardless of climate. With soils derived from shale, limestone, or igneous rocks, however, there is more truth to this contention. The surface layer of all soil profiles is for the most part light colored because of removal or concealment by carbon of iron compounds. This is called the A horizon. Where grass is or was abundant as in prairie soils, carbon is so abundant that the color is black. Next below is the B horizon, the densest and darkest in color of the sequence. Accumulation of very fine particles called colloids is greatest where subsoil drainage is poor but the soil is not entirely saturated as it is in swamps and bogs. Where drainage was good the fine particles were either carried elsewhere or were aggregated into larger ones (flocculation). The color of the B horizon is generally red, yellow, or brown, in every case brighter than either overlying or underlying material. With poor drainage deep gray or mottled gray and yellow-brown is characteristic. True all-year swamps show no definite profile, although solution of iron compounds with deposition a few feet below is common. Concretions of oxides of iron and manganese are common in or just below the B horizon. Underlying is the C horizon, described by pedologists as parent material. Geologists, however, recognize that it is part of the decomposed bed rock or original deposits which has been altered almost wholly by inorganic processes. It shows leaching of soluble minerals as well as oxidation and other chemical changes.

Climatic control of soil formation. In relatively humid warm climates the processes of laterization, podzolization, and gleization are common. With less moisture calcification, salinization, solonization, and solidization occur. In the far north under arctic conditions tundra soils are formed which are somewhat similar to the bog soils of lower latitudes, but with less organic matter. Laterization occurs in moist humid climates. Iron and aluminum oxides accumulate under some conditions in volume and purity of usable ores. Silica is dissolved. Transitional toward the cooler zones are red and yellow soils characteristic of southeastern United States. Podzolization occurs in dense forests of the far north where evaporation is slow. The A horizon is robbed of silica, alumina, and iron compounds leaving a very light colored soil known as bleicherde or bleached earth. Such soil is very poor for ordinary crops although it supports trees with deep roots. The iron oxides accumulate in the B horizon forming a hardpan known as ortstein. To the south where forests were more open gray-brown podzolic soils formed. Here the ground was less shaded and was warmer. The B horizon has less iron oxide and breaks with a starch-like fracture into blocks of ½ to 1 inch across. Such soils are decidedly more fertile for ordinary crops than are true podzols. Gleization takes place under poor subsoil drainage conditions forming a sticky, compact, rather light-colored B horizon. Where this was formed from glacial till, sand, or loess, it is called gumbotil, gumbosand, or gumboloess respectively. Some of these soils which were formed before the region was drained by erosion of valleys are now changing into silttil which is characteristic of better drained localities. Salinization, solonization, and solidization are marked by accumulation of alkaline salts in the soil and occur only in semi-arid or arid climates. Calcification, the accumulation of calcium salts, is much more important for geomorphology. Soils where calcium carbonate has accumulated near the surface are known as pedocals whereas the soils of humid regions which are being leached of that compound are termed pedalfers. The very black prairie soils, which are so widespread in central United States, are transitional. They are not abundant in other continents. The absence of trees has long been puzzling. Over wide areas the dense subsoil is known to kill off trees in wet years by excess moisture. Other areas of prairie, however, appear to suffer from drought, for tree growth is confined to valleys, including steep slopes. Owing to the marked climatic oscillations of the interior of North America it is entirely possible that alternations of too much water with periods of too little proved too much for tree growth. At the time of the coming of white men prairies were losing ground to forest and were in part continued by the Indian practice of burning the grass. Throughout the more humid prairie belt the grass vegetation brought calcium to the surface. To the west of the prairies lie the chernozem soils with more calcium, greater fertility, and less rainfall. Progressing west toward the true desert lie the chesnut, brown, and gray sierozem soils with progressively decreasing precipitation. In much of western United States calcium carbonate has accumulated in the B horizon or at the surface to such an extent that it conceals the underlying materials and has been mistaken for limestone. Such an accumulation is called caliche. Much of the calcium came from the A horizon by descending water but much was brought from below by evaporation of ascending moisture. In more arid districts, like western Australia, crusts of iron, aluminum, and manganese oxides and silica are recorded. Seasonal rainfall appears to be a large factor in the production of all such deposits.

Structure and texture of soils. Soils are in large part classified by their texture, that is the size of particles as found by mechanical analysis. This is determined in part by parent material and in part by variation of soil-forming processes. Structure, as in rocks, refers to larger features such as plates, crumbs, granules, and prisms composed of aggregates of grains. These are of great importance in erosion for it is the soil which is the first material to be affected by that process.

Depth of mantle rock. The depth to which there is unconsolidated mantle rock over solid bed rock is of great importance in geomorphology. Thickness of accumulated weathered material depends upon the factors which control chemical and physical weathering. In order to have deep disintegration of bed rock it is not only necessary that it be of a type which is readily altered, but that the agents of alteration, notably water and dissolved gases, be able to penetrate the rock to considerable depth. Under the same conditions a quartzite should display very shallow alteration to a rubble of stones and sand, whereas a granite should be softened to a much greater depth by reason of the unstable minerals which are present in it. Schists where the parallel arrangement of the minerals allows ready entry of water should show disintegration to a very considerable depth. Very striking differences in depth of decompositions are observed in glaciated districts where a dike of fine-grained granite may preserve the glacial polish adjacent to a deeply disintegrated coarse-grained variety of the same rock. In accounting for disintegration to depths of scores or even hundreds of feet many geologists have made the error of thinking only of the contest between erosion and weathering. Although it is true that the former removes the products of the latter we must realize that water cannot penetrate deeply into the rock unless there is a force to cause it to move. This force can only be different in pressure or <u>head</u> and is dependent upon the relief of the country. In flat country there is no head to cause deep underground circulation so that thick mantle rock could not originate. Most instances of deeply decomposed rocks are in regions of igneous or metamorphic rocks which have considerable relief. The disintegration must have originated with the present topography and not been inherited from a postulated time of low relief.

<u>Mass movement of unconsolidated materials</u>. The mantle rock or unconsolidated material at the surface of the earth is not everywhere in a stable condition or state of equilibrium. Movement either in large or small units takes place under several different conditions. Some kinds of mantle rock, such as loess loess, will stand for a long time in a vertical face provided only that there is not too much moisture and that the eighth is not above a certain limit. Examination of natural slopes and the sides of artificial excavations shows that most loose materials soon assume a more or less definite degree of slope known as the <u>angle of repose</u>. This slope is dependent upon the density, size, and shape of the component fragments. In places the material which descended to form this stable slope did so piece by piece. In other conditions large masses moved suddently; under still other physical controls mass movement was slow. The last is known as <u>creep</u> or where largely aided by freezing and thawing <u>solifluction</u>. In many localities particularly on slopes the <u>residium</u> of weathering rests with relatively abrupt contact upon the bed rock, not the transition which occurs only under comparatively level tracts. In some places certain kinds of bed rock are concealed completely by the weathered product of higher ground. At the contact with bed rock it is common to find that the strata bend down into the material which is in motion down hill.

<u>Talus slopes</u>. Cliffs of bed rock, formed by any process, are exposed to all kinds of weathering. Fragments of rock whose size is determined by the nature of the bed rock, including its jointing and bedding, fall from time to time and come to rest at a lower elevation. The slope they form is termed a <u>talus slope</u>. If the cliff is near to vertical rocks fall freely and attain a velocity equal to the square root of the product of twice the acceleration of gravity multiplied by the height ($v = \sqrt{2gf}$, where g = gravity and f = distance of fall). A falling stone possesses <u>kinetic energy</u> or stored work measured by one half its mass multiplied by the square of the velocity where mass is weight divided by gravity. By substitution and cancellation it is apparent that kinetic energy weight multiplied by fall, which is the same as the <u>potential energy</u> before loosening. The stone is brought to rest by the friction of the fragments which have already fallen. Their average size determines the roughness of the slope. Large stones which do not lodge between the others may roll to or beyond the bottom of the slope before coming to rest. This explains why the largest rocks commonly occur at the bottom of a talus. This is not because they fell faster than did the others. Stability is attained when the force of friction just balances the component of the weight of the stone which is parallel to the surface of the talus. Because these forces are nearly the same throughout the entire slope it follows that the angle of slope must be essentially constant although perhaps in some places the bottom may have a somewhat less degree of inclination. In this analysis we have assumed that the rock breaks into fairly large fragments which are not rapidly affected by weathering after fall. In some mountains

talus accumulates on top of snow, melting of which destroys the normal even slope. In this way ridges of loose material parallel to the foot of the cliff are formed. Angular materials may form slopes of 50 degrees. The talus of quartzite at Devils Lake, Wisconsin, slopes at about 35 degrees. Gravel comes to rest at about 35 to 40 degrees depending on the amount of sand present. Dry sand will lie at 32 to 38 degrees, but if wet the angle is much less, 22 to 25 degrees. Slopes of clay are unstable at much less angle and few attain as much as 16 degrees. These are for the most part not true talus slopes. Talus must continue to accumulate until the entire cliff is buried provided no debris is removed from the foot of the talus slope. Thus it comes about that the talus does not fill an angle at the foot of a vertical cliff but instead is a relatively thin mantle over a sloping rock surface formed by weathering back of the cliff at the top. In the case of ridges due to the outcrop of a relatively thin, resistant layer the talus slopes on opposite sides must eventually meet. If the thickness of the hard rock is constant and the elevation of the bottom of the slopes about the same the result will be a ridge top of very nearly the same elevation. Talus slopes are best developed (a) where the constituent rock is not easily altered by chemical weathering and (b) in arid climates where chemical alteration is slow.

Solids and liquids. Before we can consider the phenomena of landslides and creep it is necessary to review some of the physical properties of solids and liquids. A solid is generally thought of as a substance which under the conditions commonly met with at the surface of the earth will retain its shape indefinitely. When subjected to pressure, either of its own weight or an outside force, a solid fails by breaking or fracture. Generally the fractures are inclined at about 45 degrees to the line of application of the force. A liquid is a substance which must be placed in a confining receptacle in order to retain its shape. In this vessel it will assume a level surface if allowed to stand undisturbed. Pressures applied to a liquid are transmitted equally in all directions (Pascals principle). Yielding to pressure is always by flow and not by breaking. In nature no very sharp line can be drawn between these two kinds of substances. The mantle rock which contains in most places a large amount of finely divided soft minerals (clay) is a good example. If dry, a face of moderate height may retain its shape until broken down by weathering. However, if the pressure exceeds a certain amount, or the time is long, or weathering does not confuse the result, a slope may bulge at the bottom by flow. The pressure and time at which such movement is noticable depends largely upon the amount of water which is present. Since the amount of water varies from time to time it follows that the physical behavior is variable. The degree of fluidity of any substance is termed viscosity, which is measured by the amount of shearing stress (force) parallel to the immovable bottom which is required to produce differential movement. The unit is the poise which is the force in dynes applied to a square centimeter to produce a difference of velocity of one centimeter per second at a distance of one centimeter from the base of the fluid. Viscosity depends to a large extent on temperature and decreases with its rise. Water at 20 deg. C has a viscosity of 0.01 poise. It should be noted that viscosity determines not only force required but the time rate of change in shape. If the force and viscosity are known the time required for a given change in shape can be readily computed.

Landslides. In many places a slope either of earth (mantle rock) or bed rock which has previously been stable suddenly yields to the force of gravity.

This phenomenon is known to engineers as slope failure. It is generally the result of an unusual amount of water, freezing and thawing, or of excavation by man. Sometimes removal of material from the foot of a slope by natural causes brings about eventual failure. Yielding is in many cases along planes of weakness in the rock such as bedding planes or joints. In unconsolidated material fracture at the top of a slide is generally along a nearly vertical plane which curves outward toward the bottom of the slope. Above this line of breaking the mass settles and slides outward at the foot. In the case of true landslides, motion is accelerated at first and is then brought to a standstill by friction. In the case of the famous rock slide at Frank, Alberta, sufficient velocity was attained so that a considerable portion of the mass ascended some distance up the opposite side of the valley. The topography left by such slides is extremely irregular with many parallel fractures of irregular direction and extent. Cat-steps in loess are of this nature. The slide masses tilt in toward the higher ground leaving many undrained depressions along fractures. This serves to catch rain and lubricate the planes of movement. Although attempts have been made to analize the mathematics of the curve of fracture it is not worth while to follow them here. In the first place most, if not all, are based on unproved assumptions which do not take account of the variation in viscosity from the relatively dry surface down into the wet interior, and in the second place these slopes are not land forms.

Base failure. Another form of failure in which the entire lower part of the moving mass behaves distinctly like a fluid is known to engineers as base failure. Some of the best known examples are the famous Panama Canal slides. In these the bottom of the canal which had been under 30 feet of water rose overnight into islands. In these instances the drier upper part of the slide was carried along on top of the gluid base. Mathematical analysis appears futile in such circumstances. It seems, probable, however, that detailed study of natural slopes on the same material might have proved of value in determining a safe slope for the sides of the excavation. Equilibrium was attained when the slope component of weight of the mass was equal to the force of friction. Diversion of surface drainage appears also to have aided in drying out the slides. Somewhat similar, but perhaps more rapid, movement of saturated ground occurs during heavy rains in semi-arid regions. These are known as mudflows. Similar flows occur in mountains, including what are known as mud streams, rock glaciers, and earth flows.

Creep slopes. True creep is a very slow motion of viscous material which moves like a liquid even though it may contain many fragments of hard rock. Each layer moves parallel to and faster than that immediately below, although relative change in velocity may in some cases be greatest next to the firm bed rock where most water must accumulate. This type of movement is known as laminar flow. Although freezing and thawing doubtless aid in creep they are not the sole cause for the process which not only extends farther from the surface than does frost action but also occurs where there is no frost at any time. Water must be present to lower the viscosity of the mantle rock. It is obvious that the component of weight parallel to the slope is the motive force for creep. It is also clear that material is added to the mantle rock uniformly in proportion to distance from the top of the slope, here designated by the letter h. Velocity is then proportioned to the sine of the angle of slope which at small angles may be taken as directly proportioned to slope or tangent of angle. Now if velocity is constant, the thickness of moving mantle rock must increase down the slope uniformly and the angle of inclination be constant. But, if as is commonly observed, the thickness of mantle rock is essentially constant then, velocity must increase in direct proportion to distance h. Therefore slope must be directly proportioned to h. Total fall at distance h is equal to slope multiplied by h. Hence by substitution of s:h for slope S; we arrive at the conclusion that f: h^2. The necessary constant of proportionality must depend upon the viscosity of the mantle rock. Actual velocity should follow the law of laminar flow and be proportioned to square of depth times slope, and time required to move a certain proportionate distance would be viscosity divided by product of force times proportionate motion. It follows that the curve of a slope which is due to creep of the mantle rock is an inverted parabola. Convexity of hilltops has long been observed and the explanation proposed above was first given by Gilbert. In order to check the reliability of the conclusion recourse is taken to a well-known mathematical relation. An equation such as given above can also be written:

$$\log f = \log \text{constant} + n. \log h$$

When plotted on ordinary coordinate paper by looking up the logarithms, or directly on logarithmic coordinates, this equation is that of a straight line. The inclination of this line is proportioned to the value of the exponent of h and the intercept with the line for h = 1 gives the value of the constant. The caution must be observed that most small scale topographic maps are not accurate enough to test this law. It is also necessary to take points not too far from the crest of a divide and not along the ends of spurs where the mantle rock may move in either direction. Something should be known of the geology because marked change in bed rock or variation in thickness of the mantle rock upsets the validity of the law.

Stability of creep slopes. The development of creep slopes is also related to removal of material at a constant rate from the foot of the slope. If the material were not constantly removed stability would be attained and motion brought to a stop. Water erosion must also be minimized either by a cover of sod or a concentrate of stones. Removal of material may be due either to (a) steepening of slope because of a change in bed rock or (b) a stream. The creep slope leading down to a stream which is commonly observed in limestone country must have developed from weathering of the rock after the valley was first made by running water.

Solifluction. The term solifluction was originally applied to creep of mantle rock in subarctic climate where freezing is common and during the summer there is a constant supply of water from melting snow.

Some geologists seem, however, to have used it as a synonym for all creep. In Europe much attention has been devoted to a search for evidence of a past severe climate in regions just south of the glacial boundary, a climate similar to that of subarctic regions today, presumably due to the presence of the nearby continental ice. It is not clear, however, that a climate much different from that of today is required to explain the observed phenomena. Similar efforts have been made in this country. The unglaciated part of the Baraboo quartzite range of southern Wisconsin is almost all covered with a mantle of angular quartzite debris mixed with clay and sand. This originally spread out in low slopes at the foot of the hills. In recent time small streams have eroded this mantle, much of which is evidently residual from a former cover of sandstone and dolomite, concentrating the included boulders into "stone rivers." It is not clear, however, that any of these have crept to a notable extent but the mantle from which they were formed may very well be a relic of periglacial climate.

Stone rings and stripes. A peculiar feature of soils in the severe climates of high latitudes and high altitudes is local concentration of stones into either stripes running down the slopes or rude polygonal networks. These features are commonly ascribed to pushing together of stones by frost action, in part altered by solifluction. Such features are not strictly forms of topography and so will not be further discussed.

Technical terms. English geologists often employ the terms head, warp, trail, and coombe rock for mantle rock in which frost action is inferred as the cause of creep. Solifluction layers are also described.

Solution. Minerals which are commonly weathered by solution comprise carbonates, sulphates, and chlorides of the alkalies and alkaline earths. The commonest of these are calcite, dolomite, gypsum, and halite. Less abundant carbonates are aragonite and magnesite, the former found mainly in fossil shells, the latter in veins. Calcite and dolomite are almost insoluble in pure water but dissolve readily when dissolved carbon dioxide is present; in order of decreasing solubility are aragonite, calcite, dolomite, and magnesite. In calcite-dolomite mixtures the calcium is said to be dissolved at a rate about 24 times that of the magnesium. In the ground waters of dolomitic-limestone regions, however, the excess of calcium over magnesium is not nearly so great as that. Calcite is present in high-calcium limestones most of which have a very fine texture. These rocks are much more soluble than are dolomites and magnesian limestones. Impurities which affect solubility consist of sulphides, mainly of iron, iron replacing magnesium in dolomite, clay minerals, and chert. If the impurities are disseminated throughout the rock the effect is much less than where they are concentrated in definite strata. If the strata of shale in a limestone are impervious to water they protect the rock below from solution. In horizontal strata a shale layer may act like the roof of a house. Permeable layers of sandstone, however, admit water and retain it, thus increasing solution. Solution is not a simple process. Organic acids undoubtedly aid in solution near to the surface but break down into bicarbonates at depth. The saturation point of the underground water is determined not only by temperature but also by other substances which are present. For instance sulphates appear to reduce solubility of bicarbonates.

Rate of solution. The rate at which the limestones of a given region are being dissolved can be determined from the amount of ground water runoff and the mineralization of the underground waters. Determinations of total solids are preferable to statements of hardness or alkalinity in which the assumption is made that all the dissolved limestone is calcium carbonate. As the sulphates and chlorides as well as the aluminum compounds also came from the limestone this seems fair. Average total solids in the Nashville Basin of Tennessee is about 329 parts per million, Niagara dolomite of Wisconsin 440 p.p.m. and Galena-Platteville dolomites and limestones of Wisconsin 400 p.p.m. Computation based on the conditions in Tennessee with the assumption that the ground water runoff is 9 inches a year, work out at the removal of 214 tons of limestone from every square mile per year, equivalent to less than 1/1000 of a foot thickness. In computing the time required to form a residual mantle rock of given thickness one must not lose sight of the fact that the density of the insoluble material is probably not over half that of the parent material.

Redeposition of dissolved material. Much of the dissolved limestone does not reach the ground water at once. Solutions which enter air-filled openings lost much of their mineral content through evaporation, temperature change and pressure change. The result of deposition is the dripstone (stalactites and stalagmites) which adds so much to the beauty of many cavern. Calcite crystals are thought to have formed in water-filled passages. In considering redeposition we must also consider the factor of mass relations by which small particles are dissolved at the same time that larger ones grow. The final result of redeposition is to fill up openings which are no longer used by underground drainage.

Porosity and permeability. Porosity is defined as the percentage of volume of openings in a given volume of rock and is determined by weighing when wet and when dry. Permeability is the quantity of liquid which can be driven through a unit cube of the rock in unit time under stated pressure and viscosity. (see p. 2). The only calcareous rocks which have appreciable primary porosity are chalk, oolitic limestone, certain crystalline limestones and dolomites, and coral or shell limestone. Secondary porosity consists of openings made after the deposition of the rock. Most of these are cracks or joints although some are along bedding planes. Jointing is due only in part to regional forces; it is mainly the result of induration, drying, compaction, and relief of load of overlying material. Some secondary openings are ascribed to solution between successive periods of deposition (unconformities or disconformities). In general secondary porosity decreases rather rapidly with depth as proved by drilling for fresh water which is only rarely found in joints at depths of over 200 feet from the surface.

Underground water circulation. As soon as a limestone emerged from the sea the exposed openings would be filled with fresh water from rain. This fresh water is less dense (proportion roughly 1.0 to 1.03) than is the salt water which occupied the openings at first. Original salt water is often called connate. The fresh water floats on top of the heavier salt water and extends to such depth below sea level that the column of lighter water exactly balances that of the denser fluid. Such a condition is illustrated in such localities as Florida and the larger oceanic islands, where the land is high enough to permit such a hydrostatic balance. If the openings do not all communicate with one another, however, conditions must be extremely variable at first with no definite water table. In some places fresh water which is in excess of that required for balance escapes in submarine springs. We must bear in mind that there can be neither extensive penetration of rain water nor underground flow unless sufficient pressure head is available. Water cannot flow through underground passages without the consumption of energy by friction or loss of head. In small openings and with low velocity the flow of water is laminar and loss of head is proportioned to velocity divided by the square of diameter of a cylindrical passage, and velocity is directly related to head. In larger openings flow is turbulent, loss of head is related to square of velocity divided by diameter, and velocity is related to square root of head. For constant volume frictional resistance to laminar flow is inverse to fourth power of diameter, and for turbulent flow to the fifth power of diameter of opening. As a general thing fresh waters, whose flow is concentrated in the largest and shortest available routes, will not penetrate deeper below the surface than the level of equal pressure. It follows that extensive solution at depth must follow upon considerable uplift of the land and that cavern formation cannot take place to an important extent when the land is low and flat. As passages are enlarged by solution the shortest route to the point of outlet is most likely to be favored. Only when compelled to by lack of large direct openings will the waters take circuitous paths which lead them far below the outlet levels, be these into the sea bottom or into stream-eroded valleys. Erosion of valleys must, therefore, precede extensive cavern formation. Exploration of some of the damsites on limestone along Tennessee River has disclosed indubitable solution openings down to about 100 feet below the lowest known level of the river. Deep wells in limestone regions confirm these observations by the fact that salt water is found below fresh water not far below the level of the valley bottoms.

Topographic effects of solution. In geomorphology the primary interest is topographic effects of solution, not details of cave formation. Rain which falls upon soluble formations can either run off to streams or take a route through the earth to the same outlet. We can think of the ground as a leaky roof through which much of the rain penetrates instead of running down to the gutters. Water which enters the ground soon enlarges the lines of minimum frictional resistance. Thus intersections of two joints are enlarged into circular tubes. The larger openings along bedding planes are also made wider. Solution is aided by mechanical abrasion of particles carried in the moving water just as surface streams erode their beds. As erosion lowers the levels of the main outlet streams bedding planes which are on their level are dissolved into a complex system of caverns whose pattern is controlled by jointing. When the level of the stream reaches a lower permeable bedding plane the upper level is abandoned. These different cavern levels are connected by vertical, or near vertical, openings called wells. In Mammoth Cave, Kentucky, some of these are over 10 feet in diameter and 200 feet high. Water enters into these wells through depressions called sink holes and the sides cave into them. Locally several sinks join one another in a large enclosed depression. Some sinks receive the discharge of a surface stream. Others which have clogged contain ponds. Some are enlarged by caving of the walls. Most sinks are partly blocked up so that caverns are commonly entered via former outlets to streams rather than through sink holes. Many sinks extend up through sandstone which overlies the soluble rock. Among such may be mentioned many near Mammoth Cave, Kentucky, and Mont Lake, near Chattanooga, Tennessee. Some streams have underground channels which they follow during low water with the surface course used only during floods when the cavern is overtaxed. Outcrops of salt and gypsum-bearing formations in humid regions display many sink holes and extensive caving of overlying strata. The breccias of northern Michigan are ascribed to collapse of salt bearing strata in pre-Devonian time.

Natural bridges. Natural bridges are abundant in some regions of limestone bed rock. A few may be remnants of cavern roofs which collapsed at all other points. More commonly, however, the bridge is due to an underground leak through a narrow spur of a meandering valley. A few may be due to leakage through a waterfall.

Endpoint of solution. Solution produces in many places an extremely irregular bed rock surface which is disclosed when the overlying residual material is removed. This is because solution proceeds more rapidly along joints. Theoretically, some areas where the soluble rock is unfissured should survive after the remaining area is dissolved down to a plain. Production of a plane surface is readily possible with weathering by solution because attack is distributed over the entire area with fair regularity.

Technical terms. Most of the technical terms which have been applied to solution topography are of foreign derivation. Such topography as developed on limestone is called _karst_, possibly either from _causse_ (French from _calix_ or lime) or from the Slavic word _kras_. Most of the remaining words are from the Serbian because of the abundant literature on the limestone region northeast of the Adriatic. In that country, Jugoslavia, there is little mantle rock; this may be due to purity of the limestone, high relief or possibly seasonal rainfall. Ribbed and fluted surfaces of bare rock are called _karren_. Chasms along enlarged joints are termed _lapies_ or _bogaz_. Vertical shafts or wells are variously called _ponor_, _cenotes_, _jamas_, or _swallow-holes_. Small sink holes are _dolines_, larger ones _blind valleys_, _bournes_, _uvalas_, or _ouvalas_. Very large enclosed depressions are _polje_. Residual hills are _hums_ or _pepino hills_.

Summary. Under phenomena of weathering not only the processes but the resulting topographic forms have been considered. Exfoliation which results in the rounding of exposed rock masses which are relatively free of joints is ascribed mainly to relief of load because the overlying rock has been eroded although the hydration of feldspar by surface weathering is also a factor. Soils were considered because they play such an important role in water erosion. The various usage of the word "soil" were explained: engineers and some of the older geologists use the term as a synonym for mantle rock. Although climate is a very important control in the formation of true surficial soils parent material cannot be ignored and the material termed by pedologists by that name is by no means free from the effects of inorganic weathering. Topographic forms due to weathering consist of talus slopes, landslide slopes, creep slopes, and karst topography, the last developed upon water-soluble rocks. Talus slopes have a uniform angle because material is retained on them by friction. Creep slopes, if the moving mantle rock is of approximately uniform thickness, are shaped like an inverted parabola. Unaltered creep slopes are found where water erosion is at a minimum. Karst topography is found in humid districts where the underlying rock is readily dissolved by water. They are, like creep slopes, consequent upon prior erosion of stream valleys. It is improbable that solution passages ever extend very far below the bottoms of adjacent valleys. Characteristic features are depressions with underground drainage including lost or sunk rivers which flow at least part of the time underground. Natural bridges are mainly formed by subterranean leaks through rock spurs. Landslides, or relatively sudden mass movements of the mantle rock, cannot always be analyzed mathematically because part of the moving material behaves as a liquid at the same time the remainder acts as a solid. The slide area is left with numerous undrained depressions. A multitude of technical terms has been applied to different features of karst topography, most of which appear not only unnecessary but actually undesirable and confusing.

SECTION 4

Major relief features of the earth

Introduction. Differences of elevation on the surface of the earth are very small compared with horizontal distances and the size of the entire earth. The difference of level of the highest mountains and the greatest deeps of the ocean is only about 9 miles or roughly one four-hundredth of the earth's radius. People exaggerate hills and valleys because in ascending a slope it is necessary to do work against the force of gravity, far more effort than to walk an equal horizontal distance. Looked at broadly the surface of the earth can be divided into continents and ocean basins. The line of division between these is not necessarily the present-day shoreline for along many coasts there is a broad continental shelf under not more than a few hundred feet of water. Beyond the shelf there is a relatively abrupt descent to depths of many thousand feet. An average of 500 measurements of the continental slope given by Shepard indicates 8 feet vertical per 100 feet horizontal (8% or about $4\frac{1}{2}$ degrees or 422 feet per mile). Locally, as off parts of eastern North America, higher figures are recorded up to 700 feet per mile (13% or about $7\frac{1}{2}$ degrees). Comparing these figures with slopes of terrestial hills it is evident that were no water present the continental slope would be readily visible and be considered a major topographic feature. Another important difference between continents and ocean basins lies in the underlying bed rocks. The former are to a large extent underlain by sedimentary rocks which must have originated in rather shallow water and these lie upon granitic igneous and metamorphic rocks. With the exception of relatively small portions the ocean islands are underlain almost entirely by volcanic rocks.

Cause of elevation of continents. Many hypotheses have been advanced to account for the relatively abrupt descent from continents to ocean bottoms. Many of these theories involve assumptions as to the origin of the earth. All of them must certainly be regarded as highly speculative, except insofar as they are supported by measurements of the physical nature of the earth. Among other views may be mentioned the theory that the moon was detached from the earth leaving the basin of the Pacific Ocean. Others have tried to make out a tetrahedral pattern in the continents on opposite sides of the Atlantic to see if they were once joined and have drifted apart.

Gravity measurements. The earth, like all substances, attracts other matter. The force with which the earth attracts any given object is the weight of the latter. Since by Newton's first law of motion force = mass times acceleration we express the attraction of the earth by the time rate of change in velocity (acceleration) of an attracted object. This quantity is commonly denoted by the letter g. Mass is a property of matter which is independent of attraction; mass of a given object would be the same on the moon as it is on the earth and is equal to weight divided by g. The value of g is commonly found from the period of vibration of a pendulum but can also be found by weighing the same mass at different localities by means of a delicate spring balance called a gravimeter. Average or normal value of g is 980.665 cm/sec^2 or 32.2 feet/sec^2 but this varies not only with latitude, because the earth is not a sphere, but with density of nearby objects, as well as with altitude. Altitude change is about 3.086×10^{-5} cm/sec^2 per meter. The value of g is often expressed in gals which are equivalent to the force of one dyne acting for one second on one gram mass thus imparting unit rate of acceleration. Observations have been made at sea in submarines which can avoid the effect of waves by sinking below the surface of the water.

Conclusions from gravity measurements. When a value of g has been obtained at a given station it is necessary to correct it to sea level for comparison. Several methods of making these adjustments have been used all of them depending upon several assumptions, none of which involves the exact density of nearby materials. When the adjustment has been made there is almost always a discrepancy called an anomaly whose value and sign depend upon the particular method which was employed. The general concensus of opinion, however, is that the continents are underlain by rocks of lower density than those beneath the ocean basins. This can readily be understood when it is realized that granite is one of the commonest igneous rocks and that a considerable portion of every continent is made of relatively light sedimentary rocks. The density under continents to depth of about 6 miles has been estimated to average close to 2.67. Continental rocks are often called sial because silica and alumina are abundant in them. Observations of the value of g made in the oceans and on many oceanic islands indicate average density of about 3.0 despite the presence of a considerable thickness (several miles?) of light deep-sea sediments. These denser rocks are probably rich in ferro-magnesian minerals such as make up gabbro and basalt; they are often called sima because they are rich in silica and magnesia. The foregoing inference based on density is confirmed by the chemical composition of volcanic rocks of oceanic islands and the mainland. This is best shown around the Pacific, where with the exception of the southwest part, the lavas are basalt. Volcanoes of the continents are made of much lighter andesite.

Evidence of earthquake waves. The velocity of earthquake tremors depends directly on elasticity and inversely on density. At first sight it would seem that velocity of such waves would be less under the oceans than under continents but as a matter of fact elasticity increases more rapidly than density. Daly states that the average velocity of two different types of waves is 4.3 km/sec and 3.78 km/sec under the oceans against 3.55 km/sec and 3.08 km/sec respectively under the continents. Again we may conclude that bed rock under the oceans is really denser than is that beneath the continents.

Theory of isostacy. The data presented above seem to indicate that the continents can be regarded as floating. At a certain depth the material of the earth is believed to offer no resistance to long-continued forces, that is is behaves like a liquid. This theory is known as isostacy. Although properly in the field of structural geology we will find a number of occasions to refer to this theory in the study of geomorphology. The idea of isostacy is also supported by the observed fact that deviations of apparent from true vertical close to the bottoms of mountains indicate that the rocks beneath mountains are made of lighter-than-average material. This phonomenon is often called "deviation of the plumb line" but is actually determined by comparison of horizontal distances found by precise triangulation with those obtained by astronomical observations. The latter depend upon level vials whose reliability is affected by attraction of the mountains. Another proof of isostacy which is commonly presented is the post-glacial rise of the earth's crust which is observed in many glaciated regions. It is fair to observe, however, that many of these rising regions are localities of ancient rocks which have been repeatedly elevated throughout geologic history long before Pleistocene glaciation.

Time of adjustment. If the material of the earth behaves as a fluid under very high pressures it must possess viscosity. In applying this principle it is well to recall that strength or resistance to deformation is measured in relation to area whereas force, which is due to weight of overlying material, depends upom volume. Now with change of linear dimensions areas increase with the square and volumes with the cube. The value of viscosity of rocks is high, perhaps about 10^{22} poises. It is possible to compute time required to cause a stated change in linear dimension by the use of the formula:

$$\text{time} = \frac{3 \times \text{viscosity}}{\text{force} \times \text{relative change}}$$

Applying this to a cliff of rock 1200 feet high, it appears that a bulge at the bottom of one part in a million would take 10 years, and of one part in ten a million years. Erosion of an area might, therefore, readily be faster than change due to isostatic readjustment.

Technical terms. In reading papers on isostacy the following definitions of technical terms will prove a help. Strain is a change in shape (dimension or volume) of a solid due to an impressed stress or force. Strength pf a solid is measured by the stress needed to produce a permanent change of form (usually rupture). Elastic limit is measured by stress required to produce a strain which is no longer directly proportioned to stress. Solid is any substance which possesses strength for a short period but not necessarily for infinite time. Level of compensation is the depth within the earth where it is inferred that pressures are equal or hydrostatic. Elasticity is the ability of a solid to resume its former shape on removal of a stress. Modulus of elasticity is proportion between stress per unit area and change in linear dimension or stress required for unit change; bulk modulus or volume modulus applies to change in volume, and shear modulus is similar to viscosity.

Conclusion. Although the primary facts of isostacy, namely that the higher parts of the earth's crust are less dense than the depressed areas, appears well substantiated it is apparent that the theory has been applied rather too easily to account for phenomena which could better be explained otherwise. At best isostatic adjustment is a slow process, one that could easily be outstripped by erosion. Many alleged instances of isostatic settling of sediments are much more likely explicable by compaction. Clays may lose 50 per cent of their original volume in this way and the numerous emergences of geosynclines during sedimentation are awkward to explain under the idea of isostacy. Supposed recoil upon melting of a load of relatively light ice is equally unsubstantiated, for one might justly ask why just the last uplift of certain glaciated areas should be blamed on melting of the ice when there were no glaciers before older uplifts. It seems certain that isostacy should not be appealed to in order to account for every crustal movement!

Section 5 -- LAND FORMS DUE TO VULCANISM

Introduction. Study of land forms which are the direct result of vulcanism brings us necessarily into the field of igneous geology. Alterations of primary volcanic topography by water and ice will be postponed to later sections, and no more attention given to volcanic processes than is needed to explain the original topography due to eruptions.

Kinds of volcanic activity. The observed emergence of igneous rocks takes place under widely different conditions. Certain eruptions are relatively quiet whereas others are accompanied by violent explosions. Volcanic activity is characteristically intermittent both in time and in location. Some authorities discriminate in order of increasing violence: Hawaiian, Strombolian, Vulcanian, and Pil ean eruptions. The order is related to increase in viscosity of the erupted lavas, as well as to increasing amount of gases in the molten rock. It is not closely related to chemical composition, although in general it does agree with increase in amount of silica. It appears to be inverse to melting temperature which are highest in rocks with least silica. Not all eruptions involve expulsion of lava; some are merely explosions. Clouds of hot gas are often ejected, some of which flow down slopes like a fluid. Declining volcanic activity includes escape of steam, vapor, and gases from fumaroles. Most lavas show original temperatures of over 1000 degrees Centigrade and must therefore originate at a considerable depth below the surface.

Distribution of vulcanism. Vulcanism has occurred in many parts of the world where there are now no active volcanoes. As a general thing it is associated with movements of the earth's crust. The "ring of fire" of andesitic volcanoes which lines the shores of the Pacific Ocean occupies the site of a chain of relatively young mountains some of them still in process of formation. Not all volcanoes, however, occurred where the rocks are disturbed. The volcanic mountains of the Colorado Plateau broke through nearly horizontal sedimentary rocks. Many volcanoes occur in ocean basins where the structure of the underlying rocks cannot be determined.

Types of volcanic ejecta. Flows of lava which congeal on the surface are best crystallized in the middle where cooling is slow enough to permit this to take place. More rapid temperature reduction at top and bottom of a flow leads to fine-grained or glassy texture. Cooling under water causes rounded blocks up to several feet in diameter called pillow lava. Marine or stream deposits commonly occur between flows. The wrinkled skin of a flow is often called by the Hawaiian name pahoehoe and the broken clinker-like material is called aa. The top of a flow contains gas bubbles which leave holes, many of them filled with minerals deposited by ground water; such are termed amygdaloids. In composition lavas range with decreasing amount of silica from rhyolite and felsite through andesite to basalt and diabase. Material which was greatly foamed by escaping gases is variously named ash, clinker, pummice, and scoria. In many places these fragmental volcanic rocks were transported and somewhat sorted and stratified by water before final deposition. Such deposits are tuff or ash beds. These are volcanic-sedimentary rocks and are variously classified by different authors.

Land forms. Following Cotton we will discriminate: (a) basalt eruptions including domes, valley-floor lava fields, spatter cones, and scoria mounds, and (b) non-basaltic eruptions including domes and spines due to extrusion of partially solidified lava, cones, fragmental cones, craters, calderas, and explosion craters outside of cones.

Forms produced by basalt eruptions. Basalts are the most fluid lavas for their viscosity is much lower than that of more sileceous rocks. Viscosity ranges from 80 to 140 poises at 1400 C to several thousand poises at 1200 C. In contrast andesites have a viscosity of 150 to 160,000 at 1400 C and granites have far higher values. Mountains built by basaltic flows consist mainly of lavas. Some of the vents on the Columbia Plateau formed domes with a slope of less than 100 feet per mile or slightly over 1 degree, a slope too low to be perceived by the unaided eye. Where all lava came from a single vent increasing viscosity with cooling as distance increased led to steepening of the slope. The Hawaiian mountains have slopes near the top of only 250 feet per mile but these increase on the flanks to about 1000 feet per mile. In many places the lavas came either from fissures or from so many different vents that slopes are much confused. Very hot lavas may have locally made veritable lakes of molten rock. Filling of valleys with lava led to many stream diversions. Drainage was obstructed and where lava entered standing water it cooled as pillow lava which was very easily eroded. Notable examples of extensive lava fields of basalt are the Columbia Plateau and the Deccan flows of India. Before weathering, the tops of basaltic flows displayed in many places a ropy structure with numerous small mounds. In other places the lava was broken up by bursts of expanding gas into the spiny cavernous surface known as aa. Pressure ridges up to 25 feet high are common and in some places the pahoehoe skin was broken into fields of loose blocks. At other places the molten lava flowed out from beneath the crust leaving lava caves. Where some of these were to wide to support the roof collapse left a hollow.

Spatter cones. Where basaltic lavas were ejected violently from small vents the fragments were blown into a fire fountain. Where the particles congealed a spatter cone was formed. These are common along the fissure vents of the Craters of the Moon in the Columbia Plateau. Few are more than 50 feet high.

Scoria mounds. Mounds up to 500 feet high composed of scoriatious basalt are by no means uncommon. Examples include the larger cones of the Craters of the Moon in Idaho. The slopes range up to 35 degrees where the scoria is coarse. Local eruptions on the flanks of large basaltic volcanoes formed scoria cones.

Cones and domes of siliceous lavas. Lavas which contain a high proportion of silica have much higher viscosity than does basalt. Flows are, therefore, thicker and flow more slowly than basalt although their temperature is lower. Andesite flows congeal on slopes up to 45 degrees and these steepen outward from the vent producing a convex profile. Commonly much debris breaks loose from the terminus of a flow to roll down into a talus below. In some places the lava is so viscous that instead of making a flow it simply wells up into a dome. In other localities a solidified filling of the vent is pushed out into a spine. The famous spine of Mount Pelee, Martinique, had vertical sides. Had it not been for rapid disintegration it is thought that it would have reached an elevation of 2500 feet instead of the 1000 feet actually attained. Well-known examples of extruded domes are Puy de Dome, France, 1700 feet high, Tomichi Dome, Colorado, 2000 feet, and Lassen Peak, California, 3000 feet. These domes have no craters and are rudely cyclindrical in outline.

Fragmental volcanic cones. The chief product of eruptions of siliceous lavas is fragmental. Application of the word ash to such is misleading for they are not products of combustion. The gases which cause so much foaming with decrease of pressure at and near the surface include steam, nitrogen, carbon dioxide, carbon monoxide, hydrogen, sulphur, and chlorine. Many students of volcanoes have ascribed steam to intrusion of meteoric water but the requisite mechanism is difficult to explain. Volcanic gases are lethal by reason of their high temperature rather than their posionous ingredients. In many eruptions the gases, accompanied with hot ash, flowed down the side of the mountain. This phenomenon is called a nuee ardente and caused the great loss of life at St. Pierre, Martinique, in 1902. Large fragments called bombs are mixed with the finer ash which is widely spread by wind and water, as well as by these gas flows. Some think that the hot gases locally reach the temperature of fuison of ash, but not all regard indurated tuff (ignimbrite) as due to melting. Many large cones contain a very small percentage of flows and are predominately fragmental material. Heavy rains, in part due to rising air currents and in part to condensation of volcanic vapor, cause many mud flows of ash. In Java such flows are called lahar but the use of another technical term in a science which is already overburdened with them seems questionable. In some cases the water came from breaking of the side of a crater in an ash cone. Mounds of slid material up to 50 feet high are recorded. Cones are in many places so closely spaced that they form a confused mountain range. In other places they are concentric, one within another.

Mathematical form of ash cones. Many years ago attention was directed to the beautiful symetrical curves displayed by isolated ash cones. It has been suggested that these are related to the probability curve of the fall of debris from a single vent. Others deduced a logarithmic curve based on an inferred angle of shear in loose material or lava. Theoretical curves were then plotted and shown in comparison with actual profiles. The probability curve interpretation can be dismissed as ignoring: (a) inclined projection of fragmental material, (b) effect of wind, (c) rolling and sliding of dry ash, (d) mudflows, and (e) the enlargement of area with distance from the vent. Taking the area of the innermost circle as unity each concentric circle showing unit increase in radius adds an area which increases in arithmetical progression. The writer has plotted several profiles on logarithmic coordinates in order to determine the exponent of the equation of the curve. It is apparent that some peaks show an exponent of over 1, that is the curve is an inverted parabola. Doubtless these are lava cones. Side slopes have an exponent very close to unity and are talus or slide slopes. Lower down two different classes occur, one with exponent about .67, the other about .4. The latter form the lowest parts of slopes and are obviously ash fans and flows. The middle slopes with a higher exponent appear to be mixed flows and fragmental material. It appears that none show any relation to a shear slope, if indeed such is possible.

Craters. The crater of a volcano is a funnel-shaped expansion of the actual vent or neck. In some places, however, the vent, instead of being roughly circular, is along a fissure or rift. At the Craters of the Moon, Idaho, a series of cones occurs along such a rift. In size, craters range from little more than the diameter of the neck to huge calderas several miles across including several distinct vents. The sizes of the conduits which brought lava to the surface is known where volvanoes have been completely eroded away leaving only the lava-filled passage or volcanic neck. The inside slope of a crater cuts through the ash and flow bedding which is exposed except where sliding into the crater has obscured the relations. Some active basaltic craters contain lakes of molten lava whose level varies greatly from time to time. Calderas are distinguished from craters only by size so that there is no sharp distinction. Opinion has varied on the origin of large calderas. Some have advocated explosions such as those at Karkatoa and Katmi. Others infer subsidence, basing this conclusion on observed paucity of large masses of older lava in the ejecta. Williams concluded that Crater Lake, Oregon, was formed from a high peak and that about 17 m^3 of material have been removed in the process. The last great eruption of only a few thousand years ago he thinks accounted for 10 to 12 m^3 of ash or about 5 m^3 of rock after allowing for foaming and breaking. This would leave over 10 m^3 unaccounted for, provided the estimate of the original mountain is correct. Subsidence could be explained by (a) diminution of activity and foaming up with gas, and (b) drainage of molten rock to other vents. Intrusion of rock at depth has also been suggested but it is difficult to see how this would afford more space. Many craters which are no longer active contain lakes despite the high porosity of both flows and ash. Few such lakes overflow. Best known in this country is Crater Lake, Oregon about 2000 feet deep and 1000 feet below the rim. It contains a number of minor cones due to activity after formation of the caldera, one of which projects to form Wizard Island.

Maars. In some volcanic districts, like the Eifel, New Zealand, and Italy, there are large enclosed depressions with comparatively small amounts of volcanic material; some contain lakes. These are termed maars and are ascribed to volcanic explosions. In Hawaii rings of tuff enclosed craters up to several miles wide and are ascribed to abnormally violent ejection of fragments. In some regions of undisturbed sedimentary rocks there are isolated areas up to a few miles across where the strata are greatly disturbed by faulting and folding. These areas have been ascribed to hidden vulcanism and termed cryptovolcanic although no igneous dikes, hot springs, fumarole action, or mineralization is present to confirm the hypothesis. Topographically many form basins, although some display isolated hills of disturbed strata. Glovers Bluff, Wisconsin, Kentland, Indiana, Wells Creek Basin, Tennessee are several examples in the United States.

Summary. Land forms produced by vulcanism depend upon (a) nature of the extruded rock, and (b) amount of gas and vapor included in it. Although the melting point of lavas with a high content of silica is actually lower than that of dark-colored basic rocks the viscosity of the molten rock at a given temperature varies in the opposite direction. Basalts are the most fluid lavas. They flow out quietly with comparatively little foaming and explosive action, therefore making domes with rather gentle slopes. In other places they buried large areas beneath almost horizontal flows. On the other hand, light-colored acidic lavas are violently ejected and in large part foamed into ash and scoris. Steep-sided cones composed largely of fragmental material result. In some volcanoes partialy solidified vent-fillings have been forced up into domes and spines. Mathematical study demonstrates that convex slopes of volcanic mountains results from increase of viscosity of lava with distance from the vent, even slopes are due to sliding, slightly concave slopes to a mixture of ash with flows and the lowest slopes of many volcanoes are mudflows and washed ash. Craters and calderas differ only in size. Although some large calderas have resulted from explosion many are probably due to subsidence consequent on diminution of foaming in the molten rocks beneath the vent. Although some maars are undoubtedly due to volcanic explosions the relation of cryptovolcanic structures to vulcanism is open to question.

LAND FORMS DUE DIRECTLY TO EARTH MOVEMENTS AND IMPACT

Introduction. It has long been observed that some of the highest portions of the continents are composed of marine sediments which were, however, not deposited beneath very deep water. That there has been a change of level of at least the elevation of the present surface is thus readily demonstrable. If at the locality under discussion the strata are still essentially horizontal regional uplift is inferred, but if now in attitudes in which deposition could not occur, then local upheaval must be recognized. At most localities of uplifted marine formations the effects of subsequent erosion are apparent. The present section is confined to land forms where there has been little if any alteration since local uplift, omitting topographic forms where previous earth movement simply guided erosion by placing resistant materials in certain attitudes. Causes of earth movements, either local or regional, will not be discussed.

Classification. Local uplifts which caused definite land forms may be divided into (a) folding or bending and (b) faulting or breaking of the crust of the earth. The results on topography of impact of extra-terrestrial objects will also be discussed.

Folding. Many mountain ridges are composed of strata which dip at the same angle as the flanks slope; these are anticlinal ridges. If our attention were directed only to this coincidence of dip and slope we might conclude that folding was so recent that a remnant of an original fold has been preserved. This erroneous view was often presented to account for the remarkable anticlinal ridges of the Jura Mountains, northwest of the Alps. But there, as in many other localities, careful geological study demonstrates that the mountains exist because they are protected by a formation of rock which is more resistant to erosion than are the formations beneath the valleys along the synclines. The tops of the ridges do not preserve an original surface of the youngest strata deposited before folding. Although mountain uplift is undoubtedly still going on in many localities at the present day, erosion is so rapid compared to uplift that few remnants of such primary surfaces have been discriminated with certainty. A good example is Dominguiz Hill, south of Los Angeles, California. Here Pleistocene gravel dips at the same angle as the sides of the hill and alteration by erosion is confined to small gullies which radiate from the crest of the anticline. At some localities in that district the surface expression of an anticline is so slight that it can be recognized only by the radial stream pattern. In Texas intrusions of plastic salt into unconsolidated and semi-consolidated strata have upheaved hills up to 100 feet high and about two miles in diameter.

Faulting. A fault is defined as a break in the crust of the earth where one side has moved away from its continuation on the other. In cases where the final result is shortening the displacement is called a thrust; where expansion resulted the faulting is normal or gravity. As with topographic forms resulting from folding it is in many places difficult to tell whether what is now observed is the direct result of the earth movement or came about from the position in which that movement placed resistant masses of material. In some cases escarpments resulting from faulting have been observed in process of formation. In other instances close agreement of the escarpment with the line of displacement proves that the movement did not take place long ago. If it is possible to prove that faulting occurred later than a land surface, which is now displaced by the fault, recency of movement is safely concluded. Such is possible when a lava flow, an alluvial cone, or an old erosion surface has been displaced and the escarpment is not much eroded. If there are enclosed depressions on the downthrown side of the fault recency is assured, for such forms could in most regions not have been made by any other agency. Displaced stream courses, hanging valleys, and abnormal stream slopes may also provide similar evidence if cautiously interpreted to exclude other possible explanations. However, the characteristically straight or gently curved trace of a fault may still be preserved in the outline of an escarpment after long erosion during which topography was shaped by the relative positions of resistant formations. So also remnants of the original fault surface on spur ends, termed triangular facets, may be present on old fault escarpments which are often termed fault line scarps. Slickensided surfaces on inferred remnants of the fault surface may in some instances be due to landslides or merely uncovered by erosion of one side of a minor displacement. Lines of hot springs, inferred to occur along the fault plane, should also be interpreted with caution. The same remark applies to supposed discordance of the flank of a dissected mountain range with its internal structure. Any region may have faults of diverse ages so that obliteration of topographic effect on certain faults does not exclude the presence of another generation of younger movements. Escarpments which parallel the strike of a resistant formation are more likely simply the result of differential erosion and do not prove recent faulting. Abrupt contrast in slope between a mountain side and adjacent gravel-covered lowlands may be simply the normal contact of talus slopes with pediments or stream-worn plane surfaces resulting from prolonged erosion. Valleys whose origin is ascribed to faulting are termed grabens (German, grave) and uplifted blocks between parallel faults horsts. Caution is necessary to discover whether or not the faulting is recent for such topographic forms are readily simulated by differential erosion, conditioned on the arrangement of resistant rocks by ancient faulting. Apparent freshness of fault escarpments in glaciated districts may readily result from glacial erosion of fractured rock along one side of an ancient fault.

Examples of fault blocks. Examples of mountains which consist of young fault blocks occur in regions of recent earth movements such as western United States. Mountains in the Columbia River lavas of southern Oregon have long been placed in this class and the conclusion is still unchallenged. All along the east foot of the Sierra Nevada there is similar evidence of displaced lava flows, as well as enclosed basins which must almost certainly be due to earth movement in relatively recent time. The triangular facets of the Wasatch, on the east side of the Great Basin, appear to demonstrate young faulting which displaced a surface of considerable relief. The Great Rift Valley which runs through Palestine, the Red Sea, and much of eastern Africa, is associated with enclosed depressions of structural origin, some of which extend far below sea level.

Impact craters. At a number of localities there are depressions of considerable size which do not appear to be of volcanic origin. Best known of these is Meteor Crater (Coon Butte), Arizona, which is three-fourths of a mile in diameter and 600 feet deep. A rim of shattered, upturned limestone rises 160 feet above the adjacent plain of horizontal formations. Large blocks of limestone occur not only on the rim but scattered over the surrounding country. Sand in the bottom of the pit is partly fused and a considerable amount of meteoric iron has been found, although no single large mass has been uncovered by many years of drilling and shaft-sinking. Undisturbed rock was reached less than 1000 feet below the bottom of the crater and no igneous rock was discovered anywhere. Somewhat similar depressions are reported in Africa, Asia, and elsewhere, although none of them has been explored as thoroughly.

Processes. At first sight it may seem difficult to explain the upturned strata in the rims of such craters. But when we realize that meteorites have a velocity of 30 to 40 miles per second more than 50 times the speed of a high-power rifle bullet and compute their kinetic energy a better appreciation of the phenomena of impact is obtained. At 40 miles per second every gram of a moving object would develop 494,700 calories when brought to a standstill. Although this seems small in comparison with an estimated 2.14×10^{11} calories per gram for an atomic bomb it must be realized that a meteorite might well be vastly larger than a bomb. Not all the energy of a striking meteorite would be converted into heat, for much would be required for pulverizing rock, for sound, and for seismic waves. Nevertheless it is clear that there would be enough heat to cause a violent explosion with probable vaporization of most, if not all, of the celestial visitor. The explosion would rank as comparable to one of a high explosive rather than low-power expansion of steam and gas. The extreme shattering and fusing observed at Meteor Crater as well as the upturned strata are thus explained.

Carolina Bays. Throughout a large area centering about the Coastal Plain of South Carolina there are numerous lakes and swamps in elliptical basins up to half a mile across. Not only the shape of these basins but the northwest-southeast orientation of their longer axes is most striking. The typical basins are surrounded by rims of sand, which somewhat resemble lake beaches. When first distinguished in aerial photographs the original investigators ascribed these basins to impact of a vast shower of meteorites, possibly a comet. Each individual meteorite was thought to have penetrated some distance into the sand before exploding and throwing up the rim. For corroboration they point to unusually high values of terrestrial magnetism southeast of many of the Bays (so called from presence of bay trees). No drilling has been done to test what causes the magnetism. Alternative hypotheses relate the ellipticity of the basins to rotary currents when they contained open water. One suggestion is that the parallelism of longer axes is accounted for by gyroscopic action of the earth's rotation on currents set up by winds. The primary origin of the depressions has been ascribed to original irregularities of the sea botton, to solution or soluable material in the marine sands, and to emergence of artesian springs. Some point to transitional forms between typical elliptical bays and irregular solution depressions. Until test drilling has been done no final decision is possible, although it is true that the importance ascribed to currents and waves in such small shallow basins seems rather far-fetched.

Summary. Because folding normally occurs only at depths great enough to permit bending of firm rocks without fracture, elevations due directly to recent folding are rare. An area above a fold may be raised up so slowly that erosion destroys the uplift almost as fast as it forms. On the other hand, faulting may be sudden and many examples of new fault scarps have been recorded. Uplifted fault blocks may be either horizontal or tilted. A number of craters which display no evidence of vulcanism have been recorded and, for some at least, the hypothesis of impact of large meteorites appears well sustained. A meteorite must be unusually large to pentrate the atmosphere which normally protects us from extra-terrestrial bombardment. The origin of the Carolina Bays has been much debated without presentation of any fully satisfactory explanation.

SECTION 7 WORK OF RUNNING WATER

Introduction. The effects of rainfall upon the surface of the earth is evident over much, if not most, of its land area. Even in the driest climates some rain work is found, although it may in some places date from a time when the climate was moister than it now is. Removal of material by running water has been recognized for a long time. In the early days of geology a violent debate was carried on as to whether or not rivers made the valleys in which they now flow and many geologists of those days advocated a structural origin for many large valleys. Despite the qualitative study of the phenomena of water erosion which has been carried on for a long time, relatively few studies have been made from a quantitative standpoint. Most of these were undertaken by engineers and students of soil erosion rather than by geologists. Today there is no generally accepted mathematical explanation of erosion; many text books ignor the subject entirely and, if mentioned, often contain glaring errors and misapprehensions. The following discussion aims to clarify what is known from the studies of Gilbert, Little, Horton, and others.

Definitions. Much of the misunderstanding between engineers and geologists has been due to non-uniformity in definitions of words. In order to read any of the works intelligently it is necessary to know the following definitions. Acceleration is the time rate of change of velocity which is space traveled per unit of time. Force is what causes a mass of any substance to undergo acceleration; force = mass x acceleration. Velocity of running water may refer to average or mean velocity, V_m or to bottom velocity, V_b. Work is the product of force x distance of application. Power is the time rate of work or by substitution force x velocity. Slope of streams is fall in same unit of horizontal distance, that is the tangent of the angle of inclination from horizontal. Competence of a stream relates either to diameter or to weight of the largest particle which it is able to carry at a stated velocity and has no relation to the quantity of material it can transport or erode. Load is the actual amount which a stream moves in unit of time under stated conditions. Capacity is the upper limit of coarse debris which can be transported along the bottom of a stream under stated conditions. Hydraulic radius is the result obtained by dividing the area of a cross section of a stream channel by the length of the bottom in the same section (wetted perimeter); in channels of great width the hydraulic radius is equivalent to depth. Reynolds number is a factor by which comparisons may be made between streams of differing dimensions; it consists in multiplying a dimension (depth or radius) by velocity and dividing this product by the viscosity ($R = \frac{\text{dimension} \times \text{velocity}}{\text{viscosity}}$) always using the same system for quanities. Kinetic energy is stored work possesed by a moving object or mass of water expressed by the formula $E = \frac{1}{2}\text{mass} \times V^2$. Mass is weight divided by the acceleration of gravity, g. Kinetic energy is work the moving object will accomplish if stopped in a given distance. Kinetic energy of rotation, or fly-wheel effect, is stored work in a rotating mass at distance r from the axis of rotation; it is expressed by the formula $E = \frac{1}{2}\text{ mass} \times r^2 \times \text{ang. velocity}^2$, where velocity is expressed in radians (angles equal to a radius on the circumference) per second. Impact refers to effect of a current of water on a stationary object by reason of kinetic energy of the water. Tractive force is the force exerted on the bed of a stream by a column of water of unit cross section, expressed by product of weight of the column multiplied by the slope, that is the potential energy of the column when at the top of the slope. Terminal velocity of settling refers to the limit to falling velocity of particles in any medium, such as water, which is attained when force of resistance nearly equals weight of particle. Archemedes principle is the fact that objects submerged in a fluid loose the weight of their volume of the fluid; objects in water have an effective density which is less by one than their density in air.

Flow of water. Water has two, and very likely three, distinct methods of flow: laminar or layer over layer, turbulent or eddying, and shooting or plunging. Its behavior with respect both to erosion and transportation of debris differs greatly in these types of flow.

Laminar flow. Laminar flow has already been discussed in connection with movement of mantle rock and lava. It is present in water only when the Reynolds number is low and the bottom and sides of the channel are very smooth. It appears to be present close to the boundary of all kinds of flow. Theoretically the bottommost layer has no velocity. Erosion and transportation by laminar flow is negligable. Velocity = constant(depending on nature of the bed) x hydraulic radius ^{2} x slope ($V_m = CR^2 S$).

Turbulent flow. The major part of the flow of streams is turbulent. Breaking up of laminar flow is due to increase in the value of a couple of forces, namely force of flow and resistance of the bed. These opposed forces bring about eddies or rotating masses of water whose axes are not all parallel to the bed. This disturbance is usually visible at the surface as welling up, reversed flow, or ripples. These rotating masses of water aid greatly in erosion and transportation of sediment but absorb much kinetic energy of rotation and thus slow down the mean velocity.

Manning's formula is $V_m = C \times R^{2/3} S^{1/2}$, the constant depending upon the nature of the bottom. For British engineering units the constant is 1.486 / n, the latter a roughness factor which is generally about .04. Maximum velocity is attained a short distance below the surface and mean velocity is .82 the maximum. Bottom velocity cannot be determined directly by various methods of computing the theoretical velocity at the border of laminar flow have been attempted. The intensity of turbulence varies with different conditions and in different parts of the same cross section. It can be expressed quantitatively in several ways although difficult of actual determination. One method is to draw lines of equal velocity in a cross section of unit length. The water is then divided into a series of prisms whose apices are at the line of maximum velocity and whose bases at the bottom of the stream are all of unit area. The result obtained by dividing the potential energy of each prism (weight x slope) by the velocity gradient at any given velocity level is called by the German name austausch or mixing coefficient. Computation thus displays the inferred intensity in different parts of the section which represents the rate of energy transfer in the direction of the bed of a stream.

Shooting flow. At very high velocity the surface of a stream becomes much agitated and water moves in spurts or jets. Such shooting or plunging flow is commonly found in rapids and waterfalls. Apparently the maximum possible velocity of moving water is about 22 meters/sec and is fixed by the absorption of energy in this type of flow.

Mixed flow. At low velocity, and especially where the bed is very rough as through grass and other vegetation, water obeys a law which is intermediate between laminar and turbulent flow. Experiments with the flow of very thin sheets gives a formula V_m = Constant $D^{.9}$ $S^{.7}$ Some experiments on actual slopes covered with vegetation show a variety of exponents in the relation of quantity, Q, to depth, D.

Erosion by running water. Running water removes material from the bed by: (a) direct lifting of loose particles by either horizontal or inclined currents of water, (b) abrasion by rock particles carried in the current, (c) impact of larger rocks on the bottom utilizing their kinetic energy either of translation or rotation, and (d) collapse of vapor-filled bubbles causing violent impact of water on the bed (cavitation or water hammer). Many text books mention only the second and third methods. One might then justly ask the question as to how the water first derived its tools! It is clear that most of the course of every stream is over loose material which is not put in motion except in unusually large floods so that it is evident that loose material is really the source of most of the load. Cavitation is another method which is often overlooked. It occurs, however, only when the velocity is over 12 meters/second, a velocity attained in but few localities. When it does occur, the following formula applies.

$$\text{Pressure} = \text{velocity} \times (\text{density} \times \text{bulk modulus})^{1/2}$$

Substituting 2×10^{10} dynes/cm^2 for the bulk modulus of water and unity for density this becomes: Pressure = (lbs/in^2) = 50 x velocity (ft/sec). It is evident that repeated blows of this magnitude must be very destructive. Erosion of the bed can occur only when force of the current as exerted at the bottom exceeds the resistance of the bed material to removal. This resistance consists of two distinct factors: (a) weight of particles under water and (b) the mechanical state of packing. The former is readily found but the second is not easily determined. Mixtures of different sizes are much harder to disrupt than is a deposit of all the same sized particles. Packing is very important with small particles like clay. Another factor is the presence of stones which are too large for the water to remove and which, therefore, protect finer material beneath.

Force required to move a particle. In considering the force required to dislodge a given particle from its resting place on the bed of a stream (provided it is not broken from a larger particle) must be carefully distinguished from the force required to keep it in motion after it is started. There must be an upward component of motion at an angle of A degrees with the bottom. Force necessary to dislodge a loose particle must then equal weight x sine A (considering only weight under water). This does not take into consideration the resistance due to state of packing of the bed material. Several different methods of computing the force of the moving water which is exerted on the exposed portion of a given particle have been used. These comprise: (a) impact of the water which is proportioned to square of bottom velocity multiplied by exposed area of particle, (b) tractive force computed as pressure per unit area computed from weight of column of water of unit area multiplied by slope, (c) hydraulic lift due to difference in velocity at different depths, (d) velocity gradient expressed by kinetic energy of unit volume divided by depth, and (e) intensity of turbulence as expressed by the mixing coefficient. Rubey equates the resistance to lifting to energy of flow.

$$\text{weight} \times \text{sine } A = \text{constant} \times V_b^2 \times \text{area of cross section} \times \cos A$$

Simplifying and solving for radius of particle, R:

$$R = \frac{3 \times \text{constant} \times V^2_b}{4 \times \tan A \times \text{weight}}$$

Weight is again that under water and tan A = slope, S. It follows that diameters of particles are proportioned to square of velocity and weight of particle moved at given velocity is proportioned to sixth power of velocity. This is known as the "sixth power law" and, although correctly stated when first worked out, has been much abused as a measure of load instead of weight of a single particle which can be moved at given velocity (competence). The tractive force idea is to find the component of weight of a column of water of unit cross section which is exerted parallel to the bed. Evidently this is its weight times sine of angle of slope. Since at low angles sine and tangent are nearly the same it follows that the value of slope may be substituted for sine in the above equation. Following the same procedure as before and solving for R, we find:

$$R = \frac{3 \times \text{constant} \times \text{depth} \times \text{slope}}{4 \times \tan A \times \text{density particle}}$$

Again density is that under water. The next angle of approach is the fact which explains the lift of an airplane wing, and many other phenomena, namely that pressure sideways is least where velocity is highest. This means that a stream where velocity increases away from the bottom should exert an upward force on particles once they are lifted from the bed. This explanation appears to explain suspension of load rather than erosion. Still another measure of erosive force of unit volume was employed by Little. It is well known that loss of head (pressure) in pipes which carry water with turbulent flow is proportioned to V^2/radius (or diameter). Applied to streams this is translated to V_m^2/hydraulic radius; in the case of very wide channels this becomes depth. Since depth = quantity/velocity, substitution yields: Force = V_m^3/quantity. This method does not seem to have been used by others. The intensity of turbulence also has not been widely employed as a measure of erosive force, although it too seems to offer some advantages since it explains the rate of energy transfer.

Force required to keep particles off the bed. The force of moving water which is required to keep particles off the bed is more readily measured than is the force needed to start them. Particles over 1 mm diameter obey the impact law and their diameters are proportioned to the square of the velocity. Particles smaller than about 0.2 mm diameter are more subject to the viscous resistance to settling.

Weight = 6 x pi x radius x viscosity x velocity of settling

Solving for velocity of settling this becomes:

$$V_s = \frac{2 \times g \times R^2 \times \text{density}}{9 \times \text{viscosity}}$$

with all units in the metric system. Solving this for radius (or diameter) it will be seen that this is proportioned to square root of velocity. This relationship of diameter to settling velocity is known as Stokes Law but it is of more importance in sedimentation than in studies of erosion. Particles between .2 mm and 1.0 mm diameter obey a transitional law in which an increase of one cm/sec velocity increases the diameter of particle which can be kept from settling by about 11 per cent.

Loss of energy by streams. Every text book of physics demonstrates that could a particle slide down a frictionless inclined plane its velocity would be accelerated and attain the same value at the bottom of the incline that it would have if it had fallen through the same descent in a vertical direction. This velocity is given by the expression:

$V = (2\,g \times \text{distance of fall})^{1/2}$, fall being measured vertically. If streams of water flowed without friction their velocity would conform to this law and they could do no work on their beds. Observation demonstrates that this is not true for velocity is not accelerated but instead is related to slope. Obviously a retarding force is present and this consists partly in energy absorbed in kinetic energy of rotation and temperature rise and partly in work done on the bed and partly in keeping material off the bed. This resistance is similar to that demonstrated when an object will just move on a real inclined plane, that is when the component of weight down the face of the plane just balances the retarding force of friction. We cannot express the amount of work done on the bed because it is impracticable to measure the other losses of energy. It is true however, that the kinetic energy of the stream is approximately equal to the total retarding force.

<u>Problem of bottom velocity</u>. As with laminar flow the true bottom velocity of turbulent flow must be 0. It is thought that the bottommost layer is laminar flow. Ruby terms the velocity at the upper border of this marginal zone the <u>bed velocity</u>. Although it is impracticable to measure this velocity Ruby attempts to compute its value. Energy supplied by the current must equal energy consumed in friction on the bed. The former is measured by weight of water multiplied by its distance of fall in a given time. Distance of fall is found by multiplying mean velocity by slope by time. Density of water is unity and the angle of slope so small that it can be neglected. Energy consumed in unit length is found by multiplying the wetted perimeter of the channel by unit distance giving area. This is multiplied by frictional force times distance through it acts in given time. Frictional force is equal to square of mean velocity multiplied by a coefficient, and distance is mean velocity times time. Simplifying this equation and making necessary substitutions it appears that mean frictional force on unit area equals weight of unit volume of water multiplied by hydraulic radius times slope. Now this is the shearing stress at the bed and is equal to thickness af layer of laminar flow times viscosity. Reasoning from this premise it is concluded that the fourth power of bed velocity equals constant times square of mean velocity times hydraulic radius times slope. It is this value which should be used for computation of the impact law. It is also apparent that the coefficient of friction, which is included in the constant above, is proportioned to square of mean velocity times hydraulic radius times slope.

<u>Methods of transportation of debris</u>. Material which is kept off the bottom by vertical component of motion is called <u>suspended</u>. This material mainly obeys the law of settling velocity where diameters of particles are proportioned to square root of velocity. Larger particles obey the impact law and their diameters are proportional to the square of the velocity. The particles which cannot be suspended move in a series of jumps called <u>saltation</u>. Their progress is thus slower than the current but owing to their loss of weight under water they have little energy of impact when they return to the bottom. Thus they do not help dislodge other particles into the moving water. Still larger stones are rolled along the bottom as is proved by the rounding of water-transported pebbles. Others are pushed or keep constantly sliding into a hole which the water excavates on the down-stream side. The coarser material which is not suspended for long periods of time is known as <u>bed load</u>. The division between what is bed load and what is suspended is not constant but varies with velocity of the water. Although there is a distinct upper limit to amount of bed load with given conditions which is termed <u>capacity</u>, there does not appear to be any definite limit to the amount of suspended load. Up to 30% has been recorded. Study of the Missouri River appears to indicate that amount of suspended load varies with the square of the discharge. The explanation is probably an increase in turbulence with increase in volume which is associated with increase in velocity.

<u>Transporting power of a stream</u>. Strictly speaking capacity should be distinguished from <u>transporting power</u> of which it is the upper limit. Both quantities are measured in weight of material moved in unit width of channel in unit time. Thus both are expressed in units of power. Although many attempts have been made to express transporting power in terms of velocity of moving water there is no agreement between them. First, we must realize that this quantity will be a measure of transportation and not of erosion. Second, we find that the energy devoted to transportation is only a portion of the total energy of a stream. Third, we must see that the upper limit to transportation (capacity) is by no means a

measure of the relation of power to velocity. Fourth, we must not confuse competence with transporting power. Fifth, if bed load is considered it is the bottom velocity and not mean velocity which should be used. Sixth, there are other factors besides velocity, for instance depth or hydraulic radius determines the vertical velocity gradient and, therefore, the rate of transfer of energy to the bottom. If, however, we take the kinetic energy of unit mass of water as determined by V_m^2 as a measure of force, then power should be related to V_m^3 and in the same way with bottom velocity if we are to consider only the part of the water which is next to the bed load. Rubey used a different method of approach. He states that load per unit of channel width multiplied by the average settling rate of the material equals weight of unit volume of water times square of mean velocity times hydraulic radius times slope. $L/P\ V_s : g \times V_m^2 \times R \times S$ where P = perimeter of bed. Now since $V_b^4 : V_m^2 \times R \times S$ by substitution it appears that $L/P : V_b^4$ but as V_s (settling velocity) is also proportional to V_m then $L/P : V_b^3$ or by another substitution it also works out that L $L/P : V_m^3$. Many other expressions have been used which have been made up from the tractive force idea where force equals product of depth times slope. If we use Little's expession for force as $V_m^2/\ R$ and multiply by V_m to obtain power the expression for power is proportional to $R \times S^{3/2}$.

Critical velocity. Observation shows that it is not correct to use total velocity in any formula for there must be a certain critical velocity before any particles can be moved at all. Many of the formulas are given in a form where this value of velocity is subtracted before applying the relations given above. The energy before the attainment of this velocity is wholly absorbed in internal resistance of the water. It does not appear that this lost energy can be simply obtained by subtraction for it may well be a proportion of total energy which is not constant with change in velocity.

Overloading. Literature about streams abounds in the expression "overloaded". Consideration of hydraulic principles shows that no current can be overloaded even momentarily. Any excess load of bed material is dropped at once when conditions for its transport are not maintained. If attention is given to the suspended load there is no condition which causes overloading. But the material which is suspended at one velocity may become bed load at a lower velocity and vice versa. Streams may be underloaded because they cannot obtain enough debris for capacity but they cannot be overloaded even temporarily.

Experiments in stream transportation. The most extensive experiments in stream transportation of debris were those carried out by Gilbert. Others have been carried out by Mavis and others at the University of Iowa and at the Vicksburg experiment station of the Army Engineers. Nevin has also made some experiments, which unlike the others who used flumes, were done with glass tubes. Gilbert failed to find any single variable which was related to capacity by a simple formula. He found that with constant slope capacity is proportioned to about the third power of velocity above the minimum needed for movement of the bed. With both constant depth and constant discharge the exponents were larger. Velocity used was mean and not bottome because of the difficulty in determining the latter. An inverse relation to depth was suggested. The Iowa experiments gave an exponent of velocity above minimum which varied from 3.03 to 6.24. With a depth-slope formula the exponent was more constant and varied from 3.09 to 3.88 after subtraction of the critical value but the value of the constant varied greatly. Nevin's experiments related only to competence and not to load and the same may be true of the Vicksburg trials. Apparently little attempt has been made to consider velocity gradient as did Little. In a flume the level of maximum velocity must be depressed below what it is in a stream and in general conditions are unlike nature. The

The variability of results suggests that there are other variables which were not given enough weight.

Horizontal shape of stream beds. Few streams flow very far without making a bend. This fact is demonstrated by inspection of artificial canals which were made to carry off flood waters. After a short time these, almost without exception, begin to develop curves unless the walls are made of materials which cannot be removed. Slides of banks, fallen trees and brush, accumulation of drift-wood all serve to deflect the current from one bank against the other. Once started this process is self-perpetuating. It is most marked when velocity of the stream is relatively low. If it is high the obstruction may be swept away for the energy is enough to keep the water in a reasonably straight course; the lateral component of motion is then relatively small. This tendency to swing sideways is well known above a millpond where velocity is reduced. When a stream flows through a segment of a circle elementary physics shows that the centrifugal force of unit mass is determined by the square of velocity divided by radius of the curve; $F = mV^2/r$. Thus force is inverse to radius of curvature. Now velocity of an ordinary stream is, other things being equal, related to the square root of slope so that $V^2 : S$. By substitution we find that $F = mS/r$. Thus as a stream lengthens its course by making bends the slope is decreased and force also is decreased in direct proportion. The relative amount of force exerted by a stream on its banks as compared to that on its bed is also related to the nature of the load which it acquired during floods. If the load is coarse so that it cannot be moved at low water then almost all the force may then be lateral. As bends grow into symmetrical curves they are called meanders. Local variations in nature of the banks prevent meanders from ever becoming true segments of circles. There is no relation between meandering and stage of stream development. All that is necessary is a comparatively low velocity at the time of beginning of the meanders. Many examples of meandering streams may be observed where the bends developed as soon as the stream started to flow.

Cross section of stream channels. In a straight stream the line of maximum velocity is along the center. In a curve this line is deflected toward the outside of the bend. In a straight channel, according to the commonly accepted method of computation, there are two zones of maximum turbulence with maximum disturbance of the bottom. One is on each side of the line of maximum velocity. Since material of the bed load is given a lateral component of motion from areas of maximum turbulence toward those where that phenomenon is less marked, this tends in a straight stream to deflect bed load toward the banks and to form a minor dividing ridge in the middle directly under the maximum velocity. This process appears to initiate sand bars or subaqueous dunes which are so characteristic of streams with a sandy bed. At low water these shoals become islands and may be an important factor in producing a branching or partially braided pattern of channels. In a curve, however, one of the areas of high turbulence is deflected against the outside bank and doubtless accounts for erosion there which produces a cut bank. The other is less marked and lies next the inside. The line of maximum transport of bed load crosses over from the inside bank of one bend to the inside of the next. As it crosses the depth of water is less than it is next to the cut banks. Such shoals are known as crossings. Observation demonstrates that they are built up during floods and worn away during low water. The relation of crossings to turbulence is evident, as is the transport of material to the insides of bends. As bends grow larger they maintain about the same width of channel. The insides are progressively filled with a complex of parallel ridges commonly with swamps between. The areas just below cut banks are known as deeps; these are excavated during floods and then filled at low water. The width of a stream channel depends upon the nature of the bottom. In this connection attention should be ~~aware of~~ directed to the erosion. The tendency of a stream is to distribute its force uniformly across the bed except insofar as this is disturbed by rotational force. Banks with deep water against them are naturally most exposed to erosion. Low, shelving banks are more apt to be built up. Canals dredged with vertical sides are commonly widened by floods at the same time that the bottom is shoaled. Sandy beds are more readily altered by water than are those composed of either gravel or clay. A stream with a sandy bed has a shallower cross section and a greater width for given discharge than has a stream whose bottom is composed either of finer or of coarser particles. Moreover, it is much more apt to be divided by shoals or sandbars. A confusing fact, however, is the variability of volume which is characteristic of most rivers. The channel is no sooner adjusted to one volume than discharge changes and it must begin all over again to reshape the bed. Nevertheless, most streams seem to have a low water channel which is sharply divided from the high-water flood channel or floodplain. The change is due to the relatively infrequent floods. Most variations in discharge can be taken care of in the normal or low water channel.

Size of meanders. It is evident from the most casual examination of maps that large streams have much larger meanders than do small streams. It can also be noted that some valleys meander and that these meanders are considerably larger than are those of streams with the same discharge whose low banks are made of soft material. In some places two distinct sizes of meanders can be distinguished; one large curves of the valley, the other small meanders on the floodplain. Attempts to study the size of meanders in relation to that of streams have been directed wholly to the width of the meandering belt for the shapes of real meanders are not uniform enough to permit of quantitative measurement. Lacking data on average discharge, maximum discharge, hydraulic radius, or nature of bottom, all quantities which vary greatly in different streams, investigators measured the size of streams by width of normal channel. As seen above this is not entirely reliable, for width is related both to velocity and to nature of bottom. From the analysis of forces given above it is easy to see that as meanders grow larger the force of unit mass of water on the outside of the bends decreases because both of decrease of slope and increase of radius of curvature. Therefore, unless a stream is large in volume, the force needed to erode the cut bank is attained only with small meanders; only a large volume stream can make large meanders and still have force left to continue erosion. Meanders stop growing larger when the lateral component of total force of the stream equals the resistance of the bank to removal. Meanders also wear most on the downstream side so that in time they travel slowly downstream, a process termed sweep. Long before meanders on the soft material of a floodplain with low banks reach maximum size, or of move far downstream, they are likely to be cut off across the neck during a flood. This process does not remove all bends from the channel so that meandering simply starts over again. For this reason floodplain meanders do not last as long as do the bends of meandering valleys which are cut down into bed rock. Here both sweep and cutoffs are much retarded and meanders have time to grow very large.

Width of meander belt. Jefferson's studies of width of meander belts did not discriminate between streams of varying width. His averages were that the width of meander belt is 17.6 times that of the stream on floodplains and 30 times in the case of meandering valleys. Bates results averaged for floodplain streams with a width of 100 feet, 16 times; streams 1000 feet wide, 12 times; streams 3000 feet wide, only 11 times. Notable variation from the overall average of 11.3 times are doubtless explicable by difference in bed material. For meandering valleys the figures found by Bates were much higher: 41 times, 18 times, and 17.5 times respectively with an overall average equal to that of the largest streams. Widths were subject to the notorious inaccuracy of the older topographic maps. Although much of the difference between results on floodplains and for meandering valleys may be explicable from the factor of the longer life of the latter. Another phenomenon is that meandering valleys in many cases have swifter, narrower, gravel-bottomed streams, which differ greatly from those of floodplains. This tends to minimize the observed difference.

Misfit streams. Where a meandering valley has a floodplain within it it already has been noted that two sizes of meanders may be present. Such streams have been termed *misfit*, and many attempts made to find the cause. The phenomenon has been variously ascribed to loss of volume by reason of climatic change, diversion of headwaters, cessation of glacial drainage, or increase in underground flow through the valley filling. Instances which can be explained by the first three processes are definitely known. Diminution of flow after glaciation is well displayed in central Illinois. The loss to underflow is called Lehmann's principle but in fact is the least plausible of any. The discharge of underground water is expressed by the following formula:

$$\text{Quantity} = \frac{\text{Constant} \times \text{difference in head} \times \text{area of cross section}}{\text{length of flow}}$$

The value of the constant varies with size of grains, packing, and viscosity. For British engineering units and discharge in ft^3/min) its value at 50 F. ranges from 0.000036 in silt to 1.442 in sand with grains of 1 mm diameter. Taking the cross section of the glacial outwash fill of Wisconsin River at Muscoda as 200 by 10 000 feet, the head at 1.5 feet per mile, and the average grain size at 0.4 mm the discharge would be only 0.55 ft^3/sec. Since the lowest recorded surface discharge of the river is 2000 ft^3/sec it is evident that, even in this extraordinarily favorable instance, the underground flow is only a little over 1/4000 of the surface flow or entirely negligible! In the case of the underfit Kickapoo River it is clear that this principle cannot be of any importance whatever. What does count is that the outlet of this stream, which heads in the high rock hills of the Driftless Area, was blacked up by glacial outwash. Consequent upon this its bed was silted up with reduction of grade and formation of a floodplain in the bottom of the old meanders which were cut in rock. With low banks and lowered velocity the cause of the small modern meanders is clearly evident.

Destruction of meanders. It has already been mentioned that cutoffs do not in themselves destroy the tendency to meander. However, meanders may be destroyed when the velocity of a stream is increased. A good example of this process is the Wisconsin River between Nekoosa and Petenwell bridge. For several miles below the last rapids at Nekoosa there are perfectly-formed meanders with a slope of 0.84 ft/m and an average width of about 480 feet. Abandoned meanders and meander scars indicate that once the stream meandered much farther south than it now does. Apparently this meandering started on top of a delta which was deposited in Glacial Lake Wisconsin. Below the meanders which still survive the slope is about 1.52 ft/m. and the width averages about 700 feet. Evidently the erosion of the sandstone barrier at the Dells, many miles below, has caused an increase in slope and therefore of velocity. It is difficult to arrive at an exact figure for change in velocity because in order to compute this we should know the hydraulic radius. Since there are no important tributary streams in this part of the main river we might assume that cross sectional area is essentially constant. Assuming this the increase in width would increase the hydraulic radius by about 25 per cent. But as a matter of fact the increase in velocity should decrease the area of the cross section. For this reason we will tentatively assume that there is no change in hydraulic radius. On this assumption velocity is about one third greater below the meanders. The exact process of destruction appears to be the development of cutoffs which shorten the course of the river. Two of these are known to have been made in about 35 years. In this case increase of velocity destroyed meanders because the river bed is all loose material. The course below the meanders is not only wider but also shows a partial braiding with many more islands and sandbars than in the slower current of the meanders. Steepening of grade of a stream which is close to the level of bed rock would cause erosion of the meanders to begin thus making a meandering valley.

Effect of rotation of the earth. Objects on the surface of the earth are subject to two forces. First, the attraction of gravity pulls them toward the center of the earth; this force is resisted by the surface of the earth. Second, the centrifugal acceleration due to rotation tends to make them move relatively outward. The horizontal component of this force would make a particle move toward the equator because that is moving faster than the surface at higher latitudes. This force is resisted by the fact that the earth is already adjusted to its rotation so that the shape of the ellipsoid is just enough to counteract this force. When we consider the relative velocity of a moving object, first moving east, then one moving north, an expression for forde in each of these directions is obtained. Parts not related to rotation are neglected. The development of a formula to express the force which tends to turn a moving object to the right in the northern hemisphere and to the left in the southern is quite complicated; it is best explained in works on the winds. The force on unit mass in a horizontal plane is shown by the following:

Force = 2 x angular vel. in radians per sec x relative velocity x sin latitude

If quantities are expressed in the metric system the force is in dynes. It is 0 at the equator, 258×10^{-7} dynes at latitude 10, and 1458×10^{-7} dynes at the poles. For comparison we may take the centrifugal force developed on unit mass of water moving 200 cm/sec in a circle with a radius of 1 kilometer; this figures out at 0.4 dyne. At latitude 70 degrees force due to rotation of the earth on unit mass at same velocity is 0.0274 dyne or about 6.6%. Since this force is always in the same direction it is added to centrifugal force on right hand turns in the northern hemisphere and subtracted on left hand turns. The difference is then double the amount of the force. The force is so small, however, that it not noticeable below latitude 20 degrees but probably accounts for the greater prominence of cut banks and driftwood in right hand turns of large streams at high latitudes.

Longitudinal profile of streams. Very few streams long flow at the same volume. Very few long rivers flow all of their course over the same kind of material. Most streams have tributaries and increase in volume with distance from the source. The foregoing facts make it extremely difficult to arrive at any mathematical formula by which the longitudinal profile of a river can be expressed. Many attempts have been made to do this; some ended in absurdities such as convex profiles and others are impossible of practical application. Among the latter are equations based on amount of wear of pebbles or on size of largest particle carried. In order to derive an empirical formula it is necessary to plat the profile on either logarithmic or semi-logarithmic coordinates. The writer chose a number of streams from the numerous profiles in Water Supply Paper 44 and platted them on the former. An attempt was made to choose only streams which appear to be undisturbed by crustal movement and to avoid those which display marked irregularities in slope. As explained previously this type of platting derives an equation of the type fall equals constant times horizontal distance to a power which is less than unity. The following exponents were found: Wisconsin River below Portage .95; Black River below the Falls .92; White River, Arkansas .95. Of streams in drier climate the

Republican River showed .82 and the portion of the Arkansas through the High Plains .85. The Red River of the South is abnormal in displaying an exponent of only .55. Turning to theoretical reasoning an analysis by Rubey of the forces involved in stream transportation gives a relationship between slope, settling. velocity, load, quantity of water, and hydraulic radius according to the formula:

$$S = \left(\frac{\text{Load x settling velocity}}{\text{wt. water x } R^{1/2}} \right)^{2/3}$$

This is equivalent, other things being equal, to the slope being inverse to the cube root of the hydraulic radius. Such a formula appears inapplicable in actual practice. Wooldridge and Morgan give a formula for stream profiles which is a logarithmic curve; it is equivalent to saying that the slope is inverse to distance from the source. Somewhat better results may be secured from Little's approach He assumed that erosive force of unit mass of water is proportioned to square of mean velocity and inverse to hydraulic radius. In a wide shallow stream the latter is equivalent to depth. By the use of Mannings formula for velocity and making necessary substitutions it appears that this force is proportioned to $Q^{1/5}\ S^{9/10}$ Q = quantity of water in unit width of bed. When solved for S this becomes

$$S : F^{10/9}\ Q^{-2/9}$$

The slope of constant force would then be $f\ :Q^{7/9}$. Little derives the equation for a trapezoidal channel, which is somewhat like a natural stream channel, of $F\ :Q^{3/25}\ S^{9/10}$ which when solved for slope is: $S : F^{10/9}Q^{-30/225}$. From this the profile of uniform force is $f: Q^{195/225}$ or $F:Q^{.87}$. In order to solve any of these for fall, f, in relation to horizontal distance, h, it is necessary to make some assumptions as to relation of average discharge to distance downstream. This can only be found where drainage basins are of normal shape and where results of long-term discharge measurements are obtainable. On the assumption that $Q : h^{3/4}$ the profile of a trapezoidal channel becomes $f =$ constant x $h^{9/10}$ which is not far from the actual observations mentioned above. For that matter the result of the other formula for a wide channel becomes with the same assumption $f:h^{7/12}$ or $F:h^{.58}$ which does not agree very well with actual determinations. Were semi-logarithmic platting to yield a straight line then an equation with a variable exponent would be shown. In such an equation the constant which is applied to the exponent represents the percentage of change in each successive interval of horizontal distance. Platting of eight outwash terraces in Wisconsin gave no support to the variable exponent equation but instead yielded a constant exponent of about 0.7 or distinctly lower than that of present-day streams. The explanation of the difference from the existing conditions is undetermined; it might possibly be explained by tilting.

<u>Slope-wash or overland flow</u>. Where surface runoff does not follow channels but forms a thin sheet all over the land surface the process is called <u>slope-wash</u>, <u>overland flow</u>, <u>unconcentrated wash</u> or <u>sheet flood</u>. This process is important during both (a) initial erosion of a new land surface, and (b) in reduction of sides of valleys.

<u>Hydraulics of overland flow.</u> The two principal students of overland flow are Little and Horton. These authors used quite different lines of approach to the hydraulics of the process. Little's work seems to be wholly theoretical but Horton appears to have tried many actual experiments. These experiments demonstrated that normally overland flow is mixed, although with increase both of distance of flow and of depth it becomes wholly turbulent. On a strip down a slope which is of unit width there is a definite relation between depth, D, discharge, Q, and velocity, V. Since $Q = DV$ substitution shows that for fully turbulent flow

$Q : D^{5/3} S^{1/2}$; for purely laminar flow $Q : D^2S$. Now if we use the formula for velocity in thin sheets derived by Lewis and Neal then $Q : D^{1.9} S^{.7}$. The following development will employ their formula instead of those used by the original authors. It must first be realized that very thin sheets of water may flow in waves; these are probably the result of viscosity. Water piles up until slope is locally enough to overcome viscous resistance. Flow is then accelerated until the sheet thins; then the process starts over again. Horton felt that the waves act like a series of sudden blows and increase erosion.

Little's views. Little, as mentioned above, used the expression which gives loss of head with turbulent flow in pipes to express erosive force, namely $F: V^2{}_m/D$. Substituting $D = Q/V$ this becomes $F = V^3/Q$. Substituting the formula for mixed flow for V this becomes $F : Q^{8/19} S^{21/19}$. Solving for S, $S : F^{19/21} h^{-8/21}$ and the equation of a slope of uniform force by unit volume becomes $f : h^{13/21}$ or $f : h^{.62}$.

Horton's views. Horton employed the time-honored tractive force equation for force on the bed under a strip of unit width. This equation is also known as the depth-slope formula or DuBoys formula. It is simply the component of weight of water on unit area which is parallel to the surface. Since weight is then proportioned to depth, which is Q/V, it appears that:

$$F : D \times \sin A \text{ (where A is the angle of slope in degrees.)}$$

By substitution for value of D and taking $Q : h$ we find:

$$F : h^{10/19} S^{-7/19} \sin A$$

This expression is not readily comparable with that of Little unless we assume that for moderate slopes sin A is essentially equivalent to S (tan A). Making this substitution $F : h^{10/19} S^{12/19}$. Solving for S, $S : F^{19/12} h^{-5/6}$. This yields for a profile of uniform force $f : h^{1/6}$ or $f : h^{.167}$ or a much more concave slope than does the other approach. In making a comparison it is desirable to realize that Horton's formula gives the entire potential energy whereas Little's attempts to determine what part of total force is actually applied to the bed.

Resistance to erosion. Horton computed the force of flow at the point where erosion begins. This was expressed in pounds per foot^2 and ranges from 0.5 lb/ft^2 for newly cultivated soil to 0.5 lb/ft^2 on sod, a range of 10 times. Of course, not all this force is actually expended on the soil for much is lost in internal resistance to flow. The resistance of soil to erosion was ignored by most of the older writers. It depends upon several factors: (a) rate of infiltration, (b) physical nature of the soil, its structure as well as texture, and (c) kind of vegetation. Soils which have the finer particles aggregated into pellets or which swell when wet have a high resistance to washing compared to what would be expected from their mechanical analyses.

Belt of no erosion. One of Horton's major contributions to geomorphology is the reasoning that a certain minimum distance is required below any divide to gather sufficient water to permit the runoff to overcome the resistance of the soil to erosion. In making such computations it is evident that the actual force exerted on the soil is of no importance; what is found is simply the point at which erosion does begin. It is also evident that the width of gathering ground will vary both with rate of rainfall and with nature of the soil. Horton computed that on a 5 degree slope plowed land with rainfall rate of 0.5 in/hr. will not be eroded for 153 feet from the divide, whereas with a rate of 2.0 in/hr the belt of no erosion shrinks to only 38 feet. The corresponding results for sod are 7046 feet and 1762 feet respectively. From these figures it is fair to conclude that the belt of no erosion is a fact but that it is of little importance when the resistance of the soil to erosion is low. When resistance is high, however, the importance of the belt of no erosion is low. When resistance is high, however, the importance of the belt of no erosion cannot be exaggerated. In prairie areas it is not difficult to see that valleys do not extend to the divides. Naturally the width of the belt varies inversely with rate of rainfall so that its effect on erosion varies with the frequency of heavy rains.

Profiles developed by slope wash. All the formulas outlined above agree in so far as they indicate that profiles developed by slopewash are concave upward. The concept of the belt of no erosion shows that unless there is no vegetation such profiles do not extend to the divides. The variation of width of that belt may help to explain the convexity of many hilltops although this is not the only explanation of that phenomenon. The checking of which formula is most nearly correct must rest upon platting of actual profiles of slopes in uniform material. Unfortunately, such profiles are not at present available. In many regions the lower slopes have been covered by filling of the valleys or there is too much variation in the underlying materials. When examples are available the values of the constants of proportionality can be determined. In his analysis Little did not assume that quantity is directly proportioned to distance from the divide but instead used a rainfall equation quoted in the first section. By means of rather complex algebraic analysis he eliminated time and obtained as the exponent of a profile of uniform force 9/11, that is a much less concave slope than do the others. Reasoning that through geologic time quantity is directly related to distance the exponent becomes 7/9. Both of these results are for fully turbulent flow. Horton also used slope distances instead of horizontal distances. On moderate and low slopes this would make little difference.

Topography due to slopewash. The mathematical discussion above presupposes that material is removed uniformly from the area of slopewash. Experience shows that such is not the case. Instead, as pointed out by Horton, many parallel rill channels are formed all over the area. Each goes directly down the slope. Doubtless minute differences in resistance is in part responsible for this concentration. Another possible cause is surface tension of the water which would draw it together into threads. Once concentrated, erosion is magnified by the increase in volume. Total amount of erosion also increases with distance down the slope. Horton's figures based on experiments appear to show that increase in erosion is at a more rapid rate than his formulas indicate. Another interesting observation is that in exceptionally heavy rainfall the sod cover may be broken and rolled up leaving bare soil.

Formation of valleys. Formation of valleys on a newly-formed land surface which is steep enough to cause slopewash erosion follows upon the original rills. Certain rills become deeper than others. Overflow of the tiny divides causes concentration of water in these larger streams. This process Horton termed cross grading. In time it obliterates most of the original rills. Systems of tributary ravines develop across the original rills. Resulting slopes alter the direction of rills. The process is repeated until the entire area of a drainage basin has valleys so spaced that there is no land outside the normal width of the belt of no erosion as found along every divide. Thus it follows that the development of valleys obeys a definite mathematical law. In nature the boundaries of drainage basins are generally ovoid and the material is not uniform either in infiltration capacity or resistance to erosion.

Relation of streams to underground water. Temporary or intermittent streams, which flow only during and shortly after rain or melting of snow, normally lose some of their flow to underground water provided mantle rock and bed rock have sufficient permeability. Permanent streams are supplied in the intervals of no new supply from slopewash or surface runoff by the discharge of springs. It follows that such streams can only exist below the level at which their beds intersect the zone of saturation or water table. Exceptions to this rule are streams supplied by melting snow or the drainage of a humid area which flows through regions of less precipitation. Under these circumstances streams may be losing water to the ground water over a considerable portion of their courses. If underlying material is composed of fine particles causing low permeability this loss is not of very great magnitude.

Entrance angles of tributary streams. Since streams normally flow directly down the slope of the land rather than at an angle to it, it is obvious that the angle at which they join one another depends upon the ratio between two regional slopes. If slopes are gentle and both essentially equal the entrance angle might be of almost any value. This condition is found on flat plains. Here the angles range from 60 to 80 degrees except where interfered with by vegetation. In cases where the main valley has a low slope and the sides a moderate inclination the entrance angles are about 60 degrees. If tributaries enter from s teep side slopes into a gently sloping main stream the angle approaches 90 degrees.

Stream orders. Horton devised a system of stream orders which does not agree with that employed by some Europeans. The short unbranched streams next the headwaters (for the most part intermittent) are the first order. Those which receive tributaries of the first order are second order streams. The same process is carried on as long as necessary. In cases of doubt the stream below a junction is prolonged and the stream with the greater angle of entrance in reference to this line is taken as of the lower order. If both branches are close to the same length the shorter one is of the lower order. If both branches are close to the same length the shorter one is of the lower order. The order of a stream is unchanged throughout its length.

Drainage density. It has long been noted that many drainage basins differ greatly in the number of streams for their area. It has commonly been thought that this fact is related only to climate or to stage of development of the drainage system but Horton urges that the factors of infiltration capacity and width of the belt of no erosion have been neglected. He expresses drainage density by dividing the total length of streams in a basin by its area in square miles. Both permanent and intermittent streams must be included. The reciprocal of twice the drainage density gives approximately the average length of overland flow. Stream frequency is computed by dividing the area of the basin into the total number of streams of each order.

Drainage texture. Texture of drainage is simply another word for density. Many investigators have applied the factor of infiltration to account for coarse texture of drainage, that is a basin with relatively few streams. Although this is important, a factor which has been ignored by almost everyone is the relative resistance of both mantle rock and bed rock to erosion. Drainage basins in areas of resistant rocks are characteristically of low density or coarse texture. This can be explained by the fact that it requires a large area to gather enough water to make a valley. Areas underlain by shale or clay almost invariably have fine textured drainage. This factor explains the drainage pattern of Bad Lands. In some areas of deep mantle rock on hard bed rock shallow gullies display a fine texture although the major valleys have coarse texture.

Laws of drainage distribution. Horton worked out the following laws which govern the number of streams of different orders in a drainage basin. The proportion between the number of streams of given order in a basin to those of the next lower order is the bifurcation ratio. The number of streams of different orders in a basin approximates an inverse geometric series in which the first term is unity and the ratio is the bifurcation ratio. The average lengths of streams in a given basin approximates a direct geometric series in which the first term is the average length of streams of the first order. Horton held that these laws, which follow upon the principles of formation of successive tributaries, lend quantitative support to Playfairs Law of accordant stream junctions. He laid great stress on another relation, the ratio between the stream length ratio and the bifurcation ratio as expressing the nature of the drainage. Stream slopes were found to be in inverse geometric series thus relating them to different stream orders and proving Playfairs law. To give a complete quantitative picture of a drainage basin he listed: drainage area, order of main stream, bifurcation ratio, stream length ratio, and either length of main stream or average length of first order streams. From this the drainage density, stream frequency, etc. can be computed. Methods were worked out by which to estimate the length of first order streams where map data are inadequate. The value of all this quantitative data to geomorphology is yet to be demonstrated but its value in making definite comparisons is obvious.

Drainage patterns. Because streams are shown on maps which give no other data much attention has been devoted to drainage patterns. Most textbooks classify these into dendritic or tree-like and rectangular or trellis. Some authors also describe radial drainage such as that of a volcano, and centripetal as the drainage of a basin, where streams meet at a common point. Braided streams divide and reunite repeatedly. Branches which lead water away from a main stream are distributaries. In respect to the use of the term dendritic it is evident that the originators had in mind only the ordinary hardwood trees for all drainage patterns resemble the branching of some variety of tree! As brought out above the drainage pattern reflects relative slopes of tributaries and main streams and this is in general a result of the geologic structure of the underlying bed rock. If we study the variety of land surfaces on which streams originated it is really remarkable that there are not more distinct patterns. Branching or dendritic patterns indicate horizontal uniformity of material. Great irregularity of pattern indicates a primary surface of considerable relief such as that of the rougher phases of glacial drift or of volcanic areas. Only streams which originated on rather gently sloping but not level surfaces were preceded by the slope wash grading postulated by Horton. In many areas such as the more level portions of upraised sea bottom or glacial drift no preliminary slope wash was possible.

Relation of stream courses to geology. In many areas of disturbed rocks it is obvious that most streams are located on the outcrops of the less resistant bed rocks regardless of the position of anticlines or synclines which might conceivably have directed the primary drainage when the area was uplifted. The original or consequent courses have evidently been abandoned in favor of others which are adjusted to the nature of the bed rock and its associated mantle rock. In this process it is clear that streams which happened to be located on less resistant material deepened and widened their valleys to such an extent that in time they obliterated streams which were not so favorably located. The adjusted streams are termed subsequent. In such areas where the rock formations vary greatly in resistance to erosion the drainage pattern becomes rectangular. Tributaries from steep ridges on the outcrop of resistant formations enter the valleys along the strike of nonresistant formations nearly at right angles. However, there are exceptions to this rule. Where a considerable thickness of similar material occurs drainage is dendritic. The subsequent streams in many places cross through the ridges in water gaps, whose origin is in many places disputed.

Valleys due to recession of falls. Falls occur where (a) a stream descends a steep slope formed by another agency, (b) there is a marked change in resistance to erosion along either a near-vertical plane, or (c) soft easily erodible material occurs beneath a firm cap rock or other resistant layer. In many small falls at the head of ravines the capping material is sod or even just B horizon of the soil profile. Water which falls freely reaches a very considerable velocity, far more than is common in streams. However, there is a limit height at which the kinetic energy of the water is effective in erosion. Above a few hundred feet fall the water is broken into drops by air resistance and hence is ineffective. This is well illustrated in the high falls of the Yosemite Valley, California. In lower falls the descending water swirls around boulders and pebbles which aid in excavating a plungepool beneath the falls. Plungepools are normally filled with water but at the abandoned falls of the Columbia Plateau, Washington, several are easily observed. These depressions should not be confused with smaller potholes made by rotary motion of stones in any rapid current. Excellent examples of potholes can be seen in Interstate Park, Taylors Falls, Minnesota. They are very common along all swift streams. As the crest of a falls is worn back by undermining and falling the plungepool moves upstream. A fall normally looses height not only by breaking off of parts of the crest but also by the steep grade required for the stream in the gorge which is formed by its recession. Although excavated to the depth of the bottom of the plungepool, the gorge below is shoaled by coarse debris, in part excavated there and in part fallen from the walls. Lakes occur in abandoned plungepools. In general, however, the bottom of a valley formed by fall recession has less slope than a valley made by the same size stream by downward erosion along its entire length. A good exampoe of a valley thus formed where the falls wore back until entirely obliterated is the upper section of Grand Coulee, Washington. The width of a gorge made by falls during their recession may vary greatly if the discharge or material has not been uniform. A good example is the gorge below Niagara Falls. A widespot at the Whirlpool is due to intersection of the gorge with an old drift-filled gorge of earlier origin. Energy of fall of unit volume is naturally constant but total amount of work varies with discharge of the stream. This reduction both in power and rate of recession does not in itself leave any record in the form of the gorge. But since there is generally a relation between discharge and width of channel a shrunken stream excavates a narrower gorge than does a larger one. This also is well displayed at Niagara and vitiates its value as a "geological clock" by which early geologists sought to determine the number of years since its formation by dividing the known modern rate of recession into total length of the gorge.

Valleys due to springs. Where a large spring emerges low down on a slope the erosion is somewhat similar to that at the bottom of falls although the amount of available energy may not be anywhere near as great. The elevation of the valley head is fixed by the level at which water can emerge. Valleys of this type have been described from the Columbia Plateau where the waters emerge beneath basalt flows. They are also common in the glacial drift.

Initial formation of consequent valleys. Land surfaces originate in many ways. An area may be the bottom of a sea or lake now emerged from the waves, or it may have been made by stream, glacial, or volcanic deposition. In every case there has been a change from sedimentation in some form to erosion. This change is not necessarily due to uplift or to change in level of sea or lake; it may be due to change of climate or simply to cessation of deposition. Most theoretic reasoning has been started with the premise that a relatively flat sea bottom was upraised rather suddenly so that no significant amount of erosion occured during uplift. Furthermore, it has generally been postulated that the climate was humid and that the initial surface contained more or less irregularities so that the first drainage was imperfect. Good examples of just these conditions may be observed in the lower parts of the Atlantic and Gulf Coastal Plain, as well as on the flatter glacial drifts. Examples of irregular initial surface are found in the rougher areas of glacial drift, in volcanic districts, and in regions where uplift was due to faulting or folding. Two distinct conditions may then be present: (a) the primary surface is so flat, or so permeable, or both, that slopewash cannot occur, or (b) the original surface was altered first by overland wash before valleys were formed. We have already considered Horton's approach to the formation of valleys by successive gradings by slopewash until there is no more area left which can be thus altered. This view is supported by the observed fact that the successive dividing up of a drainage area is displayed in the bifurcation ratio. But where primary slopewash could not occur, as on the glacial plains, the origin of valleys must depend more on chance. Concentration of water would then depend upon local conditions brought about by minor irregularities of the surface. Original ponds and swamps would be abundant in the interstream areas, as is easily observed in glaciated districts. But in either case it is clear that concentrated water does form valleys of consequent streams.

Alteration of valley sides. Were erosion confined entirely to stream beds all valleys would be of the box canyon type with vertical sides. Examples of such valleys are confined to those which were formed not long ago and in which erosion is still concerned mainly with deepening the bed because of rapid flow. Examples are found in many localites where streams diverted by glacial deposition are now making canyons in the bed rock. But very slight reflection shows that vertical valley walls would be extremely unstable. They are altered by erosion of small tributary valleys, as well as by sliding and creep of the sides. Some have thought that all young valleys have convex sides because of the rapid lowering of the bed. This is certainly true in some localities, especially where creep slopes form in incoherent material. In many unconsolidated materials, such as most of the glacial drift, the angle of slope is even like that of a talus slope because it is due to sliding. Rough landslide slopes are also common, particularly where ground water emerges. Reduction of valley sides to a slope naturally exposes them to slopewash. Even if the areas between streams were at first confined to the belt of no erosion it is clear that as valleys were deepened the steep slopes along them would wear back into the formerly immune area. In other words, the slope of stability for given material and climate must start at the stream level and has no relation to the divides. Another factor, which must be considered is that the width of the belt of no erosion is variable because of occasional torrential downpours. This variability of width may easily be a factor in producing a convex divide even in areas where there is no important amount of creep. The belt of no erosion on divides cannot be of constant width until stability is attained along the valley sides by an equality of forse of erosion to resistance of material to removal. Once this condition is reached valley formation is essentially complete unless disturbed by earth movement, change of climate, or the work of man.

Base level of streams. A stream valley can be eroded no lower than the bed of the stream into which this valley debouches, nor can any stream valley be eroded more than a slight distance beneath the level of the body of standing water it reaches, or the level of a valley filled by stream deposits into which, in the case of a semi-arid climate, it ends its course. This limitation is known as base-level and its effects have long been appreciated. The accordance of most stream junctions is known as Playfair's Law. There are some exceptions to this law where the main stream is supplied from melting snows of the mountains or some other more constant water supply than the local precipitation which supplies its tributaries. Many examples of <u>hanging valleys</u> which have been unable to keep pace with the deepening of the main river are present in the Grand Canyon of the Colorado. Another cause of discordant junctions is tilting of the land along the direction of the main stream as is well shown in the Sierra Mountains of California. The velocity of the stream which flowed down the tilted surface was increased and that of the tributaries which flowed at right angles was unaffected until the main valley was deepened. Lateral motion of a stream may also cut away the lower end of a tributary.

<u>Lateral erosion of streams</u>. Lateral erosion is best developed after a stream valley attaines such a <u>grade</u> that it is able to carry off the debris which is brought to it by both tributaries and slopewash. Development of bends and true meanders then can take place because velocity is reduced to a point where the lateral component of motion is important. Most text books lay much stress on the widening of stream valleys by lateral erosion. <u>Cut banks</u> where the stream swings against the bluffs are common and doubtless account for the observed fact that hills adjoining a large stream are commonly steeper than those along small tributaries. Good examples of this are present along the Upper Mississippi valley and need only in part be accounted for by glacial floods. Such erosion serves to upset the belt of no erosion which was in former equilibrium. But the view that this is the major process of valley widening producing a wide <u>floodplain</u> underlain by bed rock at slight depth is not confirmed by examination of most stream valleys of the United States. Much more common is a considerable amount of stream deposits beneath the valley floor. Such are explicable by change of sea level, or of climate or by the indirect effects of glaciation. Lateral erosion is also limited by the factors which control the width of the meander belt as previously outlined. How far lateral erosion might extent in time is problematical.

<u>Formation of pediments</u>. Special conditions which apparently enhance the importance of lateral stream erosion are present in areas where the amount of water is not enough, or the slope is not sufficient, for the streams to transport their load to the sea. These conditions appear to be most readily attained at the bases of mountains in a semi-arid climate like that of the southwestern part of the Basin and Range province. Here the streams for the most part never reach the sea. Instead they are filling, or have filled, basins between the mountains which were originally made by earth movements. Even above the major areas of deposition, in which there is at times standing water in some places, the streams are obliged by the decrease of grade, aided to a small extent perhaps by evaporation, to lay down the coarser part of the load which they acquired in the mountains. They flow in a <u>braided</u> course on these deposits and build up their beds to such an extent that shifts of channel are of common occurrence. Under these conditions the lateral component of force is alone present. Valleys, where they reach the foot of the mountains, are widened, the ends of spurs, outlying elevations, and even the mountain face itself are cut back by lateral erosion. It has been argued that such lateral erosion is not the major cause of <u>pediments</u>, as the sloping areas of smooth bed rock thinly covered with gravel are called, because so few typical examples of cut banks occur. As a matter of fact, it is true that pediments are best developed on rocks like granite or sandstone which disintegrate into material readily moved by both streams and slopewash. It is, therefore, not to be denied that, as the interstream areas are planed down and weathered down, slopewash takes an increasingly important part in reduction of the area. This was realized long ago by the geologists who happened to witness <u>sheet floods.</u> With the scanty vegetation of semi-arid regions overland wash during occasional downpours is of greater importance in shaping the landscape than might at first be realized. Very

slight increase in rainfall would check it by increasing vegetation. Concurrently the increase of rainfall should cause the main streams to erode their beds to lower levels. Just how arid the climaté must be to allow formation of pediments is uncertain and the same remark applies to the possible extent that they might eventually attain.

Cycle of erosion. The fact that erosion progresses through a definite cycle was discovered long ago but was first widely publicized by W.M. Davis. His postulate was that uplift is relatively sudden compared with erosion. Erosion thereupon follows a definite pattern of few stream valleys at first, then more thorough dissection, followed in the end (provided no earth movement upset conditions) by reduction of divides to a gently sloping surface called a peneplain. A relatively humid climate was assumed in order to carry out this ideal progression. On the other hand Penck suggested that in some cases uplift was much slower than erosion so that the steps outlined above need not follow. Development of these concepts by their proponents was mainly philosophical rather than observational. This is particularly true in respect to the endpoint of erosion. Although many examples can be discovered, for instance in the Coastal Plain and the eroded drift plains, of the progressive development of valley systems with concurrent reduction in surviving areas of the original topography until none survives, no existing examples of peneplains of recent formation have ever been discovered. All that could be pointed out to confirm the validity of the completion of the cycle can be classed as (a) worn-down areas which are inferred to have been uplifted and eroded since peneplaination, (b) buried peneplains now exhumed in part, and (c) flat areas with rock not far below the surface found in semi-arid or seasonal rainfall areas. Some enthusiastic students actually described as young peneplains areas of lake or stream deposits where bed rock lies at considerable depths. It is probable that such errors are in large part explicable by the emphasis placed by some geologists on widespread planation by streams. Although such lateral erosion would certainly be an important factor in completion of a peneplain it is more characteristic of a pediment. The climatic conditions under which many ancient surfaces like the pre-Cambrian peneplain of North America were formed is wholly unknown. Examples of topography in Africa strongly suggest that seasonal rainfall on both sides of the equator may promote pediment formation just as well as does sporadic rainfall on mountains in the Great Basin. Certainly the numerous examples of monadnocks with very steep sides (inselberge or island mountains) appear to suggest formation by lateral erosion of streams whose level was fixed by their own deposits. Only such a process could possibly explain the steepslopes. Further discrimination of peneplains from pediments follows later in this section, as well as a discussion of the identification of remnants of erosion surfaces of different ages in the same district.

Effect of solution on peneplaination. Most discussions of peneplaination have ignored the effects of solution. On water-soluble rocks, such as limestone, this process can work over the entire exposed area at once. It is even effective to some extent under a cover of permeable rock. The result is that, unless disturbed by crustal movement or change of climate, a nearly level surface is formed.

Interruptions of the cycle of erosion. Many students have justly expressed doubt that the theoretical cycle of humid erosion could ever be brought to completion. The sedimentary record does not suggest that the lands ever remained in the same relation to sea level for more than a fraction of the probable time which should be required. Although we cannot now express in years the time which would be required for perfect peneplaination of a mountain range, we are able to measure approximately the duration of the several geologic periods by means of the study of atomic disintegration. The consensus of opinion is that the entire time since the beginning of the Cambrian is not over 500 million years. If we think of the rate of erosion slowing down markedly toward the end of a cycle, when the distance through which rainfall descended to the sea was small, it is not difficult to conclude that known geologic time is too short to permit the completion of so many cycles as have been postulated by some. If the cycles were not as complete as has been believed, then the possible number would be much increased but it is still evident that it cannot be large. For instance, it is thought that the Lower Cretaceous began not more than 120 million years ago. Yet many have thought that prior to the Upper Cretaceous of 95 million years ago, there was not only long deposition of limestone, followed by earth movements, and they by a reasonably perfect peneplain over most of the Atlantic seaboard if not all of eastern North America! The eroded rocks included not only the Lower Cretaceous limestones but also large areas of crystallines. Are we not asking too much of ordinary erosion? It is true that in parts of the Arbuckle Mountains of Oklahoma a subdued surface was eroded across tilted and folded sediments tens of thousands of feet thick, and then buried again within a fraction of the Pennsylvanian period. However, this seems to have been a local and not a regional planation, possibly a pediment formed on rocks not yet completely lithified. In spite of this somewhat startling evidence it does appear likely that most cycles of erosion could never have reached the theoretical endpoint without interruption by earth movement or change of sea level. Uplifts were both those without distortion of the sea bottom and those associated with folding. Besides these diastrophic movements we must reckon with obstruction of drainage by vulcanism, glaciation, and landsliding as well as with changes in climate. When an uplift occurs it follows that erosion with the changed baselevel will largely obliterate all record of the partial cycle before.

Terraces. When a river has stabilized its slope, or reached *grade* as many term it, a normal feature is the formation of a wide valley floored with debris, which, although in transit down the stream, must perforce be left stranded during the intervals between floods. Such a deposit floors a *floodplain* and the thickness of material above the bedrock cannot be greater than the usual flood-time depth of the channel. When the baselevel is changed or there is a change in climate either (a) the floodplain is built up with stream debris because the river cannot forward its load any longer, or (b) increase of slope accelerates the rate of erosion entrenching the stream and leaving the former floodplain as a *terrace*. The method of uplift might be with or without warping or local irregularity or might be a regional uniform tilt. As mentioned above, the last would accelerate the velocity of the stream all along its course at once causing it to form a new longitudinal profile. If an uplift without tilting, or the equivalent a change of the amount of water in the oceans, then the new profile must grow inland gradually. Irregular uplift would be a combination of the above conditions. The inland limit of the new profile has been termed a *nickpoint* and much attention has been directed to the finding of points of change of profile in a stream. Logarithmic plotting will show at once where these occur but it is not evident which are related to differences of geology and which to the start of a new cycle of erosion. The example on the middle Wisconsin River cited above is clearly due to erosion of the rock barrier at the Dells below. It is very indefinite in the detailed profile, for the change in slope of a stream is generally very gradual.

Aggradation of a valley with debris is the converse process of terrace formation. It may be due to: (a) building of a delta at the mouth of a stream, (b) climatic change, (c) obstruction of a portion of a valley by earth movement or deposition, or (d) glaciation which supplied a tremendous amount of loose material to the stream. In and near to the glaciated regions the valleys which carried glacial meltwaters were filled up to great depths with *outwash* (glacial sand and gravel). Logarithmic plotting of the profiles of a number of outwash deposits in Wisconsin yielded the equation $f : h^{7/10}$ where fall is in feet and horizontal distance in miles. The constant of proportionality varies from 9 to 25 inversely to the discharge of the stream. It is evident that with such a concave profile melting back of the ice front automatically changes the slope of the outwash streams at any given point. Readjustment then forms a terrace above the new level. Many streams carried the overflow of large lakes during ice retreat and the increased discharge caused erosion leaving terraces. The grade of many outwash deposits was changed by the melting of included ice masses which had been buried in the deposits. This change formed many terraces. During the erosion of many outwash deposits the streams were in places superimposed across rock ridges. Until eroded away these caused wide valleys above which were later entrenched into terraces. Other outwash terraces were banked against ice which on melting left them high above later streams. Deposition of outwash in the routes of glacial drainage was so rapid that it blocked up the tributary streams which headed in territory not glaciated at that time. The lower parts of these valleys at first held lakes but terracing of the outwash has in almost all places drained these. Deltas were deposited in these temporary lakes and streams which built up their beds to meet the new conditions were locally superimposed on rock spurs. Trout Falls, near Camp McCoy is of this origin. Terraces due to climatic change are perhaps the least well understood. In general, aridity should lead to excessive slopewash which would bring more material to the streams than they could carry away until the slope was increased throughout their length. European geologists think of this taking place in regions near to the continental glaciers because cold decreased vegetation. Return of more precipitation or higher temperature would increase vegetation, check slope erosion, and cause the enlarged streams to seek a new profile. In this case terraces would result. Opinion has varied with different geologists as to whether the numerous terraces and pediment levels in the southwest part of this country were due to change in climate or to uplift. It is probable that study of the profiles by methods here outlined will eventually solve this problem. In regions of folded rocks it seems reasonable to suggest that many terraces are due to stream entrenchment following upon the main stream cutting through a resistant formation.

Terrace topography. Since terraces are remnants of former higher filling in a valley, or a former wide valley adapted to a different condition of erosion, their borders represent the edge of a new lower floodplain. Where the eroding stream meandered it cut loops into the bank which are known as meander gears. Between these loops the spurs are sharp meander cusps. If, however, the stream was not meandering, or the process of erosion continued for a very long time, such cusps are absent. Terraces formed while the outwash in which they were eroded still contained many ice fragments are now filled with kettle holes and are hard to distinguish from true ice contact terraces where one side rested against stagnant glacial ice. Terraces formed by uplift or change in stream volume normally occur at corresponding elevations on both sides of the valley. Such are called paired terraces. Terraces due simply to lateral erosion of a shifting stream during down cutting are unpaired. Terraces which survived because the valley filling rested on bed rock are often termed rock defended. Terrace surfaces normally show old stream beds which can be distinguished in aerial photographs long after they carried any water. Both braided and meandering patterns may easily be discerned because of the differences in soil in the lower areas of the stream beds.

Correlation of terraces. Correct correlation of paired terraces is difficult. The surfaces were never smooth and since abandonment have been extensively altered by deposits from slope wash and wind work. The best way to match observations at different points is to construct a profile down the center line of the valley. On this correlation is rarely difficult. A further check is logarithmic platting which discloses any miscorrelations at once.

Incised meanders or meandering valleys. It has already been mentioned that some valleys have meandering courses. Two distinct explanations have been advanced to explain this fact. First, it has been suggested that during downcutting the lateral component of erosion caused what were originally minor curves to grow into large meanders. This process would be best developed where the valley walls were not as resistant to erosion as was the bed of the stream, which during low water may have been protected by the bed load dropped after the last flood. Meanders of this type were named ingrown by Rich. Second, a stream which was meandering on a floodplain (not necessarily on a peneplain, as many have supposed) might be uplifted and erosion reinstated. If bed rock was near the surface, and not at considerable depth as on Wisconsin River, the meanders would not be destroyed but would become fixed between rock walls. If these rock valley sides were sufficiently resistant that lateral growth and downstream sweep were alike retarded then the meanders would be eroded into the rock without much change of form. This type was termed intrenched by Rich. As a matter of fact, both types are often found on the same stream, if we can believe topographic maps. The insides of the bends are the criterion by which they may be distinguished. The meanders which increased in size have slipoff slopes commonly veneered with gravel, whereas the other type have practically the same slope on both sides. Intrenched meanders are abundant in the Colorado Plateau where resistant formations of rock lay not far below the ancient floodplain. By restoration of the geology the position of the level prior to uplift may be made with confidence. Such meanders definitely prove uplift of a region and the ingrown type does not, although its possibility is not denied by their evidence. Meandering valleys are almost the sole evidence of change in baselevel of some regions like the Driftless Area.

Stream patterns on floodplains. In most stream valleys there is a floodplain which is occupied only at the highest levels and is in distinct contrast with the normal low water channel. In many floodplains the border of the low water channel is higher than the area behind next to the valley wall. This feature is known as a natural levee and the low area behind is called the back swamp. The pattern of the main stream may be either braided or meandering. Braided patterns where the stream branches and reunites repeatedly are best developed where rapid deposition is taking place. Meandering streams may occur either on floodplains which are being built up or on those that are being eroded. Natural levees due to flood overflow and checking of velocity among the trees of the shore are best developed where the floodplain is being built up, for instance above a delta. Braided streams are universal on the upper part of outwash plains while still forming, in the beds of sandy rivers at low water, below breaks made by floods (crevasses) through natural levees and on deltas and alluvial fans. Within the back swamp the streams have no definite pattern but form an irregular network through the vegetation. Various explanations have been offered for the difference between meandering and braiding. The former is more characteristic of streams which have flowed for some time and hence have organized a definite channel with few islands or towheads. Braiding is apparently an indication of immaturity and rapid deposition. A peculiar feature is the ending of meandering on Mississippi River not far below New Orleans and well below the first distributaries of the delta. In this part of the river differences in water level are not great. Possibly lateral cutting is hindered by the firm clay of the natural levees. Certainly changes in route to the sea have not occurred in historic time. Another feature of floodplains with pronounced natural levees is tributary streams of the Yazoo type where access to the main river is prevented down to a locality where undercutting of the bluff is taking place. Examples of what must certainly have once been this type of stream junction before later erosion appear in the Tennessee and Cumberland near the Ohio, as well as where the Illinois reaches the Mississippi.

Deltas. Any stream which discharges into standing water is obliged to deposit its bed load at once. The suspended load may travel far before settling. Fresh water, even when muddy, is lighter than salt water and hence floats. Meltwater fresh from a glacier is less dense than is the underlying water because it is colder. The coarse material dropped at once slides down into foreset beds. Sand and gravel appear to come to rest at a slope of about 25 degrees. This abrupt descent from the nearly level top toward the lake or sea is an excellent diagnostic feature by which ancient deltas now far above the water may be discriminated. The mouth of one of the distributary streams on a delta is commonly shallow because of deposition. The passes of the Mississippi are kept open for navigation by artificial narrowing with jetties. This deepening of the bottom has caused eruptions of mud called mud-lumps which are mechanically similar to the base-failure slides of the Panama Canal.

Alluvial fans. The alluvial fan is the land equivalent of a delta. Change in original slope of the land at the foot of mountains or hills is a common cause of deposition of the bed load which was acquired higher up the streams. Although typically developed in semi-arid districts, alluvial fans can be made in any climate. Many can be so observed filling up kettles in sandy glaciated regions. Streams on fans are braided. Variation in discharge is rapid and great in most regions. Evaporation and seepage into the porous material are often regarded as important factors in deposition, but their quantitative importance is yet to be demonstrated. In California extensive water-spreading works are necessary to increase the soak-in and conserve water which would otherwise reach the sea. Restraint of streams from changing course to places less filled up is difficult but appears to have accomplished in some places by narrowing the channel to one whose competence and capacity are greater than the original braided course.

Profile of alluvial fans. Platting of several alluvial fan slopes east of Los Angeles, California yielded an expression in which f : h $^{.78}$. This is not far different from the equation of glacial outwash plains in Wisconsin. However, Krumbein platted the slope of one pronounced alluvial fan in the same region and derived the equation: elevation =,2280 e $^{-.12\ x}$ where elevations are in feet, distances in miles and e is 2.718. Replatting of data by the writer failed to confirm the general applicability of this equation, although it may be correct where the fan is well rounded and the water is constantly spreading out over a larger and larger area. The lower slopes of some volcanic mountains appear to show indices of .35 to .4 where the material is fine and the water from the mountain is spreading out. Alluvial fans are readily confused with rock-floored pediments into which they pass upstream.

Natural Bridges. The formation of natural bridges by either cavern collapse or subterranean solution channels through a meander spur has already been discussed. In insoluble rocks cutoffs have taken place both due to leakage along a joint or by lateral erosion of a spur. Some natural bridges can only be classed as freaks of weathering like towers which have not yet fallen into the talus beneath.

Drainage modifications. It has already been mentioned that as time goes on in the cycle of erosion streams come to be more closely adjusted to areas where resistance to erosion is least. There is also a progressive relocation of streams in order to secure the shortest, and therefore the steepest, route to the point of discharge. This process requires that certain areas have their drainage outlet changed. The method of change has often been called stream capture or stream piracy. Although indubitable examples of this process have been distinguished few have ever discussed the exact mechanism by which the final capture is effected. It is easy to visualize capture by lateral erosion through a narrow divide changing the point at which a tributary enters the main stream. The Greybull River, Wyoming, is supposed to have been captured by a smaller river with a lower grade. But when we recognize the validity of the belt of no erosion along divides it is hard to grasp just how the headwaters of one valley could ever wear back into another. One would think that, unless conditions for erosion differ radically on the two sides of the divide, it would prove impossible for a small intermittent stream to ever reach the bed of a large and well established river. Certainly capture of a stream on the other side of a ridge due to a tilted resistant formation appears wellnigh impossible unless aided by subterranean solution or shattered rock along a fault. In every case of recession of the head of a ravine it is obvious that there must be enough gathering ground to furnish water for erosion. If the underlying material on the divide is unconsolidated, or is pervious to water, however, it is easy to see that underground leakage would feed the lower valley long before the actual break-through. Landsliding would also aid in this process, or in some cases the divide might be so low that a flood in the stream above would overflow to the lower course. In early days geologists freely invoked warping of the land as an aid to capture but with no confirmatory evidence. Many supposed instances of capture where no abandoned course of the captured stream could be discovered are of doubtful validity. Peculiar-looking stream courses may readily be due to original irregularities of the surface which directed consequent streams. Similarity of water snails in now separate streams is of doubtful validity because migration may have been with aid of birds.

Superposition. Many stream courses which at first sight appear very peculiar in that they disregard geologic controls are evidently due to initiation of the route on top of unconformable deposits now removed by erosion. This process is known as superposition and streams of this origin may be termed superimposed. Excellent examples can be found of superposition on the Cambrian cover onto the pre-Cambrian, or by the glacial drift onto a bed rock surface. The principle of superposition has now been invoked much more widely than it once was for many of the older geologists seem to have been entirely too conservative in imagining the former extent of now-vanished formations. For instance the course of the Mississippi River on the flanks of both the Wisconsin and Ozark uplifts is much more likely due to superposition than to some more involved process.

Antecedence. Streams which held or nearly held their courses against deformation of the crust beneath them are called antecedent. This process was much invoked in early days to account for structural peculiarities of certain stream courses. Although not by any means impossible, it is clear than in many localities superposition is more probable. In fact many of the older examples, such as the Grand Canyon of the Colorado, are now definitely known to be due to superposition. On the Columbia Plateau, however, the Great Bend of the Columbia appears to be a consequent course along the edge of the basalt flows assumed before the center of the basin sank. In other places the streams of that region cross anticlinal ridges whose rise could have ponded them only temporarily.

Discrimination of peneplains from pediments. Peneplains and pediments have in common the fact that they are worn-down areas of low relief which must at one time in their history have been of much greater relief. It seems unfortunate that the idea of a plain as the endpoint of erosion in humid climate has so widely spread. This assumption demands that either (a) resistance to erosion is negligible, or (b) that geologic time is infinitely long. Because both assumptions are extremely improbable because of the known facts, many students of geomorphology have desired a change in nomenclature. Douglas Johnson suggested change of the name of the final erosional form to peneplane but this was also unfortunate for in geometry the word plane is definitely defined in a way which makes it inapplicable for use for a land form. The word surface is non-commital and we might well, were it not too late to change, substitute the term old surface or endpoint surface of humid erosion.

But everyone still speaks of "sunset" and "sunrise" although they are obvious misnomers! Another factor, often overlooked, is inconstancy of climate on the earth. When the continents were largely submerged and there were no polar ice caps the climatic belts must surely have been far different than they now are. Rainfall may have been mainly confined to the equatorial belt and to mountains. Deserts may have been far more extensive on the low lands than they now are. Certainly we should not project the existing climate of localities in middle latitudes too far into the past. For this reason it is well to review the known facts to discriminate between peneplains as ordinarily defined and pediments of presumably semi-arid climates and perhaps regions of seasonal rainfall.

Comparison of peneplains and pediments. Definition: a peneplain is the end-point of undisturbed humid climate erosion; a pediment is a sloping area with bed rock near the surface which occurs at the foot of a mountain range. Kind of rock: a peneplain should be formed on almost any kind of rock; a pediment is most rapidly formed where the bed rock breaks down into particles which are readily washed by water, for example coarse-grained granite or sandstone. Climate: although not stressed in original ideas it is clear that a peneplain of the type described by the older writers must be formed in a humid climate; a pediment, judging from existing examples, must be formed in a region where streams do not forward their load to the sea but form alluvial fans. Weathering: the bed rock under a peneplain should display a considerable amount of chemical alteration, although not to great depths as has been incorrectly assumed, because there is not enough head to cause deep underground circulation; a pediment should display bed rock altered mainly by mechanical processes. Topography: peneplains should have very gentle slopes leading down to the sea level of their time of formation and display complete adjustment of drainage to underground structure which determines disposition of the weaker rocks; pediments should have a regional slope which does not everywhere lead to sea level. Extent: peneplains must necessarily be of regional extent, merging gradually into higher land on which long-continued erosion has also left its mark; pediments may be local, passing on one hand to areas where stream deposits accumulate, on the other to the talus slopes of much higher land, and may occur in a stairway of successive levels. Subsequent tilting: because the original surface of a true peneplain must of necessity have been very gentle it follows that any planed-down area with a slope of more than a foot or two to the mile, interpreted as a peneplain, must have undergone subsequent tilting; but an inclined pediment is expectable for such areas have slopes up to many hundred feet per mile when formed. Covering deposits: streams on a peneplain might have wide floodplains but the idea that they necessarily aggrade the old surface on account of excess of disintegrated material is based on erroneous premises as to weathering. The deposits of streams would necessarily be very fine for it takes a velocity of 13 cm/sec to transport with a maximum diameter of 1 mm and about 45 cm/sec to carry pebbles of 10mm diameter, slopes which for small streams with hydraulic radius of one foot demand 0.7 ft/m and 8.2 ft/m respectively, larger streams requiring less slope. Pediments, with slopes in excess of the higher figure quoted, would have a thin coating of coarse gravel. It is probable that the confusion of surfaces of deposition with peneplains which has been general in the past is due to misunderstanding on this point. Age relations: if remains of peneplains could be found adjacent to one another and at different levels the higher must be the older; but with pediments progressive burial of a mountain range with its own debris would assuredly reverse this order of surfaces and the highest might easily, although not necessarily, be the youngest.

Survival of remnants of more than one erosion cycle. As soon as the idea of the cycle of erosion was announced enthusiastic geologists sought to apply the new tool to the interpretation of the geologic history of regions where there is a long gap in the sedimentary record. The result of this was the description of a multitude of erosion cycles, not only resulting in regional peneplaination but also surfaces which left immediately adjacent almost intact remnants of older cycles. Up to 14 such incomplete cycles were reported in one area on the basis of work with topographic maps only. The excellent preservation of some of the old surfaces, as well as their extraordinary number, led inevitably to skepticism and reexamination of the evidence. The following discussion is intended to evaluate this evidence with a view to finding whether or not alternative views were overlooked.

Criteria of past peneplaination. The best proof of the existence of ancient peneplained areas is the discovery of actual remnants. The strength of this evidence depends directly upon the size and number of remnants which could not have possibly been formed under present conditions. In this connection we must recognize that convex hilltops are not reliable; rather than remnants of an old subdued erosion surface they are more likely creep slopes preserved by the belt of no erosion. Effect of resistant rocks in protecting underlying soft material or of impervious rocks in limiting solution of limestone must also be considered in evaluating these areas of gently sloping topography. In many areas, erosion is conditioned so largely by variations of rock that cross sections drawn without geology are entirely meaningless. Regional or local truncation of tilted or folded formations has often been appealed to as conclusive evidence of peneplaination. Survival of shale within minor synclines of firm sandstone on some ridges of the Appalachaians has often been noted. In every case we must appraise the control by rock character as well as by the character of the mantle rock developed from the underlying materials. In gently inclined formations we should rather ask how it would be possible for one area to be much higher than another regardless of the structure. Would it be possible for part of a cuesta to be much higher than an adjacent portion? If it were it would surely be eroded more rapidly than the lower part. When we consider the elevation of the crest of a ridge on a steeply inclined resistant formation we should ask how it would be possible for one part of the narrow crest to be much higher than an adjacent section. Were there such local high points they should soon be lowered. For a long time the even skyline has been given as proof of regional peneplaination followed by dissection. This evenness was deduced from eye observation and not from surveys. "Distance lends enchantment to the view" is nowhere better exemplified than in this connection. Vertical differences are everywhere so small compared to horizontal distances that a distance of not many miles considerable irregularity in the skyline is invisible unless some local slopes are unusually abrupt. Eye inspection of nearby ground can also be misleading for a slope of over 100 feet per mile cannot be discriminated when there are no level or vertical objects nearby with which to make comparison. Then too, in horizontal views of the skyline distant crests and divides blend together and we tend to forget the deep valleys which lie between them. Viewed on a really accurate map, or in vertical aerial photographs, the true facts are apparent and we see such narrow divides that they could not possibly be remnants of an ancient surface. As for maps, we must at once realize that the topographers, even in recent times, were not allowed enough time to be able to visit ridge tops or even to send their rodmen to climb them. Surveying with a stadia rod is at best very tedious in timbered country and much of the contouring is by sketching. Reference to the older instruction books of the U.S. Geological Survey shows that topographers were not encouraged to climb high hills everywhere but to try to sketch a very large area from lower elevations. The effect of perspective leads to serious errors such as omission of deep valleys. A careless sketcher who works from below often records a flat-topped ridge where in fact the divide is very narrow. It is entirely unsafe to base conclusions as to old erosion surfaces on maps alone unless it is evident that the surveyor actually visited the locality. Long ago Shaler pointed out that in regions of homogeneous rocks the divides are simply the meeting point of slopes which rise from adjacent stream valleys which are spaced fairly regularly. The average depth of valleys is determined by the slope needed to carry off the runoff with its load of debris. The average depth of valleys is determined by the slope needed to carry off the runoff with its load of debris. The average slope of the valley sides is determined either by resistance of the mantle rock to erosion by slopewash or by the angle at which creep occurs. Naturally the average slope is the same on the two sides of a ridge provided vegetation is about the same and it follows that divides should be of roughly accordant level. This condition is well shown in the shale areas of the Appalachian Plateau. When it is discovered that the average elevation of divides is determined by the position of a rock formation, which is more resistant to erosion than these below, the importance to be placed on divide elevations is greatly reduced. Occurrence of regions in which the average divide level is greater than in adjacent areas with a different bed rock, the usual interpretation is that two successive peneplains were formed of which the older survived on more resistant rocks while the lower was being eroded on less resistant material. As an alternative we must certainly consider the obvious fact that areas of more resistant rock, such as sandstone or dolomite, should normally have higher divides than would an adjacent area of shale. In regions of disturbed bed rocks with widely differing resistance to erosion the harder formations almost without exception form ridges whose crests appear to the eye to be remnants of an old erosion surface to whose level all formations were once reduced. But when we ascend these ridges their tops are almost without exception discovered to be so narrow that it is absurd to think of them as being preserved intact. Surely every rock which falls must lower them. These are simply the meeting points of talus slopes whose bases on the weaker formations of the valleys start at something the same elevation. Naturally this makes the crests so nearly equal in elevation that to the eye they appear part of a once-level plain. The forces which produce these ridges work uniformly along the entire length of the flanks and careful surveys disclose a close relation of elevation of crests to geologic structure and thickness of the resistant formation. Could we justly expect any other result from long continued erosion? Need we even assume that once there was an evening-up from which the crests of today were inherited by uniform degradation? Then if we look further, there are lower even-crested ridges in the same area which are formed on formations of somewhat inferior resistance. Do we need to think of each one as a relic of a "partial peneplain" formed by lateral erosion upon the slightly weaker rocks or is their presence an expected result of erosion? Some have made much of the fact that the crests of the highest ridges of the Appalachians are lower near to the places where the major streams cross them in water gaps than they are farther away. But is this too not an expectable result since the level of the subsequent tributary valleys to which the slopes descend lowest there? The rough accordance of summit level in many mountain ranges has been questioned long ago by Daly who suggested that it may be due in part to (a) isostacy which limits the height to which mountains of given material can stand, (b) more rapid erosion above timberline, (c) greater glacial erosion of higher crests, and (d) upper limit of metamorphism or of igneous intrusions. In connection with the last the petrographers can offer testimony and in more than one bathyolith, namely those of Idaho and northern Wisconsin, they tell us that the original top was not far above the present surface or crests.

Use of Profiles . Identification of now almost destroyed "surfaces of erosion" by joining together ridge crests, shoulders, and topos of isolated hills as shown on a profile drawn from maps alone, and without geology shown, is at best extremely questionable. How could these possibly be relics of an older topography. Did erosion progress horizontally leaving higher areas intact while lower ones were worn down to baselevel? Why should we search, then, for records of old surfaces on the divides if those divides were able to survive? Perhaps some will now seize on the belt of no erosion as an answer to this contradictory attitude. If so it must be remembered that this belt applies to slopewash only, not to talus formation, landslides, and creep. Horton suggests that some areas where sloping plane surfaces exist on divides may have surviving portions of the surface which was graded by slopewash prior to the formation of the valleys.

Relative speed of channel excavation and reduction of divides. The concept was that stream valleys are widened to produce "local peneplains" and that subsequent erosion caused parallel retreat of the valley sides leaving a broad flat-bottomed valley. This is known as the treppen concept of Penck in Germany and Meyerhoff in this country. These authors presupposed without any confirmatory evidence that the deepening of valleys is very slow in comparison to widening by lateral erosion. Just how this agrees with the known fact that the debris from slope retreat must be carried down the valley and that its slope must be shaped to do this was not stated. Furthermore, it must be realized that parallel retreat of slopes is possible only when the resulting debris is removed from their bases. This is possible through (a) lateral erosion by streams or (b) the formation of pediment-like slopes below talus slopes. It has not yet been proved that either process is adequate to cause this type of slope retreat in humid regions. A possible condition for extensive valley widening is a rock, such as a coarse-grained granite, which weathers rapidly in a humid climate but is resistant to mechanical erosion. Examination of maps fails, however, to demonstrate that there is a gradation from a peneplain near the stream mouths to progressively less and less eroded topography upstream. Conclusions that such a process does occur in nature are in part due to confusion of surfaces of aggradation with peneplains and in part to the example of the Piedmont where the abruptness of the Blue Ridge escarpment tells definitely of some type of structural control. This escarpment is really very youthful for stream captures along it tell of an unstable condition of the divide.

Effect of solubility. An exception to the conditions described above for the formation of peneplains occurs when the bed rock is limestone or other soluble material. Then the processes of weathering may readily serve to destroy the divides and bring about a true peneplain before the surrounding areas have been completely degraded by mechanical processes. Good examples of limestone peneplains are found in the Appalachians.

Buried and ressurected surfaces. Some of the best examples of ancient peneplains are surfaces which have been buried under sedimentary rocks and then uplifted and ressurected from this cover. Examples are the pre-Cambrian peneplain of northern Wisconsin, Canada, and the Grand Canyon, and the pre-Cretaceous Fall Zone peneplain of the Piedmont. These areas of former mountains were eroded under unknown climatic conditions and it is also possible that the streams and seas which buried them caused considerable alteration during the process. Some of these old surfaces display chemical weathering which has often been described as part of an ancient residual soil which escaped erosion during burial. Such an origin seems most unlikely, although it is true that in Wisconsin the surface of the pre-Cambrian is much disintegrated and oxidized even where deeply buried. It seems much more likely that the chemical alteration is due to circulating waters. Waters which descended through formations of different composition might easily undergo base exchange on meeting with the feldspars of the crystallines. The subject will bear considerably more investigation. Some of the once-buried surfaces may not be peneplains or pediments but may be planes of marine erosion as will be considered in the next section.

Summary of evidences. It seems clear to the writer that a very large part of the evidence which has been presented to demonstrate that relics of several erosion cycles are present in limited areas is either questionable or invalid. Survival of such remnants is to be expected in the case of pediments but not with peneplains. Particularly objectionable as evidence are conclusions based on map profiles without showing geology.

Origins of water and wind gaps. Subsequent streams in areas of disturbed sedimentary rocks are readily understood but the places where the main streams cross through ridges due to the outcrop of resistant rock formations are less easy to account for. Four distinct explanations have been advanced by different authors for the water gaps of the Appalachians: (a) rearrangement of streams on a perfect peneplain, (b) antecedence of streams to the folding, (c) stream capture, and (d) superposition of streams on a now-vanished unconformable cover.

Development on a peneplain. The idea that the transverse streams of the Appalachians were inherited from a perfect or super-peneplain on which they (a) "lost their way" or (b) were superimposed on a thin cover of alluvial deposits or (c) were diverted by tilting was once very popular. As a matter of fact it seems theoretically impossible for such a perfect peneplain to exist. The suggestion of a cover of mantle rock changes the hypothesis to the last one, superposition. At the present time it is recognized that if a peneplain were completed on rocks of diverse hardness the adjustment of the streams would increase in perfection during its formation and that they could never disregard the underlying materials. This hypothesis is now obsolete.

Antecedent streams. It has been suggested that some, at least, of the Appalachian streams are still in the approximate locations in which they were when the folding of the rocks occured at the end of the Permian period. The sediments of the Appalachian geosyncline are thought to have been derived from the now almost-vanished content of Appalachia. The original streams should then have flowed northwest across the rising folds. For this reason the suggestion of antecedance could apply only to such rivers as the New, the French Broad, as well as other headwaters of the Tennessee. It is inapplicable to most of the streams which cross the hard rock ridges.

Stream capture. It is clear that stream capture has occurred in relatively recent time along the Blue Ridge escarpment and that it is imminent in several localities. The northern streams like the Potomac and Susquehanna, which flow direct to the Atlantic, certainly have the advantage of a steeper slope than has the Tennessee or Kanawha. But when we consider the difficulty of a small stream working back through a thick formation of resistant rock to reach a larger river on soft rock on the other side the process appears impossible. Conditions are different than in the southern Blue Ridge where the rocks are reasonably uniform. The hypothesis of capture is workable only where there is cross fault which shattered the rock of the ridge. As most water gaps do not display any offset of the ridge it would be necessary to assume in every case a fault in which movement was parallel to the dip. If there were faulting at gaps, landsliding and underground leakage of water might cause capture for then there would be no necessity for a watershed to supply a stream which would cut back through the ridge. Geologists differ greatly in conclusions as to the field evidence of faults in water gaps. Some declare that they are almost universal and others deny that there are more than a very few. Certainly many gaps in successive ridges do not line up as they should. It seems likely that since the gaps afford the best and most readily accessible exposures they have been more visited than other parts of the mountains and yet few have ever been mapped in detail as is done in oil-producing regions. The theory of capture appears rather unlikely as a general cause of gaps in the Appalachians although it may be workable in some other regions.

Superposition. It seems strange that the theory of superposition was so long neglected in the Appalachians. It has been definitely proved in the Colorado Plateau and for the localities where the Colorado crosses some of the Basin Ranges, although it will not account for the major rivers of the Columbia Plateau. The Coastal Plain sediments are not far distant from the Appalachians and seem the logical answer to their water-gap problem. However, it must be realized that it was a long time since the cover was present and that it probably rested on a surface of unknown but relatively low relief which lay well above the present ridge tops. Since the erosion of the superimposing deposits there has been time enough to cause extensive formation of subsequent valleys along the strike of non-resistant formations. First advocated by Johnson, there have been many objections to its general application. It has been pointed out that many gaps are located where the hard formations are unusually thin or make abrupt bends. However, it seems by all means the best suggestion, provided the long time since erosion of the soft covering formations is realized. The underlying ridges, which are about 100 feet high where they disappear under the Coastal Plain of Alabama, may have exerted some influence on the form of the original surface just as differential settling of glacial drift is thought to have caused reexcavation of some valleys which had been once completely filled. It must also be realized that progressive erosion into a mountain mass must certainly uncover vertical differences in structure. This is particularly true where thrust faults are present. Streams adjusted at one level are out of harmony with the formation when they have cut deeper.

Wind gaps. Wind gaps are similar to water gaps but no longer have any stream in them. It has been suggested that some wind gaps were actually due to the meeting of the heads of ravines on opposite sides of the ridge. In answer to this, such ravines do not have sufficient watershed on a narrow ridge, although they might occur where aided by fractured rock along a cross fault. Many have suggested that with gap elevations, some of which have been altered by accumulation, of talus since abandonement, record former erosion levels. The idea was that uplift of the old partial peneplain caused rapid diversion of streams leaving the gaps. However, study in Pennsylvania does not support this theory very well. During the process of erosion the number of water gaps has steadily declined in favor of windgaps thus giving no support to their original origin by stream capture.

General Summary. The subject of the work of running water is very complex. Agreement has not been reached on the mathematical relations of erosion and transportation to energy of the water. On the whole, the line of approach used by Little seems to offer the best possibilities for the computation of profiles of uniform force, both for streams and for slope wash. Use of a different formula for relation of velocity to depth and slope is advisable in the case of the latter. The reasons for variation of channel width and width of the meander belt are explained from the standpoint of hydraulics and appear satisfactory. Initiation of valleys may be either consequent on original irregularities of a new surface or may follow on primary slopewash grading as outlined by Horton. This author's discoveries of a mathematical relation of relative numbers of streams of different types (orders) appears to support his contentions. The belt along divides in which not enough water is gathered to permit erosion by slopewash cannot be neglected, nor can be the factors which alter it in time. Degradation of divides is ascribed more to creep than to slopewash because of this belt of no erosion which survives after stabilization of slopes to the point where resistance to erosion equals available force of water. The endpoint of erosion under humid climate is certainly not the peneplain of classic literature if there is vegetation to cause resistance. The formation of pediments as distinguished from peneplains is discussed with the conclusion that they are best developed in semi-arid climate and are formed chiefly by lateral erosion. They are wash-slopes and contrast sharply with the talus slopes of adjacent higher areas. Pediments can be found forming a "stairway" of successive levels but similar relations are impossible with peneplains. However, the climatic conditions under which many low-relief surfaces originated is unknown. It is pointed out that many criteria for the discrimination of ancient dissected peneplains are invalid. Superposition appears to be the best explanation of many water gaps and wind gaps are due to diversion of streams which once crossed ridges of resistant rocks.

WORK OF STANDING WATER

Introduction. The work of standing water is divisible into: (a) mechanical processes, which include both the work of waves and of currents, (b) work of organisms, and (c) the problem of submarine valleys. It is evident that this grouping includes some subjects which are not strictly under the general heading but which appear more suitable for discussion at this place than in any other.

Mechanical processes.

Origin and mechanics of waves. Wind which blows over water is retarded at the bottom by friction just as it is on land. Velocity decreases downward and probably is 0 at and near actual contact with the water. This vertical velocity gradient is in response to the transfer of energy from the moving air to the underlying water. In general, however, it does not itself cause the water to flow as a current. Instead it sets up rotational motion of water particles around horizontal axes which are at right angles to the direction of the wind. When wind first begins to blow the radius of rotation of each particle is small but as time goes on a limit appears to be attained. Each successive rotating particle is slightly out of step with the last so that the final result is a wave of oscillation. The mathematical form of such waves is that traced by a point on a radius of a rolling circle. Of course, no real circle does roll, only the particles and the radius of the hypothetical circle is larger than the actual orbit of water particles at the surface. Even after the wind has stopped blowing the wave progresses. It is then smooth and much simpler than when it is crinckled by a rising wind. At maximum size, waves are roughly half as high in feet as the velocity of the wind in statute miles per hour as ordinarily measured not far above the surface. The velocity with which particles rotate in their orbits determines the speed with which the wave progresses. Since particles revolve in a circle either the vertical component of motion obeys the laws of harmonic motion. If we let the radius of the rolling circle be R feet and the length of a wave from crest to crest be L feet then $L = 2 \text{ pi } R$. The acceleration is that of gravity, g. Applying the formulas for harmonic motion and solving for velocity in feet / second: $V = (gR)^{\frac{1}{2}}$. Substituting the value of R in terms of L, $V = (gL/2 \text{ pi})^{\frac{1}{2}}$ or substituting numerical values for the constants, $V = (5.123\ L)^{\frac{1}{2}}$. Solving this for length of a period, T seconds, then $L = 5.123\ T^2$. The relation of the height of a wave above the trough, h, to wave length, L, is not fixed but appears to vary from 19 to 39 times. The relationship of h to fetch or distance that the wind blows over open water is reported empirically as $h = 1.3 \text{ fetch}^{\frac{1}{2}}$ where the latter is measured in statute miles. (Mariners use nautical miles or knots each one of which is about 1.15 land or statute miles; they also measure depths in fathoms of 6 feet.) One reason for irregular results in measuring wave heights is the fact that a violent wind blows off the wave crests in whitecaps. The smooth waves which last after the wind, or extend outside of the area where it blows, are often called ground swell. In the open sea the distance through which winds blow sets a limit to height of waves of about 50 feet.

Energy of waves. Since wave motion in open deep water is a combination of rotational motion and vertical motion waves possess two types of energy. Mathematical analysis demonstrates that the two kinds are equal in amount. Only the energy of rise and fall is carried forward with the progresses of each wave but when a wave dashes onto a shore and the water is brought to a standstill both types must be expended. In fact, in shallow water waves become translational motion. Engineers have tried several forms of dynamoters with which to measure either impact or both impact and pressure of waves where they strike the shore. Results of these experiments are generally given in pounds per foot^2. Now the total theoretical energy of a wave in foot pounds (other dimensions in feet) is s shown by the formula Energy = $WLh^2/8\ (1 - pi^2\ h^2/2L^2)$ or substituting 64 for W. : weight of a cubic foot of salt water, and for pi, this becomes $E = 8\ Lh^2\ (1 - (1 - 4.935h^2 / L^2)$. For L=200, h=12 this is said to show an impact of 2436 pounds/ft^2. Actual records are of this order of magnitude.

Depth of wave action. The circular orbits of water particles are shown by both mathematical analysis and actual observation to decrease in radius very rapidly with depth. At a depth equal to the wave length the orbit is reduced to 1/512th. Thus one of 10 feet radius at the surface in a 400 foot wave would be only .2 inch radius at a depth of 500 feet. It is evident that although there is no theoretical limit to wave action its practical importance decreases rapidly with depth, that is its competence to disturb the bottom. For this reason pendulum observations are possible in a submarine at comparatively modest depths. The term wave base has been applied to the effective maximum depth at which waves can disturb a sand bottom.

Waves in shallow water. As waves of oscillation reach shallow water the circular orbits are believed to be changed to ellipses with the longer axes parallel to the bottom. Certain it is that the top of a wave moves forward bodily in an entirely different way that it does in deep water. The wave is retarded at the bottom and the top breaks into a confused mass of foam which rushes up a gently sloping beach until it comes to a standstill. No definite mathematical relationship has been discovered which shows the depth at which waves break. This is probably due to the fact that the bottom water is moving either as undertow, due to return of water from the beach, or as a current induced by the tides. In shallow water waves undergo refraction just as do light waves in passing from one medium to another. This is due to bottom retardation and turns the wave fronts until they are parallel to the shore. Many diagrams have been shown to demonstrate that this process also tends to concentrate waves onto headlands.

Other waves. Waves may also be due to (a) earthquakes, (b) tides, and (c) differences in atmospheric pressure. Of these only the first is important in most places. An earthquake moves a large body of water by impact producing a true wave of translation. Such waves are often called tidal but this is a misnomer. They are also called by the Japanese term tsunami. Some of these very destructive waves are known to have travelled 900 m.p.h. and to have risen over 100 feet onto the land.

Effect of waves on the shore. Waves reaching a shore must apply nearly all of their energy to it. Waves work much more constantly than does running water on land. The ocean is never still; even when smooth as glass there is still a surf from the swells of distant or past winds. In appraising its efficiency, however, it is well to realize that rocks moved by wave action are generally submerged and hence loose the weight of their volume of water. Many spectacular instances of large masses moved by wave action are also to be discounted in that the weight was not lifted but simply shifted against friction. Nevertheless, storm waves have been observed to hurl rocks high into the air and the noise of moving boulders in the surf is impressive. Repeated blows of waves do attain high pressures and may easily be associated with cavitation although this has not been recorded. Impact of stones carried in the water on stationary objects is a most important process of erosion. The result not only of direct impact but also of grinding by material moved in the breakers is to undercut the shore if it is a cliff and the depth of water is great enough to permit waves to reach the shore effectively. The process of undercutting is possible because the loosed debris is carried back by the undertow.

Currents in lakes and seas. By no means all erosion and transportation in lakes and seas is accomplished by waves. Currents of water are caused by (a) tides, (b) winds either directly or through waves, (c) density differences due either to variation in amount of sediment or in salinity, (d) entrance of fresh water, and (e) temperature differences. Of these tidal currents locally attain very high velocity on coasts where the difference of level is large. The great oceanic circulation owes its origin ultimately to the convergence of the trade winds in equatorial regions. Although much attention has been given to currents in sedimentation it is obvious that most of them are either surficial or have such low velocity that they do not disturb the bottom. In shallow water , currents along the shore are often noticed. They may be undertow which is deflected by its relation to incoming breaking waves above. Many text books describe an alongshore current set up by waves to which much importance in transportation is ascribed. It is not clear, however, that such a current is actually able to transport sand and pebbles. It can transport fine sediment but movement of coarser material is almost wholly confined to the zone of breakers.

Erosion by waves. It has already been mentioned that waves apply their energy to erosion in a zone of very limited vertical extent. The result is undermining of a shore. If the material is bed rock this produces sea caves especially in weak, thin bedded layers. Where rocks are jointed, deep coves are excavated, Stacks and islands of rock survive for a time rendering the shore very irregular. Shores of bed rock can almost everywhere be identified on a chart in this way. Where erosion is taking place in mantle rock the cliff is not vertical or overhanging because it soon slides down to the angle of repose. If large boulders are present they accumulate at the water's edge and serve to prevent further erosion. Many abandoned beaches may be recognized only by such boulder lines. A common feature of wave-eroded coasts is hanging valleys whose lower parts were cut away by the waves. The headlands undergo the most erosion because water is generally deeper off them than in bays as well as because of wave reflection toward them.

Wave terraces. Below water level the bottom is cut down to the point that the undertow can just carry off the debris which is not moved along the shore by waves which reach the coast at an angle. This subaqueous feature is called a cut terrace and may be distinguished off many abandoned shore lines. There is commonly a slight building up of the front where material carried out across is slid down into deeper water, but no important deposits of gravel occur in this situation. The cross section of a cut and built terrace is known as the profile of equilibrium and the level of its outer edge was thought to be fixed by effective wave base. It is not clear, however, that this position is long stationary for there is no exactly definable lower limit to wave work and the undertow is aided by gravity.

Subaqueous ridges. A very common feature of sandy bottoms is a succession of several parallel submerged sand ridges. Some of these are many hundreds of feet wide and the depth of water between them is a number of feet more than on the crests. Some call them low and ball. The problem of origin is unsettled. Some think of them as due to breaking of waves which reach the shore at right angles and others ascribe them to erosion of parallel currents which are induced by waves striking the shore at an angle. The second explanation is weak in that it would then be difficult to account for the number of ridges or for the fact that the sand is coarsest on the crests and not in the low places. Where currents do occur it is more likely that they are the result of the ridges and not their cause.

Alongshore transportation. The fact that material which the waves can carry is moved along shore has already been mentioned. On any sandy beach it can easily be seen that waves which come in at an angle carry sand and pebbles diagonally up onto the beach. The undertow runs directly back down the slope so that particles move in a zig-zag course in the direction which is down wind at that time. Johnson objects that the path of a particle is actually a series of inclined parabolas but the distinction is unimportant. Alongshore drift consists in many places of material which is so coarse that it is impossible to think of currents which could transport it.

Depositional changes of the shoreline. Waves tend to even up a shoreline by developing a smooth outline. The material eroded from the headlands is mainly disposed of by movement in the zone of the breakers. Large stones obey the impact law and are carried shoreward by waves which are more powerful than is the returning undertow, which transports the finer particles. Many stones which have been carried back and forth on a beach for a long time show the effect of waves , and are tabular due to shuffling instead of turning end over end as in a current. If a barrier or solid pier is built off a beach experience shows that one side is filled in and the other eroded unless storms come from both sides in equal numbers. The windward side is the one which receives sediments. The same process may be seen at a natural point. Material carried laterally from the end of a point does not form a spit into deep water for that would, where above water at all, be swept away by the next storm from the other side. Instead, the debris is carried along in the breakers to a point where the bay is shallow enough to permit start of deposition. In most places this shoaling is associated with a minor point although this is not necessary. Textbooks ordinarily describe this process of bridging the bays with wave-transported sediments as due to the outward course of the alongshore current but the process outlined above appears more logical. As spits are built out from both sides the bay is eventually enclosed. Where, however, shoal water is present only near the sides of a bay the outer end of a spit is curved back into a hook. This is due primarily to refraction of waves and not to deflection of currents. In many places islands have been joined to the mainland, or to one another, by such beach deposits often called tombolos.

Many hooks are compound showing several successive ends as the entire deposit was built out into deeper and deeper water. In the lee of many islands two spits join leaving a lagoon inside. Lagoons are also formed by the bridging off of bays with continuous bars. Where streams entering the lagoon are large enough tidal difference is enough a pass is kept open to the sea.

Offshore bars or barriers. On low, sandy coasts where the water is very shallow there are sandy ridges some distance offshore. Opinion has varied as to whether or not these were made by lateral growth of spits from distant headlands or were thrown up in place by waves which removed the material from the adjacent bottom. Some of these off Cape Hatteras are hard to account for by lateral growth alone, although in most places this process cannot be eliminated entirely. These sandy barriers are often miscalled reefs which term generally refers to a submerged rock outcrop.

Cuspate points. In many places, such as Capes Hatteras, Fear, and Canaveral, of eastern United States, sand beaches form marked cusps. Various explanations have been offered including eddies of ocean currents. It is more likely that they are barriers built out to submerged shallows from both sides.

Classification of shore lines. Many texts have defined schemes of classification of shore lines into emergent and submergent. Others have suggested primary and secondary as the major basis of separation. As a matter of observation, it is clear that all coasts of the world display phenomena of either subsidence of the land or rise of the ocean level in the form of drowned river valleys. The only exceptions are very recent shores due to organic growth or crustal movement. Shepard terms primary those shores which are due to land forces, erosion, deltas, land vegetation, glacial, and volcanic processes. He calls coasts due to wave erosion and deposition, including that of marine organisms, secondary. Another system of classification would be to discriminate shores on firm materials from those on loose deposits easily moved by waves, placing those due to organisms in a third category.

Cycle of shoreline development. The foregoing section demonstrates why it is futile to search for evidences of a cycle of shoreline development at the present time. All marine shores and inland lake shores are demonstrably young. True, progress has gone much farther in the same number of years where the material of the coast can be easily moved by waves and currents, as on the sandy Atlantic Coastal Plain, than it has on the "stern and rock-bound coast" of New England or the fiorded coast of Norway. The evident goal of shoreline development is as simple an outline as possible. To meet this condition the waves are working to wear back headlands and to fill up bays. When cut off from the sea the bays are filled with detritus from the land and the deposits of organisms which live in the quiet water. It is believed by many that barrier beaches are a temporary feature of the coast line. As evidence of this occasional outcrops of peat in the seaward faces of such bars are cited. It is possible that some of these, at least, may not be lagoon deposits of the present stand of land and sea but antedate the last rise of the waters.

Endpoint of marine erosion. Although we are unable to find good examples of a cycle of marine action at present because of recent shifts in sea level and in levels of inland lakes we may theorize over a possible endpoint of marine erosion. Many have thought that marine erosion is self-limiting in the depth it can cut into the land without change in sea level. This conclusion is based upon the assumption that the depth of the outer edge of the terrace or wave base is fixed. If such be the case the profile of equilibrium would automatically halt shore recession at a definite location inland. But it is far from clear that either such is the case or that relation of sea and land would remain constant long enough for the process to operate. In fact it is not certain that the outer edge of the submarine terrace is really located at wave base. Granted slow lowering of this level by undertow current or a slight rise in sea level in respect to the land and the limit of marine planation is greatly increased. Especially would this be true were the land first brought low by either peneplaination or pedimentation. Most papers written on this subject have been so theoretical or made up of quotations of opinions of others, which have no real meaning, that it is difficult to reach a final opinion. Certain it is that the ceaseless onslaught of the waves should in time have profound results. Moreover, waves exert more force during storms than streams ever can. Waves could plane down even the hardest and most insoluble rock such as quartzite. Most of the famous buried peneplains were buried under marine formations; how much did the oncoming sea alter their surface? Elevations which escaped the work of waves should then be steep-sided because debris was removed from their bases. Some such have been described in India. It has been often suggested that the Piedmont Plateau is a surface of marine planation because there are so few monadnocks. However, the upland extends behind some of the supposed islands and does not connect seaward with any known marine formation. It is more likely a pediment formed during a by-gone time when the climate was more arid. The top of the Baraboo quartzite range, Wisconsin, is a very gently domed plain. At places the edge carries boulder conglomerate. Projection of now-eroded formations appears to demonstrate that it was the beach at the time of deposition of an adjacent dolomite formation. It is, therefore, possible that the waves of the Ordovician sea completed the planation of the quartzite islands which had been monadnocks on the pre-Cambrian peneplain over 1000 feet lower. A somewhat similar surface on quartzite has been proved by drilling at Hartford, Wisconsin.

Pleistocene terrace problem. Many sea coasts display terraces which evidently record former levels of the oceans in respect to the lands. Many of the higher terraces, as on the Pacific Coast, are obviously deformed by subsequent orogencic movements. Some of the lower ones, however, seem to be horizontal. On the Atlantic Coastal Plain Cooke has described seven terraces which he thought to be horizontal and of world-wide extent. Flint has restudied the same area and concluded that there are only three, of which the upper one at 160 feet is limited in extent. Many of the supposed marine terraces he thought to be in fact stream floodplains but the lower terraces at 25 and 90 feet might be horizontal. Most reports of marine terraces fail to discriminate between the actual level of the water and the elevation of the cut and built terrace. Elevation figures for many vary over so wide a range that exact correlation is impossible. Too little attention has been paid to the sediments associated with the terraces. Knowledge of Pacific terraces is entirely too fragmental to permit of correlation. Postglacial uplift of the land is definitely known in and near to glaciated districts making comparisons there entirely futile. It is, therefore, too early to claim that all these terraces are horizontal and that they record changes in amount of water in the oceans related to the withdrawal in ice caps. That such a process took place is certain but just which shorelines record interglacial intervals when this water was returned to the oceans is far from assured. If the higher ones are really eustatic then it would be necessary to assume that either (a) the amount of ice carried over in continental glaciers increased in every successive interglacial interval or (b) the floor of the ocean sank during the Pleistocene lowering the level of the oceans. Stearns records evidence of just such a sinking of the bottom of the southwest Pacific Ocean. But the cause he ascirbes, namely eruption of lavas which then pressed the area down by their weight, does not appear feasible. The lava came from below and could settle no more than to refill the voids it left. It is possible that this is not the true cause but the facts of sinking are well substantiated.

The coral reef problem. The subject of the origin of coral reefs has been a subject of discussion for more than a century. According to a recent summary by Stearns the following facts are now established: (a) reef corals can live only in warm, clear water less than 200 feet deep, (b) Nullipores serve not only to bind coral skeletons together but also make reefs themselves, (c) when the Pleistocene glaciers were large sea level was lower than now (perhaps about 260 feet in the last glaciation) (d) changes in sea level have also occurred because of alteration of the shapes of oceanic basins, (e) many islands of the southwest Pacific are composed of folded continental (sialic) rocks, (f) the truly oceanic or simatic islands show no such rocks, (g) emerged Tertiary coral reefs are known, (h) atolls do not reflect the form of submerged volcanic craters, (i) atolls rest on a basement of non-coral rocks, (j) barrier reefs require a platform to start their growth, (k) growth of corals may attain 90 feet in 1000 years, (l) a reef may be killed by submergence (m) a reef may be killed by emergence, (n) reefs grow mainly on the outside, and (o) lagoons are not the product of submarine solution of calcium carbonate.

Occurrence of coral reefs. Coral reefs occur not only along coasts of other kinds of rocks as fringing and barrier reefs but also in isolated islands which rise from the depths of the ocean. Many of the latter surround a partly or wholly enclosed lagoon and are known as atolls. Two small atolls occur off the end of the Florida Keys but they are much more abundant in the Pacific Ocean. Some of the isolated islands are known to contain a volcanic core. In the Bermudas drilling discovered volcanic rocks although a test over 1100 feet deep on a Pacific island failed to find such.

Theories of coral island formation. Opinions as to the formation of coral reefs and coral islands have varied widely. Some have thought that they grew upward on top of intermittently subsiding foundations of other origin. Others held that they were formed on stationary basements and that the lagoons were made by solution. Others advocated an origin on a rising foundation, others on stationary shelves made by former erosion. Daly first discussed the control of sea level by glaciation , a theory which would make most reefs younger than the last ice age. Recently, Stearns advocated growth on any kind of basement rising or sinking, under conditions of rising sea level from any cause. The general opinion now is that only glacial control of sea level can explain the majority of reefs. The test boring put down on Funafuti was located too close to the outside of the accumulation of coral and beneath about 150 feet of coral passed through only talus outside of an older reef. Under the glacial control theory it has been claimed that volcanic islands were all planed down by waves to a uniform depth on which corals started to grow as the ice melted and slowly returned water to the oceans. This is thought to account for the uniformity of lagoon depths over considerable areas. However, this theory does not exclude other conditions and both emerged and submerged atolls have been described. It seems certain that in an area of recent vulcanism conditions must have varied widely in different island groups. Much of the argument is based upon purely theoretical reasoning based in large part upon nautical charts which do not show the land areas with much detail. It was thought by some that their failure to show many cliffs on spurs of volcanic islands inside reefs militated against the glacial control hypothesis. Since then this has been shown to be an error, and the glacial alteration of sea level is recognized as an important factor.

Work of other organisms. A number of other organisms besides corals and nullipores affect shorelines. In salt water the mangrove tree can grow where wave action is not too violent. In sheltered bays various kinds of grasses and sedges, which are tolerant of a moderate amount of salt, build salt marshes. Salt marshes on coasts where there is much tide are cut up with a complex net of branching channels through which the water runs in, then out, twice in every 24 hours. Shallow, small fresh water lakes are the habitat of a number of plants which in time may fill them up. These appear to thrive best in relatively hard water. Although most aquatic weeds do not project far above the surface some varieties like the common "bullrush" do. Remains of organisms aid in shoaling the water so that the shoreline can advance. A growth of these hinders wave action and the shore behind can then be filled in with organic deposits. A regular succession of different kinds of plants can be made out from those which thrive only in fairly open water to those of old marshes farthest inland in the old lake bed. Many shallow lakes have been entirely filled with vegetal deposits since the glaciated region was first settled by white men. A common feature of marshes is a moat between the vegetal growth of the center and the high land. These are ascribed to death of vegetation, possibly aided by burning, in dry seasons. In wet weather some of these open water areas are large enough for wave action which has made faint boulder lines along the edge. The level of water has no relation to that of adjacent lakes so that the level of these shore features has no bearing on former lake levels.

Submarine valley problem

Introduction. It has been known for a long time that the edge of the continental shelf and the continental slope are indented by deep, narrow depressions in which the water is thousands of feet deeper than it is nearby. Until the advent of echo sounding, however, few of these submarine valleys had been accurately delineated. Sounding in deep water with a wire line entails stopping the ship and a long delay in running out and winding in line. In the meantime drifting occurred and positions given on the chart are now known to have been miles out of place. Echo measurements may be made at full speed when it is easier to keep on course. Locations out of sight of land are now fixed by taut wire measurements, radio-acoustic methods, and radio bearings. Exact knowledge of depths is now important to navigators even in midocean. The result is that charts are now much more detailed than formerly and that a great increase in number of submarine valleys has resulted. Most of these new discoveries were off the coasts of the United States for this type of surveying has not advanced as far in other parts of the world. One drawback to the echo or acoustic method of sounding is that on rough bottom more than one signal is returned and the strongest echo may readily not be from the bottom under the ship but from an adjacent slope. In other words the results are apt to be a generalization.

Submarine contouring. Drawing of contours, or lines of equal depth, beneath water where the bottom cannot be seen is fraught with much change for error. If contours are simply prorated between soundings, as engineers do, the result is only a crude generalization. If drawn with some theory of interpretation in mind the result may be simply "wishful thinking". No map is worth anything which does not show all the available soundings. For the reason stated conclusions based on submarine contours have shown the "personal equation" to a marked extent. Some see in them only underwater faulting or folding, others a few valleys similar to those on the continents, other slid and slimped slopes, others conclude a multitude of small ravine-like parallel valleys like the primary rills eroded by rainwash on an earth surface.

Description of submarine valleys. Almost all submarine valleys have a steeper grade than is common on land. Few end in a delta or fan at the lower termination but instead seem to fade out gradually into indefinable irregularities of the ocean bottom. Few indent the continental shelf very many miles. Definite valleys can be traced down to several thousand but less than 10,000 feet below present sea level. In a few cases a submerged connection to an existing stream of the adjacent continent can be found; most do not have this. In rare instances the submarine valley exists inside an estuary (Congo). The sides of the valleys are known from dredging and submarine photography to be solid rock at many points. Current meter observations and bottom samples show that they are not now the location of any unusual submarine currents. Many valleys branch just like land valleys. All have a V-shaped cross sect section. Some are found on the outside of a narrow cuesta (Georges Bank).

Theories of origin. Theories of origin of submarine valleys can be divided into three major classes: (a) the depressions are not valleys at all but are of tectonic origin, (b) they are due to normal stream erosion when the continents were elevated, and (c) they are due to processes which excavated them below sea level. At date of writing there is absolutely no agreement between geologists on which theory is best.

<u>Diastropic origin</u>. The idea that submarine valleys are due to earth movements is an easy way out of the problem. Under this view the similarity to land valleys as shown on contoured maps is "purely coincidental". However, it is quite generally agreed that the winding course of many submarine valleys, their V-cross section, and the presence of tributaries are fatal to this explanation of most of the known valleys of the continental slope. Others have suggested that slumping and sliding of the soft material on this slope, especially where it is unusually steep, might easily account for the observed irregularities particularly between the deeper canyons. Parts of the deep extensions of valleys and many irregularities of the sea bottom in regions of recent disturbance are probably due to earth movements.

<u>Excavation by land rivers</u>. Many of the points of objection to the tectonic hypothesis are in favor of subaerial origin. Many examples are off of existing streams, although not connected under water. Coarse gravel has been found in the bottoms of some canyons down to 5000 feet depth. Canyons run directly down the continental slope as rivers would, and seem to be roughly related to the size of adjacent rivers on land. On the other hand, some canyons have no land extensions and those of Georges Bank have a very small watershed. The grades are certainly abnormally steep. The principal difficulty with the idea of origin on land is to account for the vast change of sea level required. In order to avoid this some have suggested that the continental shelf was tilted up, the canyons eroded, and then it was bent back again carrying them into deep water. Others have thought that the entire body of sediment has moved down the continental slope carrying the canyons with it! Possible causes for tremendous shifts in oceanic level are (a) diastrophism involving part of the ocean bottom, (b) temporary uplift of the continents, and (c) vastly magnified glacial control of amount of water. Confirmatory evidence which would support one of these startling assumptions is lacking. Even if the modest figure of 3000 feet of lowering of sea level is taken, difficulties are still present. Glacial abstraction of water involves not only much larger and thicker glaciers than those commonly thought possible but also a great increase in salinity of the remaining water.

<u>Submarine origin</u>. Geologists who were greatly impressed with the difficulties of such immense changes in relation of sea to land turned to a search for some process which could erode valleys under water. Principal suggestions comprise: (a) density currents, (b) mudflows and landslides, (c) submarine springs, and (d) earthquake waves. The first is ascribed to more muddy water than now on the continental shelves when glaciation lowered sea level a few hundred feet. Muddy water is heavier than clear water and such underwater currents are actually known, although it is admitted that they are not flowing through the canyons today and off the mouths of muddy rivers the fresh water floats. Competency of such submarine currents to erode hard rock appears open to doubt. Under-water slides have been recognized in many places, particularly after earthquakes. On the other hand the termination of the valleys inshore is unlike the basins which develop on land from sliding and there are no enclosed depressions in the bottoms of the valleys or mounds of slid material at their bottoms. Tributary valleys are hard to account for by this idea. Landsliding may have taken place, however, and might account for many minor irregularities between the canyons. Submarine fresh water springs escaping from permeable layers of the Coastal Plain sediments of the east coast of the United States are a distinct possibility, but on the California coast they are not. Nevertheless, it seems impossible for such springs to produce branching valleys in the way they could on land. Certainly the fresh water should rise and its capacity to dissolve the overlying material and leave a consistent valley during retreat of

the spring appears impossible. It has been suggested that vast waves started by earthquakes during the orogeny of the Tertiary washed up onto the continental shelf. Backwash from these might then have eroded the canyons. Just why flow should be concentrated in the way drainage is on land is unexplained. Intermittent occurrence of these waves is also against the idea. No current meter observations during the passage of waves are recorded.

<u>Summary</u>. With work of waves in standing water is included that of currents, organisms, and the problem of submarine valleys. Waves and associated currents work to toward evening up of the shoreline by destroying headlands and filling bays, the latter process greatly aided by organic deposits in the quiet water behind a bar. The mathematical relation between length of a wave and its velocity is well known but there is none between its height and its length. Given height and length, however, total energy can be computed and checked with actual measurements. It is difficult to compare total energy expended on a shore with total energy of running water on an adjacent land area. But the quantitative value of force exerted by storm waves and the continuity of wave attack lend color to the idea that wave action is more potent than running water in completing the leveling of the land. Since wave base is not a fixed depth waves can erode farther into the land than was once believed possible, so that the formation of large wave-planed surfaces cannot be denied. The level of the oceans has varied greatly particularly because of withdrawal of water to form glaciers. Because the sea is now rising upon lands all over the world the commonly used classification of shorelines is unsatisfactory. For the same reason examples of a theoretical cycle of shoreline development cannot be found. The various theories of coral reef formation are compared and the differences found to be less striking than their proponents thought. It is clear that rising waters or land subsidence were required for most reefs but the cause is unimportant; in some localities the movement has undoubtedly been tectonic but as there is other evidence of postglacial rise of sea level glacial control cannot be ignored. The problem of origin of the submarine valleys now found on most coasts is still unsettled for none of the theories thus far advanced is free of fatal defects.

WORK OF WIND

<u>Introduction</u>. Over a large part of the surface of the earth wind is a potent force in the making of land forms. Wind can perform erosion, transport material, and build deposits. Although limited in effectiveness by vegetation in the more humid climates it can do some work there at certain times of year. The following discussion is based to a large extent on the book by Bagnold.

<u>Materials carried by wind</u>. Winds reach much higher velocity than water ever does yet their competence to move material is less. This is because objects do not lose as much of the weight they would have in a vacuum in air as they do in water. Density of air is only about 1.22×10^{3} compared to 1.0 for pure water. Viscosity is about 0.17×10^{-3} compared to 0.01 for pure water. [illegible] Nevertheless, objects falling through air reach a terminal velocity when the resistance is proportioned to the square of linear dimension and the weight to the cube. For this reason terminal velocity is in a general way inverse to size of particles. Instead of classifying materials as clay, silt, and sand it is more convenient in dealing with the wind to divide only into <u>dust</u> and <u>sand</u>. Distinction is made at the upper limit of size of particles which are kept in the air by the turbulence of ordinary winds. Effective upward currents are estimated by Bagnold at about 1/5 of the forward velocity of ordinary winds. A wind of 5 m/sec (11 mi/hr) will just support particles with a diameter of 0.2 mm which is a critical line of division between dust and sand. Most wind-blown sands do not have particles smaller

than 0.08 mm. Average diameter of such sands is from 0.15 to 0.30 mm. Sands are predominantly quartz because of its abundance, hardness, and toughness.

Behavior of sand in the air. The storms which obscure the sun in dry regions lift mainly dust; this stays aloft for many days after the wind falls. Real sandstorms are dust-free except at first and the sand rarely rises as much as two meters above the surface, with clear air above. Some of the grains moved are over a millimeter in diameter. Sand moves by being driven into the air by direct impact of wind. The grains describe a parabolic trajectory and land with enough force to dislodge other grains and start them on an aerial journey. This process is not exactly the same as saltation under water but is called by the same name. Experiment proves that the grains behave just about the same as spheres with a diameter three fourths as large. Sand also moves by surface creep, which is the motion of grains which could not be forced off the ground although they were slightly moved by impact. Grains thus moved are too heavy for the wind to raise off the surface. Most sand grains are too large for true suspension.

Wind velocity. Wind blows only in a turbulent manner. Wind velocity is related to heigth above the ground, not directly but to the logarithm of the heigth. When velocities measured at different heigths are plotted on semi-logartihmic paper a straight line gradient is displayed, which reaches the 0 velocity line at a definite elevation, k, above the ground. This heigth is about 1/30th of the diameter of the sand grains on the surface or other surface roughness. Now, as with water, the direct force of the wind per unit area is equal to half the product of its density multiplied by the square of its velocity at that level. The horizontal force or drag per unit area of surface parallel to the wind is equal to the product of density of air times the square of a quantity known as the drag velocity. Now the drag velocity, V# is directly proportional to the rate of vertical increase of wind velocity compared to logarithm of the heigth, that is the tangent of the slope of velcoity lines on semi-logarithmic paper. If we measure the velocity at two heigths, one of them 10 times the lower one, the velocity difference divided by 5.75 equals V#. In general terms: V# = vel. diff./(5.75 x log-heigth diff). This relationship holds for turbulent flow of all liquids and gases. Now it is evident that when wind passes from a smooth surface to a rough one the slope of the grade on semi-log. paper is unchanged but the point of 0 velocity is raised to a higher level, k', thus reducing the wind velocity by the same amount at all levels It takes some distance over the new surface before the effect is attained at all heigths. Velocity at any heigth, z = 5.75 V# log (z/k'). A rough surface is defined as one where the Reynolds number, (V# x diam. grains)/viscosity, exceeds 3.5

Effect of sand movement on surface wind. Sand in motion alters the surface wind. When we draw on semi-log. paper the rays for several different velocity gradients all converge to a single point of 0 velocity at level k' which is about 1/30th the dimension of surface roughness. When sand begins to move the gradient lines cross at a new point which is not on the 0 velocity line but at a definite velocity. This new focus is at level k' and at a velocity at which sand just begins to move. This is termed the dynamic threshold at which sand begins to move through impact. The rasing of the level is ascribed to development of a rippled surface. The velocity ray of the original line on which the new focus lies passes through the 0 velocity point of the grade at which motion started. From these facts it is clear that velocity at any heigth above ground when sand is moving is shown by the equation: v = 5.75 V'# log (z/k') + V_t where V_t is the threshold velocity measured at heigth k', and V'# is the drag velocity when sand is moving. Loss of momentum from the moving air by reason of sand transportation may be calculated by multiplying the quanity of sand, Qs, moved in unit time in unit width by the result of dividing the average loss of velocity of grains by their average distance of travel, l. Sine the velocity with which a grain starts its flight is small it may be neglected and only the final velocity, u, need be considered. From this it follows that the drag may be expressed as: Qs x(u/l) This is equal to density x $V'\#^2$ Now it has been found that u/l closely approximates g/w where w is initial vertical component of velocity of a grain at the beginning of its path, and g is the acceleration of gravity, hence Qs x (g/w) = density x $V'\#^2$. Solving this equation for quanity, Qs = (density/g) x w x $V'\#^2$ Now making the assumption that w : V'# we can substitute for w and find that Qs = constant x $V'\#^3$ Experiment disclosed that about a quarter of total sand movement is not through saltation but is due to <u>surface creep</u> of grains which are not struck with enough force to send them into the air. The formula for sand movement was checked by wind tunnel experiments which yielded a constant of about 1.47 where value of density / g is about 1.25 x 10^{-6} It can also be demonstrated that the quantity V'# can be replaced in the equation by velocity at any given heigth above the ground less the threshold velocity thus giving the final expression:

$$\text{Quantity} = \text{constant} \times (\text{excess velocity at given heith})^3$$

It was also found that a wind of given velocity can drive sand faster over a hard, immobile surface than it can over loose sand, but no quanitative data were obtained. The above equations were derived for sand of one diameter of grains only but may be modified to meet natural conditions by altering the constant.

<u>Comparison of results for air with conditions in water.</u> Bagnold remarks that the good results he obtained in deriving and checking formulas for transportation of material by moving air are not matched by those others obtained with water. This is due in part to difficulty of observation under water but mainly to the tremendously larger loss of weight of particles in water. As a result, the reduction of velocity of a stream of water by reason of saltation is, other things being equal, less than one one thousandth that which takes place in air. The frictional drag of air on the ground may be neglected for it is so small compared to transfer of momentum by the sand load. In water the bed load extracts little momentum and water velocity is regulated almost entirely by the roughness of the bottom. Grains are not dislodged by impact under water but probably by eddies of turbulent flow. Were the bottom of a stream to remain smooth erosion would be much more that it actually is. Development of bottom roughness then acts as a limit to movement of sediment. When suspension of particles begins, however, the bottom becomes smooth again. This condition begins when 7 V# exceeds the average settling velocity of particles. When suspension is fully developed grains move along with the water like part[illegible]

<u>Suspension in air.</u> When V# exceeds one seventh of the falling velocity of grains in air it appears that suspension replaces saltation. Since this falling velocity is proportional to diameter over a wide range of grain sizes we can write as a fair approximation: quanity = $V\#^3$/ grain diameter. Bagnold suggests that the change to suspension sets in when the grain-impact method of saltation ceases to operate and that thereafter we have a condition akin to sediment transportation in water in which saltation is not the same as in air.

<u>Field experiments.</u> Bagnold carried out extensive field experiments in Libia to check the foregoing laboratory work. Here the grain sizes are more varied and the surface is nowhere trully flat. However, the checks were satisfactory.

<u>Relation of threshold velocity to grain size.</u> The threshold wind velocity is that when grains just begin to be dislodged from the surface. The angle at which a grain must be raised to get away from the ground is assumed to be close to the <u>angle of repose</u> of loose sand. It is possible to compute the required force in the same way as for streams. Equating force to resistance we find that :
$V_t\#$ = constant x (effective weight x diameter/ air density)$^{1/2}$ or in other words if other things are equal the value of the threshold velocity, $V_t\#$, varies with the square root of grain diameter. This relation holds only for diameters over 0.25 mm. The constant for these and larger grains is 0.1 compared to 0.2 in water. When the Reynolds number, V#d/ viscosity, is less than 3.5 a greater drag is required to set grains in motion. When the grain size is less than 0.2 mm the value of

the constant increases and the square root law no longer holds. In water the same change is at a diameter of about 0.6 mm. Data for force needed to start small particles is hard to obtain but it is not wholly because of cohesion that it is larger. Bagnold has observed that in aird regions with soil of fine texture there is little dust except where the surface has been disturbed artificially. He also calls attention to the clean separation of loess from sand. The critical diameters are in air 0.08 mm and in water 0.2 mm. An important factor in holding down small particles in air is their moisture content. Another is a mixture of large and small particles in which the former protect the latter. The impact threshold wind is distinguised from the fluid threshold discussed above because at it the motion of grains is maintained by the impact of falling particles alone. The wind up to the critical heigth, k', is then unchanged no matter how hard the wind blows above that level. In water there is no corresponding condition, only the fluid threshold. and there is no fixed focus of constant velocity.

Land forms produced by wind. Land forms produced by wind may be divided into first, erosional, and second, depositional. In the latter class we are here interested only in the mounds of sand known as dunes and similar large scale deposits of dust or loess.

Erosional land forms. Erosion by wind is sometimes called deflation. Topographic forms made by wind erosion include: (a) hollows or basins, sometimes called blow-outs, (b) pillars, pinnacles, and cliffs undercut by sand blown along the ground yardangs; and (c) residual portions of dried-up lake beds or playas. Hollows may be eroded by the wind in humid climates provided the soil is unfavorable to a cover of vegetation. The initial step is destruction or reduction of this protective material by drougth, fire, or the work of man. Wind then sweeps away the underlying material if it is of a nature which is readily picked up. Blow-outs are most abundant where the material is sand but some are found on shale which disintegrates into dust. Such are abundant in the Colorado Piedmont and the Wyoming Basin. Big Hollow, near Laramie, Wyoming, is 150 feet deep and about 3 by 9 miles in extent. Some of thse hollows contain lakes when there is enough rainfall. Some are limited in depth by accumulation of pebbles into a desert pavement others by reaching moist material near the water table. Large depressions in Libia and Egypt, such as the Qatara Depression, have been ascribed to wind erosion and some have applied this theory to many of the enclosed basins of the Basin and Range province of this country. Many of these are more likely due to earth movement and it seems doubtful that wind ever eroded far into solid rock. Undercut cliffs or rock shelters have been ascribed to wind erosion but old Spanish inscriptions in some of them in the southwest indicate that, if operative, the process is slow. The beds of playas or temporary lakes have been eroded by the wind in many places leaving miniature mesas capped by salts. Cracking of dried mud facilitates wind erosion. However, it is an open question whether or not the dust is removed far enough to not be washed back by the next rain.

Depositional land forms due to wind transportation. In humid regions the major sources for sand which is available for wind transportaion are beaches and river beds. When the glacial drift was newly deposited and glacial lakes disappeared large areas were free from vegetation and exposed to wind action. Under present conditions in humid regions there is an ever-present contest between vegetation and wind. In true arid climates, such as in Libia, wind has the field to itself provided the rocks weather into material which is within the power of wind to transport. There dunes reach full development in the erg. Bagnold lists: (a) sand accumulated behind obstacles, (b) true dunes divided into barchans and long mounds, (c) coarse-grained ridges or whalebacks, (d) gently undulating sand tracts, and (e) sand sheets. Rock surfaces in deserts are often called hammada. We must realize that by no means all the surface of a desert is sand.

Dunes of humid lands. Sand which is blown from areas with no vegetation is commonly deposited in a short distance to form a ridge parallel to the source. These ridges are often called transverse dunes or foredunes. Because in part of variable wind direction and in part of scanty vegetation these ridges are very unstable. Their corss section is gently sloping toward the source and sand is added in layers with a gentle dip to windward. The lee face is made by sand which slides down in a slip face(foreset bedding) at an angle of about 34 degrees. Low places or breaks in the cover of vegetation are blown out into hollows. The eroded sand is heaped around the heads of these depressions, some of which have no outlet. Similar accumulations are also formed on the lee sides of blow-outs in plains of sand. Some have called this type of dune parabolic but the name is unfortunate because there is no relation ot the mathematical form by the same name. The open ends of the crescents point toward the wind. When a group of blow-out dunes starts the earlier ones are checked when the hollow attains its maximum possible depth. With source cut off the dune becomes stabliized with a cover of vegetaion. Local failure of this starts up the process again so that a group of dunes travels slowly down wind, leaving a confused system of mounds and hollows which at first sight appear wholly without plan. In middle latitudes, away from the ocean and large lakes, winds are so variable in direction that the irregularity of dune topography defies analysis. Enclosed depressions are not all blowouts byt are made by advance of the lee faces of dunes. Lakes are present in some depressions. Blown sand also serves to help fill the lagoons behind sand beaches. Many abandoned beach lines are marked by rows of dunes.

Dunes of arid lands. The largest areas of arid climate in the world are perhaps those of the trade winds. There the constant wind direction leads to the greatest perfection of dunes. When the wind is strong sand is removed from pebbly surfaces and accumulates in sandy areas. Gentle winds cause these patches to travel downwind and the sand to be scattered. Eddies in the strong winds concentrate sand. In many places sand is blown into long stripes parallel to the wind direction. Some dunes are made of gypsum particles.

Shadow dunes. Sand accumulates in wind shadows behind obstacles such as rock cliffs or bushes. Shadow dunes which are transverse to the prevailing wind are unstable. Long dunes which grew out too far to leeward of the protection are apt to be broken up particularly by cross winds. The bedding of such dunes is parallel to the surface except at the lee end which has a slip face.

Barchan dunes. Prehaps the best known dune form is the crescent-shaped barchan whose horns are pointed down wind. These forms are best developed where the sand is rather limited in amount resting on a non-sandy basement and the winds are unidirectional. The windward face is gentle and the slip face is inside the two horns which have a minimum heigth of about 30 cm. Where barchans are closely spaced the partially protected examples are more complex in form than is normal. Enclosed depressions between them are sometimes called fulji. A system of barchans is most commonly slightly offset or en echelon. Maximum heigth of barchans is about [illegible] feet. The entire streamlined form is slowly moving downwind at a rate which may reach several centimeters per hour during storms. Layers added to the windward face are firm but the slid sand of the lee sides is very soft and unstable. Where the windward side is exposed by erosion this soft sand reaches the surface.

Seif or longitudinal dunes. In the very sandy areas of Libia, Arabia, and Australia the dunes are long ridges which parallel the direction of the prevailing trade wind. These long ridges are known as seif or longitudinal dunes. They are known to reach a heigth of over 200 meters (700 feet), a width of about 6 times their heigth, and a length of 100 kilometers (60 miles). The summits have crests from 20 to 500 meters apart or about 6 times dune heigth. The windward end of a longitudinal dune is generally broad, locally with an enclosed depression and the lee end is sharp. There is a slip face on one or both sides and a gently sloping basal plinth of firm sand which has never slid. The summits apear to be moving and the entire dune is possibly moving laterally as well as toward the lee end. The lateral spacing of seif dunes is from 1 to 10 kilometers and the corridors between them extend for long distances. Barchan dunes occur in some of them. The origin of thsese long dunes is disputed. Some demonstably occur in the lee of obstacles. Bagnold holds that they are due to the alteration and consolidation of barchans by cross winds. To the writer it seems reasonable to conclude that they are the ultimate streamlined form of minimum friction with constant wind direction.

In other words, they are extremely elongated barchans in which the two tails have been joined together.

Whalebacks. The whalebacks of the Libyan desert are sometimes called sand levees. They are from one to 3 kilometers wide and up to 3000 kilometers long with a heigth of up to 50 meters. There are discontinuous chains of dunes on the tops, at one side, or in a series side by side. Bagnold regards whalebacks as residue left by migration of either one seif dune or a series of them.

Undulating tracts. The gently undulating tracts of sand in Libia seem to occur where there is some rainfall and vegetation. They seem to be similar to many sand areas of humid or subhumid areas where dune topography is poorly defined.

Sand sheets. The areas in Libia which are termed sand sheets have a surface of coarse sand with some pebbles. The material below has layers of sand and pebbles with fine red powder below a depth of about 10 cm. Pebbles were derived from nearby bed rock. The sheets appear to be the result of longitudinal sand strips which are protected by these pebbles. Some water action may have occured.

Loess deposits. Toward the borders of the more humid regions to the lee of extensive dunes there are silt accumulations called loess. Notable examples occur in the United States southeast of the Sand Hill district of northwestern Nebraska where the dunes appear to be the result of past wind erosion of Tertiary alluvial fans at the foot of the Rocky Mountains. In China the loess lies in the lee of large desert areas. In Europe loess is most widespread in southern Russia. The so source, method of transportation and time of deposition of loess has long been disputed. Its derivation has been variously ascribed to fresh glacial drift, to stream floodplains, and to wind erosion in deserts. Transportation and deposition has been ascribed both to lakes, streams, and the wind. In time, opinion has varied from interglacial arid intervals to during or immediately after glaciation. Although loess is in many places thickest adjacent to rivers from whose beds it may have been derived, this localization of deposition might also be explained by rough, forested hilly topography which caused local still air. It is now generally thought that loess was transported and deposited by wind, although some think that the loess of the east bluffs of the Mississippi is the product of weathering of clay. The arguments cannot be discussed here except that there may be two types of loess, one late-glacial or glacial in age, the other interglacial.

Land forms due to loess. Land forms due to loess deposition are not abundant. On the east bluffs of Missouri River in Iowa loess forms hills which as viewed from the air suggest snowdrifts. Here the loess locally reaches 200 feet thickness. In most places the loess simply forms a mantle over older topography of various origins. When eroded, slopes are very steep because of its high permeability and the vertical cleavage, which some ascribe to the casts of grass roots. Landsliding on these slopes gives rise to minor terraces called catsteps.

Summary. The wind is the most potent force in shaping the landscape in truly arid regions, even though there are some traces of water-work in many of them. Both barchans and seif dunes tell of uni-directional winds, the latter prehaps the ultimate form of minimum friction between sand and wind. Barchans whose horns point down wind should not be confused with blow-out dunes of humid regions whose convex sides are exactly opposite. It is quite possible that conclusions on former wind directions in some localities are 180 degrees in error for this reason. Direction of dip of foreset bedding is a much more reliable criterion. The complex dune topography of humid regions is readily explained by the conflict of vegetation with winds of variable direction. Wind shadow dunes occur in all regions. Whalebacks and sand sheets are mainly confined to arid regions where wind has worked for a long time with little interference from vegetation. Constructional hills of loess occur on the borders of humid regions.

WORK OF ICE

Introduction. Work of ice includes the land forms made by glaciers, icebergs, and the ice of lakes. It excludes the work of ice in weathering and soil formation. The approach here is somewhat different from that used in the study of glacial geology, in that it is not concerned to any great degree with the physical nature of the deposits but is confined to processes which made land forms.

Glaciers-introduction. A glacier may be defined as a mass of ice which was formed by compaction of snow and which flows, or at some time has flowed, under the influence of gravity, that is it is drainage of precipitation in solid form. Glaciers are subdivided into (a) valley or mountain glaciers, (b) piedmont glaciers formed by the joining of several valley glaciers at the foot of the mountains, and (c) continental glaciers which cover large areas.

Origin of ice. Glacier ice is compacted recrystallized snow. Partially altered snow is often called firn and has a density of 0.72 to 0.84 A layer of not less than 100 feet of this material is found near the surface of the source areas of glaciers. Much more is probably present in polar regions. Firn is absent in the lower parts of glaciers where they are wasting away. Glacial ice below the firn contains up to 15% of included air and its density does not exceed 0.9 in contrast to 0.918 for ice made by freezing water. Ice crystals are hexagonal and are several inches in diameter.

Physics of glacial motion. The physics of glacial motion have long been misunderstood by many geologists. Near the surface, where it is under light load, ice behaves like a solid and yields to stress mainly by fracture. Under heavier pressure its physical behavior is like that of a liquid. This is no different from the phenomena of rock deformation except that it occurs under lighter stress. The only available determination of the viscosity of ice is 1.2×10^{14} poises. Demorest, who did much work on this subject, states that viscosity decreases with load. Presumably this is because pressure causes recrystalization with the structure arranged to facilitate flow to relieve the stress. We would then not expect any further change below the depth at which this process has been completed. Ice cannot exist at temperatures above 0 degrees C. The flow of ice is laminar. For a valley glacier with approximately a semi-circular cross section average velocity would be given by the following:

$$V=\frac{g \times \text{density} \times \text{sine slope} \times \text{depth}^2}{32 \times \text{viscosity}}$$

On a slope of one deg. a glacier 100 meters thick would then have an average velocity of about 3.5 cm. per day which seems to agree with actual observations. In such a thin glacier on a sloping base one layer flows over that just below. Force is the component of gravity parallel to the bottom. Rate of flow should be 0 at the base under thick ice but near the terminus the thin, rigid ice might be shoved bodily over the rock. This type of flow was termed gravity flow by Demorest. Conditions are different in a thick continental glacier. The top is rigid and is retarded where it reaches the ground at the thin outer edges. Below, the ice flows by reason of the difference of top elevation from place to place. Since yielding can only be outward the component of weight of a unit column of ice which is parallel to the top is the force for motion and velocity at depth d is shown by the following formula:

$$V = \frac{g \times \text{density} \times \text{sin slope} \times \text{depth}}{\text{viscosity}}$$

Substituting for a slope of 0.1 degree (less than 10 feet per mile) and thickness of 5000 meters the result is only about .0003 cm. per day. Bottom velocity should also be 0 except near the margin of the ice. Demorest called this extrusion flow because it is present only at depth.

Ice erosion. Ice erosion is the result of friction between moving ice and the bed. The force of friction is equal to the weight of a unit column of ice multiplied by a coefficient which depends upon nature of underlying material and not on velocity. The power, or time rate of work, of a glacier is this force multiplied by velocity. Other things being equal, the velocity of a valley glacier is related to the square of the depth. In a continental glacier this relation does not hold for the formula does not take into account the effect of spreading out of the ice towardd hte margins. If we reasonsed solely upon the formula for a valley glacier power should be related to the cube of ice thickness. However, this conclusion does not tell the whole story. Bottom velocity, except near the terminus, is 0 or close to it. Much of the energy is absorbed in internal friction. If this loss results in pressure-melting of ice followed by refreezing then no energy is lost. But if there is some permanent melting or conduction of heat to the surface energy is really lost. The greatest differential velocity of ice over rock must be near to the end of a glacier. Most geologists seem to think of erosion by ice as mainly due to grinding of rock into powder or rock flour. The word scour is often used for this type of work. A little thought will show that erosion of hard rocks in this manner would absorb an enormous amount of energy, and since the work would be spread over a very large area of ice bottom, be extremely slow. Although it is an undoubted fact that much rock flour is made by glacial action, a much more potent form of erosion of firm rock is removal in pieces. This process is termed plucking and occurs where plastic ice can flow around and freeze to rock masses which have been broken by older fractures. Such pressure-melting and refreezing takes place under thick ice. The melting point of ice is lowered about 1 degree C for every 2100 meters depth. Computation shows that with the normal heat emission from the earth the bottom of all thick glaciers must be at the pressure-controlled melting point. Under this condition very slight changes in pressure may change the ice from solid to liquid and vice versa. If cracks are present in the bed rock fragments are thus incorporated in the ice and move forward with it. This results in much more rapid erosion than would otherwise be possible. Another mode of erosion is present in valley glaciers where there is a prominent crevasse called the bergschrund at the head next to the mountain wall. Much meltwater both from the ice itself, and from banks of snow above, enters this crack and freezes. This freezing loosens many blocks of rock which are then carried off by the moving ice. This process is called sapping. The importance, and even the existance, of glacial erosion has long been debated. That such erosion is an important process in shaping of land forms, is demonstrated by (a) the vast amount of fresh material derived from bed rock in the glacial deposits, and (b) the unique topogrphy of many glaciated districts.

Deposition by ice. In considering ice deposition it is impracticable to separate the result of direct ice deposition from the work of meltwaters. Unstratified and unassorted material direct from the ice is calld till and the word drift includes both this and associated water deposits indirectly due to glaciation. Deposits of glacial streams consist of sand and gravel which bears in its nature the record of the frequent changes of volume and velocity of ice-borne streams with floating ice. In areas not long vacated by glacial ice large residual remnants were buried in the deposits and did not melt for a considerable time.

Erosional land forms of valley glaciers. The most striking and characteristic erosional land form of valley glaciers is the cirque. These bowl-shaped depressions are also known as corrie or cwm. They occur not only at the heads of mountain, valleys but also on mountain sides and frequently are found in stariways one above another. Cirques are ascribed to sapping in the bergschrund of small glaciers. Another characteristic feature of glaciated mountain valleys is a non-uniform grade with enclosed rock basins separated by intervals of abnormally steep slope. Rock basins are also present in the bottoms of many cirques. The transverse cross section of many glaciated valleys is notably U-shaped rather than the V-shape of normal stream valleys in mountains. This phenomenon is best displayed in massive igneous and metamorphic rocks and is enhanced by a filling of gravel outwash in the bottom. Unfortunately it has been termed catenary by some geologists, although it bears no no relation to the curve made by a rope or chain suspended at the ends. Instead it is explicable by the fact that the work of a glacier is spread over a wider bed tha n that of a stream carrying the same total discharge in unit time. The coasts of many glaciated mountainous regions such as Norway, Greenland, Pategonia, and Alaska are indented by many long narrow bays called fiords(fjords). Many of these are much shallower at their outlet into the sea than they are inland where depths up to 4000 feet have been recorded. Fiords form a branching system which in many places is trellis, that is adjusted to the structure of the bed rock. Many of the tributaries enter from hanging valleys; some of these have the lip under water and others give rise to spectacular falls or rapids. Fiords are now generally ascribed to glacial erosion which was less effective under thin ice near the outlet thus leaving the threshold. Hanging valleys are also common in almost all glaciated mountains although discordant junctions are by no means due only to glaciation. It is now quite generally recognized that glacial erosion on a large scale is closely related to the amount of fracturing of the bed rock and is thus controlled by regional structure. This was well shown by Matthes in Yosemite Valley, California. However, it is possible that the relationship of glacial erosive power to the cube of ice thickness is another important factor in the production of hanging valley junctions. A relatively minor topographic feature of glaciated valleys is the roche mountonee, rock knobs with a gentle slope on the stoss side toward the source of the ice and a steep slope on the opposite or lee end. Some students have ascribed these to turbulent flow of the ice but in view of its high viscosity thia is impossible. Roche mountonees are readily explicable as simply rock masses too large to be removed by plucking which were ground down on the exposed side by the moving ice; they have no necessary relation to preglacial features. Roche moutonees should not be confused with exfoliation domes. Many other irregularities of glaciated valleys are doubtless explicable by variation in amount of fracturing of the bed rock.

Cycle of mountain glacial erosion. Attempts have been made to distinguish a cycle of mountain glacial erosion similar to the cycle of stream erosion. It is true that glacial erosion varies greatly in amount in different localities depending upon the size and length of life of the glaciers. In some places only isolated cirques are present, wheras in other localities the headwalls have been worn back so far that only narrow ridges (arêtes) are left between cirques and the higher summits have been sharpened into horns. But what the next step would be is unknown, for it is clear that glaciation has been only a relatively brief episode in the history of existing mountains. It may be presumed that if glacial erosion continued long enough it would reduce the entire mountain range so much that snowfall would be decreased. Possibly this might lead to extinction of the glaciers.

Depositional land forms of valley glaciers. The depositional land forms of valley glaciers are relatively small compared to adjacent rock topography. Where the terminus of a valley glacier remained stationary at a position fixed by a balance between melting and forward motion a moraine was left in an arc across the valley. Such are known as terminal or endmoraines. Debris which fell from the mountainsides onto the ice left lateral moraines. Where tributary glaciers joined the lateral moraines coalesce into medial moraines. Many lakes are enclosed behind moraines. Valley floors both within and outside the maximum ice limit are filled with outwash deposits of sand and gravel, often called valley trains.

Erosional land forms made by glacial waters. Streams of glacial meltwater erode notches across spurs and eroded many potholes in rock. The most spectacular feature due to erosion by water is the scablands of Washington. A prolonged controversy has been waged over the origin of these areas of bare basalt with many abandoned waterfalls and only small areas of gravel. Bretz held in many papers that the scablands were eroded simultaneously by a vast flood hundreds of feet deep. No cause for such a flood could be found. Flint discovered that the gravel of what Bretz called bars is normal outwash which is too fine to have been deposited by a vast flood. He concluded that valleys across the Columbia Plateau were first filled with outwash which sloped down to Lake Lewis in the lower Columbia Valley at a grade of I3 ft/m. Neither the origin nor the cause of draining of this lake was discovered but Flint concluded that lowering of its level before the ice front was melted back brought about a change in slope of the meltwater streams to about 20 ft./m. This caused erosion of almost all of the outwash and the remnants left in the mouths of tributaries are what Bretz interpreted as bars. Allison has disputed Flint's ideas on the ground of (a) improbability of simultaneous erosion over so large an area, (b) misinterpretation of the relation of outwash to the sediments of Lake Lewis, (c) the topographic form of some of the bars which he thinks are constructional, and (d) presence in the lower Columbia Valley of high-level stream-eroded areas above the lake silt. Allison gives no general theory but suggests that the effect of ice jams in diverting the rivers has been neglected.

Erosional land forms due to continental glaciers. Erosional land forms left by continental glaciers are not as conspicious as those formed by valley glaciers. In fact, the origin of some of them has long been disputed. In making comparisons it is well to realize that continental glaciers probably had much less velocity than do mountain glaciers which rest on steeper slopes. The latter are in motion through out almost all their life wheras continental glaciers became so thin during their wastage that motion must have ceased throughout large areas. Some continental glaciers do not seem to have remained in motion long enough to make an endmoraine. It is true that many fiord coasts were once covered by continental ice but an important factor in producing these deep valleys was the presence of local glaciers both before and after every continental ice cap. For that matter, cirques occur in areas which were covered by continental ice, for instance in the White Mountains. These were certainly made by local glaciers. Major features in the United States which are generally recognized as due to erosion by continental ice are (a) the deep, youthful, glaciated valleys of the northern Appalachian Plateau, some of which contain the famous Finger Lakes, (b) the basins of the Great Lakes, and (c) some of the remarkably straight escarpments of the Great Lakes region. The Finger Lake valleys are so straight and deep, extending below sea level, that only glacial erosion can be the cause. Cooperating processes comprise (a) erosion by meltwater from the advancing ice thus lowering divides, (b) erosion by diverted streams during interglacial intervals, and (c) reversal of drainage in preglacial time from a southward course to join the subsequent valley of the Mohawk. It has been suggested that erosion was concentrated in the deeper valleys because only there was the ice thick enough for pressure-melting which allowed of plucking. The Great Lakes certainly lie in basins which are enclosed by bed rock. The problem is complicated by known earth movements which appear to be still going on. Only a small part of the depth of the deeper lakes can possibly be explained by dams of glacial drift. Both Lakes Huron and Michigan are crossed by submerged cuestas on the Devonian limestones. The deepest parts of these lakes are above Silurian salt and gypsum-bearing rocks which must certainly have been easily erided by ice. Other basins appear to be all on shale. The east end of Lake Superior is extremely irregular but the relation to the structure of the Keweenawan sediments and traps is unknown. It is impossible to account for thse basins by earth movement alone although neither the preglacial drainage nor the exact amount of glacial erosion can be determined at present. The absence of valleys across the cuestas appears to indicate that preglacial valleys were bottomed far higher than the lakes now are. Glacial erosion is also indicated by the large amount of drift south of the lakes. The simple form of the escarpments in much of the Great Lakes region tells of glacial removal of spurs and outliers. This seems to have been most marked where the soft rock is shale and where the ice moved approximately parallel to the escarpment. Roche moutonee hills occur in areas of continental glaciation.

Depositional land forms due to continental glaciers. Depositional land forms due to continental glaciation with associated meltwaters include moraines (both terminal or end, recessional, and ground), drumlins, outwash, eskers, and crevasse fillings. Terminal or endmoraines originated in the same way as those of valley glaciers. Recessional moraines have been much misunderstood. Those due to a halt in melting back of a front of moving ice are very irregular in outline. Most moraines behind the outermost or endmoraine are the result of readvances of the ice front to a regular smooth outline burying water deposits. Moraines made of stony material contain much water-sorted gravel and sand and have steep slopes. All driftt is very wet when deposited and water can escape from sand and gravel without causing extensive sliding. Where a large amount of clay and silt is present in the drift slumping leads to low slopes. In detail, many moraines consist of a complex series of minor ridges each along some temporary position of the ice margin. Many depressions are left between these ridges and others, called kettles, were made by the melting of buried ice masses. Clay moraines are comparatively inconspicious and have few depressions. Ground moraine is composed almost entirely of till which was either deposited under the ice or left behind when it melted. Where the drift is thin the present topography is a smoothed reflection of an older landscape of some other origin, either erosional or depositional. In regions where the clay content of the till is large, the ground moraine is thick enough to cover the rock topography, forms a nearly level drift or till plain. Such are common in Iowa, Illinois, and other regions where a large part of the bed rock is shale which supplied the clay to the ice. Drumlins are conspicuous only in rather stony drift. They are streamlined hills up to 200 feet high which are composed mainly of till. Their long axes are parallel to the direction of ice movement and the stoss end is the steeper. The flanks are as steep as wet till could stand. In detail the ideal form is in many instances confused by fusion of adjacent drumlins and by change of ice direction. Drumlins attained their shape because they are the form of minimum friction for material which accumulated in cracks in the bottom of moving ice some miles back of the margin. Outwash consists of the deposits of glacial meltwater streams which on leaving the ice flowed over a relatively low slope. Here they formed a braided pattern and laid down much of their load of sand and gravel. If the locality of deposition had not been occupied by ice for some considerable time the resulting topography was a smooth plain. But where residual ice masses had been left from a recent glacial invasion sediment accumulated around and above them. Melting left a confused topography with many kettles which, except for the nature of the material, resembles many terminal moraine deposits. Such rough deposits are called pitted outwash and vary from isolated pits in a plain to areas where no trace of a depositional surface is left. The areal distribution of such deposits is unlike that of moraines in that the longer dimensions are more apt to be parallel to the direction of ice motion instead of transverse to it as moraines are. Eskers represent the filling of the beds of streams which flowed between ice walls. It is hard to tell in many cases whether these streams were in tunnels or in crevasses which were open to the sky. The gravel and sand ridges are discontinuous and have irregular crests. The flanks are very steep. In some cases they have tributaries which tell definitely that deposition took place in dying, stagnant ice. Crevasse fillings are narrow ridges between kettles of pitted outwash plains. Moulin kames are conical hills due to gravel which was deposited in holes which melted through the decaying ice.

Lake deposits. Glacial ice formed the dam which enclosed many lakes. Those which were shut in front of advancing ice contained much open water although in the glacial climate both icebergs and lake ice must have been present most of the time thus damping wave action. During ice recession residual ice masses must also have been present in all low places so that many supposed lakes were simply narrow moats. The land forms made by glacial lakes differ little from the deposits of standing water described above and need no further description. In some places mounds of till within lake beds have been reported. Some of these may have been due to the overturning of icebergs.

Work of lake ice. In middle latitudes where there is a great variation in winter temperatures the lake ice can expand considerably during warm periods when snow has melted off the ice. Where shores are high the ice breaks in an expansion crack but whereever the resistance of low shores is less than the strength of the ice expansion pushes up a ridge. The coefficient of expansion of ice is 50.7 x 10^{-6} per degree centigrade or about 5 times that of steel. A kilometer of ice raised 20 deg. C. would then expand about a meter. Where ice push ridges contain many boulders they can survive wave erosion and become permanent land forms. Some examples are over 20 feet high and extend for miles. Ice push appears to be uncommon in the far north because the lakes are always covered with snow in winter and the spring thaw is so rapid that the ice soon weakens.

Summary. The work of ice was important over a large part of the northern countries. Glaciers are still present in mountains even near to the equator and in the not remote geolgic past were much more extensive all over the world. Then continental glaciers, like those still present in Greeland and Antarctica, overspread Canada and the northern United States. It has been suggested, however, that many glaciers of the present day are not survivals of these ancient ones but were formed in the Little Ice Age which began about 4000 years ago. Glacial erosion by plucking of previously fractured bed rock is far more important than grinding up of rock. Sapping or frost-loosening of rock is important in the mountains. Erosional forms, such as cirques and fiords, developed by valley glaciers are much more conspicious than are the results of erosion by continental glaciers. On the other hand, the depositional features of the continental ice and associated meltwaters are far more important than are those of the mountain ice streams. These deposits left a disorganized drainage system with many lakes and swamps and streams superimposed on the older rock topography making falls and rapids. Unlike valley glaciers the continental ice caps were thinned to the point of stagnation at many times during their recession. Some did not even leave any terminal moraines. Others underwent many periods of rejuvination through increase in snowfall and readvanced, to make a series of so-called recessional moraines which are really the record of counter-attacks against the sun which finially reduced them to their present limits.

TECHNIQUES

Introduction. Technical methods for the study of problems in geomorphology are varied. In this text a method as yet in its infancy has been attempted, namely the approach through physical controls. Either the mathematical theory is developed and then checked against facts gathered in the field or from maps or the reverse, first derivation of a formula which fits the facts followed by an analysis of the mathematics of the physical processes involved. The tools used in search for data comprise: (a) topographic maps, (b) aerial photographs, (c) profiles, and (d) block diagrams or perspective drawings of other types.

Topographic maps. Topography of an area may be shown on maps by means of (a) contours, (b) hachures, (c) shading, or (d) a combination of two or more of these methods. In this country most maps published by governmental agencies use contours alone. Unfortunately, the ability to use contour maps effectively is attained by relatively few persons. To many these maps have little meaning aside from the spot elevations of definite localities. It is now evident that in the early days of making these maps too much was expected for too little time and travel in the field. Errors and omissions due to failure to visit portions of the area are glaringly apparent not only when the locality is visited but also in aerial photographs. Not only are there mistakes in form and elevation but different streams have been joined together. Even in maps of recent date a field check almost always discloses some errors in portions which are covered with forest or are remote from roads. Many foreign maps use hachures with at most, only a few contours. Spot elevations are given on most of these maps. Some surveys have used contours for gentle slopes and hachures for cliffs and steep slopes. Various methods of shading have also been tried, some as though the light came from the upper left corner, others with density proportioned to slope as with hachures. Most of these maps are hard to read and accuracy of detail is doubtful in many instances.

Aerial photographs. Within the last 10 years most of this country has been photographed from the air largely in connection with various "New Deal" activities. In peace time these photographs have been made available to the public at reasonable cost. Aerial photographs are of three general types: (a) verticals, best for mapping and determination of areas, (b) low obliques which do not show the horizon, and (c) high obliques which include the horizon. Some methods of mapping use a cobination of the first and last types. We need not here discuss either the methods of taking aerial photographs or the details of making maps from them. It is obvious that photographs record far more detail than does any map. Except under very dense timber nothing is hidden from the aerial camera. In fact the objection is often made that photographs show too much confusing and unessential detail. Aerial photographs may be used (a) singly, in which case obliques show relief much better than do verticals, (b) joined together into mosaics which are not true maps, and which can only be made from verticals, (c) examined in pairs taken from different positions of the plane in such a way that only one is seen by each eye, that is stereoscopically, or (d) superimposed by various methods and then examined by means of special glasses so that only one is seen by each eye. Both verticals and obliques may be used in the two last ways, and so can pictures taken on the ground provided only that they were taken far enough apart and yet show the view at very nearly the same size. Several types of instruments are used to enable one to see one picture with each eye. These are known as stereoscopes but with practice many persons can dispense with all instruments and make the eyes look along parallel lines. Various methods of attaining this method have been tried. One is to relax until a single object appears double; if an effort is made the eyes automatically converge so that a single object is seen. When double vision is attained then a pair of photographs can be held up with common points from 2 to 2½ inches apart. Suddenly you will realize that three images only can be seen and that the central one, where in fact two are superimposed, is in relief. A stereoscope is essential, however, to measure differences in elevation on verticals by comparison of distance between common points which varies with elevation. The method of superimposed pictures in different color, or with light polarized at right angles, produces what is called an anaglyph. Such pictures are simply a blur until viewed through the proper kind of glasses. Maps can be made in this way, although so far only of limited areas; spot elevations or contours can be shown to give quanitative data on anaglyphs. It is clear that the use of aerial photographs opens up new fields to geomorphic study. This is particularly true where the mantle rock is thin and vegetation is sparse. Then the structure of the underlying rocks is very clear. Faults and folds may be traced accurately by the varying topographic expression and mantle rock of each kind of bed rock. Study of glacial deposits in which the surface form plays an important part is greatly facilitated by aerial photographs. Drumlins, moraines, eskers, and outwash all stand out clearly. Outwash is distinguised in many localities by the sandy soil which photographs a light color, and pitted outwash shows a mottled pattern with many rudely circular swamps and lakes. The lakes in terminal moraine are for the most part irregular in outline.

Profiles. Many students of geomorphology place much importance on profiles or cross sections. Of necessity, these are made from topographic maps and are no more accurate than they are. Profiles are drawn on cross section paper with the vertical scale considerably larger than the horizontal scale which may or may not be the same as the map. They may show only the surface along a single line, either straight, curved, or a series of lines at angles, or they may be of the projected type. If the latter two distinct methods have been employed. Barrell, in a search for ancient marine terraces in New England, drew the crests of hills beyond the front line of his profiles back so far that the result is confusing. Each summit was projected on a line at right angles to that of the section. Others have used a strip of definite width and shown the highest elevation in this strip which is present at right angles to the front line at every point. This is equivalent to showing the skyline of a model of the landscape of this strip which may be a mile or more in width. Johnson drew these profiles on cardboard, cut them out along the line and then set up each profile vertically on a map in the location of the front of each strip. The result simulated a relief map. Just how this method is better than coloring in areas of different ranges of elevation to make a layer map has never been clear to the present writer. Except where the geology of large areas is uniform any profile which does not show the underlying materials is worthless. After all, the nature of the bed rock is the predominant control of surface form and it hardly seems possible that erosion has ever continued undisturbed until all such control was destroyed. Profiles have generally been drawn in order to discover and correlati ancient, now-dissected erosion surfaces. This quest is one fraught with great chance for error. Just why should an erosion surface have a definite level? In recognizing one should we use hilltops or valley bottoms of that time? How much relief should we allow? How should we guard against the subconscious influence of the horizontal lines of the cross section paper? If surfaces were warped during uplift just what kind of curve must we assume? Are not the ranges of elevation of each surface as great or greater than the intervals between the erosion levels? How could older surfaces survive while lower ones were made close at hand unless our usual concepts of origin are mistakes? And finally, are the usual explanations of areas which are too high for a given "level" as either residuals (monadnocks) or as upwarped areas and of areas which seem too low as downwarped areas or due to subsequent erosion just too easy? Are not many of the rotvans or lines of intersection of two peneplains more likely the result of warping? Although we should not discredit all conclusions based on profiles it appears wise to accept may of them "with reservations".

Block diagrams. Block diagrams, perspective drawings, and perspective maps are of more importance in explaining than in solving problems. They are particularly adapted to an exposition of a theory of geomorphic history. The subjective element in drawing them is too great to make them an unbiased delineation of facts. Perspective maps are drawn on a map by raising the position of elevated points an amount proportional to their heigth above datum. The vertical scale used in this is always greatly exaggerated. This type of map has been employed in the several physiographic diagrams which have been published. It helps in giving an illusion of depth to show a cross section of the underground structure along the front edge. True block diagrams are intended to show the relation of underground phenomena to surface form in a certain relatively limited area. Vertical scale is almost always exaggerated. Positions in the horizontal plane may be shown either in true perspective or in isometric projection. In the former all lines which would be parallel on a map converge to a point known as the vanishing point. There is a different vanishing point for each set of parallel lines but all are located on a straight horizon line. In this type of drawing the scale decreases with distance from the front. In isometric drawings lines which are at right angles on the map are skewed to angles of 30 and 60 degrees. Distances along these lines are not altered. Various machines have been made as an aid in making these drawings because the redrawing of a map on a skewed or perspective base is laborious. All of these are hard to construct because sliding joints are required which must be hard to keep in adjustment. Relief is shown by the profiles along the sides aided by shading, hachures, and raised contours. Difficulty in drawing decreases in the order given. Drawing of hachures requires considerable practice. Published diagrams look much more complex than they really are because they were reduced in scale by photography. The illusion of distance is enhanced by making the lines fainter and nearer together for the same slope as the back of the drawing is approached. Crests are generally shown by full lines. Changing of a photography to a perspective drawing is sometimes desirable in order to bring out certain features. The work of others should be studied with a lens before attempting to do any of this kind of work.

Summary. Choice of methods for study and for showing data and interpretations in a report depends largely on experience. The trend is now definitely toward use of aerial photographs including stereo-pairs and stereo-triplets which are less expensive to reproduce than are anaglyphs. Well-drawn three-dimensional diagrams are, however, of great value and are much cheaper to publish than are photographs. The practice of publishing profiles without showing the geology is to be condemned.

GENERAL SUMMARY

In the foregoing pages the writer has attempted to approach geomorphology from the standpoint of what must ensue from given physical conditions rather than by means of purely abstract reasoning. It is true that in nature there are many variations from the ideal of homogeneous material which must be assumed for mathematical analysis. Moreover, there are many topographic forms where more than one process is at work at the same time. These facts must be kept in mind when evaluating the theoretical conclusions. It is obvious that much more field data must be accumulated in order to check the theoretical deductions. It is better to make original observations than to trust to even the best maps.

A major error in the past has undoubtedly been the assumption that the present climate of an area should be projected back into the remote past. Some seem to regard as "normal" the type of humid climate which is now widespread in middle latitudes where the majority of scientific workers live. But if we look at the relative areas of different types of climate today it is apparent that such a classification is unreal. And when we look back to the time when there were no polar ice caps, and epicontinental seas were larger, it is evident that, aside from mountains, an ever larger proportion of the lands must have had a semi-arid or arid climate during most of their history. For this reason caution is necessary in the interpretation of fossillandscapes which were long preserved under a covering and only recently exhumed. Which are true peneplains and which are pediments? Certainly those which have a gravel cover are not peneplains in the classical sense. It seems as if the emphasis that has been placed on the discrimination of remnants of ancient peneplains was, to say the least, unwise. Let us try rather to build up a foundation before we attempt this uncertain pursuit. Let us consider also the alternative interpretaions which are possible for many of the phenomena and not give too much weight simply to reputation of proponents of some of the ideas. The future will decide which interpretaion is correct but only after the necessary physical data has been gathered.

The interpretaions here presented on the discrimination of talus and creep slopes appear well supported. Their application to the slopes of volcano s made of fragmental material obviously eliminates some older attempts at mathematical analysis on the bais of shearing.

But so far the least success has attended attempts at mathematical treatment of the work of running water. It seems certain that the quanitative computations of force made by many students are misleading in that they consider only total energy instead of the small portion which is actually expended in erosion and transportation. Possibly Little's likening of the process to the loss of pressure of water in flowing thorough a pipe may lead the way to ultimate solution. In the meantime more data of grades of streams on uniform material and on slopes due to rain wash is greatly needed. Such data is hard to find and needs careful study. In this connection attention should be directed to the force opposed to erosion, resistance of material to removal. This factor has been ignored in the past. The work of Horton on this, as well as the quanitative relations of streams, is a step toward the solution of some of these important problems.

The present work has added little to problems of the work of standing water other than to point out the desirability of a new classification of shorelines.

The mechanics of the work of wind are, thanks to the careful work of Bagnold, now well in hand. Physical differences in the behavior of wind and water are now clear. Air cannot absorb so much energy in kinetic energy of rotation as does denser water. The problem of the origin of longitudinal or seif dunes is still unsettled.

In considering the work of ice, attention has been directed to the relative importance of plucking versus grinding of bed rock. The work of Demorest on the physics of glaciers points the way to solution of many problems such as the true origin of roche mountonee forms. Drumlins are now recognized as accumulations in cracks shaped to minimize resistance to moving ice. The fact that glacio-fluvial deposits consist of a mixture of sand, pebbles, and ice is also demonstrated.

The value of topography in the discrimination of structure of both sedimentary formations and lava flows has long been known. Smooth dip slopes are also important in gently inclined strata. Hogback ridges where dips are steep are harder to interpret. Overturned folds and repeated thrust fauls offer difficulties.

1-20

Climatic Controls

Introduction. Many of the materials of the crust of the earth are not in equilibrium with their environment at its surface for they were either cooled from a molten state or deposited under water. In contact with water and air they undergo changes in both physical and chemical states. The control of the nature and rapidity of these adjustments lies in the climate. We will here discuss the climates of the earth from this standpoint only, considering precipitation, together with seasonal distribution and its disposal on reaching the ground, winds, and temperatures. We must, however, not regard climate as fixed. Through geologic time changes have been very marked. because of changes in distribution of land and water plus possible changes in latitude of location

Transmission of moisture. All rain and snow which falls on the earth ultimately comes from evaporation of bodies of standing water. or the rise of plants However, its journey through the atmosphere in the form of vapor may be made in two or more stages separated by one or more intervals during which it was precipitated and then evaporated from the ground. The ultimate cause of precipitation is reduction in the vapor-carrying capacity of the air by lowering its temperature. It is a coincidence that the vapor pressure of water in millimeters of mercury at a given temperature is, within the range ordinarily met with, almost exactly the same as the maximum possible number of grams of water in a cubic meter of air. There is no necessary connection between these two quanities. Ordinarily air does not contain all the moisture it can hold at that temperature and the percentage of what it does contain of the maximum possible is termed relative humidity. Now, when air is cooled its relative humidity increases until the amount is 100 percent. The temperature at which this occurs is called the dew point. If cooled below this temperature ~~precipitation~~ condensation occurs. If the temperature is above freezing rain or fog results. If below freezing snow crystallizes directly from the vapor.

Causes of precipitation. Cooling of air which contains moisture is due primarily to ascent for temperature decreases with elevation. Rising of air is due to (a) local heating, (b) winds encountering a mountain range, and (c) winds rising over a mass of colder heavier air. The first process accounts for both local afternoon thunder showers and the tropical rainbelt which is located in the latitude of maximum solar radiation moving with the seasonal change in position of vertical sun rays. Precipitation on mountains is common throughout the world and requires no further comment. At the present the polar regions are perpetually frozen and cold. Cold air which flows south (north in the southern hemisphere) encounters the warmer air of equatorial regions. The contact of northern and southern air is sometimes called the polar front. It is irregular in outline, associated with gigantic swirls termed cyclones. Within these warm, moist air rises above the cold northern air giving rise to cyclonic precipitation, now spoken of as storms along fronts or contacts between air of different source, humidity and temperature. These areas have low atmosphere pressure and are often termed lows. If, as seems probable, in the geologic past the poles were warmer than now the width of the area through which this belt of storms migrates with the seasons was once smaller and the vigor of weather changes within it was much less than now.

Fig 1 Climates and wind belts ✓

Winds. The belt of tropical rainfall is marked by rising currents of air, low barometric pressure, and by absence of surface winds. The air which flows into this belt of calms from farther north and south and, deflected by the rotation of the earth, forms the very constant northeast and southeast trade winds. North and south of the trades descending currents occupy the tropics producing another belt of calms, here associated with relatively high atmospheric pressure. At present about half the area of each hemisphere is occupied by a belt of variable winds along the polar front whose prevailing direction is from the west. This is the belt of westerlies. The larger continents in the belt of west winds afford an exception in that the great seasonal range of temperature in the interiors causes winds to blow inward in summer and outward in winter. These temperature-controlled winds are termed monsoons.

Polar easterly winds

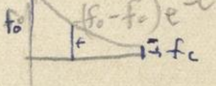
Seasonal distribution of precipitation. In few types of climate is precipitation uniform in amount from month to month of the year. The belt of tropical rains follows the apparent movement of the sun. In equatorial regions this means two rainy seasons in every 12 months, but farther north and south there is only one rainy season and the dry period is much longer and more pronounced. This is the Savanna belt. Seasonal migration of the subtropical calms brings dry weather to regions which at other times are either in the westerlies or the trades. It even causes aridity in the Mississippi Valley whenever the movement is unusually far north. In past geologic times it doubtless caused deserts like the Sahara to extend much farther north than they do now. Relative temperatures of sea and land also have an effect on seasonal changes in rainfall. In many places most rain falls when the land is cold. In regions far from the sea outflowing winter winds plus reduced evaporation spell a winter minimum.

Disposition of precipitation. Rain which falls on the surface of the ground is disposed of by (a) surface runoff into streams, (b) percolation into the ground which may or may not emerge later from springs to join the surface runoff, (c) direct evaporation from soil or free water surfaces, (d) transpiration from vegetation, (e) chemical combination in vegetation and minerals. The relative amounts of each disposition is difficult to ascertain although estimates of the first two are not particularly difficult. Exact determination of total precipitation on a watershed is inexact because of the spotty distribution of individual storms.

V done

Fig 2. changes in rate of infiltration — Horton

Percolation. Snow which melts on frozen ground must nearly all join the surface runoff. The proportion of rain falling on unfrozen ground which percolates into the soil is frequently estimated as a simple percentage of total precipitation. Because water which enters the soil must displace air between the mineral grains and causes physical changes in the soil it is obvious that this is a very inexact method. Experiments by Horton indicated that the initial rate of infiltration decreases rapidly at the beginning of a rain but attains a constant rate after from half an hour to three hours. Fig. 2. He arrived at the empirical formula:



greater than

$$f = f_c + (f_o - f_c)\,e^{-Ct}$$

where f = rate of infiltration at any time, inches per hour, f_o = initial rate, f_c = constant rate, e = 2.718, C = a constant factor, and t = time of rainfall duration in hours. Physical properties of the soil which control entrance of rain comprise (a) texture, (b) structure, (c) vegetation, (d) biologic structures such as burrows, (e) moisture content, and (f) condition of soil if cultivated, sun-cracked, etc. The value of f_c is attained only during heavy and long-continued rains thus increasing surface runoff to cause both floods and surface washing. Infiltration capacity is greatest in loose sands which display a very low proportion of surface runoff. When soil is entirely saturated with water the rate of movement is termed by Horton "transmission capacity". His conclusion was that such saturation is actually attained only in the heaviest clay soils. The relation of infiltration capacity to rainfall is vitally important to geomorphology and more exact data are needed for quanitative study of drainage.

Runoff in relation to rate of rainfall. It is evident that whenever rain falls at a rate less than that of infiltration no runoff is possible. From this it follows that no storm may ever reach the point where there is a surplus above infiltration on the divides. This accounts for the paucity of streams in areas of sandy soil. Rainfall rate generally is highest in brief storms. Various empirical formulae show this. Little used the formula:

$$r = \frac{8}{t^{\frac{1}{4}}}$$

81 min = 8/3 = 2⅔ in/hr.
16 min storm rate = 8/2 = 4 in/hr.
1 min = 8 in/hr.

where r = inches per hour and t is in minutes. Other expressions do not involve exponents.

Return of percolation to streams. Some of the water which percolated into the soil is used by plants. Much is evaporated or transpired by plants and returned to the vapor of the atmosphere. Evaporation losses in eastern U. S. range from about 18 to over 38 inches per year. Correlation with summer temperature is approximate. In the region of low rainfall the lines of equal loss cross lines of temperature at right angles because there is not enough rainfall to supply potential evaporation. Some enters into chemical combinations with minerals. The remainder becomes ground water. Much of the ground water reaches the streams either through definite springs or by seepage into streams.

done
Fig 3 Stream hydrograph

Actual stream flow. The discharge of a stream is determined by measurement of its cross section and mean velocity at several different water levels. A curve is then drawn to show relation between water level and discharge. When discharges are plotted in respect to time a very irregular curve is obtained of which the crests each correspond to a particular storm. The discharge between storms is essentially all ground water runoff provided there is no storage in lakes and swamps. From a study of surface and ground water runoff in the United States it appears that: (a) The total runoff is greatest with the highest precipitation. (b) Ground water runoff is largest in regions of porous bed rock where there is large subsurface storage. (c) Losses due to evaporation are more nearly constant than are other quantities.

channel precip

done
Fig 4 Runoff of a typical stream

Fig 5 Runoff of US USGS Visher

Summary. Climatic control of geomorphic forms is concerned chiefly with causes for variation in amount of runoff, intensity of rainfall, frequency of heavy rainfall, frequency of freezing and thawing, direction and intensity of winds, duration of frozen ground, rather than with the information given on conventional climatic maps. Some of these features will be described more fully in later sections where their bearing is more fully explained. In this section precipitation alone is considered. Water which falls on the ground is disposed of by evaporation, including water transpired by plants, by soaking into the ground beyond the reach of subsequent evaporation, and by direct runoff. Much of the water which enters the earth returns via springs and seepage to form the ground water runoff. Ground water runoff may be determined from the discharge of streams between storms. Its quanitative ratio to surface runoff depends in

Fig 6 Evap losses in US WSP 846

Fig 7 Frequency of heavy rainfalls 6" + Visher.
6" or more

part on climate but to a large extent of the geology of the watershed. Evaporation loss is related both to temperature and to total precipitation. In arid regions it disposes of almost all the precipitation.

Section 2.
Materials of the earth's surface.

Introduction. In describing the materials of the earth's surface for the purpose of accounting for the present topography the method of approach must of necessity be different from that employed in other branches of geology. We must distinguish between consolidated and unconsolidated materials. The former are rocks and owe their firm condition to either (a) the irregular shapes of constituent particles (commonly minerals) or (b) the presence of a cementing compound between the particles. The shapes of particles as well as their physical character determines the mechanical strength of the rock. The chemical composition and size of the particles determine the reaction of the rock to the chemical effects of the atmosphere and water. From the standpoint of geomorpholgy the conventional division of rocks into igneous, sedimentary, and metamorphic is almost meaninglæss. What is important is the relative resistance or durability of rocks to those forces which act upon them when they are at or near the surface. For this reason it is to reports upon building stones that we must turn for information. The ordinary geologic map ignores many of the factors of durability, such as grain size (texture), porosity, permeability, and structure, all of which by controlling both breaking strength and entrance of water are of profound influence on durability. Chemical composition, if shown on a map, is only one factor. Maps which indicate only geologic age are almost worthless for geomorphic studies. Because solid rock is commonly found beneath unconsolidated material it is often referred to as bed rock. Rocks are often divided into two great classed; (a) hard rocks of igneous and metamorphic origin mainly crystalline and (b) soft rocks, mainly sedimentary, although including some kinds of igneous rocks.

Texture. The term texture refers to the size, or range in sizes, of individual particles (commonly minerals) of a rock. It also includes their shape and arrangement. Division of igneous rocks into coarse and fine texture is generally only qualitative and no definite standards have been set up. Fragmental sedimentary rocks are classified by texture. Texture has a marked influence not only on mechanical strength but also on chemical resistance to alteration. Where the constituent particles are large it is evident that failure of a single one either by breaking or chemical change is much more important than in the case of a small particle of a fine-grained rock. Where the constituent minerals have good cleavage, as do the micas and feldspars for instance, parting along those planes of weakness extends farther in a coarse-grained than in a fine grained rock. Such failure of a rock then allows more water to penetrate causing chemical alteration. Available data also appear to indicate that fine-grained massive igneous rocks have a higher crushing strength than do coarse-grained rocks of similar composition. In general the crystalline rocks with interlocking crystals possesses much higher mechanical strength than do fragmental rocks which have been cemented together. In this connection it is well to realize that some sedimentary rocks, such as dolomite, are distinctly crystalline and hence have high crushing strength. The commoner cements are silica, calcite, dolomite, and iron oxides. In the case of cemented rocks the degree of cementation is of the first importance because it affects not only crushing strength but also the entrance of water. Sandstone, as shown on a geologic map, may vary from very well cemented with low porosity and permeability to a rock little more resistant than is loose sand. Quartzite, or sandstone cemented by quartz into a rock so hard that it breaks through the original quartz grains, has very high crushing strength as well as low porosity and permeability. Shales also vary greatly in durability as well as in mechanical strength. Some have been thoroughly compressed or have a cementing substance. Others have a high porosity and low crushing strength. All have low permeability because of the small size of individual openings and are made of minerals which are in large part resistant to chemical alteration. Although much data is available in the literature on porosities of rocks there is little on permeability except in connection with studies of underground water and petroleum, most of which have little bearing upon conditions which exist at the surface. In reference to chemical composition we must be sure to discriminate between true limestones made of calcite and dolomites or magnesian limestones because of their difference in reaction with water.

Structure. Structure refers to the larger features of rocks, the partings which divide them, including the attitude of such planes of division. Sedimentary and volcanic rocks display bedding planes which are the result of interruptions in deposition. Igneous flows are finer-grained at both top and bottom than they are in the middle where cooling was slowest. Gas bubbles are common near the top of a flow and make the rock much weaker than is the rest of the flow. Many metamorphic rocks have the crystals of readily cleavable minerals such as mica and hornblend arranged parallel producing schistosity. Others have bands of different chemical composition producing foliation. Both these factors result in weakness of the rock along definite planes. There are no definite standards of comparison in regard to the distance apart of bedding planes or other planes of weakness. The terms thick-bedded or thin-bedded are very indefinite and many geologic descriptions ignore such information. All rocks are more or less broken by planes which are due to earth movement. In some localities these planes or joints follow a more or less definite pattern in response to the forces which produced them. In other localities they are irregular in direction, inclination, spacing and continuity because caused by settling or cooling. Many lava flows, however, show regular hexagonal columns due to contraction. Standards of comparison between closely spaced jointing and widely spaced joints are wanting and many geologic reports ignore this point. In regions of disturbed sedimentary rocks the inclination of bedding planes, the position and direction of folds and faults is delineated. In making geologic maps the effect of such earth movements on the arrangement of relatively resistant bodies of rock with consequent shaping of the topography is an immense aid.

Unconsolidated deposits or mantle rock. Over most of the earth's surface there is a variable thickness of unconsolidated material above the solid bed rock. For the most part this surficial mantle rock is due to action of the atmosphere (weathering) on the underlying rock. (See Sec. 3) In other localities, such as some glaciated districts and the Coastal Plain of southeastern United States sedimentary deposits have not yet become consolidated. In these localities bed rock lies hundreds or even thousands of feet below the surface. The mantle rock in many places contains fragments of consolidated rocks which range from small granules to boulders of large size. Mantle rocks may be mixtures with a wide range in size of particles or be assorted to a narrow range of grain size; they may be massive (unstratified) or arranged in layers either of the same or of different composition and texture. In some places certain layers have been consolidated into rock. In many localities unconsolidated or semi-consolidated materials are firmest close to the surface which is exposed to the atmosphere. This phenomenon is due to evaporation of ground water leaving a cement and is known as case-hardening. It is of great importance in geomorphology.

Units for description. The lack of quantitative standards of comparison by which spacing of joints and bedding planes and grain size of igneous rocks may be compared has been noted. Porosity is expressed in per cent of voids. It is determinable from the difference between the density of a substance when dry and when fully penetrated by water. Permeability is given in many different units. In the petroleum industry the commonest is the darcy, which is measured in cubic centimeters of water at a given temperature which are forced through a section one centimeter square and one centimeter long by a pressure difference of one atmosphere in one second. Crushing strength is given in either pounds per square inch or in kilograms per square centimeter. Density is for the metric system synonymous with specific gravity, namely a comparison of weight of a specimen in air with that submerged in water where it loses the weight of its volume of water. The following table presents some data which are of interest. Reports on building stones also contain information on results of freezing and high temperatures upon specimens of different kinds of rocks.

Tables of data bearing on durability of materials

Densities and porosities (Birch)

dry range 0.8 to over 2 wet 1.4 to 2.39

		Porosity, percent	Density Dry	Density Wet
Uncondolidated Gumbo soil		54.1	1.19	1.73
Clay	range up to over	40.0-50.0	1.30-1.60	1.80-2.00
Sandy Soil	69% highest	53.2	1.25	1.78
Loess	in silt	20.0-69.4	0.8 -1.6	1.4-1.93
Silt		49.9	1.36	1.86
Sand		30.0-48.0	1.37-1.81	1.85-2.14
Gravel		20.0-37.0	1.36-2.05	1.65-2.39
Soft Rocks				
Sandstone	Range	0.9-38.0	1.60-2.68	1.99-2.73
Shale	up to nearly 45% highest	1.5-44.8	1.56-3.17	1.92-3.21
Limestone	in shale	.9-37.6	1.74-2.72	2.43-2.77
Hard rocks				
Granite	very slight	slight	2.667 give	
Gabbro		"	2.976	
Diabase		"	2.965	
Ultrabasic		"	3.370	

Crushing strengths, kg/cm^2 (Birch)

		Average	below - Range + above
Hard Rocks			
Granite	give range average — to	1480	1110-2310
Gabbro, basalt	about ~~1000~~ 500 for slate to	1800	1340-2900
Gneiss	over ~~3000~~ quartzite	1560	750-1710
Quartzite	3200	2020	1760-1180
Slate		1230	780-1650
(Buckley)			
Rhyolite, Berlin, Wis.		3210	
Granite, fine-grained, Montello, Wis.		3080 give	
Granite, coarse-grained, Pike R., Wis.		1615	

1230
780
450

range from about 100 to 2800

Soft Rocks	Average		- Range +
Limestone	960		900 -2640
Sandstone	740		630-1780
Tuff	310		210-210
Marble	1020		710-1600
Niagara dolomite up to	2800	(Buckley)	
Arkosic sandstone, Wisconsin	340	"	
Quartz-cemented sandstone, Wisconsin	830	"	

Grain sizes of sediments (actual deposits show considerable mixture of sizes)

Materials	U.S. Bureau of Soils mm	Wentworth mm
Boulders		over 256
Cobbles		64-256
Pebbles (gravel)	over 1	4-64
Granules		2-4
Sand	0.5-1	0.0625-2.0
Silt	0.005-.5	0.0039-0.0625
Clay	below 0.005	below 0.0039

Summary. In the study of rocks and other materials of the crust of the earth in relation to topographic forms the qualities considered are those which affect durability at the surface. Some of these properties have been outlined in the tables above but their relation to methods of alteration by weathering is taken up in the following section.

X

Section 3. Weathering

Introduction. The processes of weathering are all directed toward placing the physical and mineral properties of the earth's surface into harmony with their environment. Conditions at the surface are much different from those under which most of the materials originated. Outstanding results of weathering are (a) breaking up of solid rocks into small fragments, (b) chemical alteration, mainly in the direction of reduction of density, (c) chemical combination with water including solution, and (d) the formation of prevailingly simpler compounds which resist further alteration. The processes of weathering include those which are purely mechanical, those which involve chemical change, and those due to the presence of organisms.

Fig 7 8
Frequency of freezing Visher

number of particles increase as cube

Mechanical weathering. Mechanical weathering consists of reduction in size of particles of material without the aid of chemical change. Breaking up of rocks and minerals into small particles is also an accompaniment of chemical alteration. A very striking feature of breaking up of materials is the enormous increase of surface area which results. Areas are proportioned to the ~~cube~~ square of linear dimensions. Thus if we break up a single particle of a given diameter into similar shaped particles of a tenth the linear dimension the surface is increased ~~a thousand fold~~. This rapid rate of increase in surface prepares the way for the agents of chemical weathering. One of the most potent of all purely physical processes which results in breaking up of rocks and other materials is frost. Water which enters into pores, bedding planes, joints, gas bubbles, and other openings near the surface is frozen. In many regions freezing and thawing take place many times during a year. On many mountains it occurs almost every day. Expansion of water when changed into ice is estimated to give a pressure of about 150 tons per square foot or over a ton to the square inch. It is true that this pressure is well below the crushing

$\frac{n^3}{n^2} = n$

wrong — direct ratio

Fig 8 9
Refer from Visher

strength of many rocks but frost does not crush rocks. Instead it breaks them by setting up tension. Tension tests are not included in tables of physical properties although some shearing tests are made. The increase of volume by freezing is about 9 per cent and that of linear dimensions about 3 per cent. Although this seems small the effect is cumulative because when ice melts the water again fills the opening completely. The depth below the surface at which freezing occurs varies widely. In southern latitudes freezing is rare except in mountains. Going poleward the depth of winter frost increases until a normal of several feet is attained in middle latitudes. In the far north as in Siberia and Alaska the ground is permanently frozen to a great depth, locally several hundred feet, and only the surface thaws in summer. This <u>frozen ground</u> probably dates from a time of colder climate probably associated with continental glaciation. A second mechanical process on which most text books lay great stress is expansion of rocks from diurnal or seasonal increase in temperature. The following table gives some data on this subject. The figures are for linear expansion which is very near to a third of the volumetric change. It is well to recall that crystals vary in rate of expansion according to the internal arrangement of the atoms.

permafrost = active layer

Talik = unfrozen layer

Expansion in per cent, from 20 C to 100 C 80° C

Quartz	.08 to .14	according to direction in crystal	Volume .36
Hornblende	.05 to .06	" " " " "	.16
Calcite	.17 to -.05	" " " " "	.08
Orthoclase	.00 to .12	" " " " "	.12
Steel	.09 (for comparison)		

Ratio of linear expansion of rocks to temperature change

Granites and rhyolites	$8 \pm 3 \times 10^{-6}$
Andesites and diorites	7 ± 2
Basalt, gabbro, diabase	5.4 ± 1
Sandstones	10 ± 2
Quartzite	11
Limestone, marble	7 ± 4
Slates	$9 \pm .1 \times 10^{-6}$

Thermal conductivity of common rocks in watts per centimeter per degree C
(multiply by .239 to obtain calories /sec /cm^2/deg.)

Granites	$16 \text{ to } 35 \times 10^{-3}$
Diabase, basalt	14 to 30 "
Gabbro	20 to 30 "

Limestone, marble	14 to 34 x 10^{-3}
Quartzite	39 to 65 "
Sandstone	8 to 42 "
Slate	18 to 28 "
Shale	10 to 17 "
Sand, dry	2.6 (wet up to 23) x 10^{-3}
Clay, dry	2.4 (wet 9 to 16) x 10^{-3}
Snow	2.1 x 10^{-3}
Ice	22.2 x 10^{-3}
Water	5.5 x 10^{-3}
Steel (for comparison)	460 x 10^{-3}

Although the rates of expansion of rocks are less than those of many common metals there are several weak points in the argument that temperature changes do not break rocks. First, rock temperature is often much higher than adjacent air temperatures and is not recorded at weather stations. Second, the low conductivity of rock causes a much more rapid decrease in temperature (steeper temperature gradient) in rocks than in many other materials thus bringing about marked shearing stress not far below the surface. Third, the expansion of rocks is best shown at joints and other openings; if these are far apart a considerable total expansion is caused. Fourth, the differences in coefficient of expansion in different directions in crystals causes marked shearing stresses in them. Fifth, daily repetitions of temperature-induced shear may readily cause failure through fatigue. Experiments on small laboratory pieces of rock are inconclusive because of the limited total expansion and temperature gradient. It may be true, however, that temperature changes do not break up relatively small rocks. Expansion is naturally most potent in regions of large diurnal temperature variation, that is on mountains and in deserts.

Chemical weathering. Chemical weathering is defined as the work of any agent which causes changes in the composition of the molecules; it is certainly incorrect to limit the agencies concerned to purely inorganic processes. Details of the subject of chemical alteration by the atmosphere and by water with associated substances in solution are far too complex for discussion in this connection. What concerns geomorphology is mainly the alteration in physical state brought about by chemical changes. As with the work of temperature changes, including freezing, these agents cause an immense increase in surface area of particles. This results in speeding up the attack of chemical agents. Among the most active and abundant of chemical agents we may list water, oxygen, carbon dioxide, as well as acids derived either from organisms or from the alteration of sulphides. A large part of the chemical reactions of weathering result in minerals which are simpler in chemical compositions than they were before, less in density, and consequently in many cases larger in volume.

Fig 10
Exfoliation dome

Exfoliation. A result of weathering which is of much importance in geomorphology is the breaking off of concentric shells of rock, a process called exfoliation. Once regarded as due simply to temperature changes, perhaps aided by surficial chemical alteration, it is now known that the concentric fractures which are best developed in massive crystalline rocks like granite, extend too far below the surface for such an explanation. Distance between the partings increases with depth as has been noted in many granite quarries. It is treu, however, that the rounding of boulders of crystalline rocks is in part due to chemical attack from both sides of an angular projection. It is now believed that a large part of exfoliation is due to relief of load on the rock because of erosion of overlying material. It also thought the hydration of feldspar, once regarded as due to weathering, is brought about during the crystallization of the rock by the water which is then present. This probably leaves the rock under stress so that fracturing occurs upon lessening of the overlying load. Massive igneous rocks form rounded summits on account of exfoliation. World-famous exfoliation domes are Sugar Loaf in the harbor of Rio de Janeiro, Brazil, Stone Mountain, Georgia, and Half Dome in Yosemite Valley, Calfiornia. The last named has had one side removed by glacial actiom.

Other chemical changes. In general the igneous and metamorphic rocks are more susceptible to chemical alteration than are sedimentary rocks formed from the products of previous chemical weathering. An example is shale, which, although mechanically weak, is made of clay minerals resulting from chemical alteration. Exceptions to this rule are limestones, gypsum, and salt formed from material which was dissolved in water and are therefore relatively soluble under weathering conditions. The last two rarely reach the surface in humid climates. Susceptibility of silicates to weathering increases from quartz,1 through muscovite mica,2 orthoclase feldspar,3 biotite mica,4 alkalie plagioclase,5 hornblends,6 augite calcium plagioclase,7 to olivine.8

Soil formation. The word soil has been used in different ways. Students of soils (pedology) confine its application to the surficial layer, in few places much more than a foot deep, in which plants grow and other organisms thrive. Many of the older geologists, however, applied, and some still apply, the word to the entire unconsolidated material or mantle rock which overlies solid bed rock. They spoke of the "transported soils" of glaciated districts whereas if we mean only the surface layer we must recognize that its origin is essentially the same as in non-glaciated districts, namely alteration in situ of broken up rocks. It is only on floodplains and dunes that we find material which was made into true soil and then moved to another locality. Soil making involves not only the inorganic processes mentioned above but also the work of organisms. Bacteria, moulds, fungi, etc. are very abundant in soils. Among the minerals formed are many which aid in plant growth by the capacity of exchanging bases, calcium for sodium for instance. Bacteria and other organisms do not normally extend far below the surface because of adverse temperature, lack of oxygen, lack of food, and the presence of products made by other organisms which are poisonous to them. Plants possess the power to synthesize new chemical compounds taking the requisite materials from the air, water, and minerals already in the soil. On their death the decay of organic substances produces many chemical reagents which promote further mineral changes. Minerals like quartz which are extremely resistant to alteration form very unfavorable soil for the growth of plants.

Soil profiles. Provided that erosion by wind or water does not remove soil as fast as it forms we find a definite order of layers or horizons beneath the surface. The horizons differ in chemical and physical nature. At the surface alteration from original material is so marked that many of the older geologists thought that they were dealing with transported materials. The succession of layers is known as a soil profile and was first discriminated by Russian scientists. The idea was widely disseminated at first that the surface soil is determined by climate and vegetation rather than by the rock from which it was ultimately derived. Such a view may be correct in some localities but it must be realized that a sandstone which consisted wholly of quartz grains could form nothing but a sandy soil regardless of climate. With soils derived from shale, limestone, or igneous rocks, however, there is more truth to this contention. The surface layer of all soil profiles is for the most part light colored because of removal or concealment by carbon of iron compounds. This is called the A horizon. Where grass is or was abundant as in prairie soils, carbon is so abundant that the color is black. Next below is the B horizon, the densest and darkest in color of the sequence. Accumulation of very fine particles called colloids is greatest where subsoil drainage is poor but the soil is not entirely saturated as it is in swamps and bogs. Where drainage was good the fine particles were either carried elsewhere or were aggregated into larger ones (flocculation). The color of the B horizon is generally red, yellow, or brown, in every case brighter than either overlying or underlying material. With poor drainage deep gray or mottled gray and yellow-brown is characteristic. True all-year swamps show no definite profile, although solution of iron compounds with deposition a few feet below is common. Concretions of oxides of iron and manganese are common in or just below the B horizon. Underlying is the C horizon, described by pedologists as parent material. Geologists, however, recognize that it is part of the decomposed bed rock or original deposits which has been altered almost wholly by inorganic processes. It shows leaching of soluble minerals as well as oxidation and other chemical changes.

Fig 11
Soil profile
Kellogg
[illegible]
bulk density

[illegible] — climate change

Climatic control of soil formation. In relatively humid warm climates the processes of laterization, podzolization, and gleization are common. With less moisture calcification, salinization, solonization, and solidization occur. In the far north under arctic conditions tundra soils are formed which are somewhat similar to the bog soils of lower latitudes, but with less organic matter. Laterization occurs in moist humid climates. Iron and aluminum oxides accumulate under some conditions in volume and purity of usable ores. Silica is dissolved. Transitional toward the cooler zones are red and yellow soils characteristic of southeastern United States. Podzolization occurs in dense forests of the far north where evaporation is slow. The A horizon is robbed of silica, alumina, and iron compounds leaving a very light colored soil known as bleicherde or bleached earth. Such soil is very poor for ordinary crops although it supports trees with deep roots. The iron oxides accumulate in the B horizon forming a hardpan known as ortstein. To the south where forests were more open gray-brown podzolic soils formed. Here the ground was less shaded and was warmer. The B horizon has less iron oxide and breaks with a starch-like fracture into blocks of ½ to 1 inch across. Such soils are decidedly more fertile for ordinary crops than are true podzols. Gleization takes place under poor subsoil drainage conditions forming a sticky, compact, rather light-colored B horizon. Where this was formed from glacial till, sand, or loess, it is called gumbotil, gumbosand, or gumboloess respectively. Some of these soils which were formed before the region was drained by erosion of valleys are now changing into silttil which is characteristic of better drained localities. Salinization, solonization, and solidization are marked by accum-

Fig 12
Soils of US
Kellogg

ulation of alkaline salts in the soil and occur only in semi-arid or arid climates. *Calcification*, the accumulation of calcium salts, is much more important for geomorphology. Soils where calcium carbonate has accumulated near the surface are known as *pedocals* whereas the soils of humid regions which are being leached of that compound are termed *pedalfers*. The very black prairie soils, which are so widespread in central United States, are transitional. They are not abundant in other continents. The absence of trees has long been puzzling. Over wide areas the dense subsoil is known to kill off trees in wet years by excess moisture. Other areas of prairie, however, appear to suffer from drought, for tree growth is confined to valleys, including steep slopes. Owing to the marked climatic oscillations of the interior of North America it is entirely possible that alternations of too much water with periods of too little proved too much for tree growth. At the time of the coming of white men prairies were losing ground to forest and were in part continued by the Indian practice of burning the grass. Throughout the more humid prairie belt the grass vegetation brought calcium to the surface. To the west of the prairies lie the *chernozem* soils with more calcium, greater fertility, and less rainfall. Progressing west toward the true desert lie the *chesnut*, *brown*, and gray *sierozem* soils with progressively decreasing precipitation. In much of western United States calcium carbonate has accumulated in the B horizon or at the surface to such an extent that it conceals the underlying materials and has been mistaken for limestone. Such an accumulation is called *caliche*. Much of the calcium came from the A horizon by descending water but much was brought from below by evaporation of ascending moisture. In more arid districts, like western Australia, crusts of iron, aluminum, and manganese oxides and silica are recorded. Seasonal rainfall appears to be a large factor in the production of all such deposits.

calcrete

silcrete

Structure and texture of soils. Soils are in large part classified by their texture, that is the size of particles as found by mechanical analysis. This is determined in part by parent material and in part by variation of soil-forming processes. Structure, as in rocks, refers to larger features such as plates, crumbs, granules, and prisms composed of aggregates of grains. These are of great importance in erosion for it is the soil which is the first material to be affected by that process.

Depth of mantle rock. The depth to which there is unconsolidated mantle rock over solid bed rock is of great importance in geomorphology. Thickness of accumulated weathered material depends upon the factors which control chemical and physical weathering. In order to have deep disintegration of bed rock it is not only necessary that it be of a type which is readily altered, but that the agents of alteration, notably water and dissolved gases, be able to penetrate the rock to considerable depth. Under the same conditions a quartzite should display very shallow alteration to a rubble of stones and sand, whereas a granite should be softened to a much greater depth by reason of the unstable minerals which are present in it. Schists where the parallel arrangement of the minerals allows ready entry of water should show disintegration to a very considerable depth. Very striking differences in depth of decompositions are observed in glaciated districts where a dike of fine-grained granite may preserve the glacial polish adjacent to a deeply disintegrated coarse-grained variety of the same rock. In accounting for disintegration to depths of scores or even hundreds of feet many geologists have made the error of thinking only of the contest between erosion and weathering. Although it is true that the former removes the products of the latter we must realize that water cannot penetrate deeply into the rock unless there is a force to

Resistance to movement   loss: $\frac{V^2}{R}$ or $\frac{V^2}{D}$ constant Q $Q = V \cdot D$ (unit length) $V = \frac{Q}{D}$ hence constant Q loss $= \frac{1}{D^3}$

Fig 13 Depth of weathering

cause it to move. This force can only be different in pressure or <u>head</u> and is dependent upon the relief of the country. In flat country there is no head to cause deep underground circulation so that thick mantle rock could not originate. Most instances of deeply decomposed rocks are in regions of igneous or metamorphic rocks which have considerable relief. The disintegration must have originated with the present topography and not been inherited from a postulated time of low relief.

<u>Mass movement of unconsolidated materials</u>. The mantle rock or unconsolidated material at the surface of the earth is not everywhere in a stable condition or state of equilibrium. Movement either in large or small units takes place under several different conditions. Some kinds of mantle rock, such as loess ~~loess~~, will stand for a long time in a vertical face provided only that there is not too much moisture and that the highth is not above a certain limit. Examination of natural slopes and the sides of artificial excavations shows that most loose materials soon assume a more or less definite degree of slope known as the <u>angle of repose</u>. This slope is dependent upon the density, size, and shape of the component fragments. In places the material which descended to form this stable slope did so piece by piece. In other conditions large masses moved suddenly; under still other physical controls mass movement was slow. The last is known as <u>creep</u> or where largely aided by freezing and thawing (<u>solifluction</u>) on frozen ground. In many localities particularly on slopes the <u>residuum</u> of weathering rests with relatively abrupt contact upon the bed rock, not the transition which occurs only under comparatively level tracts. In some places certain kinds of bed rock are concealed completely by the weathered product of higher ground. At the contact with bed rock it is common to find that the strata bend down into the material which is in motion down hill.

Fig 14 Creep

<u>Talus slopes</u>. Cliffs of bed rock, formed by any process, are exposed to all kinds of weathering. Fragments of rock whose size is determined by the nature of the bed rock, including its jointing and bedding, fall from time to time and come to rest at a lower elevation. The slope they form is termed a <u>talus slope</u>. If the cliff is near to vertical rocks fall freely and attain a velocity equal to the square root of the product of twice the acceleration of gravity multiplied by the height ($v = \sqrt{2gf}$, where g = gravity and f = distance of fall). A falling stone possesses <u>kinetic energy</u> or stored work measured by one half its mass multiplied by the square of the velocity where mass is weight divided by gravity. By substitution and cancellation it is apparent that kinetic energy = weight multiplied by fall, which is the same as the <u>potential energy</u> before loosening. The stone is brought to rest by the friction of the fragments which have already fallen. Their average size determines the roughness of the slope. Large stones which do not lodge between the others may roll to or beyond the bottom of the slope before coming to rest. This explains why the largest rocks commonly occur at the bottom of a talus. This is not because they fell faster than did the others. Stability is attained when the force of friction just balances the component of the weight of the stone which is parallel to the surface of the talus. Because these forces are nearly the same throughout the entire slope it follows that the angle of slope must be essentially constant although perhaps in some places the bottom may have a somewhat less degree of inclination. In this analysis we have assumed that the rock breaks into fairly large fragments which are not rapidly affected by weathering after fall. In some mountains

$v = \sqrt{2gf}$ $E = \frac{1}{2}mv^2$ $= \frac{w \cdot 2gf}{2}$ $= wt \cdot f$

Fig 15 Forces in talus

Constant slope of wood

5 Afar
4 = 35
3 = 30
2 = 23
1 = 19
constant slope
θ = 19 + 5.5 D'

talus accumulates on top of snow, melting of which destroys the normal even slope. In this way ridges of loose material parallel to the foot of the cliff are formed. Angular materials may form slopes of 50 degrees. The talus of quartzite at Devils Lake, Wisconsin, slopes at about 35 degrees. Gravel comes to rest at about 35 to 40 degrees depending on the amount of sand present. Dry sand will lie at 32 to 38 degrees, but if wet the angle is much less, 22 to 25 degrees. Slopes of clay are unstable at much less angle and few attain as much as 16 degrees. These are for the most part not true talus slopes. Talus must continue to accumulate until the entire cliff is buried provided no debris is removed from the foot of the talus slope. Thus it comes about that the talus does not fill an angle at the foot of a vertical cliff but instead is a relatively thin mantle over a sloping rock surface formed by weathering back of the cliff at the top. In the case of ridges due to the outcrop of a relatively thin vertical resistant layer the talus slopes on opposite sides must eventually meet. If the thickness of the hard rock is constant and the elevation of the bottom of the slopes about the same the result will be a ridge top of very nearly the same elevation. Talus slopes are best developed (a) where the constituent rock is not easily altered by chemical weathering and (b) in arid climates where chemical alteration is slow.

Fig 15
Fig from General Geol.
Plaster experiment
Rheology

Solids and liquids. Before we can consider the phenomena of landslides and creep it is necessary to review some of the physical properties of solids and liquids. A solid is generally thought of as a substance which under the conditions commonly met with at the surface of the earth will retain its shape indefinitely. When subjected to pressure, either of its own weight or an outside force, a solid fails by breaking or fracture. Generally the fractures are inclined at about 45 degrees to the line of application of the force. A liquid is a substance which must be placed in a confining receptacle in order to retain its shape. In this vessel it will assume a level surface if allowed to stand undisturbed. Pressures applied to a liquid are transmitted equally in all directions (Pascals principle). Yielding to pressure is always by flow and not by breaking. In nature no very sharp line can be drawn between these two kinds of substances. The mantle rock which contains in most places a large amount of finely divided soft minerals (clay) is a good example. If dry, a face of moderate height may retain its shape until broken down by weathering. However, if the pressure exceeds a certain amount, or the time is long, or weathering does not confuse the result, a slope may bulge at the bottom by flow. The pressure and time at which such movement is noticable depends largely upon the amount of water which is present. Since the amount of water varies from time to time it follows that the physical behavior is variable. The degree of fluidity of any substance is termed viscosity, which is measured by the amount of shearing stress (force) parallel to the immovable bottom which is required to produce differential movement. The unit is the poise which is the force in dynes applied to a square centimeter to produce a difference of velocity of one centimeter per second at a distance of one centimeter from the base of the fluid. Viscosity depends to a large extent on temperature and decreases with its rise. Water at 20 deg. C has a viscosity of 0.01 poise. It should be noted that viscosity determines not only force required but the time rate of change in shape. If the force and viscosity are known the time required for a given change in shape can be readily computed.

Landslides. In many places a slope either of earth (mantle rock) or bed rock which has previously been stable suddenly yields to the force of gravity.

here: Change of a saturated loose sand to close packing frees water and makes the mass a fluid. Shaking loosens water even in packed sand - causes shear

Fig 17 Fractures in slide

This phenomenon is known to engineers as slope failure. It is generally the result of an unusual amount of water, freezing and thawing, or of excavation by man. Sometimes removal of material from the foot of a slope by natural causes brings about eventual failure. Yielding is in many cases along planes of weakness in the rock such as bedding planes or joints. In unconsolidated material fracture at the top of a slide is generally along a nearly vertical plane which curves outward toward the bottom of the slope. Above this line of breaking the mass settles and slides outward at the foot. In the case of true landslides, motion is accelerated at first and is then brought to a standstill by friction. In the case of the famous rock slide at Frank, Alberta, sufficient velocity was attained so that a considerable portion of the mass ascended some distance up the opposite side of the valley. The topography left by such slides is extremely irregular with many parallel fractures of irregular direction and extent. Cat-steps in loess are of this nature. The slide masses tilt in toward the higher ground leaving many undrained depressions along fractures. This serves to catch rain and lubricate the planes of movement. Although attempts have been made to analize the mathematics of the curve of fracture it is not worth while to follow them here. In the first place most, if not all, are based on unproved assumptions which do not take account of the variation in viscosity from the relatively dry surface down into the wet interior, and in the second place these slopes are not land forms.

Fig 18 Frank Alt. slide

Base failure. Another form of failure in which the entire lower part of the moving mass behaves distinctly like a fluid is known to engineers as base failure. Some of the best known examples are the famous Panama Canal slides. In these the bottom of the canal which had been under 30 feet of water rose overnight into islands. In these instances the drier upper part of the slide was carried along on top of the fluid base. Mathematical analysis appears futile in such circumstances. It seems, probable, however, that detailed study of natural slopes on the same material might have proved of value in determining a safe slope for the sides of the excavation. Equilibrium was attained when the slope component of weight of the mass was equal to the force of friction. Diversion of surface drainage appears also to have aided in drying out the slides. Somewhat similar, but perhaps more rapid, movement of saturated ground occurs during heavy rains in semi-arid regions. These are known as mudflows. Similar flows occur in mountains, including what are known as mud streams, rock glaciers, and earth flows.

Fig 19 Panama slide

Creep slopes. True creep is a very slow motion of viscous material which moves like a liquid even though it may contain many fragments of hard rock. Each layer moves parallel to and faster than that immediately below, although relative change in velocity may in some cases be greatest next to the firm bed rock where most water must accumulate. This type of movement is known as laminar flow. Although freezing and thawing doubtless aid in creep they are not the sole cause for the process which not only extends farther from the surface than does frost action but also occurs where there is no frost at any time. Water must be present to lower the viscosity of the mantle rock. It is obvious that the component of weight parallel to the slope is the motive force for creep. It is also clear that material is added to the mantle rock uniformly in proportion to distance from the top of the slope, here designated by the letter h. Velocity is then proportioned to the sine of the angle of slope which at small angles may be taken as directly proportioned to slope or tangent of angle. Now if velocity is constant, the thickness of moving mantle rock must increase down the slope uniformly and the angle of inclination be constant. But, if as is commonly

Fig 20 propeller & bog plate

equilibrium form not proved

observed, the thickness of mantle rock is essentially constant then, velocity must increase in direct proportion to distance h. Therefore slope must be directly proportioned to h. Total fall at distance h is equal to average slope multiplied by h. Hence by substitution of s:h for slope S, we arrive at the conclusion that f: h^2. The necessary constant of proportionality must depend upon the viscosity of the mantle rock. Actual velocity should follow the law of laminar flow and be proportioned to square of depth times slope, and time required to move a certain proportionate distance would be viscosity divided by product of force times proportionate motion. It follows that the curve of a slope which is due to creep of the mantle rock is an inverted parabola. Convexity of hilltops has long been observed and the explanation proposed above was first given by Gilbert. In order to check the reliability of the conclusion recourse is taken to a well-known mathematical relation. An equation such as given above can also be written:

average slope $\frac{h}{2}$

$s = \frac{df}{dh}$ $\quad f = sh$

$$\log f = \log \text{constant} + n. \log h$$

$df = s\,dh$ $\quad df = s_1 h\,dh$ $\quad f = \int_0^h s_1 h\,dh = \frac{h^2}{2}$

$y = \text{const}\,K^{x}$ is

When plotted on ordinary coordinate paper by looking up the logarithms, or directly on logarithmic coordinates, this equation is that of a straight line. The inclination of this line is proportioned to the value of the exponent of h and the intercept with the line for h = 1 gives the value of the constant. The caution must be observed that most small scale topographic maps are not accurate enough to test this law. It is also necessary to take points not too far from the crest of a divide and not along the ends of spurs where the mantle rock may move in either direction. Something should be known of the geology because marked change in bed rock or variation in thickness of the mantle rock upsets the validity of the law. original form.

Stability of creep slopes. The development of creep slopes is also related to removal of material at a constant rate from the foot of the slope. If the material were not constantly removed stability would be attained and motion brought to a stop. Water erosion must also be minimized either by a cover of sod or a concentrate of stones. Removal of material may be due either to (a) steepening of slope because of a change in bed rock or (b) a stream. The creep slope leading down to a stream which is commonly observed in limestone country must have developed from weathering of the rock after the valley was first made by running water.

Solifluction. The term solifluction was originally applied to creep of mantle rock in subarctic climate where freezing is common and during the summer there is a constant supply of water from melting snow. or frozen ground

Some geologists seem, however, to have used it as a synonym for all creep. In Europe much attention has been devoted to a search for evidence of a past severe climate in regions just south of the glacial boundary, a climate similar to that of subarctic regions today, presumably due to the presence of the nearby continental ice. It is not clear, however, that a climate much different from that of today is required to explain the observed phenomena. Similar efforts have been made in this country. The unglaciated part of the Baraboo quartzite range of southern Wisconsin is almost all covered with a mantle of angular quartzite debris mixed with clay and sand. This originally spread out in low slopes at the foot of the hills. In recent time small streams have eroded this mantle, much of which is evidently residual from a former cover of sandstone and dolomite, concentrating the included boulders into "stone rivers." It is not clear, however, that any of these have crept to a notable extent but the mantle from which they were formed may very well be a relic of periglacial climate.

Stone rings and stripes. A peculiar feature of soils in the severe climates of high latitudes and high altitudes is local concentration of stones into either stripes running down the slopes or rude polygonal networks. These features are commonly ascribed to pushing together of stones by frost action, in part altered by solifluction. Such features are not strictly forms of topography and so will not be further discussed.

Technical terms. English geologists often employ the terms head, warp, trail, and coombe rock for mantle rock in which frost action is inferred as the cause of creep. Solifluction layers are also described.

Solution. Minerals which are commonly weathered by solution comprise carbonates, sulphates, and chlorides of the alkalies and alkaline earths. The commonest of these are calcite, dolomite, gypsum, and halite. Less abundant carbonates are aragonite and magnesite, the former found mainly in fossil shells, the latter in veins. Calcite and dolomite are almost insoluble in pure water but dissolve readily when dissolved carbon dioxide is present; in order of decreasing solubility are aragonite, calcite, dolomite, and magnesite. In calcite-dolomite mixtures the calcium is said to be dissolved at a rate about 24 times that of the magnesium. In the ground waters of dolomitic-limestone regions, however, the excess of calcium over magnesium is not nearly so great as that. Calcite is present in high-calcium limestones, most of which have a very fine texture. These rocks are much more soluble than are dolomites and magnesian limestones. Impurities which affect solubility consist of sulphides, mainly of iron, iron replacing magnesium in dolomite, clay minerals, and chert. If the impurities are disseminated throughout the rock the effect is much less than where they are concentrated in definite strata. If the strata of shale in a limestone are impervious to water they protect the rock below from solution. In horizontal strata a shale layer may act like the roof of a house. Permeable layers of sandstone, however, admit water and retain it, thus increasing solution. Solution is not a simple process. Organic acids undoubtedly aid in solution near to the surface but break down into bicarbonates at depth. The saturation point of the underground water is determined not only by temperature but also by other substances which are present. For instance sulphates appear to reduce solubility of bicarbonates.

Rate of solution. The rate at which the limestones of a given region are being dissolved can be determined from the amount of ground water runoff and the mineralization of the underground waters. Determinations of total solids are preferable to statements of hardness or alkalinity in which the assumption is made that all the dissolved limestone is calcium carbonate. As the sulphates and chlorides as well as the aluminum compounds also came from the limestone this seems fair. Average total solids in the Nashville Basin of Tennessee is about 329 parts per million, Niagara dolomite of Wisconsin 440 p.p.m. and Galena-Platteville dolomites and limestones of Wisconsin 400 p.p.m. Computation based on the conditions in Tennessee with the assumption that the ground water runoff is 9 inches a year, work out at the removal of 214 tons of limestone from every square mile per year, equivalent to less than 1/1000 of a foot thickness. In computing the time required to form a residual mantle rock of given thickness one must not lose sight of the fact that the density of the insoluble material is probably not over half that of the parent material.

Redeposition of dissolved material. Much of the dissolved limestone does not reach the ground water at once. Solutions which enter air-filled openings lost much of their mineral content through evaporation, temperature change and pressure change. The result of deposition is the dripstone (stalactites and stalagmites) which adds so much to the beauty of many caverns. Calcite crystals are thought to have formed in water-filled passages. In considering redeposition we must also consider the factor of mass relations by which small particles are dissolved at the same time that larger ones grow. The final result of redeposition is to fill up openings which are no longer used by underground drainage.

Porosity and permeability. Porosity is defined as the percentage of volume of openings in a given volume of rock and is determined by weighing when wet and when dry. Permeability is the quantity of liquid which can be driven through a unit cube of the rock in unit time under stated pressure and viscosity. (see p. 5). The only calcareous rocks which have appreciable primary porosity are chalk, oolitic limestone, certain crystalline limestones and dolomites, and coral or shell limestone. Secondary porosity consists of openings made after the deposition of the rock. Most of these are cracks or joints although some are along bedding planes. Jointing is due only in part to regional forces; it is mainly the result of induration, drying, compaction, and relief of load of overlying material. Some secondary openings are ascribed to solution between successive periods of deposition (unconformities or disconformities). In general secondary porosity decreases rather rapidly with depth as proved by drilling for fresh water which is ohly rarely found in joints at depths of over 200 feet from the surface.

Underground water circulation. As soon as a limestone emerged from the sea the exposed openings would be filled with fresh water from rain. This fresh water is less dense (proportion roughly 1.0 to 1.03) than is the salt water which
Fig 21
occupied the openings at first. Original salt water is often called connate. The fresh water floats on top of the heavier salt water and extends to such
Fresh on salt water
depth below sea level that the column of lighter water exactly balances that of the denser fluid. Such a condition is illustrated in such localities as Florida and the larger oceanic islands, where the land is high enough to permit such an hydrostatic balance. If the openings do not all communicate with one another, however, conditions must be extremely variable at first with no definite water table. In some places fresh water which is in excess of that required for balance escapes in submarine springs. We must bear in mind that there can be neither extensive penetration of rain water nor underground flow unless sufficient pressure head is available. Water cannot flow through underground passages without the consumption of energy by friction or loss of head. In small openings and with low velocity the flow of water is laminar and lose of head is proportioned to velocity divided by the square of diameter of a cylindrical passage, and velocity is directly related to head. In larger openings flow is turbulent, loss of head is related to square of velocity divided by diameter, and velocity is related to square root of head. For constant volume frictional resistance to laminar flow is inverse to fourth power of diameter, and for turbulent flow to the fifth power of diameter of opening. As a general thing fresh
Fig 22
waters, whose flow is concentrated in the largest and shortest available routes, will not penetrate deeper below the surface than the level of equal pressure.
circulation of water cave forms
It follows that extensive solution at depth must follow upon considerable uplift of the land and that cavern formation cannot take place to an important extent when the land is low and flat. As passages are enlarged by solution the shortest route to the point of outlet is most likely to be favored. Only when

flat

turbulent $V = \sqrt{DS}$ and $V^2 = DS$ $S = \frac{V^2}{D}$ and $V = \frac{Q}{A} = \frac{Q}{D}$ hence $S = \frac{Q^2}{D^3}$ loss inverse to D^3

laminar $V = D^2 S$ $S = \frac{V}{D^2}$ $V = \frac{Q}{D}$ = hence $S = \frac{Q}{D^3}$ same

circular $A = \pi R^2$ hence for turbulent $S = \frac{Q^2}{R^5}$

(use A for D) " laminar $S = \frac{Q}{R^4}$ not important compelled to by lack of large direct openings will the waters take circuitous paths which lead them far below the outlet levels, be these into the sea bottom or into stream-eroded valleys. Erosion of valleys must, therefore, precede extensive cavern formation. Exploration of some of the damsites on limestone along Tennessee River has disclosed indubitable solution openings down to about 100 feet below the lowest known level of the river. Deep wells in limestone regions confirm these observations by the fact that salt water is found below fresh water not far below the level of the valley bottoms.

Fig 23

Topographic effects of solution. In geomorphology the primary interest is topographic effects of solution, not details of cave formation. Rain which falls upon soluble formations can either run off to streams or take a route through the earth to the same outlet. We can think of the ground as a leaky roof through which much of the rain penetrates instead of running down to the gutters. Water which enters the ground soon enlarges the lines of minimum frictional resistance. Thus intersections of two joints are enlarged into circular tubes. The larger openings along bedding planes are also made wider. Solution is aided by mechanical abrasion of particles carried in the moving water just as surface streams erode their beds. As erosion lowers the levels of the main outlet streams bedding planes which are on their level are dissolved into a complex system of caverns whose pattern is controlled by jointing. When the level of the stream reaches a lower permeable bedding plane the upper level is abandoned. These different cavern levels are connected by vertical, or near vertical, openings called wells. In Mammoth Cave, Kentucky, some of these are over 10 feet in diameter and 200 feet high. Water enters into these wells through depressions called sink holes and the sides cave into them. Locally several sinks join one another in a large enclosed depression. Some sinks receive the discharge of a surface stream. Others which have clogged contain ponds. Some are enlarged by caving of the walls. Most sinks are partly blocked up so that caverns are commonly entered via former outlets to streams rather than through sink holes. Many sinks extend up through sandstone which overlies the soluble rock. Among such may be mentioned many near Mammoth Cave, Kentucky, and Mont Lake, near Chattanooga, Tennessee. Some streams have underground channels which they follow during low water with the surface course used only during floods when the cavern is overtaxed. Outcrops of salt and gypsum-bearing formations in humid regions display many sink holes and extensive caving of overlying strata. The breccias of northern Michigan are ascribed to collapse of salt bearing strata in pre-upper Devonian time.

Fig 26 Nat Bridge

Natural bridges. Natural bridges are abundant in some regions of limestone bed rock. A few may be remnants of cavern roofs which collapsed at all other points. More commonly, however, the bridge is due to an underground leak through a narrow spur of a meandering valley. A few may be due to leakage through a waterfall.

Fig 27-28

Endpoint of solution. Solution produces in many places an extremely irregular bed rock surface which is disclosed when the overlying residual material is removed. This is because solution proceeds more rapidly along joints. Theoretically, some areas where the soluble rock is unfissured should survive after the remaining area is dissolved down to a plain. Production of a plane surface is readily possible with weathering by solution because attack is distributed over the entire area with fair regularity. stream level

Technical terms. Most of the technical terms which have been applied to solution topography are of foreign derivation. Such topography as developed on limestone is called <u>karst</u>, possibly either from <u>causse</u> (French from <u>calix</u> or lime) or from the Slavic word <u>kras</u>. Most of the remaining words are from the Serbian because of the abundant literature on the limestone region northeast of the Adriatic. In that country, Jugoslavia, there is little mantle rock; this may be due to purity of the limestone, high relief or possibly seasonal rainfall. Ribbed and fluted surfaces of bare rock are called <u>karren</u>. Chasms along enlarged joints are termed <u>lapies</u> or <u>bogaz</u>. Vertical shafts or wells are variously called <u>ponor</u>, <u>cenotes</u>, <u>jamas</u>, or <u>swallow-holes</u>. Small sink holes are <u>dolines</u>, larger ones <u>blind valleys</u>, <u>bournes</u>, <u>uvalas</u>, or <u>ouvalas</u>. Very large enclosed depressions are <u>polje</u>. Residual hills are <u>hums</u> or <u>pepino hills</u>.

<u>Summary</u>. Under phenomena of weathering not only the processes but the resulting topographic forms have been considered. Exfoliation which results in the rounding of exposed rock masses which are relatively free of joints is ascribed mainly to relief of load because the overlying rock has been eroded although the hydration of feldspar by surface weathering is also a factor. Soils were considered because they play such an important role in water erosion. The various usage of the word "soil" were explained: engineers and some of the older geologists use the term as a synonym for mantle rock. Although climate is a very important control in the formation of true surficial soils parent material cannot be ignored and the material termed by pedologists by that name is by no means free from the effects of inorganic weathering. Topographic forms due to weathering consist of talus slopes, landslide slopes, creep slopes, and karst topography, the last developed upon water-soluble rocks. Talus slopes have a uniform angle because material is retained on them by friction. Creep slopes, if the moving mantle rock is of approximately uniform thickness, are shaped like an inverted parabola. Unaltered creep slopes are found where water erosion is at a minimum. Karst topography is found in humid districts where the underlying rock is readily dissolved by water. They are, like creep slopes, consequent upon prior erosion of stream valleys. It is improbable that solution passages ever extend very far below the bottoms of adjacent valleys. Characteristic features are depressions with underground drainage including lost or sunk rivers which flow at least part of the time underground. Natural bridges are mainly formed by subterranean leaks through rock spurs. Landslides, or relatively sudden mass movements of the mantle rock, cannot always be analyzed mathematically because part of the moving material behaves as a liquid at the same time the remainder acts as a solid. The slide area is left with numerous undrained depressions. A multitude of technical terms has been applied to different features of karst topography, most of which appear not only unnecessary but actually undesirable and confusing.

Wisconsin average of Ca 53.5 ppm $\frac{53.5}{40} = 1.33$ ratio about 85% —
Mg 27 " $\frac{27}{24} = 1.12$

For Niagara only wts $CaCO_3$ 55 $\frac{55}{100} = .55$ or very nearly equivalent
$MgCO_3$ 44 $\frac{44}{84} = .525$

Republican River showed .82 and the portion of the Arkansas through the High Plains .85. The Red River of the South is abnormal in displaying an exponent of only .55. Turning to theoretical reasoning an analysis by Rubey of the forces involved in stream transportation gives a relationship between slope, settling velocity, load, quantity of water, and hydraulic radius according to the formula:

$$S = \left(\frac{\text{Load x settling velocity}}{\text{wt. water x } R^{1/2}} \right)^{2/3}$$

This is equivalent, other things being equal, to the slope being inverse to the cube root of the hydraulic radius. Such a formula appears inapplicable in actual practice. Wooldridge and Morgan give a formula for stream profiles which is a logarithmic curve; it is equivalent to saying that the slope is inverse to distance from the source. Somewhat better results may be secured from Little's approach He assumed that erosive force of unit mass of water is proportioned to square of mean velocity and inverse to hydraulic radius. In a wide shallow stream the latter is equivalent to depth. By the use of Mannings formula for velocity and making necessary substitutions it appears that this force is proportioned to $Q^{1/5} S^{9/10}$ Q = quantity of water in unit width of bed. When solved for S this becomes

$$S : F^{10/9} Q^{-2/9}$$

The slope of constant force would then be $f : Q^{7/9}$. Little derives the equation for a trapezoidal channel, which is somewhat like a natural stream channel, of $F : Q^{3/25} S^{9/10}$ which when solved for slope is: $S : F^{10/9} Q^{-30/225}$. From this the profile of uniform force is $f : Q^{195/225}$ or $F : Q^{.87}$. In order to solve any of these for fall, f, in relation to horizontal distance, h, it is necessary to make some assumptions as to relation of average discharge to distance downstream. This can only be found where drainage basins are of normal shape and where results of long-term discharge measurements are obtainable. On the assumption that $Q : h^{3/4}$ the profile of a trapezoidal channel becomes $f = \text{constant x } h^{9/10}$, which is not far from the actual observations mentioned above. For that matter the result of the other formula for a wide channel becomes with the same assumption $f : h^{7/12}$ or $F : h^{.58}$ which does not agree very well with actual determinations. Were semi-logarithmic platting to yield a straight line then an equation with a variable exponent would be shown. In such an equation the constant which is applied to the exponent represents the percentage of change in each successive interval of horizontal distance. Platting of eight outwash terraces in Wisconsin gave no support to the variable exponent equation but instead yielded a constant exponent of about 0.7 or distinctly lower than that of present-day streams. The explanation of the difference from the existing conditions is undetermined; it might possibly be explained by tilting.

Exam to here 1950, 1958

<u>Slope-wash or overland flow</u>. Where surface runoff does not follow channels but forms a thin sheet all over the land surface the process is called <u>slope-wash</u>, <u>overland flow</u>, <u>unconcentrated wash</u> or <u>sheet flood</u>. This process is important during both (a) initial erosion of a new land surface, and (b) in reduction of sides of valleys.

<u>Hydraulics of overland flow.</u> The two principal students of overland flow are Little and Horton. These authors used quite different lines of approach to the hydraulics of the process. Little's work seems to be wholly theoretical but Horton appears to have tried many actual experiments. These experiments demonstrated that normally overland flow is mixed, although with increase both of distance of flow and of depth it becomes wholly turbulent. On a strip down a slope which is of unit width there is a definite relation between depth, D, discharge, Q, and velocity, V. Since Q = DV substitution shows that for fully turbulent flow

.67 .58 .62 .7 .78 .82 .87 .90

Rewrite this page

$Q : D^{5/3}\ S^{1/2}$; for purely laminar flow $Q : D^2S$. Now if we use the formula for velocity in thin sheets derived by Lewis and Neal then $Q : D^{1.9}\ S^{.7}$. The following development will employ their formula instead of those used by the original authors. It must first be realized that very thin sheets of water may flow in waves; these are probably the result of viscosity. Water piles up until slope is locally enough to overcome viscous resistance. Flow is then accelerated until the sheet thins; then the process starts over again. Horton felt that the waves act like a series of sudden blows and increase erosion.

Little's views. Little, as mentioned above, used the expression which gives loss of head with turbulent flow in pipes to express erosive force, namely $F: V^2{}_m/D$. Substituting $D = Q/V$ this becomes $F = V^3/Q$. Substituting the formula for mixed flow for V this becomes $F : Q^{8/19}\ S^{21/19}$. Solving for S, $S : F^{19/21}\ h^{-8/21}$ and the equation of a slope of uniform force by unit volume becomes $f : h^{13/21}$ or $f : h^{.62}$.

Horton's views. Horton employed the time-honored tractive force equation for force on the bed under a strip of unit width. This equation is also known as the depth-slope formula or DuBoys formula. It is simply the component of weight of water on unit area which is parallel to the surface. Since weight is then proportioned to depth, which is Q/V, it appears that:

$$F : D \times \sin A \text{ (where A is the angle of slope in degrees.)}$$

By substitution for value of D and taking Q : h we find:

$$F : h^{10/19}\ S^{-7/19}\ \sin A$$

This expression is not readily comparable with that of Little unless we assume that for moderate slopes sin A is essentially equivalent to S (tan A). Making this substitution $F : h^{10/19}\ S^{12/19}$. Solving for S, $S : F^{19/12}\ h^{-5/6}$. This yields for a profile of uniform force $f : h^{1/6}$ or $f : h^{.167}$ or a much more concave slope than does the other approach. In making a comparison it is desirable to realize that Horton's formula gives the entire potential energy whereas Little's attempts to determine what part of total force is actually applied to the bed.

Resistance to erosion. Horton computed the force of flow at the point where erosion begins. This was expressed in pounds per foot² and ranges from 0.5 lb/ft² for newly cultivated soil to 0.5 lb/ft² on sod, a range of 10 times. Of course, not all this force is actually expended on the soil for much is lost in internal resistance to flow. The resistance of soil to erosion was ignored by most of the older writers. It depends upon several factors: (a) rate of infiltration, (b) physical nature of the soil, its structure as well as texture, and (c) kind of vegetation. Soils which have the finer particles aggregated into pellets or which swell when wet have a high resistance to washing compared to what would be expected from their mechanical analyses.

Belt of no erosion. One of Horton's major contributions to geomorphology is the reasoning that a certain minimum distance is required below any divide to gather sufficient water to permit the runoff to overcome the resistance of the soil to erosion. In making such computations it is evident that the actual force exerted on the soil is of no importance; what is found is simply the point at which erosion does begin. It is also evident that the width of gathering ground will vary both with rate of rainfall and with nature of the soil. Horton computed that on a 5 degree slope plowed land with rainfall rate of 0.5 in/hr. will not be eroded for 153 feet from the divide, whereas with a rate of 2.0 in/hr the belt of no erosion

Fig 63 from Horton Fig 14

shrinks to only 38 feet. The corresponding results for sod are 7046 feet and 1762 feet respectively. From these figures it is fair to conclude that the belt of no erosion is a fact but that it is of little importance when the resistance of the soil to erosion is low. When resistance is high, however, the importance of the ~~belt of no erosion is low. When resistance is high, however, the importance of~~ the belt of no erosion cannot be exaggerated. In prairie areas it is not difficult to see that valleys do not extend to the divides. Naturally the width of the belt varies inversely with rate of rainfall so that its effect on erosion varies with the frequency of heavy rains.

Profiles developed by slope wash. All the formulas outlined above agree in so far as they indicate that profiles developed by slopewash are concave upward. The concept of the belt of no erosion shows that unless there is no vegetation such profiles do not extend to the divides. The variation of width of that belt may help to explain the convexity of many hilltops although this is not the only explanation of that phenomenon. The checking of which formula is most nearly correct must rest upon platting of actual profiles of slopes in uniform material. Unfortunately, such profiles are not at present available. In many regions the lower slopes have been covered by filling of the valleys or there is too much variation in the underlying materials. When examples are available the values of the constants of proportionality can be determined. In his analysis Little did not assume that quantity is directly proportioned to distance from the divide but instead used a rainfall equation quoted in the first section. By means of rather complex algebraic analysis he eliminated time and obtained as the exponent of a profile of uniform force 9/11, that is a much less concave slope than do the others. Reasoning that through geologic time quantity is directly related to distance the exponent becomes 7/9. Both of these results are for fully turbulent flow. Horton also used slope distances instead of horizontal distances. On moderate and low slopes this would make little difference.

Fig 64 show normal profile
see footnote S.A.
c=1
.82
error
.78

Topography due to slopewash. The mathematical discussion above presupposes that material is removed uniformly from the area of slopewash. Experience shows that such is not the case. Instead, as pointed out by Horton, many parallel rill channels are formed all over the area. Each goes directly down the slope. Doubtless minute differences in resistance is in part responsible for this concentration. Another possible cause is surface tension of the water which would draw it together into threads. Once concentrated, erosion is magnified by the increase in volume. Total amount of erosion also increases with distance down the slope. Horton's figures based on experiments appear to show that increase in erosion is at a more rapid rate than his formulas indicate. Another interesting observation is that in exceptionally heavy rainfall the sod cover may be broken and rolled up leaving bare soil.

Exam to here 1948 — 49-50

Formation of valleys. Formation of valleys on a newly-formed land surface which is steep enough to cause slopewash erosion follows upon the original rills. Certain rills bedome deeper than others. Overflow of the tiny divides causes concentration of water in these larger streams. This process Horton termed cross grading. In time it obliterates most of the original rills. Systems of tributary ravines develop across the original rills. Resulting slopes alter the direction of rills. The process is repeated until the entire area of a drainage basin has valleys so spaced that there is no land outside the normal width of the belt of no erosion as found along every divide. Thus it follows that the development of valleys obeys a definite mathematical law. In nature the boundaries of drainage basins are generally ovoid and the material is not uniform either in infiltration capacity or resistance to erosion.

Fig 65 Horton Figs 22 23
Fig 66
43
43 exam to here

Relation of streams to underground water. Temporary or intermittent streams, which flow only during and shortly after rain or melting of snow, normally lose some of their flow to underground water provided mantle rock and bed rock have sufficient permeability. Permanent streams are supplied in the intervals of no new supply from slopewash or surface runoff by the discharge of springs. It follows that such streams can only exist below the level at which their beds intersect the zone of saturation or water table. Exceptions to this rule are streams supplied by melting snow or the drainage of a humid area which flows through regions of less precipitation. Under these circumstances streams may be losing water to the ground water over a considerable portion of their courses. If underlying material is composed of fine particles causing low permeability this loss is not of very great magnitude.

67 Fig 67 Horton Fig 32

Entrance angles of tributary streams. Since streams normally flow directly down the slope of the land rather than at an angle to it, it is obvious that the angle at which they join one another depends upon the ratio between two regional slopes. If slopes are gentle and both essentially equal the entrance angle might be of almost any value. This condition is found on flat plains. Here the angles range from 60 to 80 degrees except where interfered with by vegetation. In cases where the main valley has a low slope and the sides a moderate inclination the entrance angles are about 60 degrees. If tributaries enter from steep side slopes into a gently sloping main stream the angle approaches 90 degrees. Fig 69

Fig 68 69

Stream orders. Horton devised a system of stream orders which does not agree with that employed by some Europeans. The short unbranched streams next the headwaters (for the most part intermittent) are the first order. Those which receive tributaries of the first order are second order streams. The same process is carried on as long as necessary. In cases of doubt the stream below a junction is prolonged and the stream with the greater angle of entrance in reference to this line is taken as of the lower order. If both branches are close to the same length the shorter one is of the lower order. If both branches are close to the same length the shorter one is of the lower order. The order of a stream is unchanged throughout its length.

70 Fig 70 Horton Fig 7 dendritic

Drainage density. It has long been noted that many drainage basins differ greatly in the number of streams for their area. It has commonly been thought that this fact is related only to climate or to stage of development of the drainage system but Horton urges that the factors of infiltration capacity and width of the belt of no erosion have been neglected. He expresses drainage density by dividing the total length of streams in a basin by its area in square miles. Both permanent and intermittent streams must be included. The reciprocal of twice the drainage density gives approximately the average length of overland flow. Stream frequency is computed by dividing the area of the basin into the total number of streams of each order.

see 70 on

Fig 71 Horton Fig 10 71

Drainage texture. Texture of drainage is simply another word for density. Many investigators have applied the factor of infiltration to account for coarse texture of drainage, that is a basin with relatively few streams. Although this is important, a factor which has been ignored by almost everyone is the relative resistance of both mantle rock and bed rock to erosion. Drainage basins in areas of resistant rocks are characteristically of low density or coarse texture. This can be explained by the fact that it requires a large area to gather enough water to make a valley. Areas underlain by shale or clay almost invariably have fine textured drainage. This factor explains the drainage pattern of Bad Lands. In some areas of deep mantle rock on hard bed rock shallow gullies display a fine texture although the major valleys have coarse texture.

Laws of drainage distribution. Horton worked out the following laws which govern the number of streams of different orders in a drainage basin. The proportion between the number of streams of given order in a basin to those of the next lower order is the bifurcation ratio. The number of streams of different orders in a basin approximates an inverse geometric series in which the first term is unity and the ratio is the bifurcation ratio. The average lengths of streams in a given basin approximates a direct geometric series in which the first term is the average length of streams of the first order. Horton held that these laws, which follow upon the principles of formation of successive tributaries, lend quantitative support to Playfairs Law of accordant stream junctions. He laid great stress on another relation, the ratio between the stream length ratio and the bifurcation ratio as expressing the nature of the drainage. Stream slopes were found to be in inverse geometric series thus relating them to different stream orders and proving Playfairs law. To give a complete quantitative picture of a drainage basin he listed: drainage area, order of main stream, bifurcation ratio, stream length ratio, and either length of main stream or average length of first order streams. From this the drainage density, stream frequency, etc. can be computed. Methods were worked out by which to estimate the length of first order streams where map data are inadequate. The value of all this quantitative data to geomorphology is yet to be demonstrated but its value in making definite comparisons is obvious.

Drainage patterns. Because streams are shown on maps which give no other data much attention has been devoted to drainage patterns. Most textbooks classify these into dendritic or tree-like and rectangular or trellis.[76] Some authors also describe radial drainage such as that of a volcano, and centripetal as the drainage of a basin, where streams meet at a common point. Braided streams[77] divide and reunite repeatedly. Branches which lead water away from a main stream are distributaries.[78] In respect to the use of the term dendritic it is evident that the originators had in mind only the ordinary hardwood trees for all drainage patterns resemble the branching of some variety of tree! As brought out above the drainage pattern reflects relative slopes of tributaries and main streams and this is in general a result of the geologic structure of the underlying bed rock. If we study the variety of land surfaces on which streams originated it is really remarkable that there are not more distinct patterns. Branching or dendritic patterns indicate horizontal uniformity of material. Great irregularity of pattern indicates a primary surface of considerable relief such as that of the rougher phases of glacial drift or of volcanic areas. Only streams which originated on rather gently sloping but not level surfaces were preceded by the slope wash grading postulated by Horton. In many areas such as the more level portions of upraised sea bottom or glacial drift no preliminary slope wash was possible.

Relation of stream courses to geology. In many areas of disturbed rocks it is obvious that most streams are located on the outcrops of the less resistant bed rocks regardless of the position of anticlines or synclines which might conceivably have directed the primary drainage when the area was uplifted. The original or consequent courses have evidently been abandoned in favor of others which are adjusted to the nature of the bed rock and its associated mantle rock. In this process it is clear that streams which happened to be located on less resistant material deepened and widened their valleys to such an extent that in time they obliterated streams which were not so favorably located. The adjusted streams are termed subsequent. In such areas where the rock formations vary greatly in resistance to erosion the drainage pattern becomes rectangular. Tributaries from steep ridges on the outcrop of resistant formations enter the valleys along the strike of nonresistant formations nearly at right angles. However, there are exceptions to this rule. Where a considerable thickness of similar material occurs drainage is dendritic. The subsequent streams in many places cross through the ridges in water gaps, whose origin is in many places disputed.

Valleys due to recession of falls. Falls occur where (a) a stream descends a steep slope formed by another agency, (b) there is a marked change in resistance to erosion along either a near-vertical plane, or (c) soft easily erodible material occurs beneath a firm cap rock or other resistant layer. In many small falls at the head of ravines the capping material is sod or even just B horizon of the soil profile. Water which falls freely reaches a very considerable velocity, far more than is common in streams. However, there is a limit height at which the kinetic energy of the water is effective in erosion. Above a few hundred feet fall the water is broken into drops by air resistance and hence is ineffective. This is well illustrated in the high falls of the Yosemite Valley, California. In lower falls the descending water swirls around boulders and pebbles which aid in excavating a plungepool beneath the falls. Plungepools are normally filled with water but at the abandoned falls of the Columbia Plateau, Washington, several are easily observed. These depressions should not be confused with smaller potholes made by rotary motion of stones in any rapid current. Excellent examples of potholes can be seen in Interstate Park, Taylors Falls, Minnesota. They are very common along all swift streams. As the crest of a falls is worn back by undermining and falling the plungepool moves upstream. A fall normally looses height not only by breaking off of parts of the crest but also by the steep grade required for the stream in the gorge which is formed by its recession. Although excavated to the depth of the bottom of the plungepool, the gorge below is shoaled by coarse debris, in part excavated there and in part fallen from the walls. Lakes occur in abandoned plungepools. In general, however, the bottom of a valley formed by fall recession has less slope than a valley made by the same size stream by downward erosion along its entire length. A good example of a valley thus formed where the falls wore back until entirely obliterated is the upper section of Grand Coulee, Washington. The width of a gorge made by falls during their recession may vary greatly if the discharge or material has not been uniform. A good example is the gorge below Niagara Falls. A wide spot at the Whirlpool is due to intersection of the gorge with an old drift-filled gorge of earlier origin. Energy of fall of unit volume is naturally constant but total amount of work varies with discharge of the stream. This reduction both in power and rate of recession does not in itself leave any record in the form of the gorge. But since there is generally a relation between discharge and width of channel a shrunken stream excavates a narrower gorge than does a larger one. This also is well displayed at Niagara and vitiates its value as a "geological clock" by which early geologists sought to determine the number of years since its formation by dividing the known modern rate of recession into total length of the gorge.

Valleys due to springs. Where a large spring emerges low down on a slope the erosion is somewhat similar to that at the bottom of falls although the amount of available energy may not be anywhere near as great. The elevation of the valley head is fixed by the level at which water can emerge. Valleys of this type have been described from the Columbia Plateau where the waters emerge beneath basalt flows. They are also common in the glacial drift.

Initial formation of consequent valleys. Land surfaces originate in many ways. An area may be the bottom of a sea or lake now emerged from the waves, or it may have been made by stream, glacial, or volcanic deposition. In every case there has been a change from sedimentation in some form to erosion. This change is not necessarily due to uplift or to change in level of sea or lake; it may be due to change of climate or simply to cessation of deposition. Most theoretic reasoning has been started with the premise that a relatively flat sea bottom was upraised rather suddenly so that no significant amount of erosion occured during uplift. Furthermore, it has generally been postulated that the climate was humid and that the initial surface contained more or less irregularities so that the first drainage was imperfect. Good examples of just these conditions may be observed in the lower parts of the Atlantic and Gulf Coastal Plain, as well as on the flatter glacial drifts. Examples of irregular initial surface are found in the rougher areas of glacial drift, in volcanic districts, and in regions where uplift was due to faulting or folding. Two distinct conditions may then be present: (a) the primary surface is so flat, or so permeable, or both, that slopewash cannot occur, or (b) the original surface was altered first by overland wash before valleys were formed. We have already considered Horton's approach to the formation of valleys by successive gradings by slopewash until there is no more area left which can be thus altered. This view is supported by the observed fact that the successive dividing up of a drainage area is displayed in the bifurcation ratio. But where primary slopewash could not occur, as on the glacial plains, the origin of valleys must depend more on chance. Concentration of water would then depend upon local conditions brought about by minor irregularities of the surface. Original ponds and swamps would be abundant in the interstream areas, as is easily observed in glaciated districts. But in either case it is clear that concentrated water does form valleys of consequent streams.

Alteration of valley sides. Were erosion confined entirely to stream beds all valleys would be of the box canyon type with vertical sides. Examples of such valleys are confined to those which were formed not long ago and in which erosion is still concerned mainly with deepening the bed because of rapid flow. Examples are found in many localites where streams diverted by glacial deposition are now making canyons in the bed rock. But very slight reflection shows that vertical valley walls would be extremely unstable. They are altered by erosion of small tributary valleys, as well as by sliding and creep of the sides. Some have thought that all young valleys have convex sides because of the rapid lowering of the bed. This is certainly true in some localities, especially where creep slopes form in incoherent material. In many unconsolidated materials, such as most of the glacial drift, the angle of slope is even like that of a talus slope because it is due to sliding. Rough landslide slopes are also common, particularly where ground water emerges. Reduction of valley sides to a slope naturally exposes them to slopewash. Even if the areas between streams were at first confined to the belt of no erosion it is clear that as valleys were deepened the steep slopes along them would wear back into the formerly immune area. In other words, the slope of stability for given material and climate must start at the stream level and has no relation to the divides. Another factor, which must be considered is that the width of the belt of no erosion is variable because of occasional torrential downpours. This variability of width may easily be a factor in producing a convex divide even in areas where there is no important amount of creep. The belt of no erosion on divides cannot be of constant width until stability is attained along the valley sides by an equality of force of erosion to resistance of material to removal. Once this condition is reached valley formation is essentially complete unless disturbed by earth movement, change of climate, or the work of man.

Fig 81 change of belt of no erosion

Fig 82 convex sides

Base level of streams. A stream valley can be eroded no lower than the bed of the stream into which this valley debouches, nor can any stream valley be eroded more than a slight distance beneath the level of the body of standing water it reaches, or the level of a valley filled by stream deposits in which, in the case of a semi-arid climate, it ends its course. This limitation is known as base-level and its effects have long been appreciated. The accordance of most stream junctions is known as Playfair's Law. There are some exceptions to this law where the main stream is supplied from melting snows of the mountains or some other more constant water supply than the local precipitation which supplies its tributaries. Many examples of hanging valleys which have been unable to keep pace with the deepening of the main river are present in the Grand Canyon of the Colorado. Another cause of discordant junctions is tilting of the land along the direction of the main stream as is well shown in the Sierra Mountains of California. The velocity of the stream which flowed down the tilted surface was increased and that of the tributaries which flowed at right angles was unaffected until the main valley was deepened. Lateral motion of a stream may also cut away the lower end of a tributary. Fig 83

Lateral erosion of streams. Lateral erosion is best developed after a stream valley attaines such a grade that it is able to carry off the debris which is brought to it by both tributaries and slopewash. Development of bends and true meanders then can take place because velocity is reduced to a point where the lateral component of motion is important. Most text books lay much stress on the widening of stream valleys by lateral erosion. Cut banks where the stream swings against the bluffs are common and doubtless account for the observed fact that hills adjoining a large stream are commonly steeper than those along small tributaries. Good examples of this are present along the Upper Mississippi valley and need only in part be accounted for by glacial floods. Such erosion serves to upset the belt of no erosion which was in former equilibrium. But the view that this is the major process of valley widening producing a wide floodplain underlain by bed rock at slight depth is not confirmed by examination of most stream valleys of the United States. Much more common is a considerable amount of stream deposits beneath the valley floor. Such are explicable by change of sea level, or of climate or by the indirect effects of glaciation. Lateral erosion is also limited by the factors which control the width of the meander belt as previously outlined. How far lateral erosion might extent in time is problematical.

Formation of pediments. Special conditions which apparently enhance the importance of lateral stream erosion are present in areas where the amount of water is not enough, or the slope is not sufficient, for the streams to transport their load to the sea. These conditions appear to be most readily attained at the bases of mountains in a semi-arid climate like that of the southwestern part of the Basin and Range province. Here the streams for the most part never reach the sea. Instead they are filling, or have filled, basins between the mountains which were originally made by earth movements. Even above the major areas of deposition, in which there is at times standing water in some places, the streams are obliged by the decrease of grade, aided to a small extent perhaps by evaporation, to lay down the coarser part of the load which they acquired in the mountains. They flow in a braided course on these deposits and build up their beds to such an extent that shifts of channel are of common occurrence. Under these conditions the lateral component of force is alone present. Valleys, where they reach the foot of the mountains, are widened, the ends of spurs, outlying elevations, and even the mountain face itself are cut back by lateral erosion. It has been argued that such lateral erosion is not the major cause of pediments, as the sloping areas of smooth bed rock thinly covered with gravel are called, because so few typical examples of cut banks occur. As a matter of fact, it is true that pediments are best developed on rocks like granite or sandstone which disintegrate into material readily moved by both streams and slopewash. It is, therefore, not to be denied that, as the interstream areas are planed down and weathered down, slopewash takes an increasingly important part in reduction of the area. This was realized long ago by the geologists who happened to witness sheet floods. With the scanty vegetation of semi-arid regions overland wash during occasional downpours is of greater importance in shaping the landscape than might at first be realized. Very slight increase in rainfall would check it by increasing vegetation. Concurrently the increase of rainfall should cause the main streams to erode their beds to lower levels. Just how arid the climatè must be to allow formation of pediments is uncertain and the same remark applies to the possible extent that they might eventually attain.

Cycle of erosion. The fact that erosion progresses through a definite *cycle* was discovered long ago but was first widely publicized by W.M. Davis. His postulate was that uplift is relatively sudden compared with erosion. Erosion thereupon follows a definite pattern of few stream valleys at first, then more thorough dissection, followed in the end (provided no earth movement upset conditions) by reduction of divides to a gently sloping surface called a *peneplain*. A relatively humid climate was assumed in order to carry out this ideal progression. On the other hand Penck suggested that in some cases uplift was much slower than erosion so that the steps outlined above need not follow. Development of these concepts by their proponents was mainly philosophical rather than observational. This is particularly true in respect to the endpoint of erosion. Although many examples can be discovered, for instance in the Coastal Plain and the eroded drift plains, of the progressive development of valley systems with concurrent reduction in surviving areas of the original topography until none survives, no existing examples of peneplains of recent formation have ever been discovered. All that could be pointed out to confirm the validity of the completion of the cycle can be classed as (a) worn-down areas which are inferred to have been uplifted and eroded since peneplaination, (b) buried peneplains now exhumed in part, and (c) flat areas with rock not far below the surface found in semi-arid or seasonal rainfall areas. Some enthusiastic students actually described as young peneplains areas of lake or stream deposits where bed rock lies at considerable depths. It is probable that such errors are in large part explicable by the emphasis placed by some geologists on widespread planation by streams. Although such lateral erosion would certainly be an important factor in completion of a peneplain it is more characteristic of a pediment. The climatic conditions under which many ancient surfaces like the pre-Cambrian peneplain of North America were formed is wholly unknown. Examples of topography in Africa strongly suggest that seasonal rainfall on both sides of the equator may promote pediment formation just as well as does sporadic rainfall on mountains in the Great Basin. Certainly the numerous examples of *monadnocks* with very steep sides (*inselberge* or *island mountains*) appear to suggest formation by lateral erosion of streams whose level was fixed by their own deposits. Only such a process could possibly explain the steepslopes. Further discrimination of peneplains from pediments follows later in this section, as well as a discussion of the identification of remnants of erosion surfaces of different ages in the same district.

Effect of solution on peneplaination. Most discussions of peneplaination have ignored the effects of solution. On water-soluble rocks, such as limestone, this process can work over the entire exposed area at once. It is even effective to some extent under a cover of permeable rock. The result is that, unless disturbed by crustal movement or change of climate, a nearly level surface is formed.

Interruptions of the cycle of erosion. Many students have justly expressed doubt that the theoretical cycle of humid erosion could ever be brought to completion. The sedimentary record does not suggest that the lands ever remained in the same relation to sea level for more than a fraction of the probable time which should be required. Although we cannot now express in years the time which would be required for perfect peneplaination of a mountain range, we are able to measure approximately the duration of the several geologic periods by means of the study of atomic disintegration. The consensus of opinion is that the entire time since the beginning of the Cambrian is not over 500 million years. If we think of the rate of erosion slowing down markedly toward the end of a cycle, when the distance through which rainfall descended to the sea was small, it is not difficult to conclude that known geologic time is too short to permit the completion of so many cycles as have been postulated by some. If the cycles were not as complete as has been believed, then the possible number would be much increased but it is still evident that it cannot be large. For instance, it is thought that the Lower Cretaceous began not more than 120 million years ago. Yet many have thought that prior to the Upper Cretaceous of 95 million years ago, there was not only long deposition of limestone, followed by earth movements, and they by a reasonably perfect peneplain over most of the Atlantic seaboard if not all of eastern North America. The eroded rocks included not only the Lower Cretaceous limestones but also large areas of crystallines. Are we not asking too much of ordinary erosion? It is true that in parts of the Arbuckle Mountains of Oklahoma a subdued surface was eroded across tilted and folded sediments tens of thousands of feet thick, and then buried again within a fraction of the Pennsylvanian period. However, this seems to have been a local and not a regional planation, possibly a pediment formed on rocks not yet completely lithified. In spite of this somewhat startling evidence it does appear likely that most cycles of erosion could never have reached the theoretical endpoint without interruption by earth movement or change of sea level. Uplifts were both those without distortion of the sea bottom and those associated with folding. Besides these diastrophic movements we must reckon with obstruction of drainage by vulcanism, glaciation, and landsliding as well as with changes in climate. When an uplift occurs it follows that erosion with the changed baselevel will largely obliterate all record of the partial cycle before.

Fig 93

Terraces. When a river has stabilized its slope, or reached grade as many term it, a normal feature is the formation of a wide valley floored with debris, which, although in transit down the stream, must perforce be left stranded during the intervals between floods. Such a deposit floors a floodplain and the thickness of material above the bedrock cannot be greater than the usual flood-time depth of the channel. When the baselevel is changed or there is a change in climate either (a) the floodplain is built up with stream debris because the river cannot forward its load any longer, or (b) increase of slope accelerates the rate of erosion entrenching the stream and leaving the former floodplain as a terrace. The method of uplift might be with or without warping or local irregularity or might be a regional uniform tilt. As mentioned above, the last would accelerate the velocity of the stream all along its course at once causing it to form a new longitudinal profile. If an uplift without tilting, or the equivalent a change of the amount of water in the oceans, then the new profile must grow inland gradually. Irregular uplift would be a combination of the above conditions. The inland limit of the new profile has been termed a nickpoint and much attention has been directed to the finding of points of change of profile in a stream. Logarithmic plotting will show at once where these occur but it is not evident which are related to differences of geology and which to the start of a new cycle of erosion. The example on the middle Wisconsin River cited above is clearly due to erosion of the rock barrier at the Dells below. It is very indefinite in the detailed profile, for the change in slope of a stream is generally very gradual.

Aggradation of a valley with debris is the converse process of terrace formation. It may be due to: (a) building of a delta at the mouth of a stream, (b) climatic change, (c) obstruction of a portion of a valley by earth movement or deposition, or (d) glaciation which supplied a tremendous amount of loose material to the stream. In and near to the glaciated regions the valleys which carried glacial meltwaters were filled up to great depths with outwash (glacial sand and gravel). Logarithmic plotting of the profiles of a number of outwash deposits in Wisconsin yielded the equation $f : h^{7/10}$ where fall is in feet and horizontal

See fig 62

distance in miles. The constant of proportionality varies from 9 to 25 inversely to the discharge of the stream. It is evident that with such a concave profile melting back of the ice front automatically changes the slope of the outwash streams at any given point. Readjustment then forms a terrace above the new level. Many streams carried the overflow of large lakes during ice retreat and the increased discharge caused erosion leaving terraces. The grade of many outwash deposits was changed by the melting of included ice masses which had been buried in the deposits. This change formed many terraces. During the erosion of many outwash deposits the streams were in places superimposed across rock ridges. Until eroded away these caused wide valleys above which were later entrenched into terraces. Other outwash terraces were banked against ice which on melting left them high above later streams. Deposition of outwash in the routes of glacial drainage was so rapid that it blocked up the tributary streams which headed in territory not glaciated at that time. The lower parts of these valleys at first held lakes but terracing of the outwash has in almost all places drained these. Deltas were deposited in these temporary lakes and streams which built up their beds to meet the new conditions were locally superimposed on rock spurs. Trout Falls, near Camp McCoy is of this origin. Terraces due to climatic change are perhaps the least well understood. In general, aridity should lead to excessive slopewash which would bring more material to the streams than they could carry away until the slope was increased throughout their length. European geologists think of this taking place in regions near to the continental glaciers because cold decreased vegetation. Return of more precipitation or higher temperature would increase vegetation, check slope erosion, and cause the enlarged streams to seek a new profile. In this case terraces would result. Opinion has varied with different geologists as to whether the numerous terraces and pediment levels in the southwest part of this country were due to change in climate or to uplift. It is probable that study of the profiles by methods here outlined will eventually solve this problem. In regions of folded rocks it seems reasonable to suggest that many terraces are due to stream entrenchment following upon the main stream cutting through a resistant formation.

Terrace topography. Since terraces are remnants of former higher filling in a valley, or a former wide valley adapted to a different condition of erosion, See Fig 94 their borders represent the edge of a new lower floodplain. Where the eroding stream meandered it cut loops into the bank which are known as meander scars. Between these loops the spurs are sharp meander cusps. If, however, the stream was not meandering, or the process of erosion continued for a very long time, such cusps are absent. Terraces formed while the outwash in which they were eroded still contained many ice fragments are now filled with kettle holes and are hard to distinguish from true ice contact terraces where one side rested against stagnant Fig 98 glacial ice. Terraces formed by uplift or change in stream volume normally occur at corresponding elevations on both sides of the valley. Such are called paired terraces. Terraces due simply to lateral erosion of a shifting stream during Fig 99 down cutting are unpaired. Terraces which survived because the valley filling rested on bed rock are often termed rock defended. Terrace surfaces normally show old stream beds which can be distinguished in aerial photographs long after they carried any water. Both braided and meandering patterns may easily be discerned because of the differences in soil in the lower areas of the stream beds.

Correlation of terraces. Correct correlation of paired terraces is difficult. The surfaces were never smooth and since abandonment have been extensively altered by deposits from slope wash and wind work. The best way to match observations at different points is to construct a profile down the center line of the valley. On this correlation is rarely difficult. A further check is logarithmic platting which discloses any miscorrelations at once.

Incised meanders or meandering valleys. It has already been mentioned that some valleys have meandering courses. Two distinct explanations have been advanced to explain this fact. First, it has been suggested that during downcutting the lateral component of erosion caused what were originally minor curves to grow into large meanders. This process would be best developed where the valley walls were not as resistant to erosion as was the bed of the stream, which during low water may have been protected by the bed load dropped after the last flood. Meanders of this type were named ingrown by Rich. Second, a stream which was meandering on a floodplain (not necessarily on a peneplain, as many have supposed) might be uplifted and erosion reinstated. If bed rock was near the surface, and not at considerable depth as on Wisconsin River, the meanders would not be destroyed but would become fixed between rock walls. If these rock valley sides were sufficiently resistant that lateral growth and downstream sweep were alike retarded then the meanders would be eroded into the rock without much change of form. This type was termed intrenched by Rich. As a matter of fact, both types are often found on the same stream, if we can believe topographic maps. The insides of the bends are the criterion by which they may be distinguished. The meanders which increased in size have slipoff slopes commonly veneered with gravel, whereas the other type have practically the same slope on both sides. Intrenched meanders are abundant in the Colorado Plateau where resistant formations of rock lay not far below the ancient floodplain. By restoration of the geology the position of the level prior to uplift may be made with confidence. Such meanders definitely prove uplift of a region and the ingrown type does not, although its possibility is not denied by their evidence. Meandering valleys are almost the sole evidence of change in baselevel of some regions like the Driftless Area.

Here

Stream patterns on floodplains. In most stream valleys there is a floodplain which is occupied only at the highest levels and is in distinct contrast with the normal low water channel. In many floodplains the border of the low water channel is higher than the area behind next to the valley wall. This feature is known as a natural levee and the low area behind is called the back swamp. The pattern of the main stream may be either braided or meandering. Braided patterns where the stream branches and reunites repeatedly are best developed where rapid deposition is taking place. Meandering streams may occur either on floodplains which are being built up or on those that are being eroded. Natural levees due to flood overflow and checking of velocity among the trees of the shore are best developed where the floodplain is being built up, for instance above a delta. Braided streams are universal on the upper part of outwash plains while still forming, in the beds of sandy rivers at low water, below breaks made by floods (crevasses) through natural levees and on deltas and alluvial fans. Within the back swamp the streams have no definite pattern but form an irregular network through the vegetation. Various explanations have been offered for the difference between meandering and braiding. The former is more characteristic of streams which have flowed for some time and hence have organized a definite channel with few islands or towheads. Braiding is apparently an indication of immaturity and rapid deposition. A peculiar feature is the ending of meandering on Mississippi River not far below New Orleans and well below the first distributaries of the delta. In this part of the river differences in water level are not great. Possibly lateral cutting is hindered by the firm clay of the natural levees. Certainly changes in route to the sea have not occurred in historic time. Another feature of floodplains with pronounced natural levees is tributary streams of the Yazoo type where access to the main river is prevented down to a locality where undercutting of the bluff is taking place. Examples of what must certainly have once been this type of stream junction before later erosion appear in the Tennessee and Cumberland near the Ohio, as well as where the Illinois reaches the Mississippi.

Fig 100
AAPG 23-1159
Fig 101
Fig 102 from Russell
See before Fig.
Fig 103
Fig 104

Fig 72 Delta

<u>Deltas</u>. Any stream which discharges into standing water is obliged to deposit its bed load at once. The suspended load may travel far before settling. Fresh water, even when muddy, is lighter than salt water and hence floats. Meltwater fresh from a glacier is less dense than is the underlying water because it is colder. The coarse material dropped at once slides down into <u>foreset beds</u>. Sand and gravel appear to come to rest at a slope of about 25 degrees. This abrupt descent from the nearly level top toward the lake or sea is an excellent diagnostic feature by which ancient deltas now far above the water may be discriminated. The mouth of one of the distributary streams on a delta is commonly shallow because of deposition. The <u>passes</u> of the Mississippi are kept open for navigation by artificial narrowing with jetties. This deepening of the bottom has caused eruptions of mud called <u>mud-lumps</u> which are mechanically similar to the base-failure slides of the Panama Canal.

105 Fig all fan

<u>Alluvial fans</u>. The <u>alluvial fan</u> is the land equivalent of a delta. Change in original slope of the land at the foot of mountains or hills is a common cause of deposition of the bed load which was acquired higher up the streams. Although typically developed in semi-arid districts, alluvial fans can be made in any climate. Many can be so observed filling up kettles in sandy glaciated regions. Streams on fans are braided. Variation in discharge is rapid and great in most regions. Evaporation and seepage into the porous material are often regarded as important factors in deposition, but their quantitative importance is yet to be demonstrated. In California extensive water-spreading works are necessary to increase the soak-in and conserve water which would otherwise reach the sea. Restraint of streams from changing course to places less filled up is difficult but appears to have accomplished in some places by narrowing the channel to one whose competence and capacity are greater than the original braided course. Q:h

$f : Q^{.87}$ $f : h^{.5}$

106 Fig

<u>Profile of alluvial fans</u>. Platting of several alluvial fan slopes east of Los Angeles, California yielded an expression in which $f : h^{.78}$. This is not far different from the equation of glacial outwash plains in Wisconsin. However, Krumbein platted the slope of one pronounced alluvial fan in the same region and derived the equation: elevation $= 2280\, e^{-.12x}$ where elevations are in feet, distances in miles and e is 2.718. Replatting of data by the writer failed to confirm the general applicability of this equation, although it may be correct where the fan is well rounded and the water is constantly spreading out over a larger and larger area. The lower slopes of some volcanic mountains appear to show indices of .35 to .4 where the material is fine and the water from the mountain is spreading out. Alluvial fans are readily confused with rock-floored pediments into which they pass upstream. also in Africa (Natal)

$Q : h^{.46}$ $b : h^{.4}$

107 Fig natural bridge

<u>Natural Bridges</u>. The formation of natural bridges by either cavern collapse or subterranean solution channels through a meander spur has already been discussed. In insoluble rocks cutoffs have taken place both due to leakage along a joint or by lateral erosion of a spur. Some natural bridges can only be classed as freaks of weathering like towers which have not yet fallen into the talus beneath.

108 Fig capture

<u>Drainage modifications</u>. It has already been mentioned that as time goes on in the cycle of erosion streams come to be more closely adjusted to areas where resistance to erosion is least. There is also a progressive relocation of streams in order to secure the shortest, and therefore the steepest, route to the point of discharge. This process requires that certain areas have their drainage outlet changed. The method of change has often been called <u>stream capture</u> or <u>stream piracy</u>. Although indubitable examples of this process have been distinguished few have ever discussed the exact mechanism by which the final capture is effected. It is easy to visualize capture by lateral erosion through a narrow divide changing the point at which a tributary enters the main stream. The Greybull River,

109 Greybull R AJS 231:373

$D : wt^{1/3} \quad wt : D^3$

$D : h^{-3/10} \qquad D^3 : \left(\frac{1}{h^{3/10}}\right)^3$

$wt : h^{-1/10}$

Wyoming, is supposed to have been captured by a smaller river with a lower grade. But when we recognize the validity of the belt of no erosion along divides it is hard to grasp just how the headwaters of one valley could ever wear back into another. One would think that, unless conditions for erosion differ radically on the two sides of the divide, it would prove impossible for a small intermittent stream to ever reach the bed of a large and well established river. Certainly capture of a stream on the other side of a ridge due to a tilted resistant formation appears wellnigh impossible unless aided by subterranean solution or shattered rock along a fault. In every case of recession of the head of a ravine it is obvious that there must be enough gathering ground to furnish water for erosion. If the underlying material on the divide is unconsolidated, or is pervious to water, however, it is easy to see that underground leakage would feed the lower valley long before the actual break-through. Landsliding would also aid in this process, or in some cases the divide might be so low that a flood in the stream above would overflow to the lower course. In early days geologists freely invoked warping of the land as an aid to capture but with no confirmatory evidence. Many supposed instances of capture where no abandoned course of the captured stream could be discovered are of doubtful validity. Peculiar-looking stream courses may readily be due to original irregularities of the surface which directed consequent streams. Similarity of water snails in now separate streams is of doubtful validity because migration may have been with aid of birds.

Superposition. Many stream courses which at first sight appear very peculiar in that they disregard geologic controls are evidently due to initiation of the route on top of unconformable deposits now removed by erosion. This process is known as superposition and streams of this origin may be termed superimposed. Excellent examples can be found of superposition on the Cambrian cover onto the pre-Cambrian, or by the glacial drift onto a bed rock surface. The principle of superposition has now been invoked much more widely than it once was for many of the older geologists seem to have been entirely too conservative in imagining the former extent of now-vanished formations. For instance the course of the Mississippi River on the flanks of both the Wisconsin and Ozark uplifts is much more likely due to superposition than to some more involved process.

Antecedence. Streams which held or nearly held their courses against deformation of the crust beneath them are called antecedent. This process was much invoked in early days to account for structural peculiarities of certain stream courses. Although not by any means impossible, it is clear than in many localities superposition is more probable. In fact many of the older examples, such as the Grand Canyon of the Colorado, are now definitely known to be due to superposition. On the Columbia Plateau, however, the Great Bend of the Columbia appears to be a consequent course along the edge of the basalt flows assumed before the center of the basin sank. In other places the streams of that region cross anticlinal ridges whose rise could have ponded them only temporarily.

Discrimination of peneplains from pediments. Peneplains and pediments have in common the fact that they are worn-down areas of low relief which must at one time in their history have been of much greater relief. It seems unfortunate that the idea of a plain as the endpoint of erosion in humid climate has so widely spread. This assumption demands that either (a) resistance to erosion is negligible, or (b) that geologic time is infinitely long. Because both assumptions are extremely improbable because of the known facts, many students of geomorphology have desired a change in nomenclature. Douglas Johnson suggested change of the name of the final erosional form to peneplane but this was also unfortunate for in geometry the word plane is definitely defined in a way which makes it inapplicable for use for a land form. The word surface is non-commital and we might well, were it not too late to change, substitute the term old surface or endpoint surface of humid erosion.

But everyone still apeaks of "sunset" and "sunrise" although they are obvious misnomers! Another factor, often overlooked, is inconstancy of climate on the earth. When the continents were largely submerged and there were no polar ice caps the climatic belts must surely have been far different than they now are. Rainfall may have been mainly confined to the equatorial belt and to mountains. Deserts may have been far more extensive on the low lands than they now are. Certainly we should not project the existing climate of localities in middle latitudes too far into the past. For this reason it is well to review the known facts to discriminate between peneplains as ordinarily defined and pediments of presumably semi-arid climates and perhaps regions of seasonal rainfall.

Comparison of peneplains and pediments. Definition: a peneplain is the end-point of undisturbed humid climate erosion; a pediment is a sloping area with bed rock near the surface which occurs at the foot of a mountain range. Kind of rock: a peneplain should be formed on almost any kind of rock; a pediment is most rapidly formed where the bed rock breaks down into particles which are readily washed by water, for example coarse-grained granite or sandstone. Climate: although not stressed in original ideas it is clear that a peneplain of the type described by the older writers must be formed in a humid climate; a pediment, judging from existing examples, must be formed in a region where streams do not forward their load to the sea but form alluvial fans. Weathering: the bed rock under a peneplain should display a considerable amount of chemical alteration, although not to great depths as has been incorrectly assumed, because there is not enough head to cause deep underground circulation; a pediment should display bed rock altered mainly by mechanical processes. Topography: peneplains should have very gentle slopes leading down to the sea level of their time of formation and display complete adjustment of drainage to underground structure which determines disposition of the weaker rocks; pediments should have a regional slope which does not everywhere lead to sea level. Extent: peneplains must necessarily be of regional extent, merging gradually into higher land on which long-continued erosion has also left its mark; pediments may be local, passing on one hand to areas where stream deposits accumulate, on the other to the talus slopes of much higher land, and may occur in a stairway of successive levels. Subsequent tilting: because the original surface of a true peneplain must of necessity have been very gentle it follows that any planed-down area with a slope of more than a foot or two to the mile, interpreted as a peneplain, must have undergone subsequent tilting; but an inclined pediment is expectable for such areas have slopes up to many hundred feet per mile when formed. Covering deposits: streams on a peneplain might have wide floodplains but the idea that they necessarily aggrade the old surface on account of excess of disintegrated material is based on erroneous premises as to weathering. The deposits of streams would necessarily be very fine for it takes a velocity of 13 cm/sec to transport with a maximum diameter of 1 mm and about 45 cm/sec to carry pebbles of 10mm diameter, slopes which for small streams with hydraulic radius of one foot demand 0.7 ft/m and 8.2 ft/m respectively, larger streams requiring less slope. Pediments, with slopes in excess of the higher figure quoted, would have a thin coating of coarse gravel. It is probable that the confusion of surfaces of deposition with peneplains which has been general in the past is due to misunderstanding on this point. Age relations: if remains of peneplains could be found adjacent to one another and at different levels the higher must be the older; but with pediments progressive burial of a mountain range with its own debris would assuredly reverse this order of surfaces and the highest might easily, although not necessarily, be the youngest.

Survival of remnants of more than one erosion cycle. As soon as the idea of the cycle of erosion was announced enthusiastic geologists sought to apply the new tool to the interpretation of the geologic history of regions where there is a long gap in the sedimentary record. The result of this was the description of a multitude of erosion cycles, not only resulting in regional peneplaination but also surfaces which left immediately adjacent almost intact remnants of older cycles. Up to 14 such incomplete cycles were reported in one area on the basis of work with topographic maps only. The excellent preservation of some of the old surfaces, as well as their extraordinary number, led inevitably to skepticism and reexamination of the evidence. The following discussion is intended to evaluate this evidence with a view to finding whether or not alternative views were overlooked.

Fig

Criteria of past peneplaination. The best proof of the existence of ancient peneplained areas is the discovery of actual remnants. The strength of this evidence depends directly upon the size and number of remnants which could not have possibly been formed under present conditions. In this connection we must recognize that convex hilltops are not reliable; rather than remnants of an old subdued erosion surface they are more likely creep slopes preserved by the belt of no erosion. Effect of resistant rocks in protecting underlying soft material or of impervious rocks in limiting solution of limestone must also be considered in evaluating these areas of gently sloping topography. In many areas, erosion is conditioned so largely by variations of rock that cross sections drawn without geology are entirely meaningless. Regional or local truncation of tilted or folded formations has often been appealed to as conclusive evidence of peneplaination. Survival of shale within minor synclines of firm sandstone on some ridges of the Appalachaians has often been noted. In every case we must appraise the control by rock character as well as by the character of the mantle rock developed from the underlying materials. In gently inclined formations we should rather ask how it would be possible for one area to be much higher than another regardless of the structure. Would it be possible for part of a cuesta to be much higher than an adjacent portion? If it were it would surely be eroded more rapidly than the lower part. When we consider the elevation of the crest of a ridge on a steeply inclined resistant formation we should ask how it would be possible for one part of the narrow crest to be much higher than an adjacent section. Were there such local high points they should soon be lowered. For a long time the even skyline has been given as proof of regional peneplaination followed by dissection. This evenness was deduced from eye observation and not from surveys. "Distance lends enchantment to the view" is nowhere better exemplified than in this connection. Vertical differences are everywhere so small compared to horizontal distances that a distance of not many miles considerable irregularity in the skyline is invisible unless some local slopes are unusually abrupt. Eye inspection of nearby ground can also be misleading for a slope of over 100 feet per mile cannot be discriminated when there are no level or vertical objects nearby with which to make comparison. Then too, in horizontal views of the skyline distant crests and divides blend together and we tend to forget the deep valleys which lie between them. Viewed on a really accurate map, or in vertical aerial photographs, the true facts are apparent and we see such narrow divides that they could not possibly be remnants of an ancient surface. As for maps, we must at once realize that the topographers, even in recent times, were not allowed enough time to be able to visit ridge tops or even to send their rodmen to climb them. Surveying with a stadia rod is at best very tedious in timbered country and much of the contouring is by sketching. Reference to the older instruction books of the U.S. Geological Survey shows that topographers were not encouraged to climb high hills everywhere but to try to sketch a very large area from lower elevations. The effect of perspective leads to serious errors such as omission of deep valleys. A careless sketcher who works from below often records a flat-topped ridge where in fact the divide is very narrow. It is entirely unsafe to base conclusions as to old erosion surfaces on maps alone unless it is evident that the surveyor actually visited the locality. Long ago Shaler pointed

Fig 112

Fig 113

out that in regions of homogeneous rocks the divides are simply the meeting point of slopes which rise from adjacent stream valleys which are spaced fairly regularly. The average depth of valleys is determined by the slope needed to carry off the run-off with its load of debris. The average depth of valleys is determined by the slope needed to carry off the runoff with its load of debris. The average slope of the valley sides is determined either by resistance of the mantle rock to erosion by slopewash or by the angle at which creep occurs. Naturally the average slope is the same on the two sides of a ridge provided vegetation is about the same and follows that divides should be of roughly accordant level. This condition is well shown in the shale areas of the Appalachian Plateau. When it is discovered that the average elevation of divides is determined by the position of a rock formation, which is more resistant to erosion than these below, the importance to be placed on divide elevations is greatly reduced. Occurrence of regions in which the average divide level is greater than in adjacent areas with a different bed rock, the usual interpretation is that two successive peneplains were formed of which the older survived on more resistant rocks while the lower was being eroded on less resistant material. As an alternative we must certainly consider the obvious fact that areas of more resistant rock, such as sandstone or dolomite, should normally have higher divides than would an adjacent area of shale. In regions of disturbed bed rocks with widely differing resistance to erosion the harder formations almost without exception for ridges whose crests appear to the eye to be remnants of an old erosion surface to whose level all formations were once reduced. But when we ascend these ridges their tops are almost without exception discovered to be so narrow that it is absurd to think of them as being preserved intact. Surely every rock which falls must lower them. These are simply the meeting points of talus slopes whose bases on the weaker formations of the valleys start at something the same elevation. Naturally this makes the crests so nearly equal in elevation that to the eye they appear part of a once-level plain. The forces which produce these ridges work uniformly along the entire length of the flanks and careful surveys disclose a close relation of elevation of crests to geologic structure and thickness of the resistant formation. Could we justly expect any other result from long continued erosion? Need we even assume that once there was an evening-up from which the crests of today were inherited by uniform degradation? Then if we look further, there are lower even-crested ridges in the same area which are formed on formations of somewhat inferior resistance. Do we need to think of each one as a relic of a "partial peneplain" formed by lateral erosion upon the slightly weaker rocks or is their presence an expected result of erosion? Some have made much of the fact that the crests of the highest ridges of the Appalachians are lower near to the places where the major streams cross them in water gaps than they are farther away. But is this too not an expectable result since the level of the subsequent tributary valleys to which the slopes descend lowest there? The rough accordance of summit level in many mountain ranges has been questioned long ago by Daly who suggested that it may be due in part to (a) isostacy which limits the height to which mountains of given material can stand, (b) more rapid erosion above timberline, (c) greater glacial erosion of higher crests, and (d) upper limit of metamorphism or of igneous intrusions. In connection with the last the petrographers can offer testimony and in more than one bathyolith, namely those of Idaho and northern Wisconsin, they tell us that the original top was not far above the present surface or crests.

Use of Profiles. Identification of now almost destroyed "surfaces of erosion" by joining together ridge crests, shoulders, and topos of isolated hills as shown on a profile drawn from maps alone, and without geology shown, is at best extremely questionable. How could these possible be relics of an older topography. Did erosion progress horizontally leaving higher areas intact while lower ones were worn down to baselevel? Why should we search, then, for records of old surfaces on the divides if those divides were able to survive? Perhaps some will now seize on the belt of no erosion as an answer to this contradictory attitude. If so it must be remembered that this belt applies to slopewash only, not to talus formation, landslides, and creep. Horton suggests that some areas where sloping plane surfaces exist on divides may have surviving portions of the surface which was graded by slopewash prior to the formation of the valleys.

Relative speed of channel excavation and reduction of divides. The concept was that stream valleys are widened to produce "local peneplains" and that subsequent erosion caused parallel retreat of the valley sides leaving a broad flat-bottomed valley. This is known as the treppen concept of Penck in Germany and Meyerhoff in this country. These authors presupposed without any confirmatory evidence that the deepening of valleys is very slow in comparison to widening by lateral erosion. Just how this agrees with the known fact that the debris from slope retreat must be carried down the valley and that its slope must be shaped to do this was not stated. Furthermore, it must be realized that parallel retreat of slopes is possible only when the resulting debris is removed from their bases. This is possible through (a) lateral erosion by streams or (b) the formation of pediment-like slopes below talus slopes. It has not yet been proved that either process is adequate to cause this type of slope retreat in humid regions. A possible condition for extensive valley widening is a rock, such as a coarse-grained granite, which weathers rapidly in a humid climate but is resistant to mechanical erosion. Examination of maps fails, however, to demonstrate that there is a gradation from a peneplain near the stream mouths to progressively less and less eroded topography upstream. Conclusions that such a process does occur in nature are in part due to confusion of surfaces of aggradation with peneplains and in part to the example of the Piedmont where the abruptness of the Blue Ridge escarpment tells definitely of some type of structural control. This escarpment is really very youthful for stream captures along it tell of an unstable condition of the divide.

Effect of solubility. An exception to the conditions described above for the formation of peneplains occurs when the bed rock is limestone or other soluble material. Then the processes of weathering may readily serve to destroy the divides and bring about a true peneplain before the surrounding areas have been completely degraded by mechanical processes. Good examples of limestone peneplains are found in the Appalachians.

Buried and ressurected surfaces. Some of the best examples of ancient peneplains are surfaces which have been buried under sedimentary rocks and then uplifted and ressurected from this cover. Examples are the pre-Cambrian peneplain of northern Wisconsin, Canada, and the Grand Canyon, and the pre-Cretaceous Fall Zone peneplain of the Piedmont. These areas of former mountains were eroded under unknown climatic conditions and it is also possible that the streams and seas which buried them caused considerable alteration during the process. Some of these old surfaces display chemical weathering which has often been described as part of an ancient residual soil which escaped erosion during burial. Such an origin seems most unlikely, although it is true that in Wisconsin the surface of the pre-Cambrian is much disintegrated and oxidized even where deeply buried. It seems much more likely that the chemical alteration is due to circulating waters. Waters which descended through formations of different composition might easily undergo base exchange on meeting with the feldspars of the crystallines. The subject will bear considerably more investigation. Some of the once-buried surfaces may not be peneplains or pediments but may be planes of marine erosion as will be considered in the next section.

See before

Summary of evidences. It seems clear to the writer that a very large part of the evidence which has been presented to demonstrate that relics of several erosion cycles are present in limited areas is either questionable or invalid. Survival of such remnants is to be expected in the case of pediments but not with peneplains. Particularly objectionable as evidence are conclusions based on map profiles without showing geology.

Origins of water and wind gaps. Subsequent streams in areas of disturbed sedimentary rocks are readily understood but the places where the main streams cross through ridges due to the outcrop of resistant rock formations are less easy to account for. Four distinct explanations have been advanced by different authors for the water gaps of the Appalachians: (a) rearrangement of streams on a perfect peneplain, (b) antecedence of streams to the folding, (c) stream capture, and (d) superposition of streams on a now-vanished unconformable cover.

Development on a peneplain. The idea that the transverse streams of the Appalachians were inherited from a perfect or super-peneplain on which they (a) "lost their way" or (b) were superimposed on a thin cover of alluvial deposits or (c) were diverted by tilting was once very popular. As a matter of fact it seems theoretically impossible for such a perfect peneplain to exist. The suggestion of a cover of mantle rock changes the hypothesis to the last one, superposition. At the present time it is recognized that if a peneplain were completed on rocks of diverse hardness the adjustment of the streams would increase in perfection during its formation and that they could never disregard the underlying materials. This hypothesis is now obsolete.

Antecedent streams. It has been suggested that some, at least, of the Appalachian streams are still in the approximate locations in which they were when the folding of the rocks occured at the end of the Permian period. The sediments of the Appalachian geosyncline are thought to have been derived from the now almost-vanished content of Appalachia. The original streams should then have flowed northwest across the rising folds. For this reason the suggestion of antecedance could apply only to such rivers as the New, the French Broad, as well as other headwaters of the Tennessee. It is inapplicable to most of the streams which cross the hard rock ridges.

Stream capture. It is clear that stream capture has occurred in relatively recent time along the Blue Ridge escarpment and that it is imminent in several localities. The northern streams like the Potomac and Susquehanna, which flow direct to the Atlantic, certainly have the advantage of a steeper slope than has the Tennessee or Kanawha. But when we consider the difficulty of a small stream working back through a thick formation of resistant rock to reach a larger river on soft rock on the other side the process appears impossible. Conditions are different than in the southern Blue Ridge where the rocks are reasonably uniform. The hypothesis of capture is workable only where there is a cross fault which shattered the rock of the ridge. As most water gaps do not display any offset of the ridge it would be necessary to assume in every case a fault in which movement was parallel to the dip. If there were faulting at gaps, landsliding and underground leakage of water might cause capture for then there would be no necessity for a watershed to supply a stream which would cut back through the ridge. Geologists differ greatly in conclusions as to the field evidence of faults in water gaps. Some declare that they are almost universal and others deny that there are more than a very few. Certainly many gaps in successive ridges do not line up as they should. It seems likely that since the gaps afford the best and most readily accessible exposures they have been more visited than other parts of the mountains and yet few have ever been mapped in detail as is done in oil-producing regions. The theory of capture appears rather unlikely as a general cause of gaps in the Appalachians although it may be workable in some other regions.

Superposition. It seems strange that the theory of superposition was so long neglected in the Appalachians. It has been definitely proved in the Colorado Plateau and for the localities where the Colorado crosses some of the Basin Ranges, although it will not account for the major rivers of the Columbia Plateau. The Coastal Plain sediments are not far distant from the Appalachians and seem the logical answer to their water-gap problem. However, it must be realized that it was a long time since the cover was present and that it probably rested on a surface of unknown but relatively low relief which lay well above the present ridge tops. Since the erosion of the superimposing deposits there has been time enough to cause extensive formation of subsequent valleys along the strike of non-resistant formations. First advocated by Johnson, there have been many objections to its general application. It has been pointed out that many gaps are located where the hard formations are unusually thin or make abrupt bends. However, it seems by all means the best suggestion, provided the long time since erosion of the soft covering formations is realized. The underlying ridges, which are about 100 feet high where they disappear under the Coastal Plain of Alabama, may have exerted some influence on the form of the original surface just as differential settling of glacial drift is thought to have caused reexcavation of some valleys which had been once completely filled. It must also be realized that progressive erosion into a mountain mass must certainly uncover vertical differences in structure. This is particularly true where thrust faults are present. Streams adjusted at one level are out of harmony with the formation when they have cut deeper.

Fig 19 Subsequent valley and watergap

Wind gaps. Wind gaps are similar to water gaps but no longer have any stream in them. It has been suggested that some wind gaps were actually due to the meeting of the heads of ravines on opposite sides of the ridge. In answer to this, such ravines do not have sufficient watershed on a narrow ridge, although they might occur where aided by fractured rock along a cross fault. Many have suggested that with gap elevations, some of which have been altered by accumulation, of talus since abandonement, record former erosion levels. The idea was that uplift of the old partial peneplain caused rapid diversion of streams leaving the gaps. However, study in Pennsylvania does not support this theory very well. During the process of erosion the number of water gaps has steadily declined in favor of windgaps thus giving no support to their original origin by stream capture.

General Summary. The subject of the work of running water is very complex. Agreement has not been reached on the mathematical relations of erosion and transportation to energy of the water. On the whole, the line of approach used by Little seems to offer the best possibilities for the computation of profiles of uniform force, both for streams and for slope wash. Use of a different formula for relation of velocity to depth and slope is advisable in the case of the latter. The reasons for variation of channel width and width of the meander belt are explained from the standpoint of hydraulics and appear satisfactory. Initiation of valleys may be either consequent on original irregularities of a new surface or may follow on primary slopewash grading as outlined by Horton. This author's discoveries of a mathematical relation of relative numbers of streams of different types (orders) appears to support his contentions. The belt along divides in which not enough water is gathered to permit erosion by slopewash cannot be neglected, nor can be the factors which alter it in time. Degradation of divides is ascribed more to creep than to slopewash because of this belt of no erosion which survives after stabilization of slopes to the point where resistance to erosion equals available force of water. The endpoint of erosion under humid climate is certainly not the peneplain of classic literature if there is vegetation to cause resistance. The formation of pediments as distinguished from peneplains is discussed with the conclusion that they are best developed in semi-arid climate and are formed chiefly by lateral erosion. They are wash-slopes and contrast sharply with the talus slopes of adjacent higher areas. Pediments can be found forming a "stairway" of successive levels but similar relations are impossible with peneplains. However, the climatic conditions under which many low-relief surfaces originated is unknown. It is pointed out that many criteria for the discrimination of ancient dissected peneplains are invalid. Superposition appears to be the best explanation of many water gaps and wind gaps are due to diversion of streams which once crossed ridges of resistant rocks.

last 118

WORK OF STANDING WATER Sec. 8

Introduction. The work of standing water is divisible into: (a) mechanical processes, which include both the work of waves and of currents, (b) work of organisms, and (c) the problem of submarine valleys. It is evident that this grouping includes some subjects which are not strictly under the general heading but which appear more suitable for discussion at this place than in any other.

Mechanical processes.

Origin and mechanics of waves. Wind which blows over water is retarded at the bottom by friction just as it is on land. Velocity decreases downward and probably is 0 at and near actual contact with the water. This vertical velocity gradient is in response to the transfer of energy from the moving air to the underlying water. In general, however, it does not itself cause the water to flow as a current. Instead it sets up rotational motion of water particles around horizontal axes which are at right angles to the direction of the wind. When wind first begins to blow the radius of rotation of each particle is small but as time goes on a limit appears to be attained. Each successive rotating particle is slightly out of step with the last so that the final result is a wave of oscillation. The mathematical form of such waves is that traced by a point on a radius of a rolling circle. Of course, no real circle does roll, only the particles and the radius of the hypothetical circle is larger than the actual orbit of water particles at the surface. Even after the wind has stopped blowing the wave progresses. It is then smooth and much simpler than when it is crinckled by a rising wind. At maximum size, waves are roughly half as high in feet as the velocity of the wind in statute miles per hour as ordinarily measured not far above the surface. The velocity with which particles rotate in their orbits determines the speed with which the wave progresses. Since particles revolve in a circle ~~either~~ the vertical component of motion obeys the laws of harmonic motion. If we let the radius of the rolling circle be R feet and the length of a wave from crest to crest be L feet then $L = 2\pi R$. The acceleration is that of gravity, g. Applying the formulas for harmonic motion and solving for velocity in feet / second: $V = (gR)^{\frac{1}{2}}$. Substituting the value of R in terms of L, $V = (gL/2\pi)^{\frac{1}{2}}$ or substituting numerical values for the constants, $V = (5.123\ L)^{\frac{1}{2}}$. Solving this for length of a period, T seconds, then $L = 5.123\ T^2$. The relation of the height of a wave above the trough, h, to wave length, L, is not fixed but appears to vary from 19 to 39 times. The relationship of h to fetch or distance that the wind blows over open water is reported empirically as $h = 1.3\ \text{fetch}^{\frac{1}{2}}$ where the latter is measured in statute miles. (Mariners use nautical miles or knots each one of which is about 1.15 land or statute miles; they also measure depths in fathoms of 6 feet.) One reason for irregular results in measuring wave heights is the fact that a violent wind blows off the wave crests in whitecaps. The smooth waves which last after the wind, or extend outside of the area where it blows, are often called ground swell. In the open sea the distance through which winds blow sets a limit to height of waves of about 50 feet.

L=200 h=12 $\frac{64 \times 144}{8} = 1152$

Energy of waves. Since wave motion in open deep water is a combination of rotational motion and vertical motion waves possess two types of energy. Mathematical analysis demonstrates that the two kinds are equal in amount. Only the energy of rise and fall is carried forward with the progress~~s~~ of each wave but when a wave dashes onto a shore and the water is brought to a standstill both types must be expended. In fact, in shallow water waves become translational motion. Engineers have tried several forms of dynameters with which to measure either impact or both impact and pressure of waves where they strike the shore. Results of these experiments are generally given in pounds per foot^2. Now the total theoretical energy of a wave in foot pounds (other dimensions in feet) is s shown by the formula Energy = $WLh^2/8\ (1 - \pi^2 h^2/2L^2)$ or substituting 64 for W, weight of a cubic foot of salt water, and for pi, this becomes $E = 8\ Lh^2$ $(1 - 4.935h^2 / L^2)$. For L=200, h=12 this is said to show an impact of 2436 pounds/ft^2. Actual records are of this order of magnitude.

W = 8x200x144 = 230,400

Depth of wave action. The circular orbits of water particles are shown by both mathematical analysis and actual observation to decrease in radius very rapidly with depth. At a depth equal to the wave length the orbit is reduced to 1/512th. Thus one of 10 feet radius at the surface in a 400 foot wave would be only .2 inch radius at a depth of 500 feet. It is evident that although there is no theoretical limit to wave action its practical importance decreases rapidly with depth, that is its competence to disturb the bottom. For this reason pendulum observations are possible in a submarine at comparatively modest depths. The term wave base has been applied to the effective maximum depth at which waves can disturb a sand bottom. see 121A

Waves in shallow water. As waves of oscillation reach shallow water the circular orbits are believed to be changed to ellipses with the longer axes parallel to the bottom. Certain it is that the top of a wave moves forward bodily in an entirely different way that it does in deep water. The wave is retarded at the bottom and the top breaks into a confused mass of foam which rushes up a gently sloping beach until it comes to a standstill. No definite mathematical relationship has been discovered which shows the depth at which waves break. This is probably due to the fact that the bottom water is moving either as undertow, due to return of water from the beach, or as a current induced by the tides. In shallow water waves undergo refraction just as do light waves in passing from one medium to another. This is due to bottom retardation and turns the wave fronts until they are parallel to the shore. Many diagrams have been shown to demonstrate that this process also tends to concentrate waves onto headlands.

Fig 121 Fig 122 Fig 123

Other waves. Waves may also be due to (a) earthquakes, (b) tides, and (c) differences in atmospheric pressure. Of these only the first is important in most places. An earthquake moves a large body of water by impact producing a true wave of translation. Such waves are often called tidal but this is a misnomer. They are also called by the Japanese term tsunami. Some of these very destructive waves are known to have travelled 900 m.p.h. and to have risen over 100 feet onto the land.

Effect of waves on the shore. Waves reaching a shore must apply nearly all of their energy to it. Waves work much more constantly than does running water on land. The ocean is never still; even when smooth as glass there is still a surf from the swells of distant or past winds. In appraising its efficiency, however, it is well to realize that rocks moved by wave action are generally submerged and hence loose the weight of their volume of water. Many spectacular instances of large masses moved by wave action are also to be discounted in that the weight was not lifted but simply shifted against friction. Nevertheless, storm waves have been observed to hurl rocks high into the air and the noise of moving boulders in the surf is impressive. Repeated blows of waves do attain high pressures and may easily be associated with cavitation although this has not been recorded. Impact of stones carried in the water on stationary objects is a most important process of erosion. The result not only of direct impact but also of grinding by material moved in the breakers is to undercut the shore if it is a cliff and the depth of water is great enough to permit waves to reach the shore effectively. The process of undercutting is possible because the loosed debris is carried back by the undertow.

Currents in lakes and seas. By no means all erosion and transportation in lakes and seas is accomplished by waves. Currents of water are caused by (a) tides, (b) winds either directly or through waves, (c) density differences due either to variation in amount of sediment or in salinity, (d) entrance of fresh water, and (e) temperature differences. Of these tidal currents locally attain very high velocity on coasts where the difference of level is large. The great oceanic circulation owes its origin ultimately to the convergence of the trade winds in equatorial regions. Although much attention has been given to currents in sedimentation it is obvious that most of them are either surficial or have such low velocity that they do not disturb the bottom. In shallow water, currents along the shore are often noticed. They may be undertow which is deflected by its relation to incoming breaking waves above. Many text books describe an alongshore current set up by waves to which much importance in transportation is ascribed. It is not clear, however, that such a current is actually able to transport sand and pebbles. It can transport fine sediment but movement of coarser material is almost wholly confined to the zone of breakers.

Fig 125

Erosion by waves. It has already been mentioned that waves apply their energy to erosion in a zone of very limited vertical extent. The result is undermining of a shore. If the material is bed rock this produces sea caves especially in weak, thin bedded layers. Where rocks are jointed, deep coves are excavated. Stacks and islands of rock survive for a time rendering the shore very irregular. Shores of bed rock can almost everywhere be identified on a chart in this way. Where erosion is taking place in mantle rock the cliff is not vertical or overhanging because it soon slides down to the angle of repose. If large boulders are present they accumulate at the water's edge and serve to prevent further erosion. Many abandoned beaches may be recognized only by such boulder lines. A common feature of wave-eroded coasts is hanging valleys whose lower parts were cut away by the waves. The headlands undergo the most erosion because water is generally deeper off them than in bays as well as because of wave reflection toward them.

Fig 126

Wave terraces. Below water level the bottom is cut down to the point that the undertow can just carry off the debris which is not moved along the shore by waves which reach the coast at an angle. This subaqueous feature is called a cut terrace and may be distinguished off many abandoned shore lines. There is commonly a slight building up of the front where material carried out across is slid down into deeper water, but no important deposits of gravel occur in this situation. The cross section of a cut and built terrace is known as the profile of equilibrium and the level of its outer edge was thought to be fixed by effective wave base. It is not clear, however, that this position is long stationary for there is no exactly definable lower limit to wave work and the undertow is aided by gravity.

127 Fig

Subaqueous ridges. A very common feature of sandy bottoms is a succession of several parallel submerged sand ridges. Some of these are many hundreds of feet

Fig 128
129 after Evans

wide and the depth of water between them is a number of feet more than on the crests. Some call them <u>low and ball</u>. The problem of origin is unsettled. Some think of them as due to breaking of waves which reach the shore at right angles and others ascribe them to erosion of parallel currents which are induced by waves striking the shore at an angle. The second explanation is weak in that it would then be difficult to account for the number of ridges or for the fact that the sand is coarsest on the crests and not in the low places. Where currents do occur it is more likely that they are the result of the ridges and not their cause.

<u>Alongshore transportation</u>. The fact that material which the waves can carry is moved along shore has already been mentioned. On any sandy beach it can easily be seen that waves which come in at an angle carry sand and pebbles diagonally up onto the beach. The undertow runs directly back down the slope so that particles move in a zig-zag course in the direction which is down wind at that time. Johnson objects that the path of a particle is actually a series of inclined parabolas but the distinction is unimportant. Alongshore drift consists in many places of material which is so coarse that it is impossible to think of currents which could transport it. See fig 125

<u>Depositional changes of the shoreline</u>. Waves tend to even up a shoreline by developing a smooth outline. The material eroded from the headlands is mainly disposed of by movement in the zone of the breakers. Large stones obey the impact law and are carried shoreward by waves which are more powerful than is the returning undertow, which transports the finer particles. Many stones which have been carried back and forth on a beach for a long time show the effect of waves , and are tabular due to shuffling instead of turning end over end as in a current. If a barrier or solid pier is built off a beach experience shows that one side is filled in and the other eroded unless storms come from both sides in equal numbers. The windward side is the one which receives sediments. The same process may be seen at a natural point. Material carried laterally from the end of a point does not form a <u>spit</u> into deep water for that would, where above water at all, be swept away by the next storm from the other side. Instead, the debris is carried along in the breakers to a point where the bay is shallow enough to permit start of deposition. In most places this shoaling is associated with a minor point although this is not necessary. Textbooks ordinarily describe this process of bridging the bays with wave-transported sediments as due to the outward course of the alongshore current but the process outlined above appears more logical. As spits are built out from both sides the bay is eventually enclosed. Where, however, shoal water is present only near the sides of a bay the outer end of a spit is curved back into a <u>hook</u>. This is due primarily to refraction of waves and not to deflection of currents. In many places islands have been joined to the mainland, or to one another, by such beach deposits often called <u>tombolos</u>.

Fig 130

Fig 131

Many hooks are compound showing several successive ends as the entire deposit was built out into deeper and deeper water. In the lee of many islands two spits join leaving a <u>lagoon</u> inside. Lagoons are also formed by the bridging off of bays with continuous <u>bars</u>. Where streams entering the lagoon are large enough, tidal difference is enough a pass is kept open to the sea.

Fig 132

Fig 133

<u>Offshore bars or barriers</u>. On low, sandy coasts where the water is very shallow there are sandy ridges some distance offshore. Opinion has varied as to whether or not these were made by lateral growth of spits from distant headlands or were thrown up in place by waves which removed the material from the adjacent bottom. Some of these off Cape Hatteras are hard to account for by lateral growth alone, although in most places this process cannot be eliminated entirely. These sandy <u>barriers</u> are often miscalled <u>reefs</u> which term generally refers to a submerged rock outcrop.

Fig 134

Fig 135

Cuspate points. In many places, such as Capes Hatteras, Fear, and Canaveral, of eastern United States, sand beaches form marked cusps. Various explanations have been offered including eddies of ocean currents. It is more likely that they are barriers built out to submerged shallows from both sides. See also Fig 135

Classification of shore lines. Many texts have defined schemes of classification of shore lines into emergent and submergent. Others have suggested primary and secondary as the major basis of separation. As a matter of observation, it is clear that all coasts of the world display phenomena of either subsidence of the land or rise of the ocean level in the form of drowned river valleys. The only exceptions are very recent shores due to organic growth or crustal movement. Shepard terms primary those shores which are due to land forces, erosion, deltas, land vegetation, glacial, and volcanic processes. He calls coasts due to wave erosion and deposition, including that of marine organisms, secondary. Another system of classification would be to discriminate shores on firm materials from those on loose deposits easily moved by waves, placing those due to organisms in a third category.

Cycle of shoreline development. The foregoing section demonstrates why it is futile to search for evidences of a cycle of shoreline development at the present time. All marine shores and inland lake shores are demonstrably young. True, progress has gone much farther in the same number of years where the material of the coast can be easily moved by waves and currents, as on the sandy Atlantic Coastal Plain, than it has on the "stern and rock-bound coast" of New England or the fiorded coast of Norway. The evident goal of shoreline development is as simple an outline as possible. To meet this condition the waves are working to wear back headlands and to fill up bays. When cut off from the sea the bays are filled with detritus from the land and the deposits of organisms which live in the quiet water. It is believed by many that barrier beaches are a temporary feature of the coast line. As evidence of this occasional outcrops of peat in the seaward faces of such bars are cited. It is possible that some of these, at least, may not be lagoon deposits of the present stand of land and sea but antedate the last rise of the waters.

Endpoint of marine erosion. Although we are unable to find good examples of a cycle of marine action at present because of recent shifts in sea level and in levels of inland lakes we may theorize over a possible endpoint of marine erosion. Many have thought that marine erosion is self-limiting in the depth it can cut into the land without change in sea level. This conclusion is based upon the assumption that the depth of the outer edge of the terrace or wave base is fixed. If such be the case the profile of equilibrium would automatically halt shore recession at a definite location inland. But it is far from clear that either such is the case or that relation of sea and land would remain constant long enough for the process to operate. In fact it is not certain that the outer edge of the submarine terrace is really located at wave base. Granted slow lowering of this level by undertow current or a slight rise in sea level in respect to the land and the limit of marine planation is greatly increased. Especially would this be true were the land first brought low by either peneplaination or pedimentation. Most papers written on this subject have been so theoretical or made up of quotations of opinions of others, which have no real meaning, that it is difficult to reach a final opinion. Certain it is that the ceaseless onslaught of the waves should in time have profound results. Moreover, waves exert more force during storms than streams ever can. Waves could plane down even the hardest and most insoluble rock such as quartzite. Most of the famous buried peneplains were buried under marine formations; how much did the oncoming sea alter their surface? Elevations which escaped the work of waves should then be steep-sided because debris was removed from their bases. Some such have been described in India. It has been often suggested that the Piedmont Plateau is a surface of marine planation because there are so few monadnocks. However, the upland extends behind some of the supposed islands and does not connect seaward with any known marine formation. It is more likely a pediment formed during a by-gone time when the climate was more arid. The top of the Baraboo quartzite range, Wisconsin, is a very gently domed plain. At places the edge carries boulder conglomerate. Projection of now-eroded formations appears to demonstrate that it was the beach at the time of deposition of an adjacent dolomite formation. It is, therefore, possible that the waves of the Ordovician sea completed the planation of the quartzite islands which had been monadnocks on the pre-Cambrian peneplain over 1000 feet lower. A somewhat similar surface on quartzite has been proved by drilling at Hartford, Wisconsin.

Fig. 137

X

<u>Pleistocene terrace problem</u>. Many sea coasts display terraces which evidently record former levels of the oceans in respect to the lands. Many of the higher terraces, as on the Pacific Coast, are obviously deformed by subsequent orogencic movements. Some of the lower ones, however, seem to be horizontal. On the Atlantic Coastal Plain Cooke has described seven terraces which he thought to be horizontal and of world-wide extent. Flint has restudied the same area and concluded that there are only three, of which the upper one at 160 feet is limited in extent. Many of the supposed marine terraces he thought to be in fact stream floodplains but the lower terraces at 25 and 90 feet might be horizontal. Most reports of marine terraces fail to discriminate between the actual level of the water and the elevation of the cut and built terrace. Elevation figures for many vary over so wide a range that exact correlation is impossible. Too little attention has been paid to the sediments associated with the terraces. Knowledge of Pacific terraces is entirely too fragmental to permit of correlation. Postglacial uplift of the land is definitely known in and near to glaciated districts making comparisons there entirely futile. It is, therefore, too early to claim that all these terraces are horizontal and that they record changes in amount of water in the oceans related to the withdrawal in ice caps. That such a process took place is certain but just which shorelines record interglacial intervals when this water was returned to the oceans is far from assured. If the higher ones are really <u>eustatic</u> then it would be necessary to assume that either (a) the amount of ice carried over in continental glaciers increased in every successive interglacial interval or (b) the floor of the ocean sank during the Pleistocene lowering the level of the oceans. Stearns records evidence of just such a sinking of the bottom of the southwest Pacific Ocean. But the cause he ascirbes, namely eruption of lavas which they pressed the area down by their wight, does not appear feasible. The lava came from below and could settle no more than to refill the voids it left. It is possible that this is not the true cause but the facts of sinking are well substantiated.

<u>The coral reef problem.</u> The subject of the origin of <u>coral</u> <u>reefs</u> has been a subject of discussion for more than a century. According to a recent summary by Stearns the following facts are now established: (a) reef corals can live only in warm, clear water less than 200 feet deep, (b) Nullipores serve not only to bind coral skeletons together but also make reefs themselves, (c) when the Pleistocene glaciers were large sea level was lower than now (perhaps about 260 feet in the last glaciation) (d) changes in sea level have also occurred because of alteration of the shapes of oceanic basins, (e) many islands of the southwest Pacific are composed of folded continental (sialic) rocks, (f) the truly oceanic or simatic islands show no such rocks, (g) emerged Tertiary coral reefs are known, (h) atolls do not reflect the form of submerged volcanic craters, (i) atolls rest on a basement of non-coral

rocks, (j) barrier reefs require a platform to start their growth, (k) growth of corals may attain 90 feet in 1000 years, (l) a reef may be killed by submergence (m) a reef may be killed by emergence, (n) reefs grow mainly on the outside, and (o) lagoons are not the product of submarine solution of calcium carbonate.

Occurrence of coral reefs. Coral reefs occur not only along coasts of other kinds of rocks as fringing and barrier reefs but also in isolated islands which rise from the depths of the ocean. Many of the latter surround a partly or wholly enclosed lagoon and are known as atolls. Two small atolls occur off the end of the Florida Keys but they are much more abundant in the Pacific Ocean. Some of the isolated islands are known to contain a volcanic core. In the Bermudas drilling discovered volcanic rocks although a test over 1100 feet deep on a Pacific island failed to find such. Results at Bikini

Theories of coral island formation. Opinions as to the formation of coral reefs and coral islands have varied widely. Some have thought that they grew upward on top of intermittently subsiding foundations of other origin. Others held that they were formed on stationary basements and that the lagoons were made by solution. Others advocated an origin on a rising foundation, others on stationary shelves made by former erosion. Daly first discussed the control of sea level by glaciation , a theory which would make most reefs younger than the last ice age. Recently, Stearns advocated growth on any kind of basement, rising or sinking, under conditions of rising sea level from any cause. The general opinion now is that only glacial control of sea level can explain the majority of reefs. The test boring put down on Funafuti was located too close to the outside of the accumulation of coral and beneath about 150 feet of coral passed through only talus outside of an older reef. Under the glacial control theory it has been claimed that volcanic islands were all planed down by waves to a uniform depth on which corals started to grow as the ice melted and slowly returned water to the oceans. This is thought to account for the uniformity of lagoon depths over considerable areas. However, this theory does not exclude other conditions and both emerged and submerged atolls have been described. It seems certain that in an area of recent vulcanism conditions must have varied widely in different island groups. Much of the argument is based upon purely theoretical reasoning based in large part upon nautical charts which do not show the land areas with much detail. It was thought by some that their failure to show many cliffs on spurs of volcanic islands inside reefs militated against the glacial control hypothesis. Since then this has been shown to be an error, and the glacial alteration of sea level is recognized as an important factor.

Work of other organisms. A number of other organisms besides corals and nullipores affect shorelines. In salt water the mangrove tree can grow where wave action is not too violent. In sheltered bays various kinds of grasses and sedges, which are tolerant of a moderate amount of salt, build salt marshes. Salt marshes on coasts where there is much tide are cut up with a complex net of branching channels through which the water runs in, then out, twice in every 24 hours. Shallow, small fresh water lakes are the habitat of a number of plants which in time may fill them up. These appear to thrive best in relatively hard water. Although most aquatic weeds do not project far above the surface some varieties like the common "bullrush" do. Remains of organisms aid in shoaling the water so that the shoreline can advance. A growth of these hinders wave action and the shore behind can then be filled in with organic deposits. A regular succession of different kinds of plants can be made out from those which thrive only in fairly open water to those of old marshes farthest inland in the old lake bed. Many shallow lakes have been entirely filled with vegetal deposits since the glaciated region was first settled by white men. A common feature of marshes is a moat between the vegetal growth of the center and the high land. These are ascribed to death of vegetation, possibly aided by burning, in dry seasons. In wet weather some of these open water areas are large enough for wave action which has made faint boulder lines along the edge. The level of water has no relation to that of adjacent lakes so that the level of these shore features has no bearing on former lake levels.

Submarine valley problem.

Introduction. It has been known for a long time that the edge of the continental shelf and the continental slope are indented by deep, narrow depressions in which the water is thousands of feet deeper than it is nearby. Until the advent of echo sounding, however, few of these submarine valleys had been accurately delineated. Sounding in deep water with a wire line entails stopping the ship and a long delay in running out and winding in line. In the meantime drifting occurred and positions given on the chart are now known to have been miles out of place. Echo measurements may be made at full speed when it is easier to keep on course. Locations out of sight of land are now fixed by taut wire measurements, radio-acoustic methods, and radio bearings. Exact knowledge of depths is now important to navigators even in midocean. The result is that charts are now much more detailed than formerly and that a great increase in number of submarine valleys has resulted. Most of these new discoveries were off the coasts of the United States for this type of surveying has not advanced as far in other parts of the world. One drawback to the echo or acoustic method of sounding is that on rough bottom more than one signal is returned and the strongest echo may readily not be from the bottom under the ship but from an adjacent slope. In other words the results are apt to be a generalization.

Submarine contouring. Drawing of contoumrs, or lines of equal depth, beneath water where the bottom cannot be seen is fraught with much change for error. If contours are simply prorated between soundings, as engineers do, the result is only a crude generalization. If drawn with some theory of interpretation in mind the result may be simply "wishful thinking". No map is worth anything which does not show all the available soundings. For the reason stated conclusions based on submarine contours have shown the "personal equation" to a marked extent. Some see in them only underwater faulting or folding, others a few valleys similar to those on the continents, other slid and slimped slopes, others conclude a multitude of small ravine-like parallel valleys like the primary rills eroded by rainwash on an earth surface.

Description of submarine valleys. Almost all submarine valleys have a steeper grade than is common on land. Few end in a delta or fan at the lower termination but instead seem to fade out gradually into indefinable irregularities of the ocean bottom. Few indent the continental shelf very many miles. Definite valleys can be traced down to several thousand but less than 10,000 feet below present sea level. In a few cases a submerged connection to an existing stream of the adjacent continent can be found; most do not have this. In rare instances the submarine valley exists inside an estuary (Congo). The sides of the valleys are known from dredging and submarine photography to be solid rock at many points. Current meter observations and bottom samples show that they are not now the location of any unusual submarine currents. Many valleys branch just like land valleys. All have a V-shaped cross ~~secti~~ section. Some are found on the outside of a narrow cuesta (Georges Bank).

Theories of origin. Theories of origin of submarine valleys can be divided into three major classes: (a) the depressions are not valleys at all but are of tectonic origin, (b) they are due to normal stream erosion when the continents were elevated, and (c) they are due to processes which excavated them below sea level. At date of writing there is absolutely no agreement between geologists on which theory is best.

Diastropic origin. The idea that submarine valleys are due to earth movements is an easy way out of the problem. Under this view the similarity to land valleys as shown on contoured maps is "purely coincidental". However, it is quite generally agreed that the winding course of many submarine valleys, their V-cross section, and the presence of tributaries are fatal to this explanation of most of the known valleys of the continental slope. Others have suggested that slumping and sliding of the soft material on this slope, especially where it is unusually steep, might easily account for the observed irregularities particularly between the deeper canyons. Parts of the deep extensions of valleys and many irregularities of the sea bottom in regions of recent disturbance are probably due to earth movements.

Excavation by land rivers. Many of the points of objection to the tectonic hypothesis are in favor of subaerial origin. Many examples are off of existing streams, although not connected under water. Coarse gravel has been found in the bottoms of some canyons down to 5000 feet depth. Canyons run directly down the continental slope as rivers would, and seem to be roughly related to the size of adjacent rivers on land. On the other hand, some canyons have no land extensions and those of Georges Bank have a very small watershed. The grades are certainly abnormally steep. The principal difficulty with the idea of origin on land is to account for the vast change of sea level required. In order to avoid this some have suggested that the continental shelf was tilted up, the canyons eroded, and then it was bent back again carrying them into deep water. Others have thought that the entire body of sediment has moved down the continental slope carrying the canyons with it! Possible causes for tremendous shifts in oceanic level are (a) diastrophism involving part of the ocean bottom, (b) temporary uplift of the continents, and (c) vastly magnified glacial control of amount of water. Confirmatory evidence which would support one of these startling assumptions is lacking. Even if the modest figure of 3000 feet of lowering of sea level is taken, difficulties are still present. Glacial abstraction of water involves not only much larger and thicker glaciers than those commonly thought possible but also a great increase in salinity of the remaining water.

Submarine origin. Geologists who were greatly impressed with the difficulties of such immense changes in relation of sea to land turned to a search for some process which could erode valleys under water. Principal suggestions comprise: (a) density currents, (b) mudflows and landslides, (c) submarine springs, and (d) earthquake waves. The first is ascribed to more muddy water than now on the continental shelves when glaciation lowered sea level a few hundred feet. Muddy water is heavier than clear water and such underwater currents are actually known, although it is admitted that they are not flowing through the canyons today and off the mouths of muddy rivers the fresh water floats. Competency of such submarine currents to erode hard rock appears open to doubt. Under-water slides have been recognized in many places, particularly after earthquakes. On the other hand the termination of the valleys inshore is unlike the basins which develop on land from sliding and there are no enclosed depressions in the bottoms of the valleys or mounds of slid material at their bottoms. Tributary valleys are hard to account for by this idea. Landsliding may have taken place, however, and might account for many minor irregularities between the canyons. Submarine fresh water springs escaping from permeable layers of the Coastal Plain sediments of the east coast of the United States are a distinct possibility, but on the California coast they are not. Nevertheless, it seems impossible for such springs to produce branching valleys in the way they could on land. Certainly the fresh water should rise and its capacity to dissolve the overlying material and leave a consistent valley during retreat of the spring appears impossible. It has been suggested that vast waves started by earthquakes during the orogeny of the Tertiary washed up onto the continental shelf. Backwash from these might then have eroded the canyons. Just why flow should be concentrated in the way drainage is on land is unexplained. Intermittent occurrence of these waves is also against the idea. No current meter observations during the passage of waves are recorded.

Summary. With work of waves in standing water is included that of currents, organisms, and the problem of submarine valleys. Waves and associated currents work to toward evening up of the shoreline by destroying headlands and filling bays, the latter process greatly aided by organic deposits in the quiet water behind a bar. The mathematical relation between length of a wave and its velocity is well known but there is none between its height and its length. Given height and length, however, total energy can be computed and checked with actual measurements. It is difficult to compare total energy expended on a shore with total energy of running water on an adjacent land area. But the quantitative value of force exerted by storm waves and the continuity of wave attack lend color to the idea that wave action is more potent than running water in completing the leveling of the land. Since wave base is not a fixed depth waves can erode farther into the land than was once believed possible, so that the formation of large wave-planed surfaces cannot be denied. The level of the oceans has varied greatly particularly because of withdrawal of water to form glaciers. Because the sea is now rising upon lands all over the world the commonly used classification of shorelines is unsatisfactory. For the same reason examples of a theoretical cycle of shoreline development cannot be found. The various theories of coral reef formation are compared and the differences found to be less striking than their proponents thought. It is clear that rising waters or land subsidence were required for most reefs but the cause is unimportant; in some localities the movement has undoubtedly been tectonic but as there is other evidence of postglacial rise of sea level glacial control cannot be ignored. The problem of origin of the submarine valleys now found on most coasts is still unsettled for none of the theories thus far advanced is free of fatal defects.

WORK OF WIND

Introduction. Over a large part of the surface of the earth wind is a potent force in the making of land forms. Wind can perform erosion, transport material, and build deposits. Although limited in effectiveness by vegetation in the more humid climates it can do some work there at certain times of year. The following discussion is based to a large extent on the book by Bagnold.

Materials carried by wind. Winds reach much higher velocity than water ever does yet their competence to move material is less. This is because objects do not lose as much of the weight they would have in a vacuum in air as they do in water. Density of air is only about 1.22×10^{-3} compared to 1.0 for pure water. Viscosity is about 0.17×10^{-3} compared to 0.01 for pure water. Nevertheless, objects falling through air reach a terminal velocity when the resistance is proportioned to the square of linear dimension and the weight to the cube. For this reason terminal velocity is in a general way inverse to size of particles. Instead of classifying materials as clay, silt, and sand it is more convenient in dealing with the wind to divide only into dust and sand. Distinction is made at the upper limit of size of particles which are kept in the air by the turbulence of ordinary winds. Effective upward currents are estimated by Bagnold at about 1/5 of the forward velocity of ordinary winds. A wind of 5 m/sec (11 mi/hr) will just support particles with a diameter of 0.2 mm which is a critical line of division between dust and sand. Most wind-blown sands do not have particles smaller

compare with fig 114

than 0.08 mm. Average diameter of such sands is from 0.15 to 0.30 mm. Sands are predominantly quartz because of its abundance, hardness, and toughness.

Behavior of sand in the air. The storms which obscure the sun in dry regions lift mainly dust; this stays aloft for many days after the wind falls. Real sandstorms are dust-free except at first and the sand rarely rises as much as two meters above the surface, with clear air above. Some of the grains moved are over a millimeter in diameter. Sand moves by being driven into the air by direct impact of wind. The grains describe a parabolic trajectory and land with enough force to dislodge other grains and start them on an aerial journey. This process is not exactly the same as saltation under water but is called by the same name. Experiment proves that the grains behave just about the same as spheres with a diameter three fourths as large. Sand also moves by surface creep, which is the motion of grains which could not be forced off the ground although they were slightly moved by impact. Grains thus moved are too heavy for the wind to raise off the surface. Most sand grains are too large for true suspension.

Wind velocity. Wind blows only in a turbulent manner. Wind velocity is related to heigth above the ground, not directly but to the logarithm of the heigth. When velocities measured at different heigths are plotted on semi-logartihmic paper a straight line gradient is displayed, which reaches the 0 velocity line at a definite elevation, k, above the ground. This heigth is about 1/30th of the diameter of the sand grains on the surface or other surface roughness. Now, as with water, the direct force of the wind per unit area is equal to half the product of its density multiplied by the square of its velocity at that level. The horizontal force or drag per unit area of surface parallel to the wind is equal to the product of density of air times the square of a quanity known as the drag velocity. Now the drag velocity, $V\#$ is directly proportional to the rate of vertical increase of wind velocity compared to logarithm of the heigth, that is the tangent of the slope of velocity lines on semi-logarithmic paper. If we measure the velocity at two heigths, one of them 10 times the lower one, the velocity difference divided by 5.75 equals $V\#$. In general terms: $V\# = \text{vel. diff.}/(5.75 \times \text{log-heigth diff})$. This relationship holds for turbulent flow of all liquids and gases. Now it is evident that when wind passes from a smooth surface to a rough one the slope of the grade on semi-log. paper is unchanged but the point of 0 velocity is raised to a higher level, k', thus reducting the wind velocity by the same amount at all levels It takes some distance over the new surface before the effect is attained at all heigths. Velocity at any heigth, $z = 5.75\ V\# \log (z/k)$. A rough surface is defined as one where the Reynolds number, $(V\# \times \text{diam. grains})/\text{viscosity}$, exceeds 3.5

Effect of sand movement on surface wind. Sand in motion alters the surface wind. When we draw on semi-log. paper the rays for several different velocity gradients all converge to a single point of 0 velocity at level k which is about 1/30th the dimension of surface roughness. When sand begins to move the gradient lines cross at a new point which is not on the 0 velocity line but at a definite velocity. This new focus is at level k' and at a velocity at which sand just begins to move. This is termed the dynamic threshold at which sand begins to move through impact. The rasing of the level is ascribed to development of a rippled surface. The velocity ray of the original line on which the new focus lies passes through the 0 velocity point of the grade at which motion started. From these facts it is clear that velocity at any heigth above ground when sand is moving is shown by the equation: $v = 5.75\ V'\# \log (z/k') + V_t$ where V_t is the threshold velocity measured at heigth k', and $V'\#$ is the drag velocity when sand is moving. Loss of momentum from the moving air by reason of sand transportation may be calculated by multiplying the quanity of sand, Qs, moved in unit time in unit width by the result of dividing the average loss of velocity of grains by their average distance of travel, l. Sine the velocity with which a grain starts its flight is small it may be neglected and only the final velocity, u, need be considered. From this it follows l = distance of travel

that the drag may be expressed as: Qs x(u/l) This is equal to density x $V'\#^2$ Now it has been found that u/l closely approximates g/w where w is initial vertical component of velocity of a grain at the beginning of its path, and g is the acceleration of gravity, hence Qs x (g/w) = density x $V'\#^2$. Solving this equation for quanity, Qs = (density/g) x w x $V'\#^2$ Now making the assumption that w : $V'\#$ we can substitute for w and find that Qs = constant x $V'\#^3$ Experiment disclosed that about a quarter of total sand movement is not through saltation but is due to surface creep of grains which are not struck with enough force to send them into the air. The formula for sand movement was checked by wind tunnel experiments which yielded a constant of about 1.47 where value of density / g is about 1.25 x 10^{-6} It can also be demonstrated that the quantity $V'\#$ can be replaced in the equation by velocity at any given heigth above the ground less the threshold velocity thus giving the final expression:

Fig 152 153

$$\text{Quantity} = \text{constant} \times (\text{excess velocity at given height})^3$$

It was also found that a wind of given velocity can drive sand faster over a hard, immobile surface than it can over loose sand, but no quanitative data were obtained. The above equations were derived for sand of one diameter of grains only but may be modified to meet natural conditions by altering the constant.

Comparison of results for air with conditions in water. Bagnold remarks that the good results he obtained in deriving and checking formulas for transpotation of material by moving air are not matched by those others obtained with water. This is due in part to difficulty of observation under water but mainly to the tremendously larger loss of weight of particles in water. As a result, the reduction of velocity of a stream of water by reason of saltation is, other things being equal, less than one one thousandth that which takes place in air. The frictional drag of air on the ground may be neglected for it is so small compared to transfer of momentum by the sand load. In water the bed load extracts little momentum and water velocity is regulated almost entirely by the roughness of the bottom. Grains are not dislodged by impact under water but probably by eddies of turbulent flow. Were the bottom of a stream to remain smooth erosion would be much more that it actually is. Development of bottom roughness then acts as a limit to movement of sediment. When suspension of particles begins, however, the bottom becomes smooth again. This condition begins when 7 $V\#$ exceeds the average settling velocity of particles. When suspension is fully developed grains move along with the water like part of it.

Suspension in air. When $V\#$ exceeds one seventh of the falling velocity of grains in air it appears that suspension replaces saltation. Since this falling velocity is proportional to diameter over a wide range of grain sizes we can write as a fair approximation: quanity = $V\#^3$/ grain diameter. Bagnold suggests that the change to suspension sets in when the grain-impact method of saltation ceases to operate and that thereafter we have a condition akin to sediment transportation in water in which saltation is not the same as in air.

Field experiments. Bagnold carried out extensive field experiments in Libia to check the foregoing laboratory work. Here the grain sizes are more varied and the surface is nowhere trully flat. However, the checks were satisfactory.

Relation of threshold velocity to grain size. The threshold wind velocity is that when grains just begin to be dislodged from the surface. The angle at which a grain must be raised to get away from the ground is assumed to be close to the angle of repose of loose sand. It is possible to compute the required force in the same way as for streams. Equating force to resistance we find that : $V_t\#$ = constant x (effective weight x diameter/ air density)$^{1/2}$ or in other words if other things are equal the value of the threshold velocity, $V_t\#$, varies with the square root of grain diameter. This relation holds only for diameters over 0.25 mm. The constant for these and larger grains is 0.1 compared to 0.2 in water. When the Reynolds number, $V\#d$/ viscosity, is less than 3.5 a greater drag is required to set grains in motion. When the grain size is less than 0.2 mm the value of

Fig 154

the constant increases and the square root law no longer holds. In water the same change is at a diameter of about 0.6 mm. Data for force needed to start small particles is hard to obtain but it is not wholly because of cohesion that it is larger. Bagnold has observed that in aird regions with soil of fine texture there is little dust except where the surface has been disturbed artificially. He also calls attention to the clean separation of loess from sand. The critical diameters are in air 0.08 mm and in water 0.2 mm. An important factor in holding down small particles in air is their moisture content. Another is a mixture of large and small particles in which the former protect the latter. The impact threshold wind is distinguised from the fluid threshold discussed above because at it the motion of grains is maintained by the impact of falling particles alone. The wind up to the critical heigth, k', is then unchanged no matter how hard the wind blows above that level. In water there is no corresponding condition, only the fluid threshold. and there is no fixed focus of constant velocity.

Land forms produced by wind. Land forms produced by wind may be divided into first, erosional, and second, depositional. In the latter class we are here interested only in the mounds of sand known as dunes and similar large scale deposits of dust or loess.

Erosional land forms. Erosion by wind is sometimes called deflation. Topographic forms made by wind erosion include: (a) hollows or basins, sometimes called blow-outs, (b) pillars, pinnacles, and cliffs undercut by sand blown along the ground yardangs, and (c) residual portions of dried-up lake beds or playas. Hollows may be eroded by the wind in humid climates provided the soil is unfavorable to a cover of vegetation. The initial step is destruction or reduction of this protective material by drougth, fire, or the work of man. Wind then sweeps away the underlying material if it is of a nature which is readily picked up. Blow-outs are most abundant where the material is sand but some are found on shale which disintegrates into dust. Such are abundant in the Colorado Piedmont and the Wyoming Basin. Big Hollow, near Laramie, Wyoming, is 150 feet deep and about 3 by 9 miles in extent. Some of thse hollows contain lakes when there is enough rainfall. Some are limited in depth by accumulation of pebbles into a desert pavement others by reaching moist material near the water table. Large depressions in Libia and Egypt, such as the Qattara Depression, have been ascribed to wind erosion and some have applied this theory to many of the enclosed basins of the Basin and Range province of this country. Many of these are more likely due to earth movement and it seems doubtful that wind ever eroded far into solid rock. Undercut cliffs or rock shelters have been ascribed to wind erosion but old Spanish inscriptions in some of them in the southwest indicate that, if operative, the process is slow. The beds of playas or temporary lakes have been eroded by the wind in many places leaving miniature mesas capped by salts. Cracking of dried mud facilitates wind erosion. However, it is an open question whether or not the dust is removed far enough to not be washed back by the next rain.

Depositional land forms due to wind transportation. In humid regions the major sources for sand which is available for wind transportaion are beaches and river beds. When the glacial drift was newly deposited and glacial lakes disappeared large areas were free from vegetation and exposed to wind action. Under present conditions in humid regions there is an ever-present contest between vegetation and wind. In true arid climates, such as in Libia, wind has the field to itself provided the rocks weather into material which is within the power of wind to transport. There dunes reach full development in the erg. Bagnold lists: (a) sand accumulated behind obstacles, (b) true dunes divided into barchans and long mounds, (c) coarse-grained ridges or whalebacks, (d) gently undulating sand tracts, and (e) sand sheets, Rock surfaces in deserts are often called hammada. We must realize that by no means all the surface of a desert is sand.

Dunes of humid lands. Sand which is blown from areas with no vegetation is commonly deposited in a short distance to form a ridge parallel to the source. These ridges are often called transverse dunes or foredunes. Because in part of variable wind direction and in part of scanty vegetation these ridges are very unstable. Their corss section is gently sloping toward the source and sand is added in layers with a gentle dip to windward. The lee face is made by sand which slides down in a slip face(foreset bedding) at an angle of about 34 degrees. Low places or breaks in the cover of vegetation are blown out into hollows. The eroded sand is heaped around the heads of these depressions, some of which have no outlet. Similar accumulations are also formed on the lee sides of blow-outs in plains of sand. Some have called this type of dune parabolic but the name is unfortunate because there is no relation to the mathematical form by the same name. The open ends of the crescents point toward the wind. When a group of blow-out dunes starts the earlier ones are checked when the hollow attains its maximum possible depth. With source cut off the dune becomes stabliized with a cover of vegetaion. Local failure of this starts up the process again so that a group of dunes travels slowly down wind, leaving a confused system of mounds and hollows which at first sight appear wholly without plan. In middle latitudes, away from the ocean and large lakes, winds are so variable in direction that the irregularity of dune topography defies analysis. Enclosed depressions are not all blowouts byt are made by advance of the lee faces of dunes. Lakes are present in some depressions. Blown sand also serves to help fill the lagoons behind sand beaches. Many abandoned beach lines are marked by rows of dunes.

Fig 160

Dunes of arid lands. The largest areas of arid climate in the world are prehaps those of the trade winds. There the constant wind direction leads to the greatest perfection of dunes. When the wind is strong sand is removed from pebbly surfaces and accumulates in sandy areas. Gentle winds cause these patches to travel downwind and the sand to be scattered. Eddies in the strong winds concentrate sand. In many places sand is blown into long stripes parallel to the wind direction. Some dunes are made of gypsum particles.

Shadow dunes. Sand accumulates in wind shadows behind obstacles such as rock cliffs or bushes. Shadow dunes which are transverse to the prevailing wind are unstable. Long dunes which grew out too far to leeward of the protection are apt to be broken up particularly by cross winds. The bedding of such dunes is parallel to the surface except at the lee end which has a slip face.

Fig 161

Barchan dunes. Prehaps the best known dune form is the crescent-shaped barchan whose horns are pointed down wind. These forms are best developed where the sand is rather limited in amount resting on a non-sandy basement and the winds are unidirectional. The windward face is gentle and the slip face is inside the two horns which have a minimum heigth of about 30 cm. Where barchans are closely spaced the partially protected examples are more complex in form than is normal. Enclosed depressions between them are sometimes called fulji. A system of barchans is most commonly slightly offset or en echelon. Maximum heigth of barchans is about 100 feet. The entire streamlined form is slowly moving downwind at a rate which may reach several centimeters per hour during storms. Layers added to the windward face are firm but the slid sand of the lee sides is very soft and unstable. Where the windward side is exposed by erosion this soft sand reaches the surface.

Fig 162

Fig 163 section

Seif or longitudinal dunes. In the very sandy areas of Libia, Arabia, and Australia the dunes are long ridges which parallel the direction of the prevailing trade wind. These long ridges are known as seif or longitudinal dunes. They are known to reach a heigth of over 200 meters (700 feet), a width of about 6 times their heigth, and a length of 100 kilometers (60 miles). The summits have crests from 20 to 500 meters apart or about 6 times dune heigth. The windward end of a longitudinal dune is generally broad, locally with an enclosed depression and the lee end is sharp. There is a slip face on one or both sides and a gently sloping basal plinth of firm sand which has never slid. The summits apear to be moving and the entire dune is possibly moving laterally as well as toward the lee end. The lateral spacing of seif dunes is from 1 to 10 kilometers and the corridors between them extend for long distances. Barchan dunes occur in some of them. The origin of thsese long dunes is disputed. Some demonstably occur in the lee of obstacles. Bagnold holds that they are due to the alteration and consolidation of barchans by cross winds. To the writer it seems reasonable to conclude that they are the ultimate streamlined form of minimum friction with constant wind direction.

Fig 164

Fig 165 x-section

In other words, they are extremely elongated barchans in which the two tails have been joined together.

Fig 166

<u>Whalebacks</u>. The whalebacks of the Libyan desert are sometimes called sand levees. They are from one to 3 kilometers wide and up to 300 kilometers long with a heigth of up to 50 meters. There are discontinuous chains of dunes on the tops, at one side, or in a series side by side. Bagnold regards whalebacks as residue left by migration of either one seif dune or a series of them.

<u>Undulating tracts</u>. The gently undulating tracts of sand in Libia seem to occur where there is some rainfall and vegetation. They seem to be similar to many sand areas of humid or subhumid areas where dune topography is poorly defined.

Fig 167

<u>Sand sheets</u>. The areas in Libia which are termed sand sheets have a surface of coarse sand with some pebbles. The material below has layers of sand and pebbles with fine red powder below a depth of about 10 cm. Pebbles were derived from near-by bed rock. The sheets appear to be the result of longitudinal sand strips which are protected by these pebbles. Some water action may have occured.

Fig 168

<u>Loess deposits</u>. Toward the borders of the more humid regions to the lee of extensive dunes there are silt accumulations called <u>loess</u>. Notable examples occur in the United States southeast of the Sand Hill district of northwestern Nebraska where the dunes appear to be the result of past wind erosion of Tertiary alluvial fans at the foot of the Rocky Mountains. In China the loess lies in the lee of large desert areas. In Europe loess is most widespread in southern Russia. The so source, method of transportation and time of deposition of loess has long been disputed. Its derivation has been variously ascribed to fresh glacial drift, to stream floodplains, and to wind erosion in deserts. Transportation and deposition has been ascribed both to lakes, streams, and the wind. In time, opinion has varied from interglacial arid intervals to during or immediately after glaciation. Although loess is in many places thickest adjacent to rivers from whose beds it may have been derived, this localization of deposition might also be explained by rough, forested hilly topography which caused local still air. It is now generally thought that loess was transported and deposited by wind, although some think that the loess of the east bluffs of the Mississippi is the product of weathering of clay. The arguments cannot be discussed here except that there may be two types of loess, one late-glacial or glacial in age, the other interglacial.

Fig 169

<u>Land forms due to loess</u>. Land forms due to loess deposition are not abundant. On the east bluffs of Missouri River in Iowa loess forms hills which as viewed from the air suggest snowdrifts. Here the loess locally reaches 200 feet thickness. In most places the loess simply forms a mantle over older topography of various origins. When eroded, slopes are very steep because of its high permeability and the vertical cleavage, which some ascribe to the casts of grass roots. Landsliding on these slopes gives rise to minor terraces called <u>catsteps</u>.

<u>Summary</u>. The wind is the most potent force in shaping the landscape in truly arid regions, even though there are some traces of water-work in many of them. Both barchans and seif dunes tell of uni-directional winds, the latter prehaps the ultimate form of minimum friction between sand and wind. Barchans whose horns point down wind should not be confused with blow-out dunes of humid regions whose convex sides are exactly opposite. It is quite possible that conclusions on former wind directions in some localities are 180 degrees in error for this reason. Direction of dip of foreset bedding is a much more reliable criterion. The complex dune topography of humid regions is readily explained by the conflict of vegetation with winds of variable direction. Wind shadow dunes occur in all regions. Whalebacks and sand sheets are mainly confined to arid regions where wind has worked for a long time with little interference from vegetation. Constructional hills of loess occur on the borders of humid regions.

WORK OF ICE

Introduction. Work of ice includes the land forms made by glaciers, icebergs, and the ice of lakes. It excludes the work of ice in weathering and soil formation. The approach here is somewhat different from that used in the study of glacial geology, in that it is not concerned to any great degree with the physical nature of the deposits but is confined to processes which made land forms.

Glaciers-introduction. A glacier may be defined as a mass of ice which was formed by compaction of snow and which flows, or at some time has flowed, under the influence of gravity, that is it is drainage of precipitation in solid form. Glaciers are subdivided into (a) valley or mountain glaciers, (b) piedmont glaciers formed by the joining of several valley glaciers at the foot of the mountains, and (c) continental glaciers which cover large areas.

Origin of ice. Glacier ice is compacted recrystallized snow. Partially altered snow is often called firn and has a density of 0.72 to 0.84 A layer of not less than 100 feet of this material is found near the surface of the source areas of glaciers. Much more is probably present in polar regions. Firn is absent in the lower parts of glaciers where they are wasting away. Glacial ice below the firn contains up to 15% of included air and its density does not exceed 0.9 in contrast to 0.918 for ice made by freezing water. Ice crystals are hexagonal and are several inches in diameter.

Physics of glacial motion. The physics of glacial motion have long been misunderstood by many geologists. Near the surface, where it is under light load, ice behaves like a solid and yields to stress mainly by fracture. Under heavier pressure its physical behavior is like that of a liquid. This is no different from the phenomena of rock deformation except that it occurs under lighter stress. The only available determination of the viscosity of ice is 1.2×10^{14} poises. Demorest, who did much work on this subject, states that viscosity decreases with load. Presumably this is because pressure causes recrystalization with the structure arranged to facilitate flow to relieve the stress. We would then not expect any further change below the depth at which this process has been completed. Ice cannot exist at temperatures above 0 degrees C. The flow of ice is laminar. For a valley glacier with approximately a semi-circular cross section average velocity would be given by the following:

$$V = \frac{g \times \text{density} \times \text{sine slope} \times \text{depth}^2}{32 \times \text{viscosity}}$$

On a slope of one deg. a glacier 100 meters thick would then have an average velocity of about 3.5 cm. per day, which seems to agree with actual observations. In such a thin glacier on a sloping base one layer flows over that just below. Force is the component of gravity parallel to the bottom. Rate of flow should be 0 at the base under thick ice but near the terminus the thin, rigid ice might be shoved bodily over the rock. This type of flow was termed gravity flow by Demorest. Conditions are different in a thick continental glacier. The top is rigid and is retarded where it reaches the ground at the thin outer edges. Below, the ice flows by reason of the difference of top elevation from place to place. Since yielding can only be outward the component of weight of a unit column of ice which is parallel to the top is the force for motion and velocity at depth d is shown by the following formula:

$$V = \frac{g \times \text{density} \times \text{sin slope} \times \text{depth}^2}{2 \times \text{viscosity}}$$

Substituting for a slope of 0.1 degree (less than 10 feet per mile) and thickness of 5000 meters the result is only about ~~.0003~~ 1.138 cm. per day. Bottom velocity should also be 0 except near the margin of the ice. Demorest called this extrusion flow because it is present only at depth.

Ice erosion. Ice erosion is the result of friction between moving ice and the bed. The force of friction is equal to the weight of a unit column of ice multiplied by a coefficient which depends upon nature of underlying material and not on velocity. The power, or time rate of work, of a glacier is this force multiplied by velocity. Other things being equal, the velocity of a valley glacier is related to surface

the square of the depth. In a continental glacier this relation does not hold for the formula does not take into account the effect of spreading out of the ice toward hte margins. If we reasonsed solely upon the formula for a valley glacier power should be related to the cube of ice thickness. However, this conclusion does not tell the whole story. Bottom velocity, except near the terminus, is 0 or close to it. Much of the energy is absorbed in internal friction. If this loss results in pressure-melting of ice followed by refreezing then no energy is lost. But if there is some permanent melting or conduction of heat to the surface energy is really lost. The greatest differential velocity of ice over rock must be near to the end of a glacier. Most geologists seem to think of erosion by ice as mainly due to grinding of rock into powder or rock flour. The word scour is often used for this type of work. A little thought will show that erosion of hard rocks in this manner would absorb an enormous amount of energy, and since the work would be spread over a very large area of ice bottom, be extremely slow. Although it is an undoubted fact that much rock flour is made by glacial action, a much more potent form of erosion of firm rock is removal in pieces. This process is termed plucking and occurs where plastic ice can flow around and freeze to rock masses which have been broken by older fractures. Such pressure-melting and refreezing takes place under thick ice. The melting point of ice is lowered about 1 degree C for every 2100 meters depth. Computation shows that with the normal heat emission from the earth the bottom of all thick glaciers must be at the pressure-controlled melting point. Under this condition very slight changes in pressure may change the ice from solid to liquid and vice versa. If cracks are present in the bed rock fragments are thus incorporated in the ice and move forward with it. This results in much more rapid erosion than would otherwise be possible. Another mode of erosion is present in valley glaciers where there is a prominent crevasse called the bergschrund at the head next to the mountain wall. Much meltwater both from the ice itself, and from banks of snow above, enters this crack and freezes. This freezing loosens many blocks of rock which are then carried off by the moving ice. This process is called sapping. The importance, and even the existance, of glacial erosion has long been debated. That such erosion is an important process in shaping of land forms, is demonstrated by (a) the vast amount of fresh material derived from bed rock in the glacial deposits, and (b) the unique topogrphy of many glaciated districts.

Deposition by ice. In considering ice deposition it is impracticable to separate the result of direct ice deposition from the work of meltwaters. Unstratified and unassorted material direct from the ice is called till and the word drift includes both this and associated water deposits indirectly due to glaciation. Deposits of glacial streams consist of sand and gravel which bears in its nature the record of the frequent changes of volume and velocity of ice-borne streams with floating ice. In areas not long vacated by glacial ice large residual remnants were buried in the deposits and did not melt for a considerable time.

Erosional land forms of valley glaciers. The most striking and characteristic erosional land form of valley glaciers is the cirque. These bowl-shaped depressions are also known as corrie or cwm. They occur not only at the heads of mountain valleys but also on mountain sides and frequently are found in stariways one above another. Cirques are ascribed to sapping in the bergschrund of small glaciers. Another characteristic feature of glaciated mountain valleys is a non-uniform grade with enclosed rock basins separated by intervals of abnormally steep slope. Rock basins are also present in the bottoms of many cirques. The transverse cross section of many glaciated valleys is notably U-shaped rather than the V-shape of normal stream valleys in mountains. This phenomenon is best displayed in massive igneous and metamorphic rocks and is enhanced by a filling of gravel outwash in the bottom. Unfortunately it has been termed catenary by some geologists, although it bears no no relation to the curve made by a rope or chain suspended at the ends. Instead it is explicable by the fact that the work of a glacier is spread over a wider bed than that of a stream carrying the same total discharge in unit time. The coasts of many glaciated mountainous regions such as Norway, Greenland, Pategonia, and Alaska are indented by many long narrow bays called fiords(fjords). Many of these are much shallower at their outlet into the sea than they are inland where depths up to 4000 feet have been recorded. Fiords form a branching system which in many places is trellis, that is adjusted to the structure of the bed rock. Many of the tributaries enter from hanging valleys; some of these have the lip under water and others give rise to spectacular falls or rapids. Fiords are now generally ascribed to glacial erosion which was less effective under thin ice near the outlet thus leaving the threshold. Hanging valleys are also common in almost all glaciated mountains although discordant junctions are by no means due only to glaciation. It is now quite generally recognized that glacial erosion on a large scale is closely related to the amount of fracturing of the bed rock and is thus controlled by regional structure. This was well shown by Matthes in Yosemite Valley, California. However, it is possible that the relationship of glacial erosive power to the cube of ice thickness is another important factor in the production of hanging valley junctions. A relatively minor topographic feature of glaciated valleys is the roche moutonee, rock knobs with a gentle slope on the stoss side toward the source of the ice and a steep slope on the opposite or lee end. Some students have ascribed these to turbulent flow of the ice but in view of its high viscosity this is impossible. Roche moutonees are readily explicable as simply rock masses too large to be removed by plucking which were ground down on the exposed side by the moving ice; they have no necessary relation to preglacial features. Roche moutonees should not be confused with exfoliation domes. Many other irregularities of glaciated valleys are doubtless explicable by variation in amount of fracturing of the bed rock.

Cycle of mountain glacial erosion. Attempts have been made to distinguish a cycle of mountain glacial erosion similar to the cycle of stream erosion. It is true that glacial erosion varies greatly in amount in different localities depending upon the size and length of life of the glaciers. In some places only isolated cirques are present, wheras in other localities the headwalls have been worn back so far that only narrow ridges (aretes) are left between cirques and the higher summits have been sharpened into horns. But what the next step would be is unknown, for it is clear that glaciation has been only a relatively brief episode in the history of existing mountains. It may be presumed that if glacial erosion continued long enough it would reduce the entire mountain range so much that snowfall would be decreased. Possibly this might lead to extinction of the glaciers.

Depositional land forms of valley glaciers. The depositional land forms of valley glaciers are relatively small compared to adjacent rock topography. Where the terminus of a valley glacier remained stationary at a position fixed by a balance between melting and forward motion a moraine was left in an arc across the valley. Such are known as terminal or endmoraines. Debris which fell from the mountainsides onto the ice left lateral moraines. Where tributary glaciers joined the lateral moraines coalesce into medial moraines. Many lakes are enclosed behind moraines. Valley floors both within and outside the maximum ice limit are filled with outwash deposits of sand and gravel, often called valley trains.

Erosional land forms made by glacial waters. Streams of glacial meltwater eroded notches across spurs and eroded many potholes in rock. The most spectacular feature due to erosion by water is the scablands of Washington. A prolonged controversy has been waged over the origin of these areas of ~~xxxx~~ bare basalt with many abandoned waterfalls and only small areas of gravel. Bretz held in many papers that the scablands were eroded simultaneously by a vast flood, hundreds of feet deep. No cause for such a flood could be found. Flint discovered that the gravel of what Bretz called bars is normal outwash which is too fine to have been deposited by a vast flood. He concluded that valleys across the Columbia Plateau were first filled with outwash which sloped down to Lake Lewis in the lower Columbia Valley at a grade of 13 ft/m. Neither the origin nor the cause of draining of this lake was discovered but Flint concluded that lowering of its level before the ice front was melted back brought about a change in slope of the meltwater streams to about 20 ft./m. This caused erosion of almost all of the outwash and the remnants left in the mouths of tributaries are what Bretz interpreted as bars. Allison has disputed Flint's ideas on the ground of (a) improbability of simultaneous erosion over so large an area, (b) misinterpretation of the relation of outwash to the sediments of Lake Lewis, (c) the topographic form of some of the bars which he thinks are constructional, and (d) presence in the lower Columbia Valley of high-level stream-eroded areas above the lake silt. Allison gives no general theory but suggests that the effect of ice jams in diverting the rivers has been neglected.

<u>Erosional land forms due to continental glaciers</u>. Erosional land forms left by continental glaciers are not as conspicious as those formed by valley glaciers. In fact, the origin of some of them has long been disputed. In making comparisons it is well to realize that continental glaciers probably had much less velocity than do mountain glaciers which rest on steeper slopes. The latter are in motion throughout almost all their life wheras continental glaciers became so thin during their wastage that motion must have ceased throughout large areas. Some continental glaciers do not seem to have remained in motion long enough to make an endmoraine. It is true that many fiord coasts were once covered by continental ice but an important factor in producing these deep valleys was the presence of local glaciers both before and after every continental ice cap. For that matter, cirques occur in areas which were covered by continental ice, for instance in the White Mountains. These were certainly made by local glaciers. Major features in the United States which are generally recognized as due to erosion by continental ice are (a) the deep, youthful, glaciated valleys of the northern Appalachian Plateau, some of which contain the famous Finger Lakes, (b) the basins of the Great Lakes, and (c) some of the remarkably straight escarpments of the Great Lakes region. The Finger Lake valleys are so straight and deep, extending below sea level, that only glacial erosion can be the cause. Cooperating processes comprise (a) erosion by meltwater from the advancing ice thus lowering divides, (b) erosion by diverted streams during interglacial intervals, and (c) reversal of drainage in preglacial time from a southward course to join the subsequent valley of the Mohawk. It has been suggested that erosion was concentrated in the deeper valleys because only there was the ice thick enough for pressure-melting which allowed of plucking. The Great Lakes certainly lie in basins which are enclosed by bed rock. The problem is complicated by known earth movements which appear to be still going on. Only a small part of the depth of the deeper lakes can possibly be explained by dams of glacial drift. Both Lakes Huron and Michigan are crossed by submerged cuestas on the Devonian limestones. The deepest parts of these lakes are above Silurian salt and gypsum-bearing rocks which must certainly have been easily erided by ice. Other basins appear to be all on shale. The east end of Lake Superior is extremely irregular but the relation to the structure of the Keweenawan sediments and traps is unknown. It is impossible to account for thse basins by earth movement alone although neither the preglacial drainage nor the exact amount of glacial erosion can be determined at present. The absence of valleys across the cuestas appears to indicate that preglacial valleys were bottomed far higher than the lakes now are. Glacial erosion is also indicated by the large amount of drift south of the lakes. The simple form of the escarpments in much of the Great Lakes region tells of glacial removal of spurs and outliers. This seems to have been most marked where the soft rock is shale and where the ice moved approximately parallel to the escarpment. Roche moutonee hills occur in areas of continental glaciation.

<u>Depositional land forms due to continental glaciers</u>. Depositional land forms due to continental glaciation with associated meltwaters include moraines (both terminal or end, <u>recessional</u>, and <u>ground</u>), <u>drumlins</u>, outwash, <u>eskers</u>, and <u>crevasse fillings</u>. Terminal or endmoraines originated in the same way as those of valley glaciers. Recessional moraines have been much misunderstood. Those due to a halt in melting back of a front of moving ice are very irregular in outline. Most moraines behind the outermost or endmoraine are the result of readvances of the ice front to a regular smooth outline burying water deposits. Moraines made of stony material contain much water-sorted gravel and sand and have steep slopes. All driftt is very wet when deposited and water can escape from sand and gravel without causing extensive sliding. Where a large amount of clay and silt is present in the

drift slumping leads to low slopes. In detail, many moraines consist of a complex series of minor ridges each along some temporary position of the ice margin. Many depressions are left between these ridges and others, called kettles, were made by the melting of buried ice masses. Clay moraines are comparatively inconspicious and have few depressions. Ground moraine is composed almost entirely of till which was either deposited under the ice or left behind when it melted. Where the drift is thin the present topography is a smoothed reflection of an older landscape of some other origin, either erosional or depositional. In regions where the clay content of the till is large, the ground moraine is thick enough to cover Fig the rock topography, forms a nearly level drift or till plain. Such are common in Iowa, Illinois, and other regions where a large part of the bed rock is shale + lake deposit which supplied the clay to the ice. Drumlins are conspicuous only in rather stony drift. They are streamlined hills up to 200 feet high which are composed mainly of till. Their long axes are parallel to the direction of ice movement and the stoss end is the steeper. The flanks are as steep as wet till could stand. In detail the ideal form is in many instances confused by fusion of adjacent drumlins and by change of ice direction. Drumlins attained their shape because they are the form of minimum friction for material which accumulated in cracks in the bottom of moving ice some miles back of the margin. Outwash consists of the deposits of glacial meltwater streams which on leaving the ice flowed over a relatively low slope. Here they formed a braided pattern and laid down much of their load of sand and gravel. If the locality of deposition had not been occupied by ice for some considerable time the resulting topography was a smooth plain. But where residual ice masses had been left from a recent glacial invasion sediment accumulated around and above them. Melting left a confused topography with many kettles which, except for the nature of the material, resembles many terminal moraine deposits. Such rough deposits are called pitted outwash and vary from isolated pits in a plain Fig to areas where no trace of a depositional surface is left. The areal distribution of such deposits is unlike that of moraines in that the longer dimensions are more apt to be parallel to the direction of ice motion instead of transverse to it as moraines are. Eskers represent the filling of the beds of streams which flowed between ice walls. It is hard to tell in many cases whether these streams were in tunnels or in crevasses which were open to the sky. The gravel and sand ridges are discontinuous and have irregular crests. The flanks are very steep. In some cases they have tributaries which tell definitely that deposition took place in dying, stagnant ice. Crevasse fillings are narrow ridges between kettles of pitted outwash plains. Moulin kames are conical hills due to gravel which was deposited Fig in holes which melted through the decaying ice.

Lake deposits. Glacial ice formed the dam which enclosed many lakes. Those which were shut in front of advancing ice contained much open water although in the glacial climate both icebergs and lake ice must have been present most of the time thus damping wave action. During ice recession residual ice masses must also have been present in all low places so that many supposed lakes were simply narrow moats. The land forms made by glacial lakes differ little from the deposits of standing water described above and need no further description. In some places mounds of till within lake beds have been reported. Some of these may have been due to the overturning of icebergs.

Work of lake ice. In middle latitudes where there is a great variation in winter temperatures the lake ice can expand considerably during warm periods when snow has melted off the ice. Where shores are high the ice breaks in an expansion crack but whereever the resistance of low shores is less than the strength of the ice expansion pushes up a ridge. The coefficient of expansion of ice is 50.7×10^{-6} per degree centigrade or about 5 times that of steel. A kilometer of ice Fig raised 20 deg. C. would then expand about a meter. Where ice push ridges contain many boulders they can survive wave erosion and become permanent land forms. Some examples are over 20 feet high and extend for miles. Ice push appears to be uncommon in the far north because the lakes are always covered with snow in winter and the spring thaw is so rapid that the ice soon weakens.

Summary. The work of ice was important over a large part of the northern countries. Glaciers are still present in mountains even near to the equator and in the not remote geolgic past were much more extensive all over the world. Then continental glaciers, like those still present in Greenland and Antarctica, overspread Canada and the northern United States. It has been suggested, however, that many glaciers of the present day are not survivals of these ancient ones but were formed in the Little Ice Age which began about 4000 years ago. Glacial erosion by plucking of previously fractured bed rock is far more important than grinding up of rock. Sapping or frost-loosening of rock is important in the mountains. Erosional forms, such as cirques and fiords, developed by valley glaciers are much more conspicuous than are the results of erosion by continental glaciers. On the other hand, the depositional features of the continental ice and associated meltwaters are far more important than are those of the mountain ice streams. These deposits left a disorganized drainage system with many lakes and swamps and streams superimposed on the older rock topography making falls and rapids. Unlike valley glaciers the continental ice caps were thinned to the point of stagnation at many times during their recession. Some did not even leave any terminal moraines. Others underwent many periods of rejuvination through increase in snowfall and readvanced to make a series of so-called recessional moraines which are really the record of counter-attacks against the sun which finially reduced them to their present limits.

TECHNIQUES

Introduction. Technical methods for the study of problems in geomorphology are varied. In this text a method as yet in its infancy has been attempted, namely the approach through physical controls. Either the mathematical theory is developed and then checked against facts gathered in the field or from maps or the reverse, first derivation of a formula which fits the facts followed by an analysis of the mathematics of the physical processes involved. The tools used in search for data comprise: (a) topographic maps, (b) aerial photographs, (c) profiles, and (d) block diagrams or perspective drawings of other types.

Topographic maps. Topography of an area may be shown on maps by means of (a) contours, (b) hachures, (c) shading, or (d) a combination of two or more of these methods. In this country most maps published by governmental agencies use contours alone. Unfortunately, the ability to use contour maps effectively is attained by relatively few persons. To many these maps have little meaning aside from the spot elevations of definite localities. It is now evident that in the early days of making these maps too much was expected for too little time and travel in the field. Errors and omissions due to failure to visit portions of the area are glaringly apparent not only when the locality is visited but also in aerial photographs. Not only are there mistakes in form and elevation but different streams have been joined together. Even in maps of recent date a field check almost always discloses some errors in portions which are covered with forest or are remote from roads. Many foreign maps use hachures with at most only a few contours. Spot elevations are given on most of these maps. Some surveys have used contours for gentle slopes and hachures for cliffs and steep slopes. Various methods of shading have also been tried, some as though the light came from the upper left corner, others with density proportioned to slope as with hachures. Most of these maps are hard to read and accuracy of detail is doubtful in many instances.

Aerial photographs. Within the last 10 years most of this country has been photographed from the air largely in connection with various "New Deal" activities. In peace time these photographs have been made available to the public at reasonable cost. Aerial photographs are of three general types: (a) verticals, best for mapping and determination of areas, (b) low obliques which do not show the horizon, and (c) high obliques which include the horizon. Some methods of mapping use a combination of the first and last types. We need not here discuss either the methods of taking aerial photographs or the details of making maps from them. It is obvious that photographs record far more detail than does any map. Except under very dense timber nothing is hidden from the aerial camera. In fact the objection is often made that photographs show too much confusing and unessential detail. Aerial photographs may be used (a) singly, in which case obliques show relief much better than do verticals, (b) joined together into mosaics which are not true maps, and which can only be made from verticals, (c) examined in pairs taken from different positions of the plane in such a way that only one is seen by each eye, that is stereoscopically, or (d) superimposed by various methods and then examined by means of special glasses so that only one is seen by each eye. Both verticals and obliques may be used in the two last ways, and so can pictures taken on the ground provided only that they were taken far enough apart and yet show the view at very nearly the same size. Several types of instruments are used to enable one to see one picture with each eye. These are known as stereoscopes but with practice many persons can dispense with all instruments and make the eyes look along parallel lines. Various methods of attaining this method have been tried. One is to relax until a single object appears double; if an effort is made the eyes automatically converge so that a single object is seen. When double vision is attained then a pair of photographs can be held up with common points from 2 to 2½ inches apart. Suddenly you will realize that three images only can be seen and that the central one, where in fact two are superimposed, is in relief. A stereoscope is essential, however, to measure differences in elevation on verticals by comparison of distance between common points which varies with elevation. The method of superimposed pictures in different color, or with light polarized at right angles, produces what is called an anaglyph. Such pictures are simply a blur until viewed through the proper kind of glasses. Maps can be made in this way, although so far only of limited areas; spot elevations or contours can be shown to give quanitative data on anaglyphs. It is clear that the use of aerial photographs opens up new fields to geomorphic study. This is particularly true where the mantle rock is thin and vegetation is sparse. Then the structure of the underlying rocks is very clear. Faults and folds may be traced accurately by the varying topographic expression and mantle rock of each kind of bed rock. Study of glacial deposits in which the surface form plays an important part is greatly facilitated by aerial photographs. Drumlins, moraines, eskers, and outwash all stand out clearly. Outwash is distinguised in many localities by the sandy soil which photographs a light color, and pitted outwash shows a mottled pattern with many rudely circular swamps and lakes. The lakes in terminal moraine are for the most part irregular in outline.

Profiles. Many students of geomorphology place much importance on profiles or cross sections. Of necessity, these are made from topographic maps and are no more accurate than they are. Profiles are drawn on cross section paper with the vertical scale considerably larger than the horizontal scale which may or may not be the same as the map. They may show only the surface along a single line, either straight, curved, or a series of lines at angles, or they may be of the projected type. If the latter two distinct methods have been employed. Barrell, in a search for ancient marine terraces in New England, drew the crests of hills beyond the front line of his profiles back so far that the result is confusing. Each summit was projected on a line at right angles to that of the section. Others have used a strip of definite width and shown the highest elevation in this strip which is present at right angles to the front line at every point. This is equivalent to showing the skyline of a model of the landscape of this strip which may be a mile or more in width. Johnson drew these profiles on cardboard, cut them out along the line and then set up each profile vertically on a map in the location of the front of each strip. The result simulated a relief map. Just how this method is better than coloring in areas of different ranges of elevation to make a layer map has never been clear to the present writer. Except where the geology of large areas is uniform any profile which does not show the underlying materials is worthless. After all, the nature of the bed rock is the predominant control of surface form and it hardly seems possible that erosion has ever continued undisturbed until all such control was destroyed. Profiles have generally been drawn in order to discover and correlati ancient, now-dissected erosion surfaces. This quest is one fraught with great chance for error. Just why should an erosion surface have a definite level? In recognizing one should we use hilltops or valley bottoms of that time? How much relief should we allow? How should we guard against the subconscious influence of the horizontal lines of the cross section paper? If surfaces were warped during uplift just what kind of curve must we assume? Are not the ranges of elevation of each surface as great or greater than the intervals between the erosion levels? How could older surfaces survive while lower ones were made close at hand unless our usual concepts of origin are mistakes? And finally, are the usual explanations of areas which are too high for a given "level" as either residuals (monadnocks) or as upwarped areas and of areas which seem too low as downwarped areas or due to subsequent erosion just too easy? Are not many of the morvans or lines of intersection of two peneplains more likely the result of warping? Although we should not discredit all conclusions based on profiles it appears wise to accept may of them "with reservations".

Block diagrams. Block diagrams, perspective drawings, and perspective maps are of more importance in explaining than in solving problems. They are particularly adapted to an exposition of a theory of geomorphic history. The subjective element in drawing them is too great to make them an unbiased delineation of facts. Perspective maps are drawn on a map by raising the position of elevated points an amount proportional to their heigth above datum. The vertical scale used in this is always greatly exaggerated. This type of map has been employed in the several physiographic diagrams which have been published. It helps in giving an illusion of depth to show a cross section of the underground structure along the front edge. True block diagrams are intended to show the relation of underground phenomena to surface form in a certain relatively limited area. Vertical scale is almost always exaggerated. Positions in the horizontal plane may be shown either in true perspective or in isometric projection. In the former all lines which would be parallel on a map converge to a point known as the vanishing point. There is a different vanishing point for each set of parallel lines but all are located on a straight horizon line. In this type of drawing the scale decreases with distance from the front. In isometric drawings lines which are at right angles on the map are skewed to angles of 30 and 60 degrees. Distances along these lines are not altered. Various machines have been made as an aid in making these drawings because the redrawing of a map on a skewed or perspective base is laborious. All of these are hard to construct because sliding joints are required which must be hard to keep in adjustment. Relief is shown by the profiles along the sides aided by shading, hachures, and raised contours. Difficulty in drawing decreases in the order given. Drawing of hachures requires considerable practice. Published diagrams look much more complex than they really are because they were reduced in scale by photography. The illusion of distance is enhanced by making the lines fainter and nearer together for the same slope as the back of the drawing is approached. Crests are generally shown by full lines. Changing of a photography to a perspective drawing is sometimes desirable in order to bring out certain features. The work of others should be studied with a lens before attempting to do any of this kind of work.

Summary. Choice of methods for study and for showing data and interpretations in a report depends largely on experience. The trend is now definitely toward use of aerial photographs including stereo-pairs and stereo-triplets which are less expensive to reproduce than are anaglyphs. Well-drawn three-dimensional diagrams are, however, of great value and are much cheaper to publish than are photographs. The practice of publishing profiles without showing the geology is to be condemned.

GENERAL SUMMARY

In the foregoing pages the writer has attempted to approach geomorphology from the standpoint of what must ensue from given physical conditions rather than by means of purely abstract reasoning. It is true that in nature there are many variations from the ideal of homogeneous material which must be assumed for mathematical analysis. Moreover, there are many topographic forms where more than one process is at work at the same time. These facts must be kept in mind when evaluating the theoretical conclusions. It is obvious that much more field data must be accumulated in order to check the theoretical deductions. It is better to make original observations than to trust to even the best maps.

A major error in the past has undoubtedly been the assumption that the present climate of an area should be projected back into the remote past. Some seem to regard as "normal" the type of humid climate which is now widespread in middle latitudes where the majority of scientific workers live. But if we look at the relative areas of different types of climate today it is apparent that such a classification is unreal. And when we look back to the time when there were no polar ice caps, and epicontinental seas were larger, it is evident that, aside from mountains, an ever larger proportion of the lands must have had a semi-arid or arid climate during most of their history. For this reason caution is necessary in the interpretation of fossillandscapes which were long preserved under a covering and only recently exhumed. Which are true peneplains and which are pediments? Certainly those which have a gravel cover are not peneplains in the classical sense. It seems as if the emphasis that has been placed on the discrimination of remnants of ancient peneplains was, to say the least, unwise. Let us try rather to build up a foundation before we attempt this uncertain pursuit. Let us consider also the
t alternative interpretations which are possible for many of the phenomena and not
t give too much weight simply to reputation of proponents of some of the ideas.
t The future will decide which interpretation is correct but only after the necessary physical data has been gathered.

The interpretations here presented on the discrimination of talus and creep slopes appear well supported. Their application to the slopes of volcano s made of fragmental material obviously eliminates some older attempts at mathematical analysis on the bais of shearing.

But so far the least success has attended attempts at mathematical treatment of the work of running water. It seems certain that the quanitative computations of force made by many students are misleading in that they consider only total energy instead of the small portion which is actually expended in erosion and transportation. Possibly Little's likening of the process to the loss of pressure of water in flowing thorough a pipe may lead the way to ultimate solution. In the meantime more data of grades of streams on uniform material and on slopes due to rain wash is greatly needed. Such data is hard to find and needs careful study. In this connection attemtion should be directed to the force opposed to erosion, resistance of material to removal. This factor has been ignored in the past. The work of Horton on this, as well as the quanitative relations of streams, is a step toward the solution of some of these important problems.

The present work has added little to problems of the work of standing water other than to point out the desirability of a new classification of shorelines.

The mechanics of the work of wind are, thanks to the careful work of Bagnold, now well in hand. Physical differences in the behavior of wind and water are now clear. Air cannot absorb so much energy in kinetic energy of rotation as does denser water. The problem of the origin of longitudinal or seif dunes is still unsettled.

In considering the work of ice, attention has been directed to the relative importance of plucking versus grinding of bed rock. The work of Demorest on the physics of glaciers points the way to solution of many problems such as the true origin of roche mountonee forms. Drumlins are now recognized as accumulations in cracks shaped to minimize resistance to moving ice. The fact that glacio-fluvial deposits consist of a mixture of sand, pebbles, and ice is also demonstrated.

The value of topography in the discrimination of structure of both sedimentary formations and lava flows has long been known. Smooth dip slopes are also important in gently inclined strata. Hogback ridges where dips are steep are harder to interpret. Overturned folds and repeated thrust fauls offer difficulties.

2

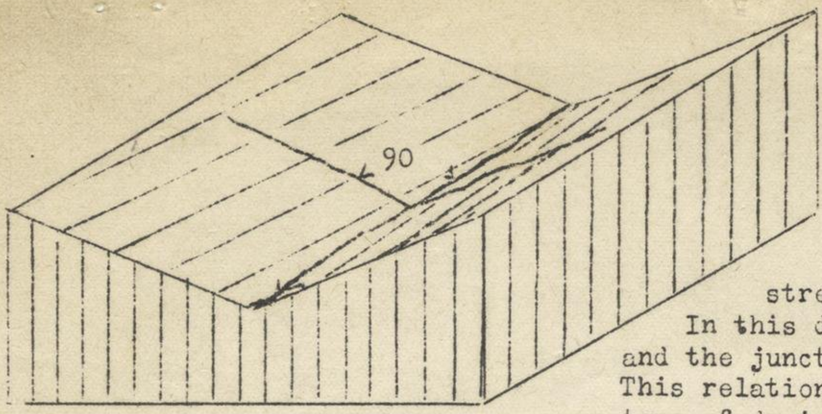


Fig. 69, p. 44

Entrance angles of tributary streams. Horton concludes that cosine of thie angle is measured by ration of slopes (tangent of angles) of main stream divided by slope of tributary. In this diagrams this ratio is very small and the junction angle is almost 90 degrees. This relationship is common in <u>trellis</u> type of drainage pattern. Slopes shown by contours. Streams on slopes are at 90 deg. to contours. G. S. A. B. 56: 349

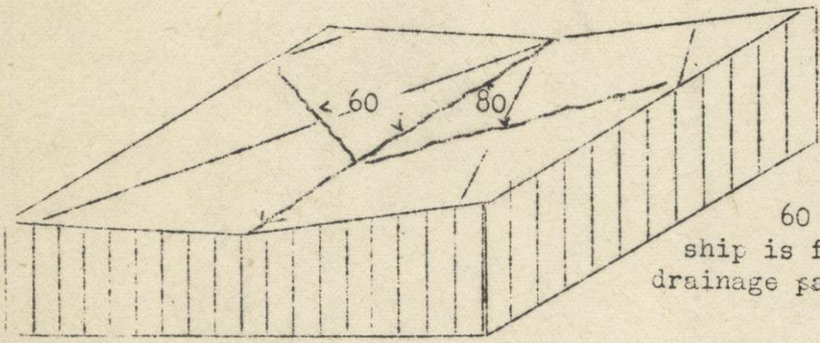
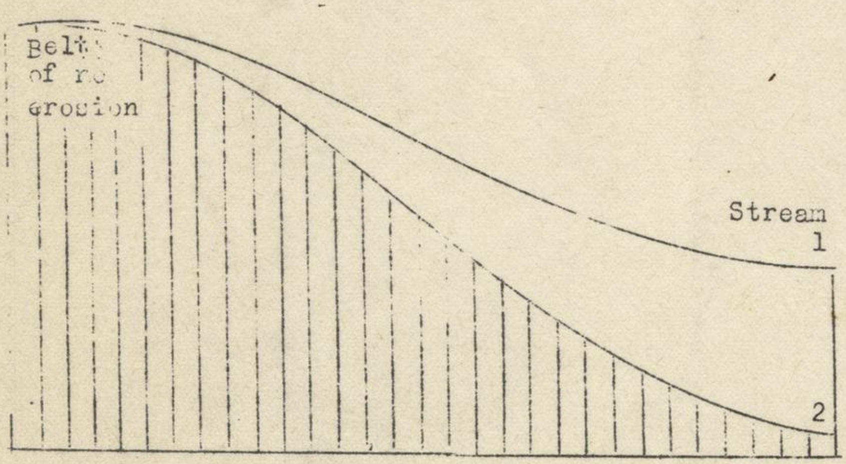


Fig. 68, p. 44

Entrance angles of streams where the difference in slopes is not great gives angles of 60 to 80 deg rees. This relationship is found in so-called <u>dendritic</u> drainage patterns.



Fig, 81, p. 47

Lowering of the stream level from 1 to 2 would reduce width of "belt of no erosion" by increas e of slope. Width of this belt also varies with intensity of rainfall. The net result of both factors is to round off the upper slopes into convex profiles like t hose due to creep.

Belt of no erosion

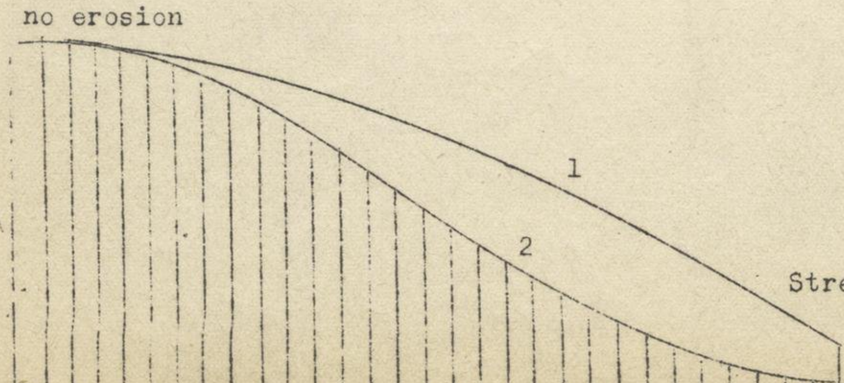


Fig. 82, p. 47

Lowerin of the stream bed more rapidly than slope wash can shape the valley side should cause a convex profile as shown above. he lower compound s lope ought to be formed later when the stream is not lowering its bed so rapidly.

N

See G. S. A. B.
56: 297

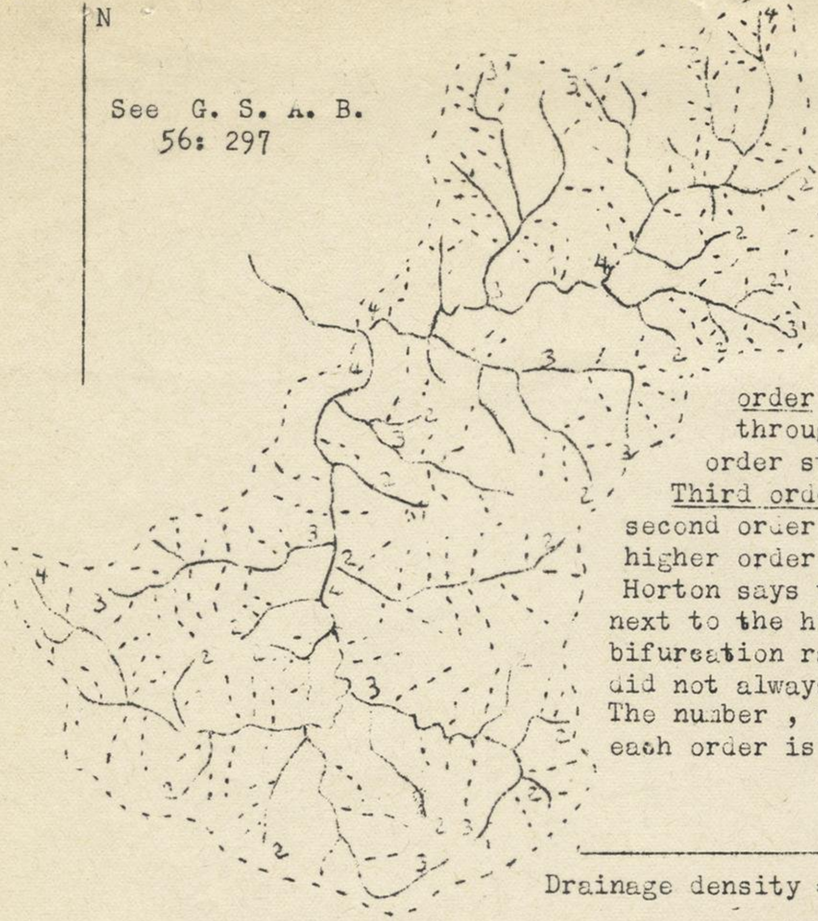


Fig. 70, p. 44

Part of bas~in of Hiwassee R., Georgia, showing Horton's system of stream orders. Bed rock is schist, but the drainage is "dendritic" or "branchwork". Horton starts with unbranched streams which are <u>first order</u>. , Streams which are joined by these are <u>second order</u> but the numbering is carried through to their headwaters. First order streams are dotted in this map. <u>Third order</u> streams have one or more second order tributaries and so one each higher order being kept up to the head. Horton says that the number of streams of next to the highest order determines the bifurcation ratio but in his computations he did not always follow this rule exactly. The number , length, and average length of each order is shown in the table.

Order	No.	Length	Av.l.
1	146	72m	0.49m
2	32	41	1.28
3	9	32.8	3.65
4	2	24.6	12.3

189

Drainage density = 2.06 miles

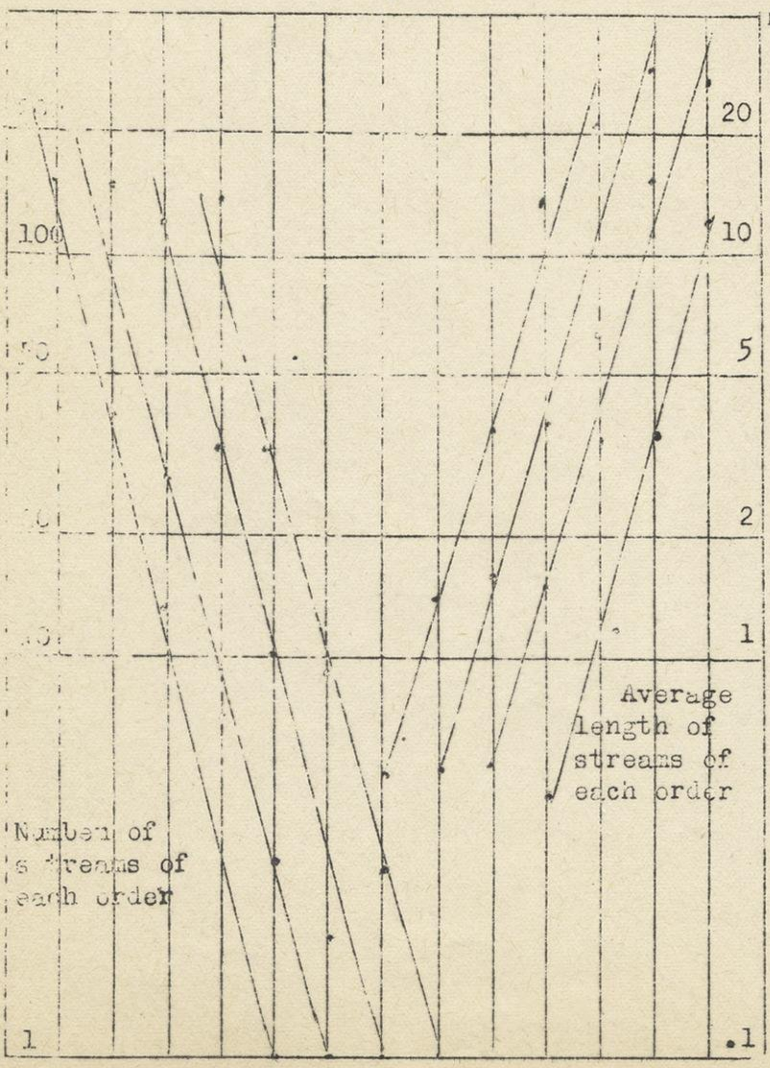


Fig. 71, p. 44

Mathematical relations of number of strea~ms of each order and average length of each order after Horton. (p. 303) By platting the logarithms of these quantities it appears that both follow geometric progression. At left the <u>bifurcation ratio</u> is raised to powers which are found by subtracting t he number of stream order from the order of the highest order stream in the basin. Horton used for this value 3.45, 3.0, and 3.15 respectively for the three streams at the left. For the Hiwasee, which was added to the original diagram, the figure appears to be about 3.45

The relation of stream lengths follows a similar rule where the <u>stream lengt~h ratio</u> is used. This is found from actual ratio of lengths of two successive orders. It is raised to a power which is one less than the order and multiplied by the length of first order streams. Horton gives values of 2.70, 2.85, and 2.92 for first three at left. The Hiwasee should have about 2.62 Shown at right.

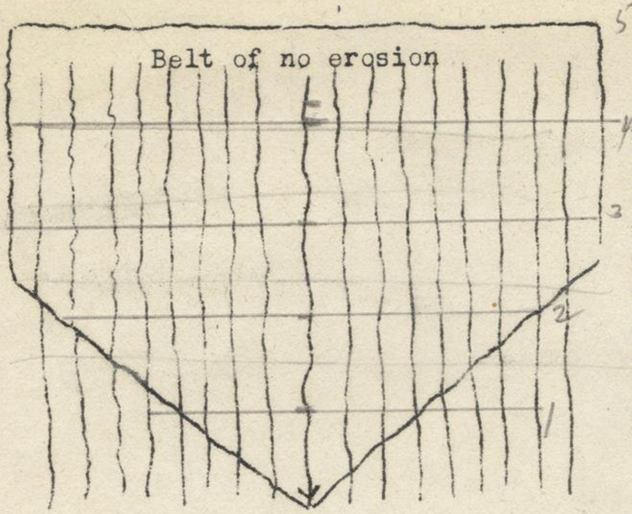


Fig. 73, p. 45
Initial stage in formation of rill valleys by natural concentration of slope or sheet wash. One rill in the center developed faster than the others which successively joined it until the diagonal divides in lower part of the area developed. Then the land above these was worn down until it sloped to the one central valley. The idea is Horton's but drawing is modified.

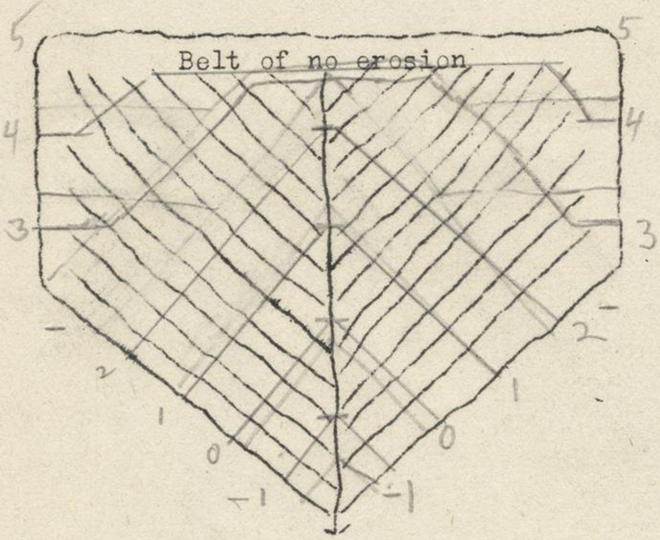


Fig. 74, p. 45
A new set of rills developed on this gently sloping surface on both sides of the main strea m. The central one on each side absorbed the drainage of the others and there are now two tributaries to the main stream. Note development of belt of no erosion along the lateral divides.

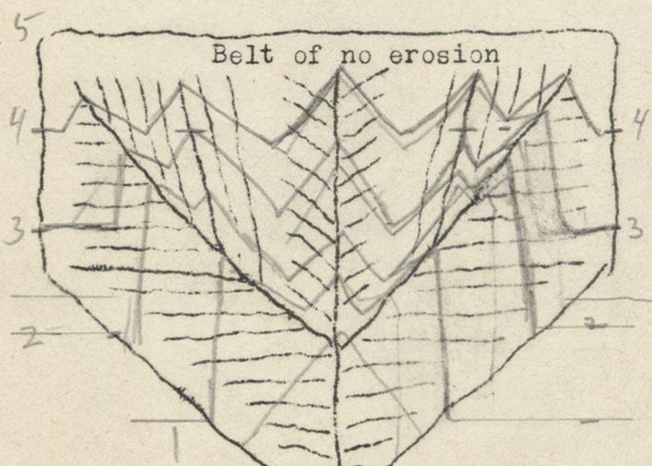


Fig. 75, p. 45
The process outlined for Fig. 74 has now divided up the areas tributary to the two new streams and led to formation of four new tributaries. Belts of no erosion were thus formed along the new divides. Horton held that this process of subdivision of drainage areas would continue until the area of no erosion (except in exceptional storms) embraced all the interstream spaces. The process was held to explain the observed fact that the number of streams of each order follows an inverse geometric progression. In the diagrams (which do not complete the process) we have in increasing order of streams, 1, 2, 4. The bifurcation ratio is 2.

It seems probable that such stream development would occur mainly on regions with some original slope. It is not readily observed in most areas of young drainage such as the glacial, alluvial, and marine plains of the United States.

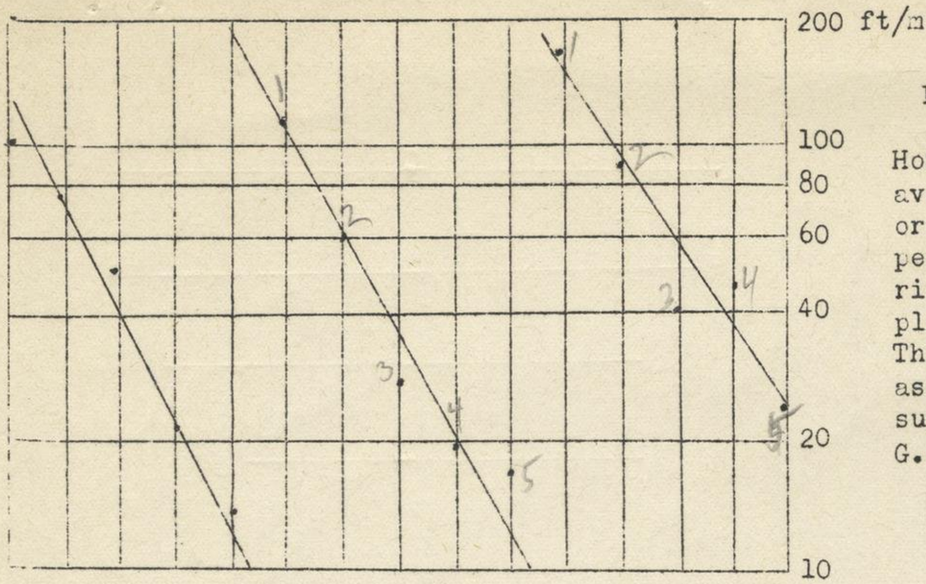


Fig. 72, p. 45

Horton's diagram showing average slope of different orders of streams in feet per mile. A rough geometric progression is shown by platting logarithms of slopes. This relationship was cited as lending quanitative support to Playfairs Law. G. S. A. B. 56: 295

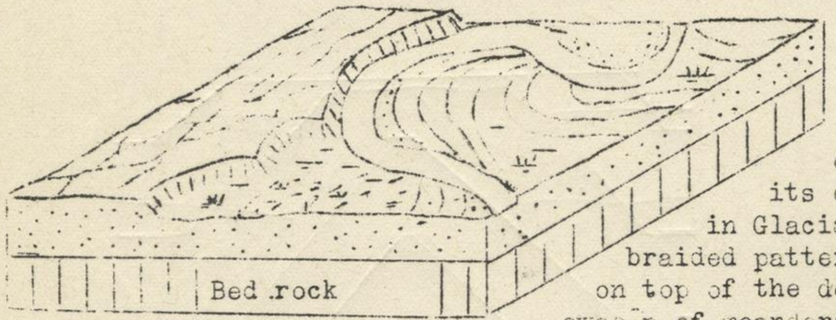


Fig. 101, p. 52

Lateral growth of a meander of upper Wisconsin River during its erosion into a delta deposited in Glacial Lake Wisconsin. Note the braided pattern of the stream when flowing on top of the delta. evidence of downstream <u>sweep</u> of meanders, and the <u>meander scars</u> separated by sharp <u>cusps</u>. Drawn from an aerial photograph.

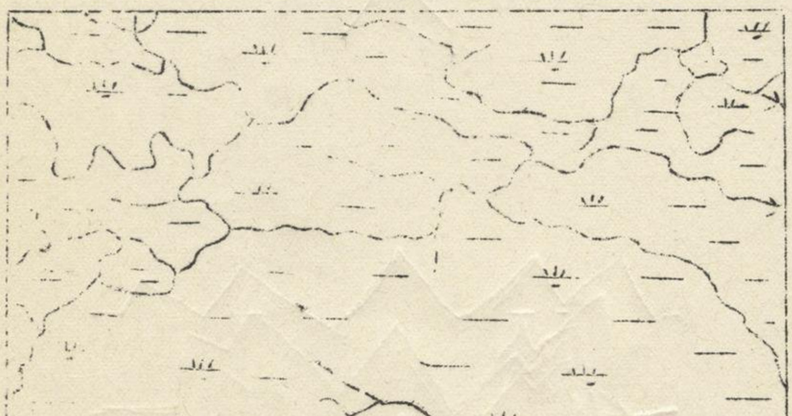


Fig. 102, p. 45

Network stream pattern in Louisiana swamps. The only dry land is on natural levees. Russell, A. A. P. G. 23: 1218

$y = n \log x + \log C$

$y = C \cdot x^n$

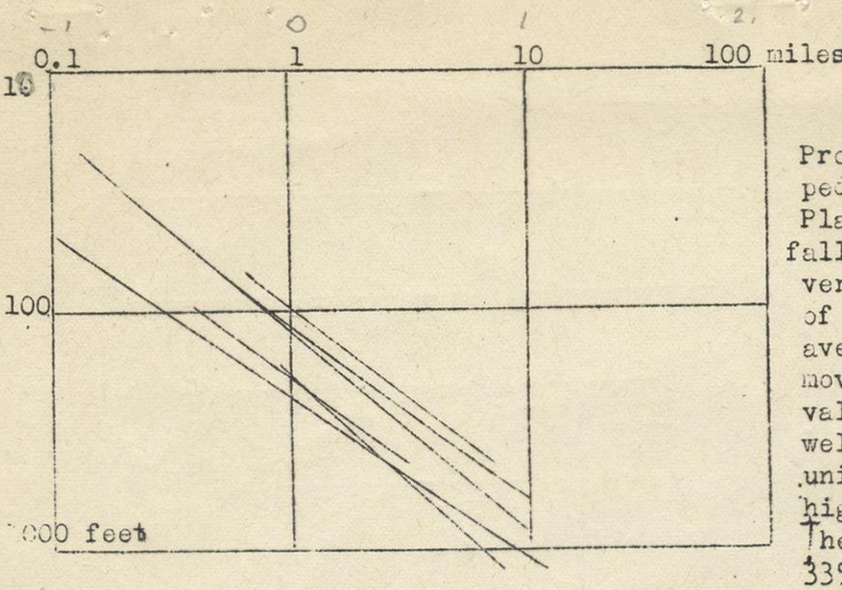
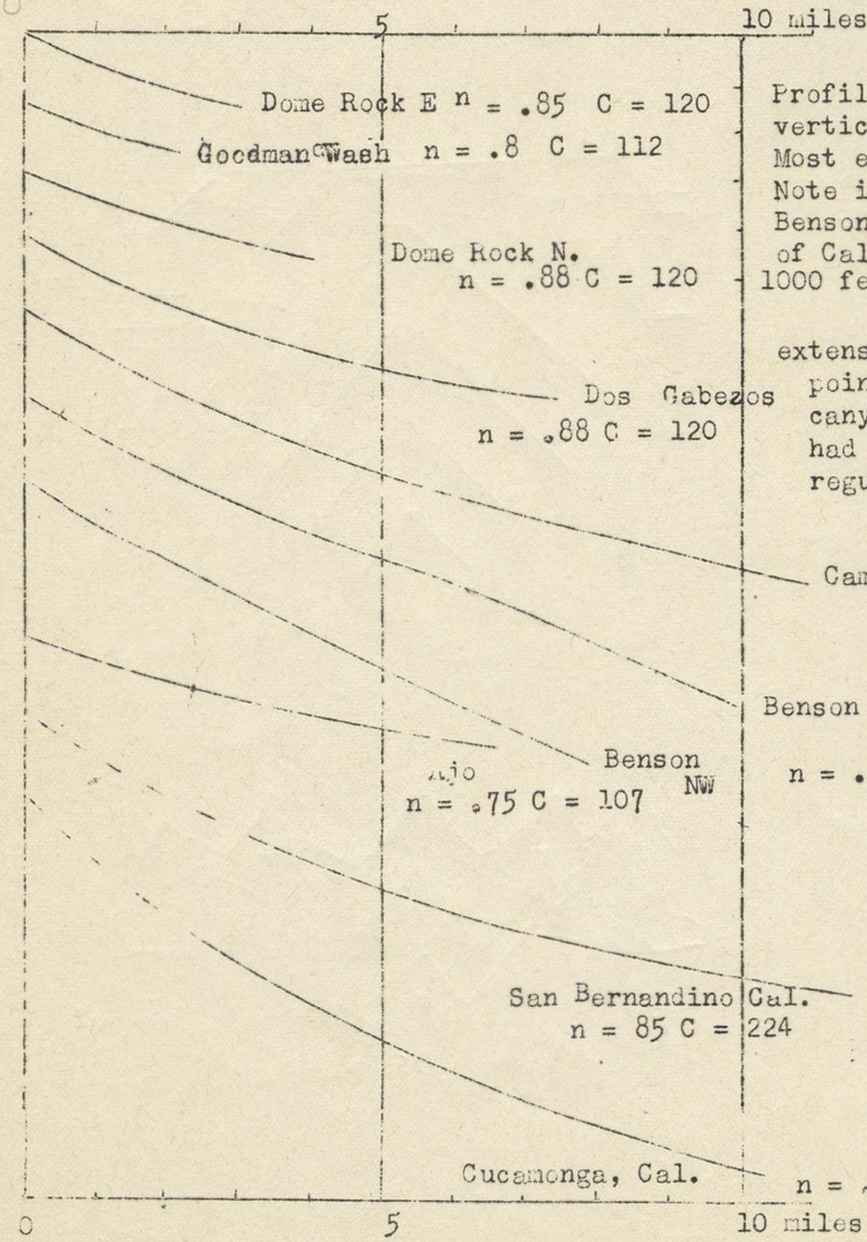


Fig. 106, p. 53

Profiles of alluvial fans and pediments from topographic maps. Platting of horizontal distance and fall to log scale demonstrates a very close agreement although parts of a few examples depart from the average. This may be due to earth movement or to erosion. Average value, n = 0.79 which agrees well with Little's profile of uniform erosive force. It is much higherthan for outwash plains. The constant ranges from 107 to 339, also higher than for outwash.



Profiles on linear scale with vertical exaggerated 17.5 times. Most examples from Arizona maps. Note irregularity in those from Benson quadrangle. Upper parts of California examples were also not in agreement with the average although extension of the curve fits with points of origin at mouths of canyons. Many points of origin had to be chosen at top of the regular slope.

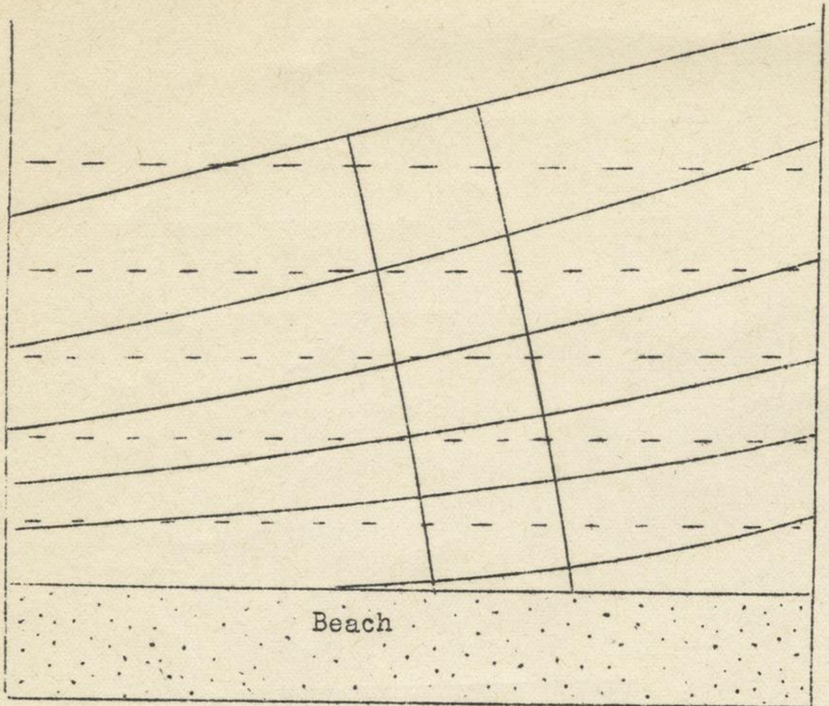


Fig. 123, p. 62

Refraction of waves on straight shoreline. After Munk and Traylor, J. G. 55: 3. Depth contours dotted, wave crests solid, lines normal to waves are called <u>orthogonals</u>. As waves enter s hallow water the circular orbits of particles of water change into ellipses. This slows the wave and causes refraction in the same way that light is affected when it passes from a medium with a given velocity to one with a different velocity. The orthogonals bound areas of constant energy.

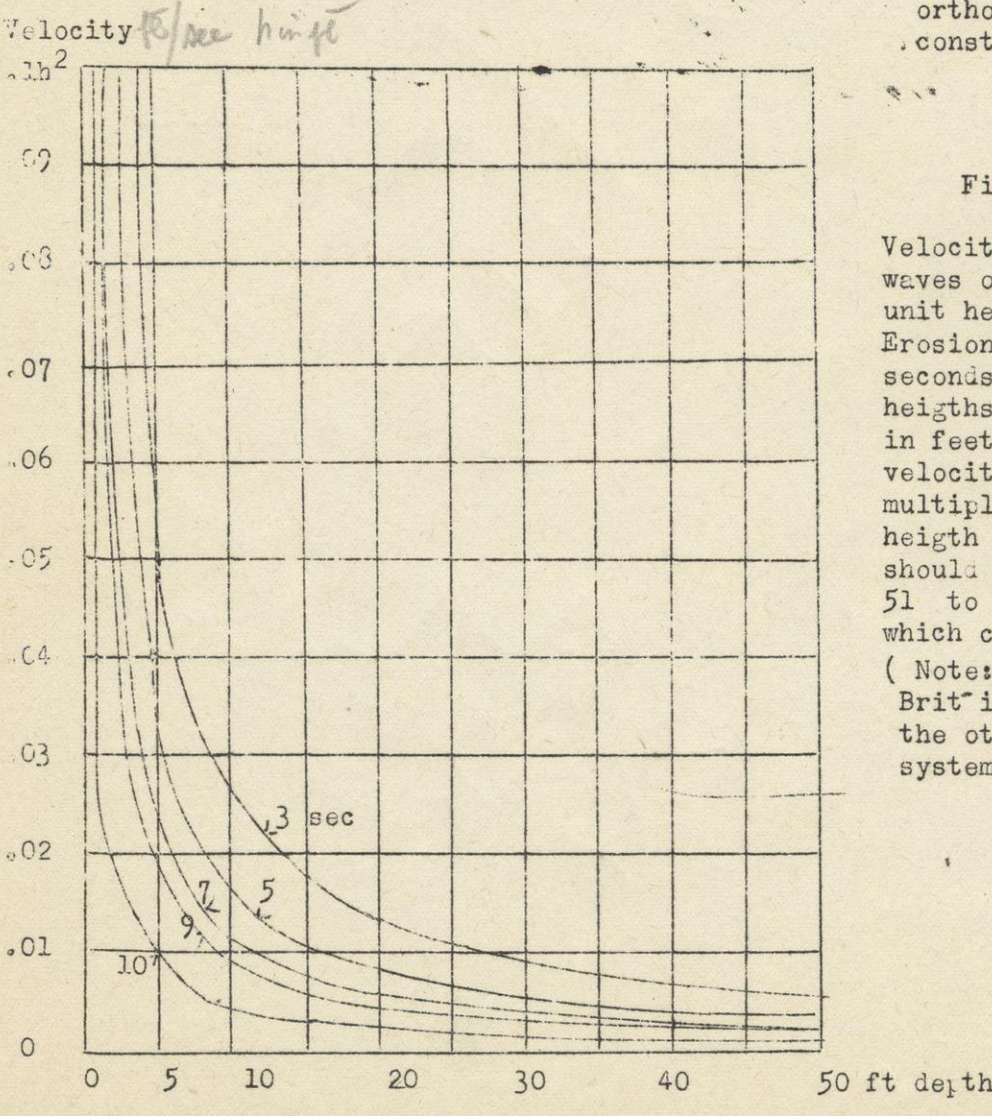
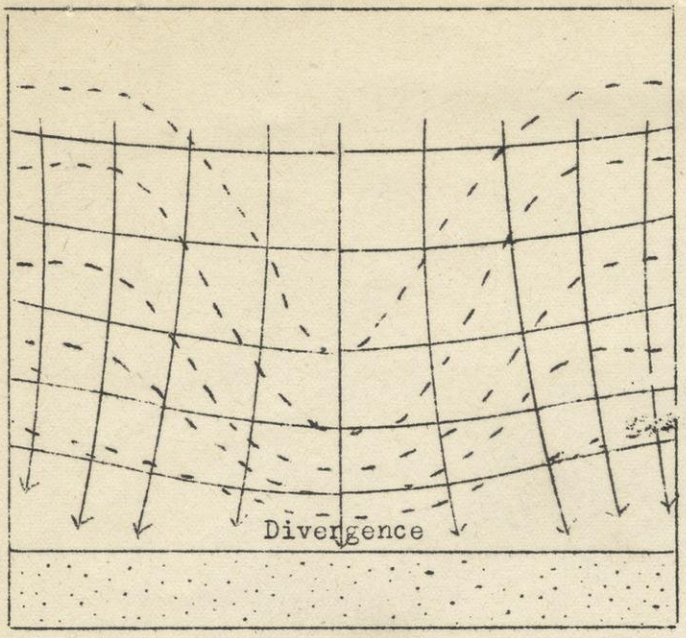


Fig. 122, p. 62

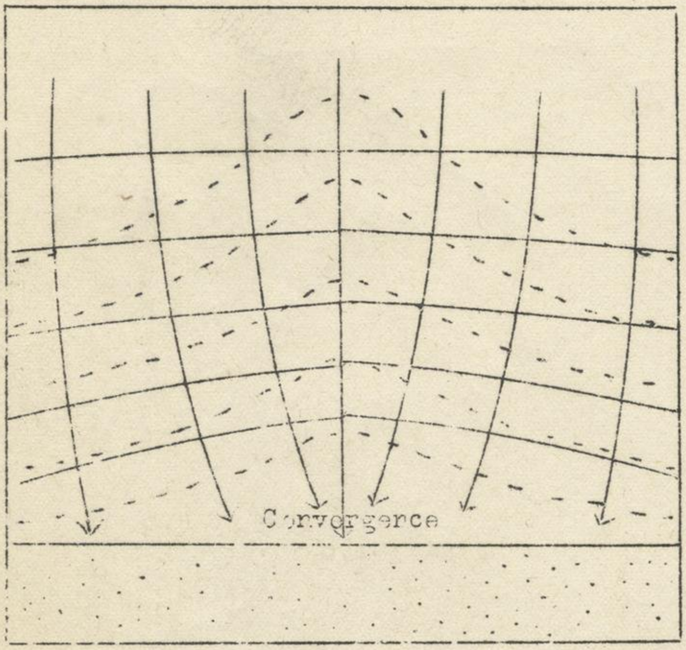
Velocity of undertow for waves of different periods and unit heigth. After Beach Erosion Board. Periods in seconds, depths in feet, heigths in feet, velocities in feet per second. To obtain velocities for larger waves multiply by square of wave heigth in feet. Results should be compared with Fig. 51 to find sizes of material which can be transported. (Note: this figure is in British Engineering units, the other in the metric system.)



Beach

Fig. 124, p. 62

Refraction of waves on coasts with different forms of bottom. After Munk and Traylor. Depth contours dotted, wave fronts and orthogonals solid. Where the last diverge energy is spread over a wider area and surf is lower. The reverse occurs where they converge. Deep water near shore produces a minimum area of wave work. The converse occurs where a shoal extends seaward as in the lower diagram.



Beach

Convergence of waves over a shoal may result in building out the shore into a cuspate cape. See Fig. 136.

Energy of a wave in deep water is wholly potential and is shown by 1/8 of product of unit weight of water times square of heigth. In the breakers both potential and kinetic energy are exerted following the formula:

$$E= \frac{W}{L}\left(\frac{4\ \text{depth.heigth}}{3}\right)^{\frac{3}{2}}$$

where W = unit weight of water and l = wave length. Depth of breaking is 1.28 times ehigth of breakers.

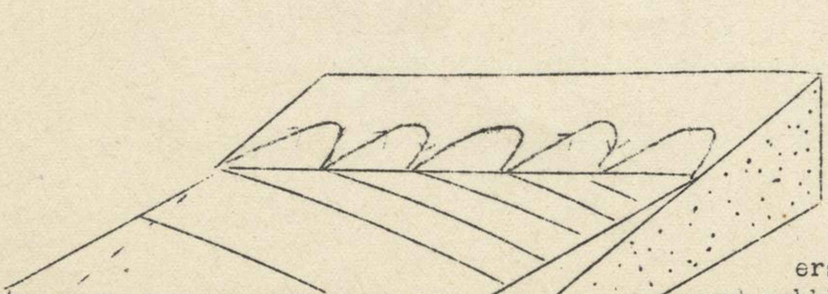


Fig. 125, p. 64

Transportation of material on beach in zone of breakers. Each wave carries sand and pebbles up in the direction of its advance. The returning undertow moves it directly down the slope. Thus material is advanced downwind along the beach. This proces s appears to be much more important than currents offshore.

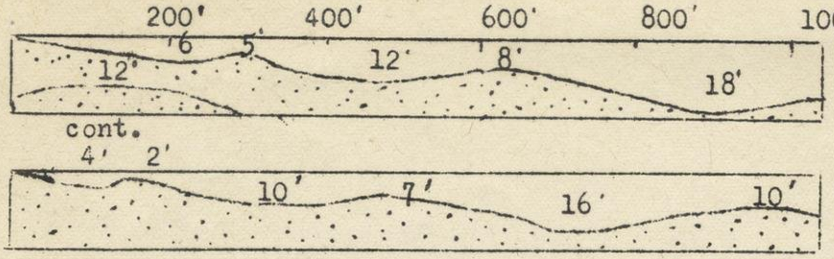


Fig. 128, p. 63

Subaqueous ridges after Evans, J. G. 48: 490 Why some call these "low and ball" is not known. Examples from Lake Michigan. Averages: feet from shore, feet depth, 237, 4.1; 602, 8; 1094, 12. Distances seem to have a ratio of 1, $2\frac{1}{2}$, 5; depths about .7 power o distance. Emerged ridges of this kind could cause serious error in determining former water levels.

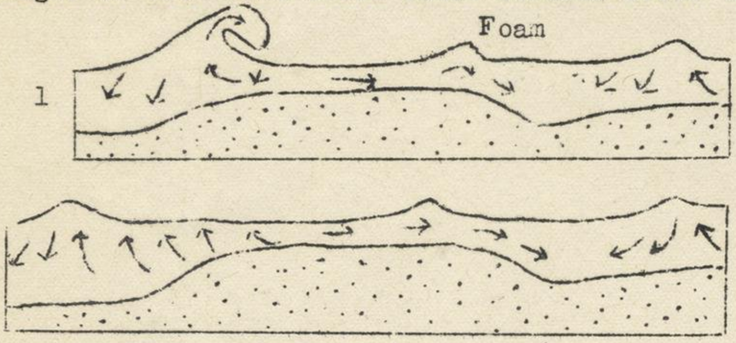


Fig. 129, p. 64

Evans' interpretation of currents in breakers above subaqueous ridges. Observation shows that waves break over each ridge and that intervals decrease toward shore. Relation to size of waves is not known.

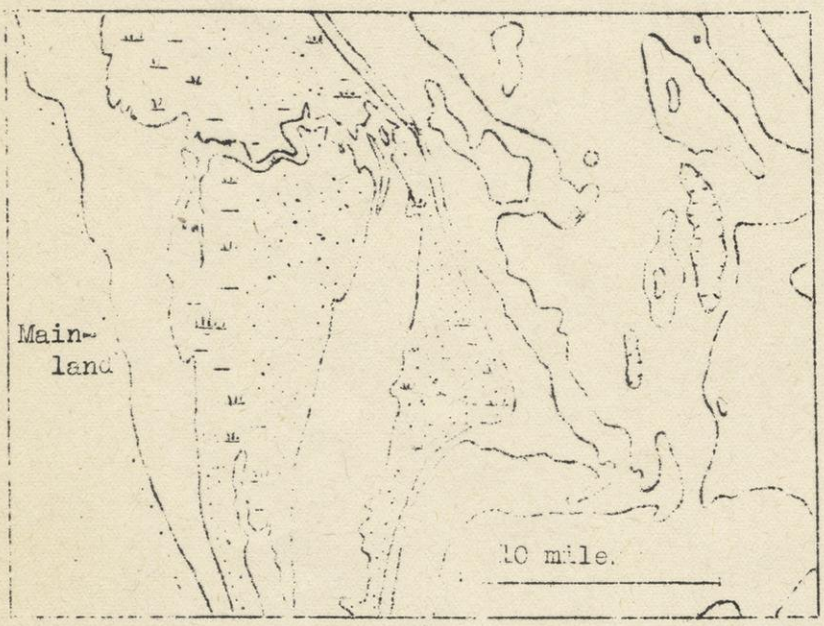


Fig. 136, p. 65

Cape Canaveral, Florida, after Coast Chart 161. Contour interval 20 feet. Such capes have commonly been ascribed to currents but association with shoals strongly suggests that wave refraction is an important factor. Much of the dry land is composed of dunes

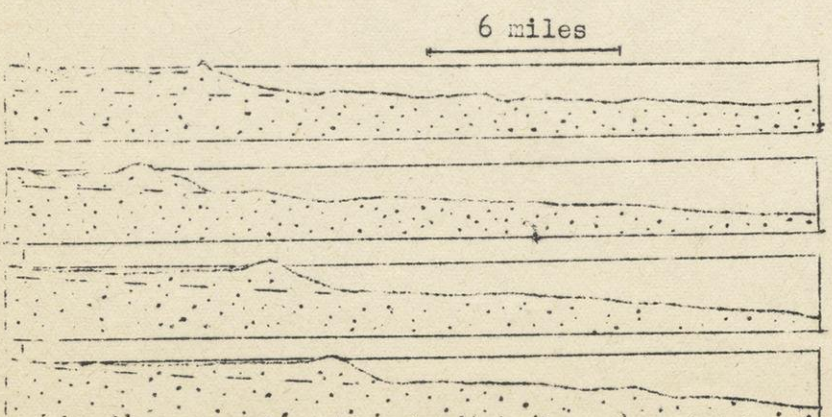


Fig. 135, p. 64

Cross sections of offshore barrier ridges after Johnson. Vertical scale exaggerated x 79.2 Dotted lines is projection of average slope of sea bottom outside the barrier. Only in the last (from Texas) does this line reach the inner side of the lagoon. In this case the barrier might have been built by transportaion along shore; all others were thrown up from the sea.

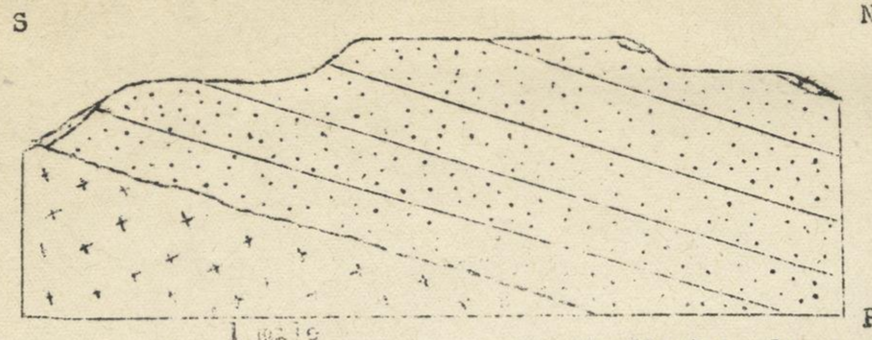


Fig. 137, p. 66

Happy Hill, near Baraboo, Wis. showing possible levels of ancient marine planation. The lower terrace corresponds to the Prairie du Chien dolomite, the highest to the Platteville dolomite. Both have conglomerate at the margins. Vertical scale exaggerated about 5 times.

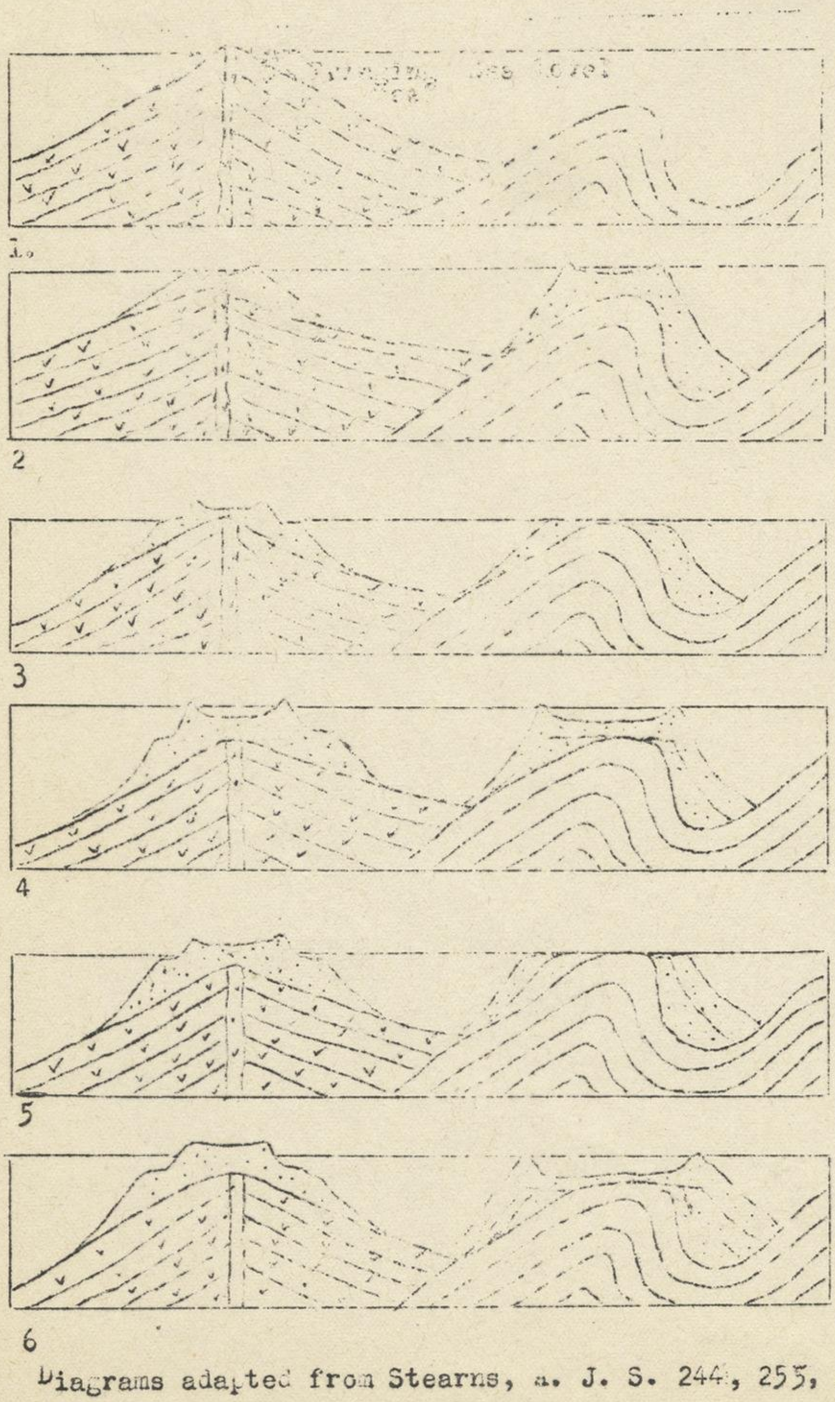


Fig. 138, p. 67.

1. Ideal diagrams showing stages in development of atolls on volcanic and anticlinal structures, latter still too deep for coral growth.

2. The volcano has sunk and the anticline risen so that both are capped by atolls.

3. Sea level has fallen due to glaciation. A beach terrace has been eroded into the left atoll and the atoll at right has been truncated by wave action.

4. Sea level has risen in an interglacial interval so that new atolls have formed on both shoals.

5. During the following glaciation the level of the sea is low. The left atoll is now emerged and that at right has been truncated.

6. Present conditions are intermediate between glacial and interglacial. The lft atoll is still high and a new one has formed at right.

Diagrams adapted from Stearns, A. J. S. 244, 255, 257.

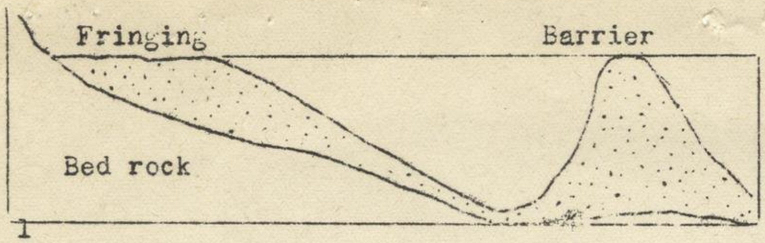
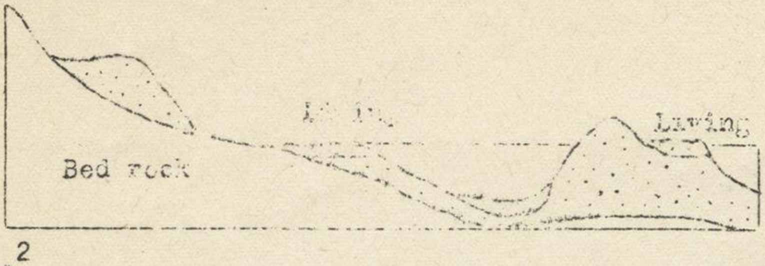


Fig. 139, p. 67

Stages in growth of fringing and barrier reefs, after Stearns, A. J. S. 244: 25 9



Emergence of old reefs due to glacial lowering of sea level. Formation of new reefs.

Sea level higher than present in an interglacial interval. New Reefs formed.

Conditions at present

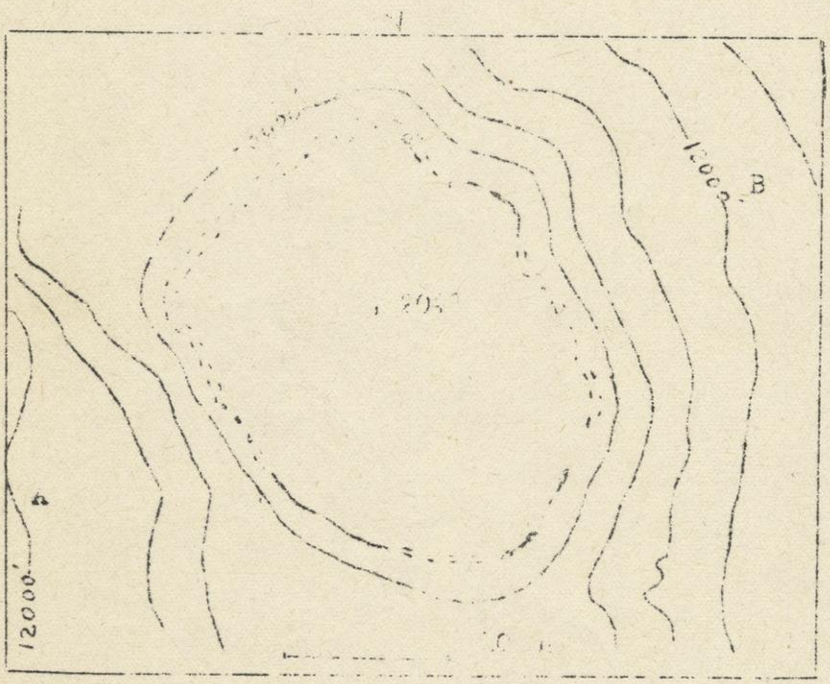
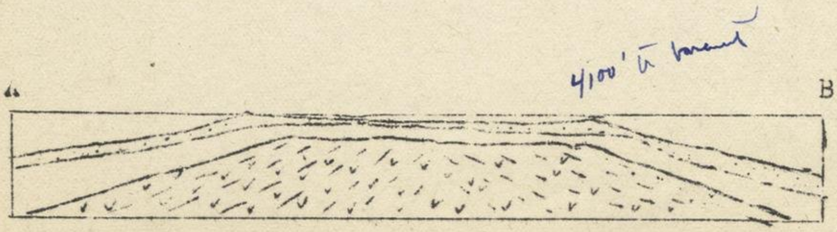


Fig. 140, p. 67

Eniwetok Atoll , Marshall Ids. after Nugent, G. S. A. B. 57: 747. Contour interval 500 fathoms= 3000 ft . Only a part of the reef rises above high tide level. Section A-B drawn from contours with probable geolog y added from geophysical exploration at Bikini Atoll. This discloed a zone to depth 2000' with velocity 7000 ft/sec.; this is coral. Underlying is 3500 to 11000' with velocity 11,000 ft./sec.; this may be fragmental volcanic material. B elow that velocity is 17,000 ft/sec apparently in volcanic rocks.

obsolete



obsolete

Vertical and horizontal scales the same

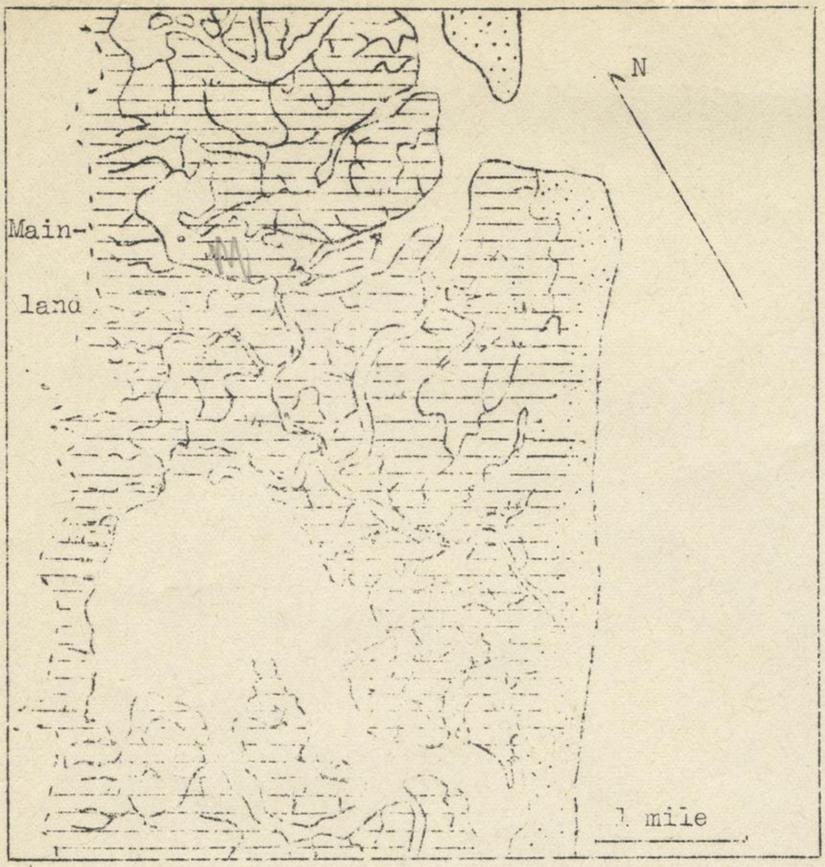


Fig. 141, p. 67

Salt marsh behind barrier beach, New Jersey. After Chart 123. The complex stream pattern is due to tidal flow except in the case of mainland drainage. This tidal flow also keeps open the break in the barrier.

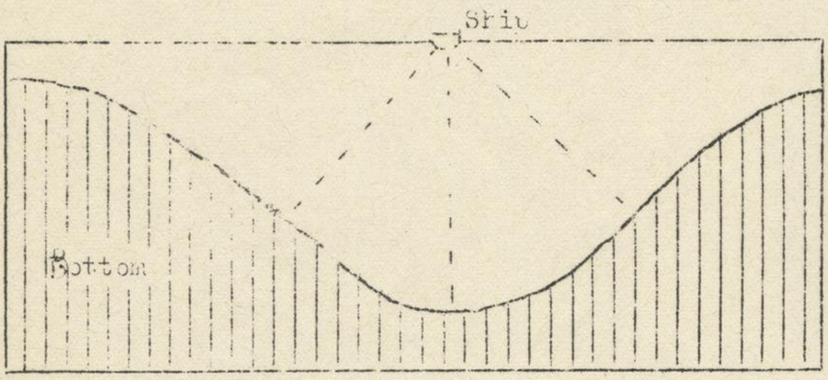


Fig. 142, p. 68

Echo sounding with supersonic vibrations. Over rough bottom more than one signal may be returned. Motion of the ship prevents very close beaming of signals.

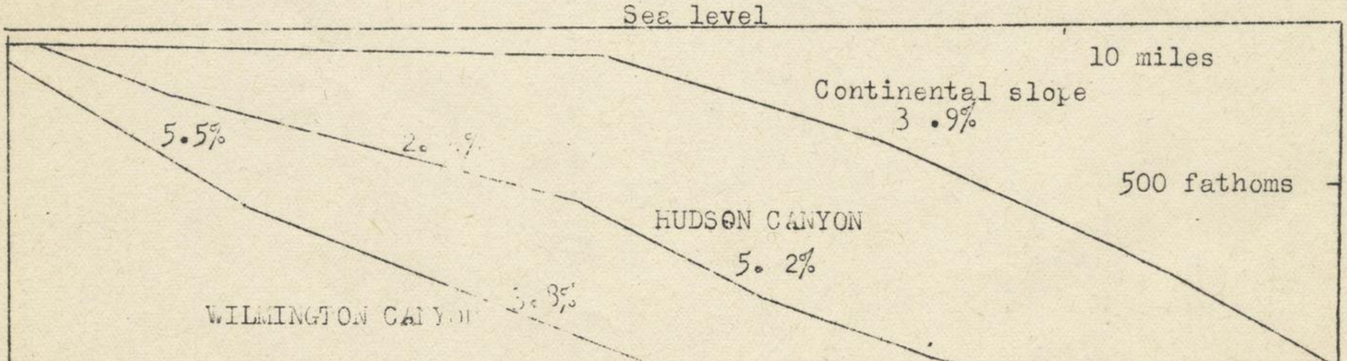


Fig. 146, p. 68. After Veatch and Smith, Spec. Pub. 7: 16 Longitudinal profiles of two submarine canyons of Atlantic coast. Note high and irregular grades.

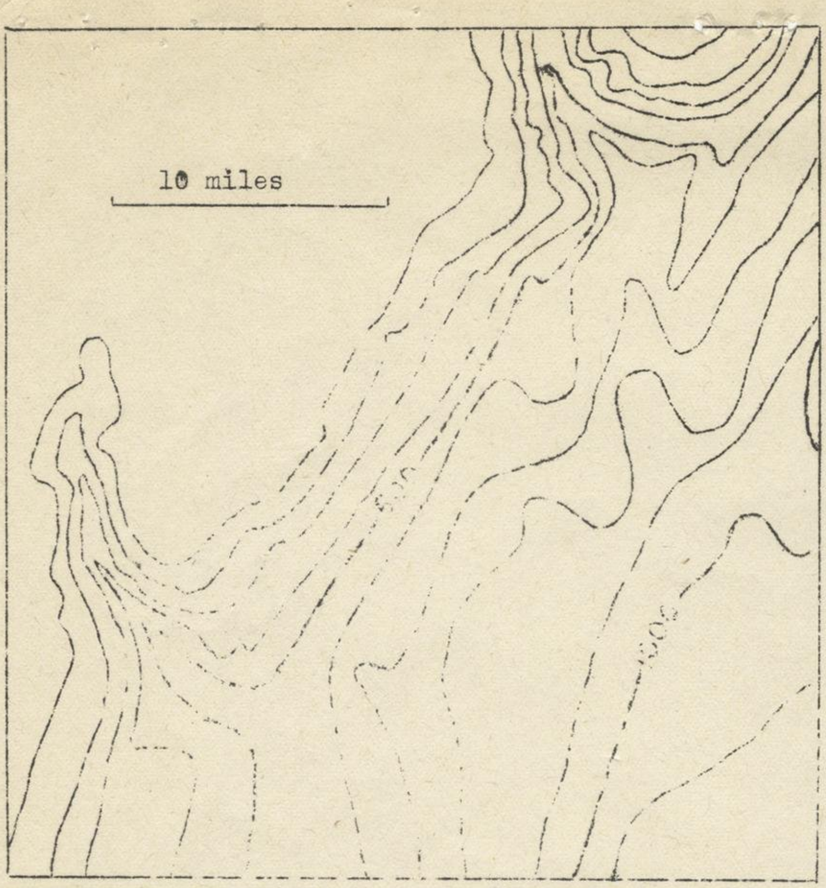


Fig. 143, p. 68

Edge of part of continental shelf, Atlantic coast. After Veatch and Smith, G. S. A. Spec. Pub. 7: 76
Contours drawn mechanically from soundings (not shown). Contour interval = 100 fathoms or 600 feet.

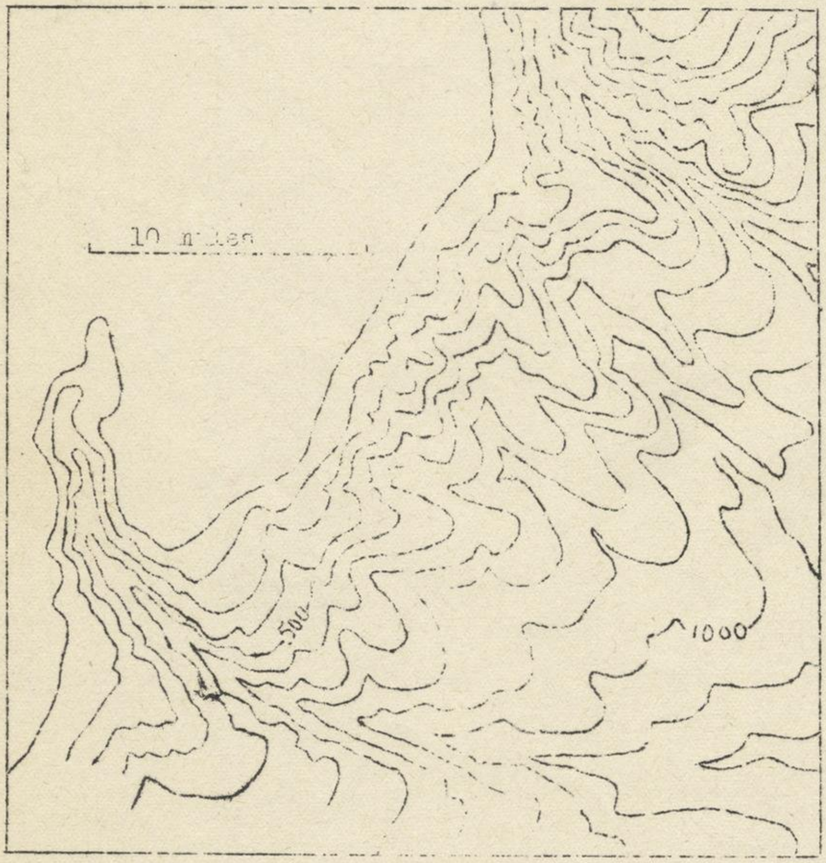


Fig. 144, p. 68

The same area as in Fig. 143 but contours were drawn with the idea that the continental slope was gullied by land streams when sea level was low. Contour interval same.

Fig. 145, p. 68
After Jones, Geogr. Jour. 97
A test from another area. Dashed lines from Veatch's contours; dotted line from Jone's mechanical contours; solid line from a new line of soundings run across older lines.

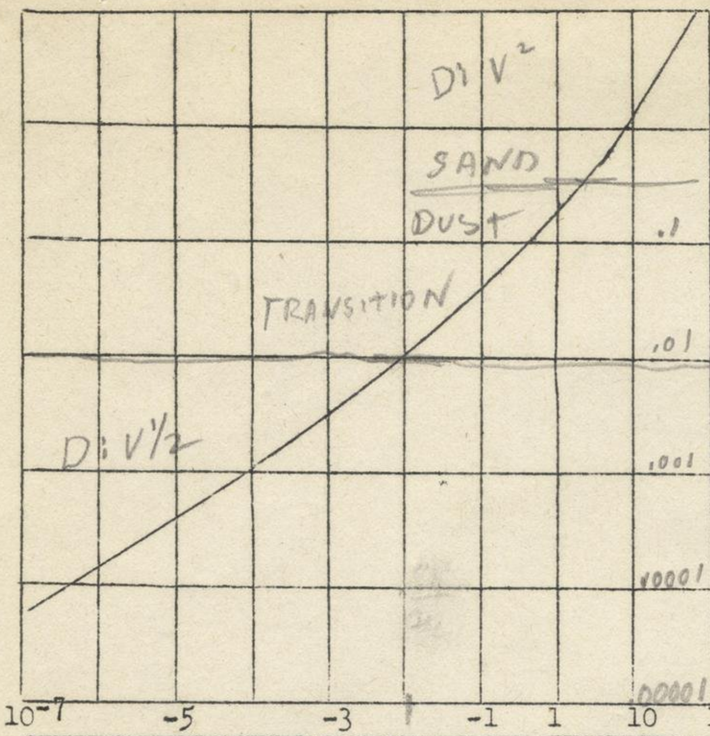


Fig. 148, p. 70

Falling velocity of different sizes of grains in air. After Bagnold, p.1. Redrawn to compare with Fig. 51. As with water the resistance to falling produces a terminal velocity. Relationship of air resistance to size of particle is not uniform but varies with diameter . Bagnold places the diameter of 0.2 mm as the line of division between sand and dust which is much more readily sustained in air. Diameter of sand grains is proportional to square of velocity, same as in water.

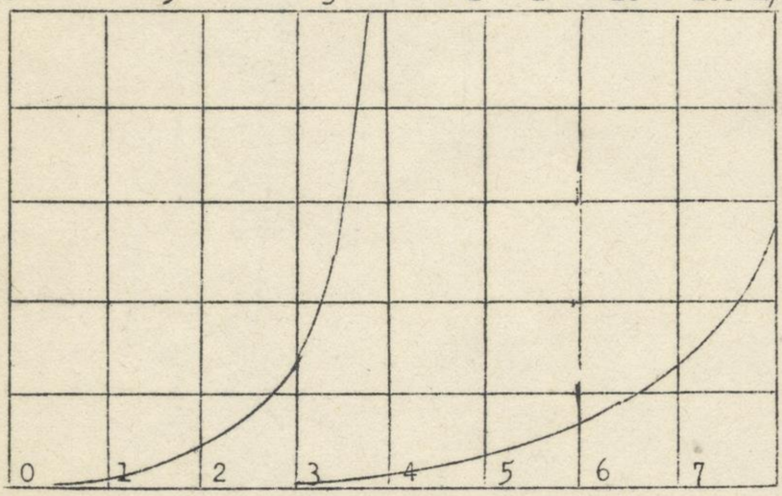


Fig. 149, p. 71

Vertical distribution of wind velocities platted to ordinary scale. After Bagnold, p. 48
Two examples shown.

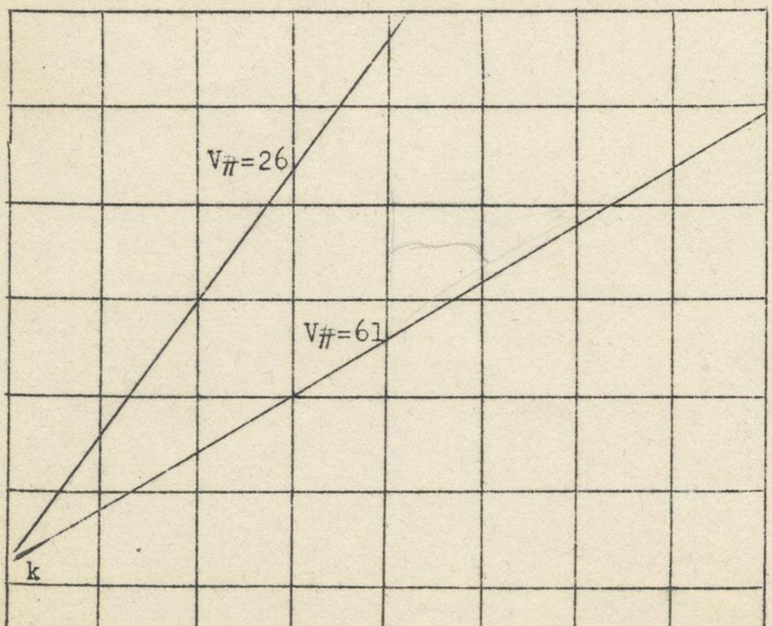


Fig. 150, p. 71

Vertical distribution of wind velocities platted to semi-log. scale. Note that both examples now form straight lines which meet at a definite distance above the ground, k. k = 1/30 diam. of sand grains on surface. Bagnold rates force of wind by quantity $V_{\#}$ which is result of dividing tan of angle made by line with the vertical by 5.75

vel off log str[illegible]

Drag on unit area = density $.V_{\#}^2$
Velocity at heigth z = 5.75 $V_{\#}$. log z/k (in cm./sec.)
also $V_{\#} = (\text{drag/density})^{\frac{1}{2}}$
After Bagnold p. 49

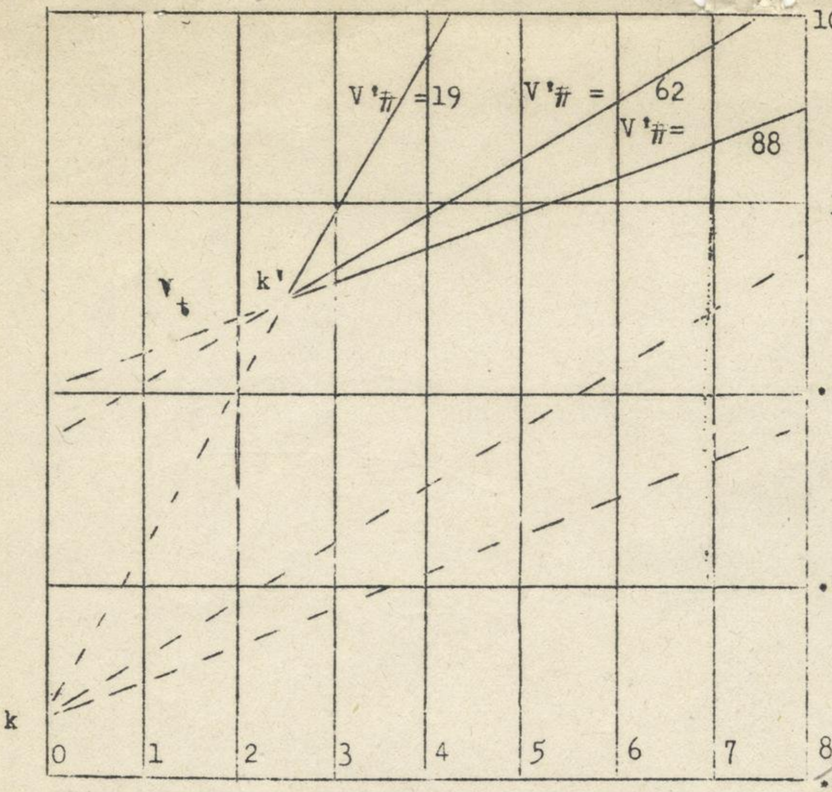


Fig. 151, p. 71

Change in condition after movement of sand begins. After Bagnold, p. 60
Broken lines show wind velocities with no sand moving. Point of no velocity is than at heigth k. However, above a certain value of V#, here 22, there is a slight surface movement; this border velocity is the <u>fluid</u> or <u>static</u> <u>threshold</u>. After sand movement is maintained by impact of falling grains the level of 0 velocity rises to k' which lies at the threshold velocity V_t. The new borderline is called <u>impact</u> or <u>dynamic</u> <u>threshold</u>. V# becomes <u>V'#</u> above this velocity

Impact threshold √ : diam.$^{\frac{1}{2}}$

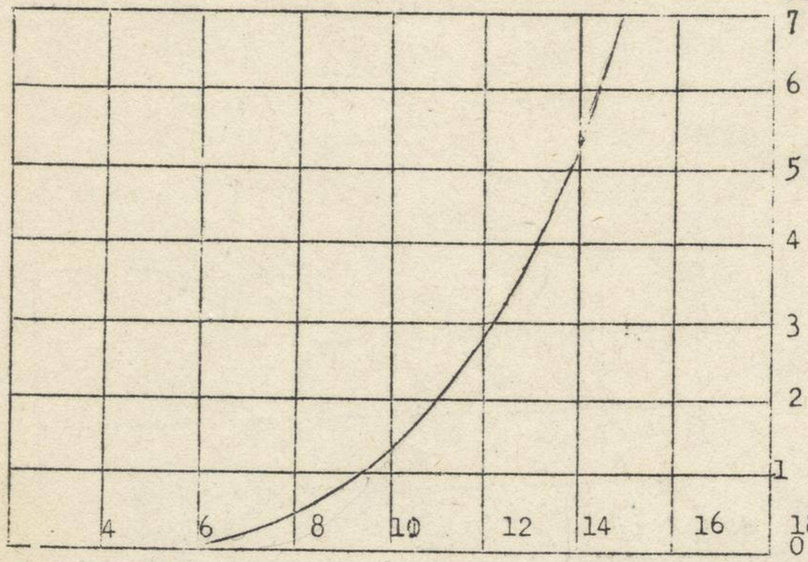


Fig. 152, p. 72

Average flow of dune sand in metric tons/ meter width/ hour in relation to wind at 1 meter heigth. Computed from formula quant. = $5.2 \times 10^{-4}(V-V_t)^3$

$p = \frac{W}{t} = \frac{FS}{t} = FV$

$F : V^2$

hence $p = (V^2 - V_t^2)^3$

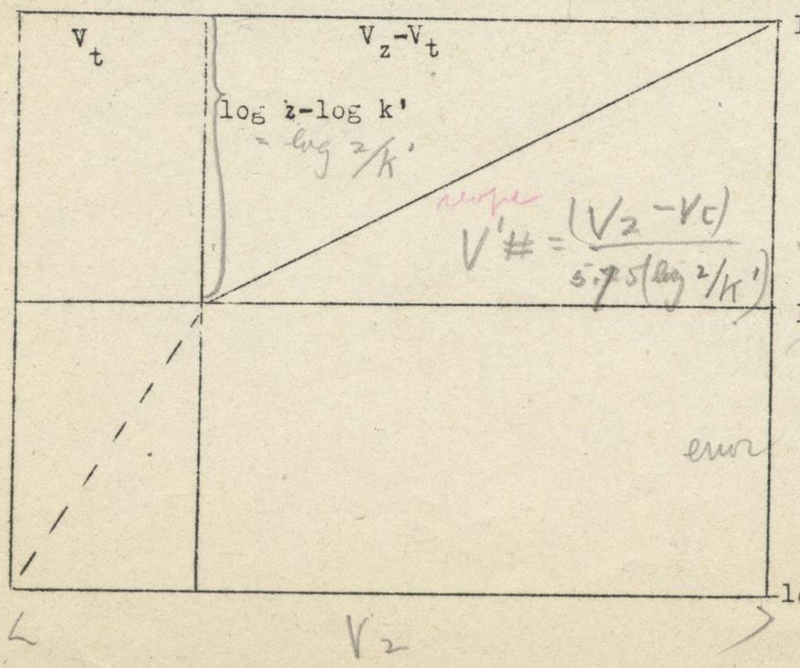


Fig. 153, p. 72

Bagnold's proof of substitution of velcity in excess of thres hold velocity (V_z-V_t) at a given heigth for V'#.

error

V'# = (V_z-V_t)/ 5.75(log z-log k')

or 0.174 (log base z/k')(V_z-V_t) error check

Since the first term is a constant for given values of z and k' the substitution can be made.
For z = 1 meter, k' = 1 cm the value is 6.58×10^{-4}

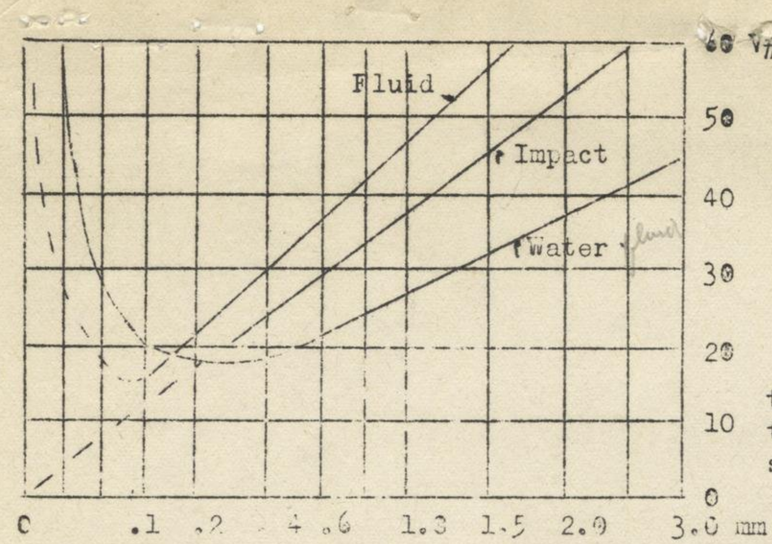


Fig. 154, p. 73

Relation of fluid and impact threshold in air to size of grains, also relation of fluid threshold in water to velocity in cm/sec. After Bagnold, pp. 88-89 Diameters in square root scale. Note departures of both fluid thresholds in air and water from the square root relationship when small particles are considered.

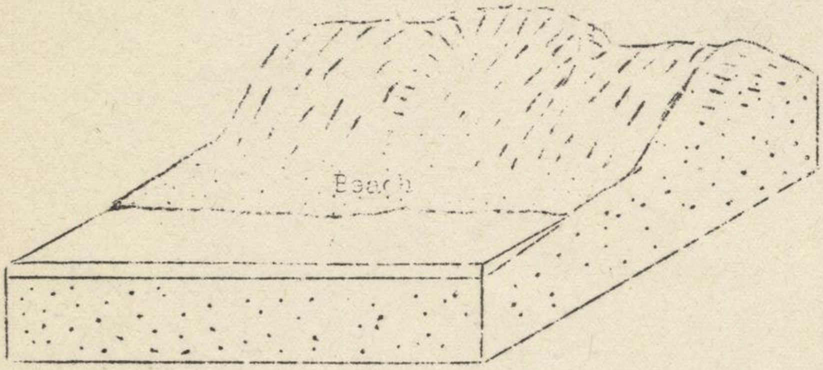


Fig. 159, p. 73

Foredune along a beach with a portion blown out through local destruction of vegetation.

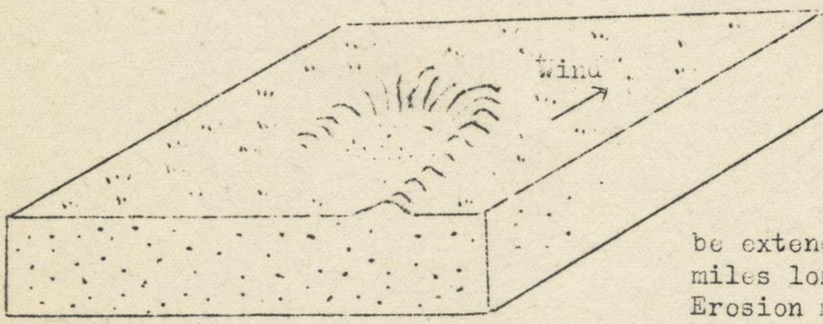


Fig. 160, p. 74

Blowout dune formed by wind excavation in a local area where vegetation was destroyed. After formation such dunes can be extended into longitudinal ridges many miles long. These resemble seif dunes. Erosion may be checked by accumulation of pebbles in the blowout.

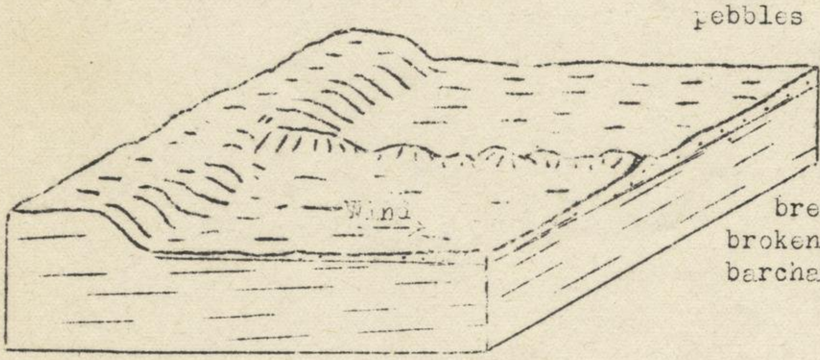


Fig. 161 , p. 74

Wind shadow dune in lee of break in escarpment. Such dunes are broken by cross winds and pass into barchan dunes.

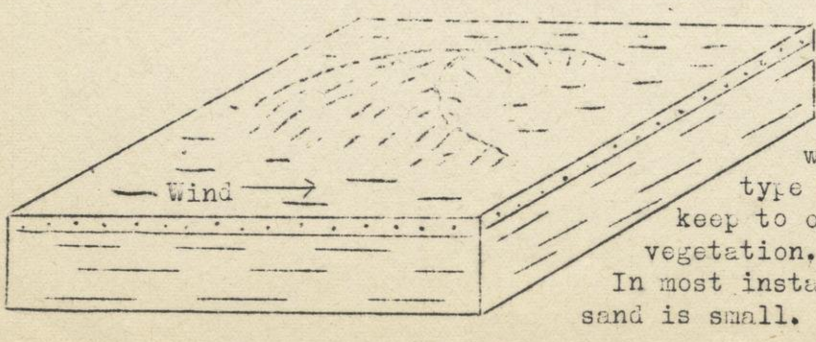


Fig. 1 62, p. 74

B archan or crescenti dune with steep face downwind. This type of dune forms only where winds keep to only one direction and there is no vegetation. They are rare in humid lands. In most instances they occur where amount of sand is small.

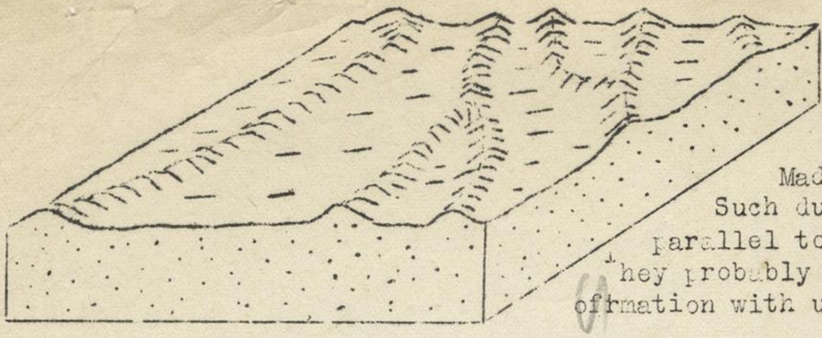


Fig. 164, p. 74

Seif or longitudinal dunes. After aerial photograph by Madigan, Geogr. Rev.. 26: 211 Such dunes, which here trend NNW are parallel to direction of prevailing wind. They probably represent the endpoint of dune ofrmation with unidirectional winds.

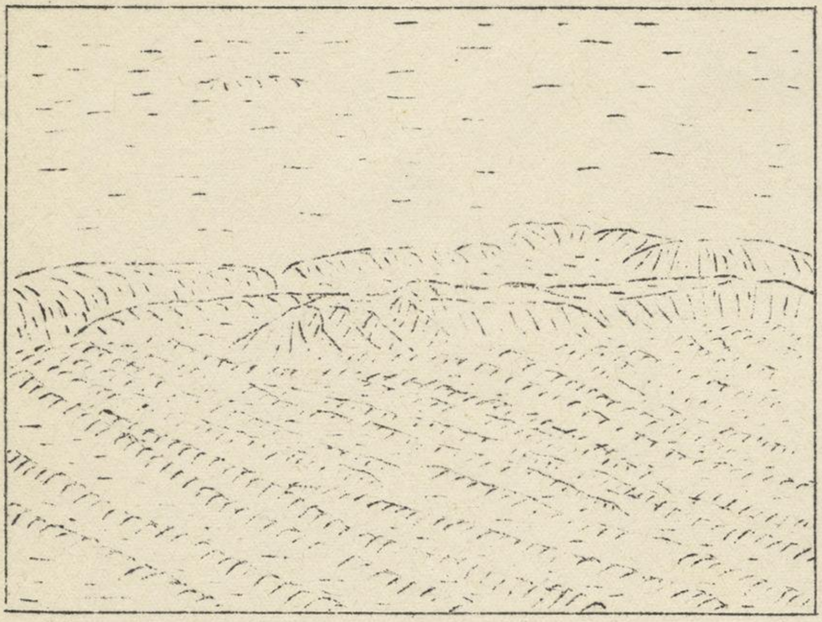


Fig. 1 67, p. 74

Low oblique aerial photo. of seif dunes in Australia. Here the long ridges end at foot of a mountain range. Looking NE. Madigan, Geogr. Rev. 26: 221

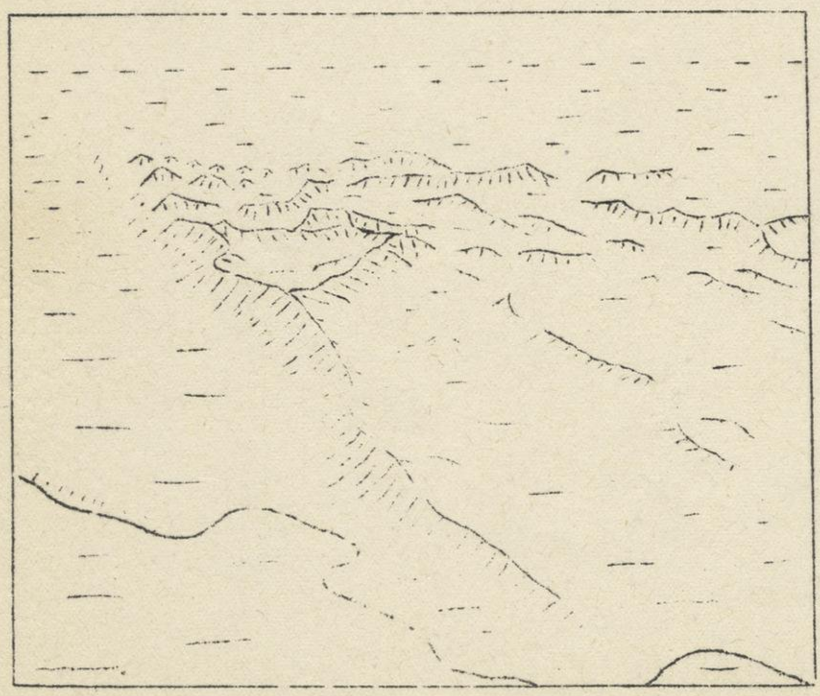
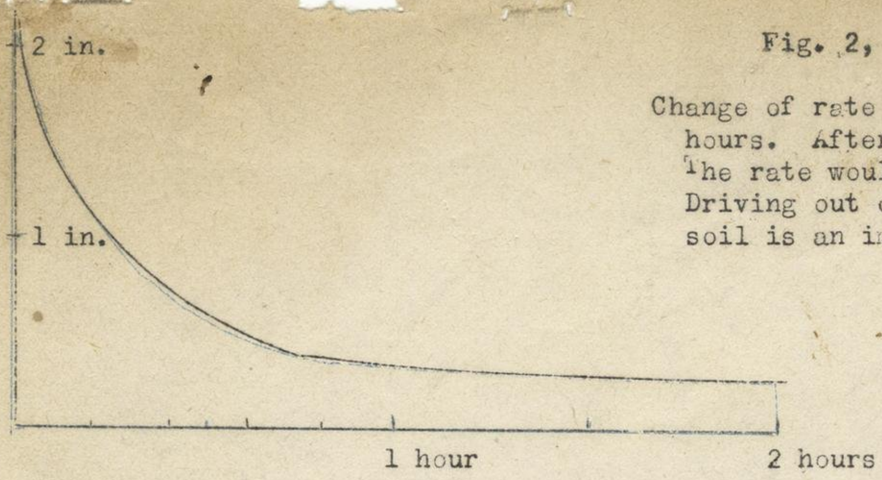


Fig. 1169, p. 75

High oblique aerial view of loess hills on east bluffs of Missouri River, Iow a. After Iowa Geological Survey, Pleis. of Iowa: 166. Many of thse hills have been considerably eroded and their sides have slumped into cat steps.

Fig. 2, p. 2

old



Change of rate of infiltration with time in hours. After Horton, G. S. A. B. 56: 307 The rate would never reach 0 in most soils. Driving out of air from the pores of the soil is an important factor.

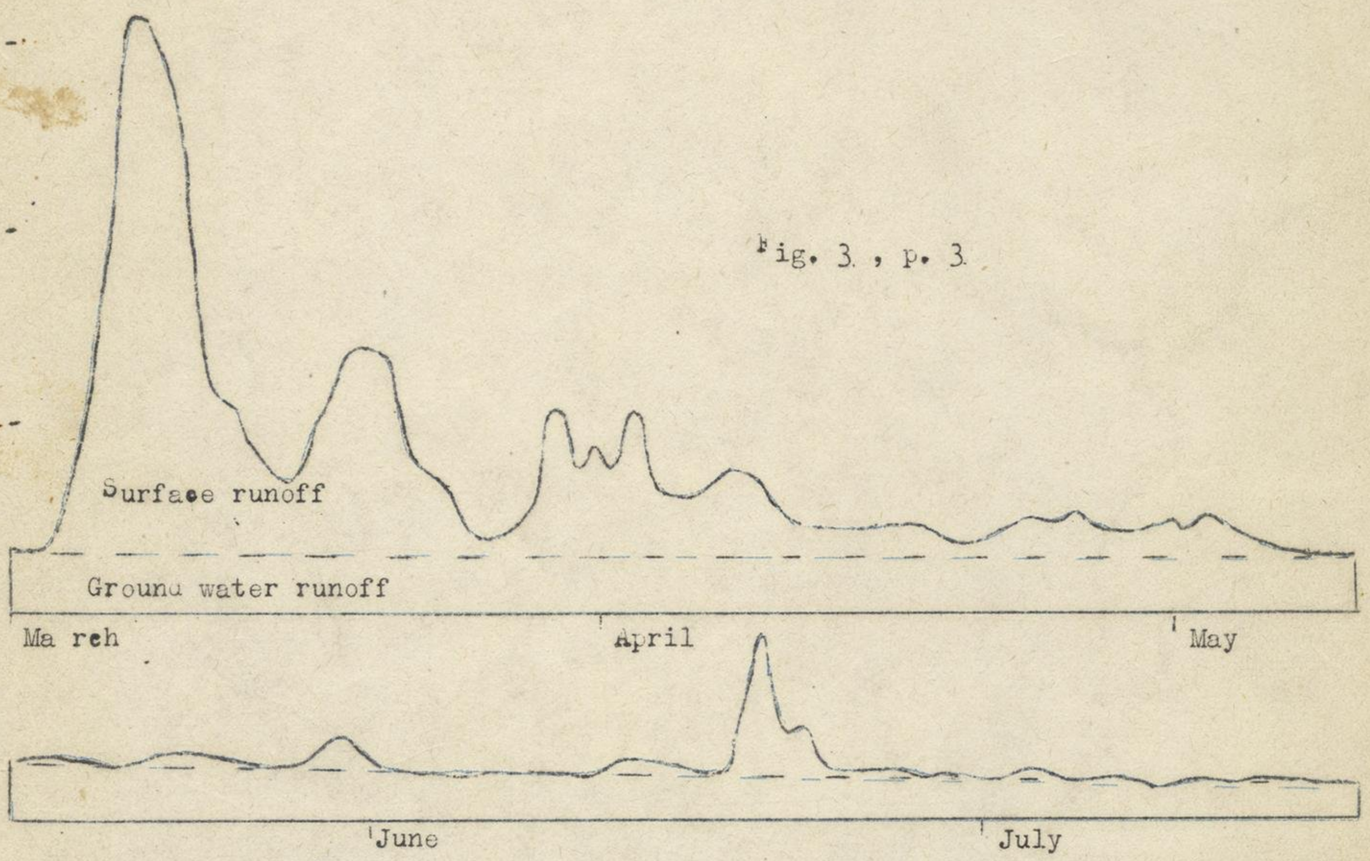
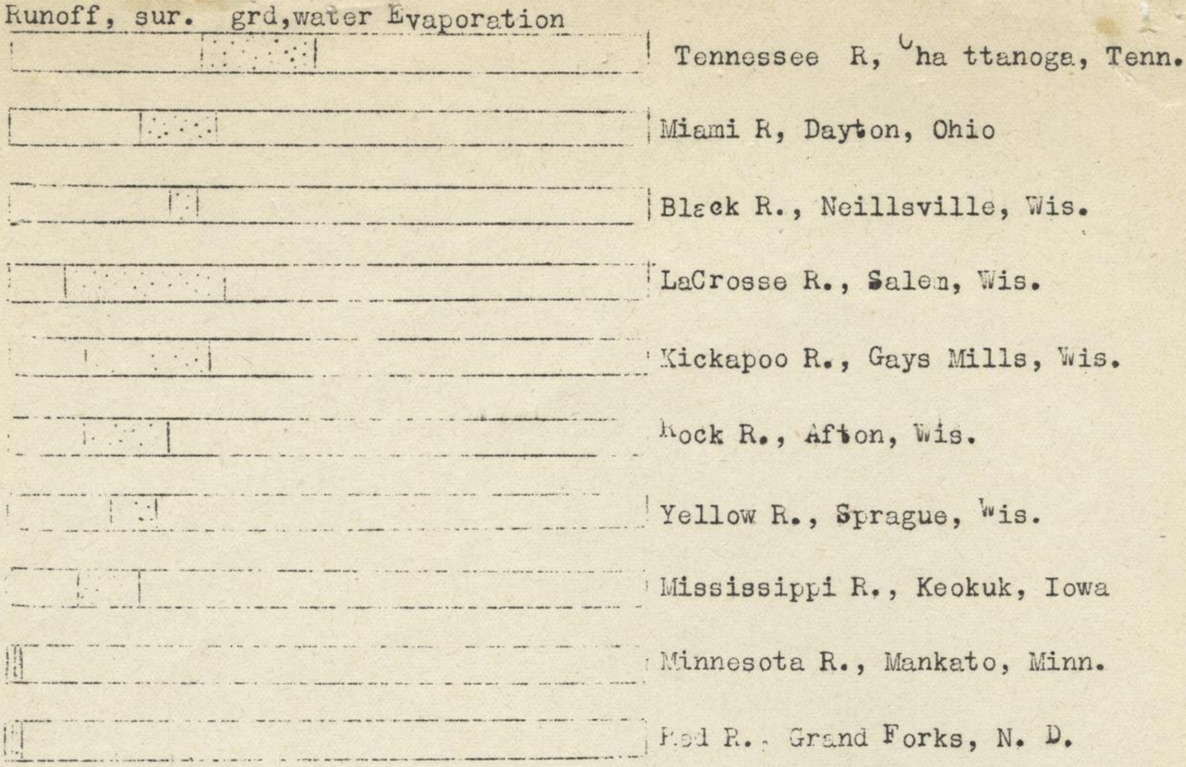


Fig. 3 , p. 3

Discharge in ft.^{3}/sec (second feet) of Kickapoo River, Wisconsin, for part of 1937, after U. S. G. S. Water Sup. Paper 825. See also Fig. 4. Ground water runoff from springs and storage in stream bed is estimated as that below dotted line. Note irregularity of flood discharge; this is characteristic of most rivers. At present these floods bring down much debris from the adjacent hills. Much of this is deposited on the high-water channel floor or <u>floodplain</u>. Some also accumulates in the low-water channel. This channel meanders in small loops. The entire valley of the Kicapoo also meanders in some sections of its course. Thus we have two distinct sizes of meanders in the same stream, one large and rock bound, the other small and on a flood plain of loose materials. Thus the Kickapoo is an <u>underfit</u> stream.



Relative percentage of surface runoff, ground water runoff and evaporation in typical streams. BELOW, the same expressed in inches of total precipitation.

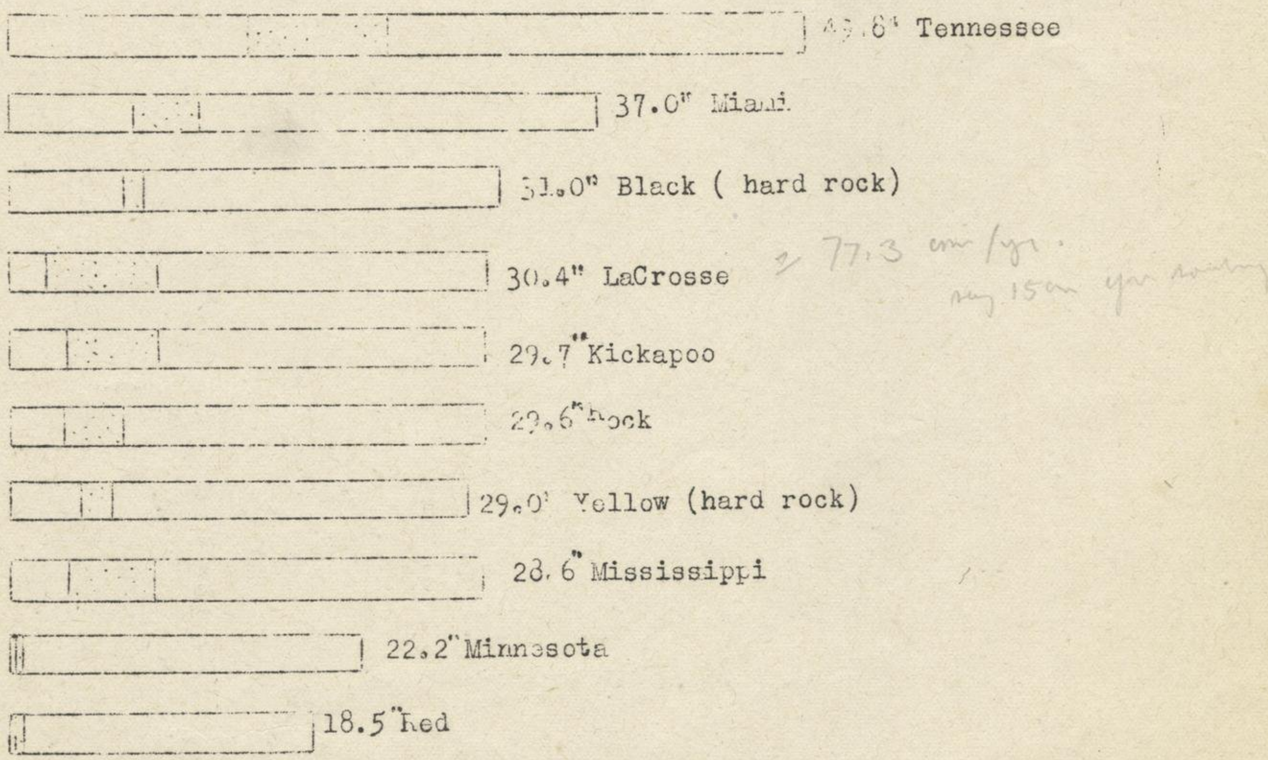
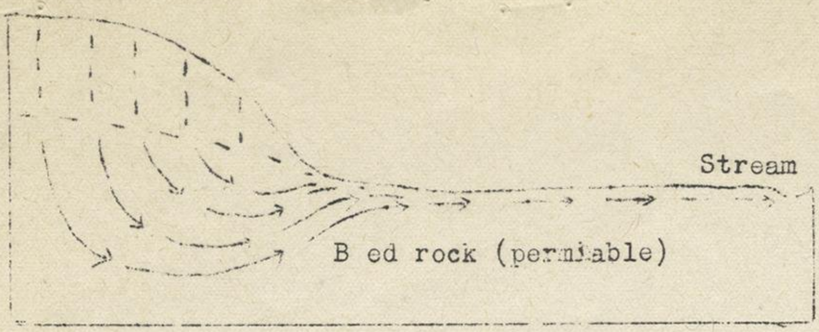


Fig. 4, p. 3 Relation of surface runoff, ground water runoff and evaporation in typical rivers of the United States. U. S. Geol. Survey Water Supply Paper 772. In all basins evaporation is the largest item. This includes water used by vegetation and in chemical combination of weathering. Most soil evaporation takes place when the air is dry and surface runoff occurs immediately after rain begins when air is moist. Note relatively low values for ground water runoff in the hard rock basins of Black and Yellow Rivers.



Water table shown by dotted line

Fig. 1 3, p. 13

Relation of depth of weathering to topography. Under low ground there is not enough pressure to cause waters to go far below the surface unless a deep path is the only o ne possible.

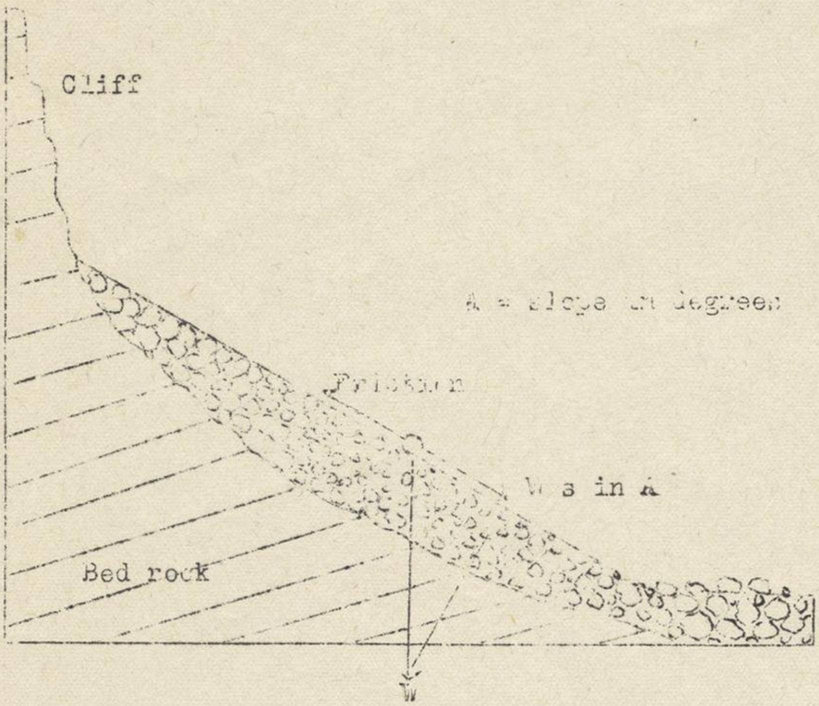


Fig. 15, p. 13

Forces involved in talus slope. The component of weight of a rock which is along the average slope is balanced by friction. Rocks which are larger than average are not retarded as much by irregularities of t he slope and so may roll out some distance from the base of the even slope. Note that weathering back of the cliff makes the talus a relatively thin veneer over the bed rock.

In the talus shown weathering of the talus fragments has not altered the original slope.

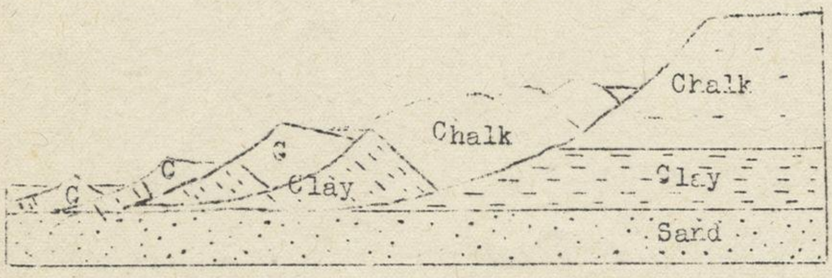


Fig. 17, p. 15

La ndslides in wave-cut cliff of chalk overlying clay and sand, England. After Ward The lines of fracture curve outward. Engineers usually treat them as sections of cylinders with horizontal axes above and outside the slid mass. Such analysis cannot be exact. The clay when saturated with water must behave as a fluid whereas the chalk is a solid. Note ponds between the slid blocks which serve to lubricate the pla nes of sliding. Lowest block is in the sea and will be removed by wa ve work.

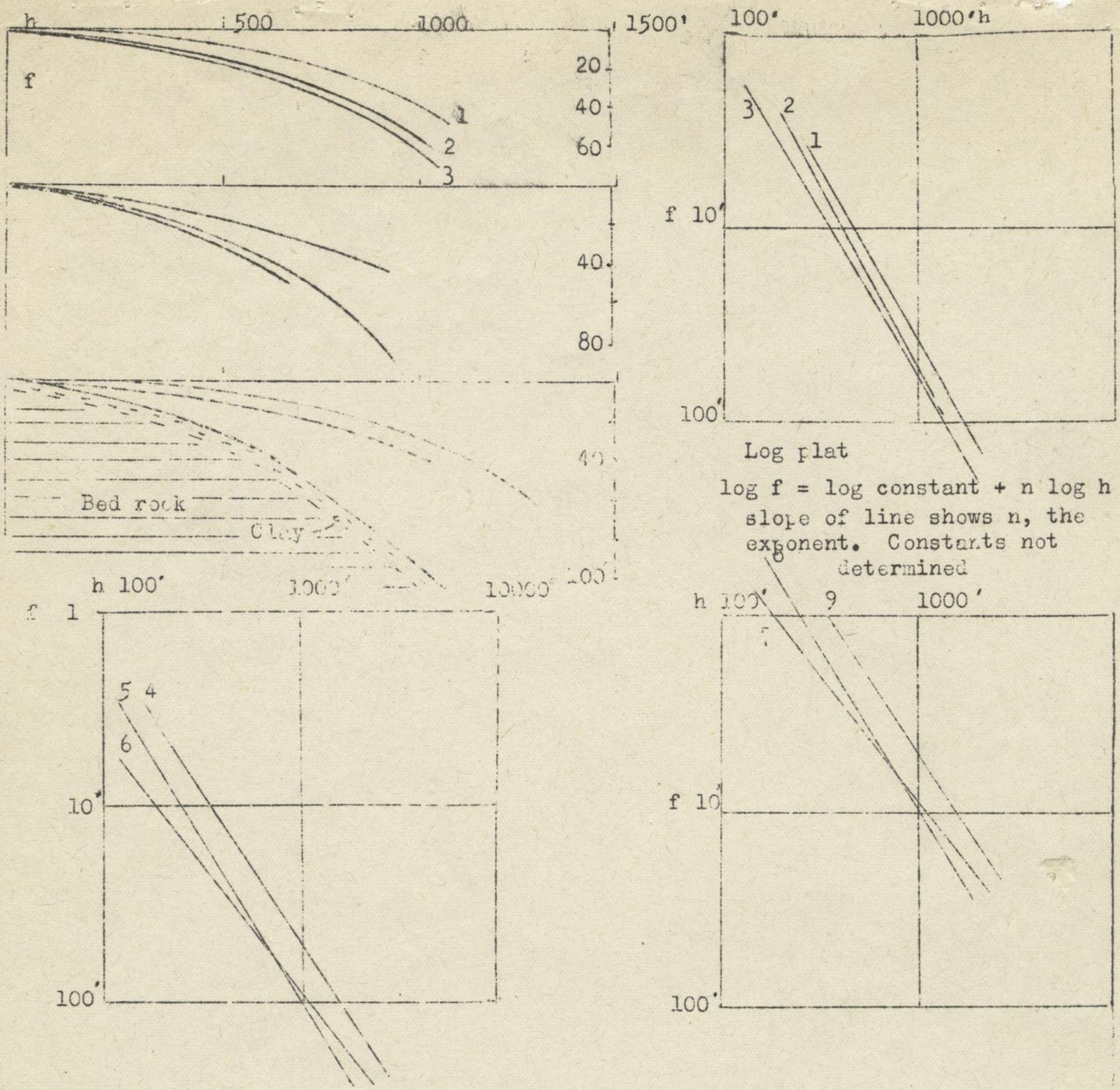


Fig. 20, p. 15

Ma thematical analysis of creep slopes, southwestern Wisconsin.

Upper left: profiles with vertical scale x 5 It is known that in most of these slopes the mantle rock is essentially uniform in thickness. Most of the area was originally grassland (prairie) and close to the divides there was very little a lteration by running water. The profile of stability is then determined by creep. The original slope was probably developed by alteration of sides of stream valleys by weathering which reached its maximum close to the steeper slope thus rounding off the corner and preparing t he way for creep which extends beyond the z one of slope wash and talus formation. Since material is added to the moving mantle or clay and stone a ll the way down the slope its velocity must increase progressively away from the divide is the thickness is constant. Force producing motion is the component of weight along the slope. Slope must also increase in proportion to distance, h, from the divide. Fall, f, must then be related t o the s quare of distance from the divide f: h^2 a s shown either by algebra or by integration of slope and horizontal distance. df/dh = h f:h^2
Average exponent of the 9 slopes shown is 1. 75

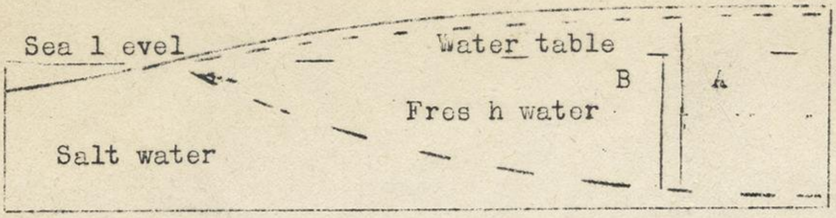


Fig. 21, p. 18
Fresh underground water floating on salt connate water.
A.1.0 = B . 1.03 to reach hydrostatic balance

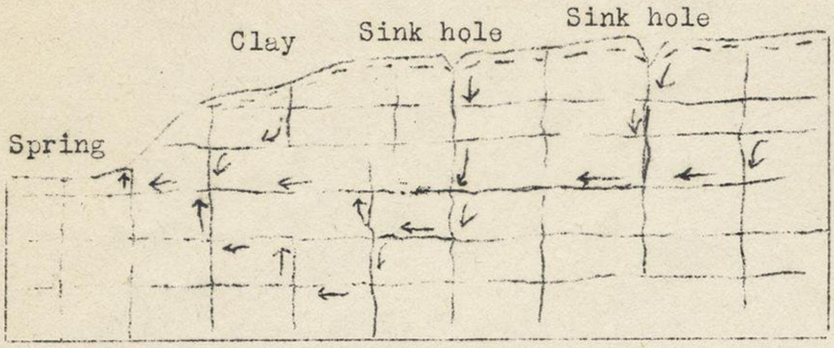


Fig. 22, p. 18
Circulation of underground water through joints and bedding planes of limestone. The paths which offer least resistance to flow are followed and enlarged by solution. Only connate salt water occurs b elow the lowest paths of flow of fresh water

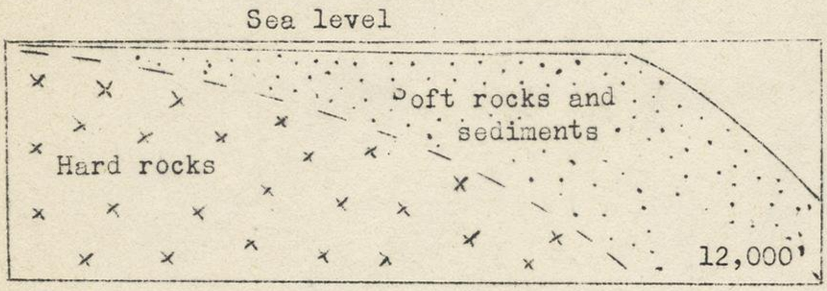


Fig. 29, p. 21
Continental shelf of eastern North America. After Ewing, G. S. A. B. 48: 804
Width about 190 miles

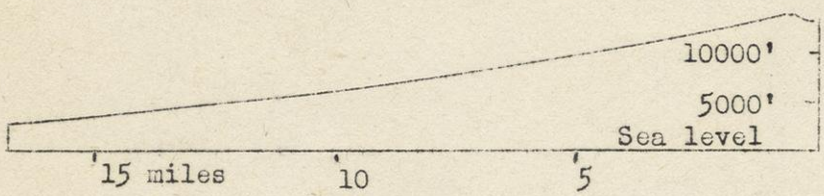


Fig. 30, p. 24
Mauna Loa, a basalt cone

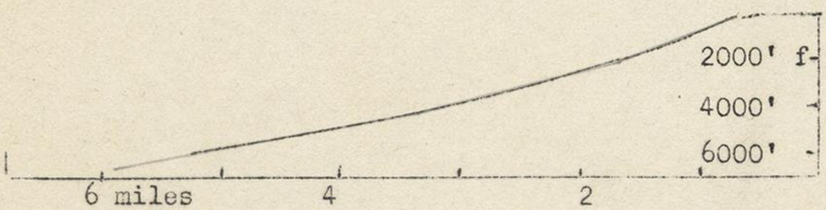
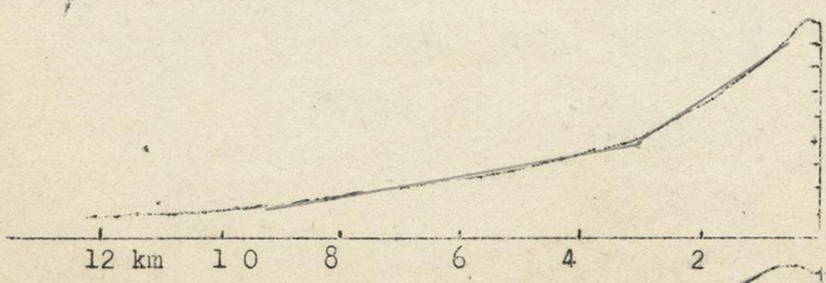
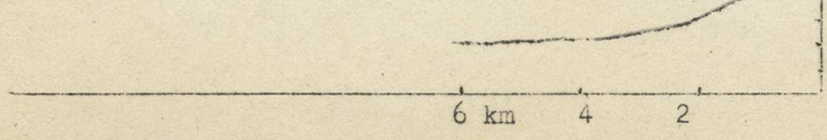


Fig. 35, p. 26
Mt. Hood, andesitic volcano
Measurements of fall, f, from summit



Fujiyama, ash cone
f in meters

2000 meters



Vesuvius

2000 m

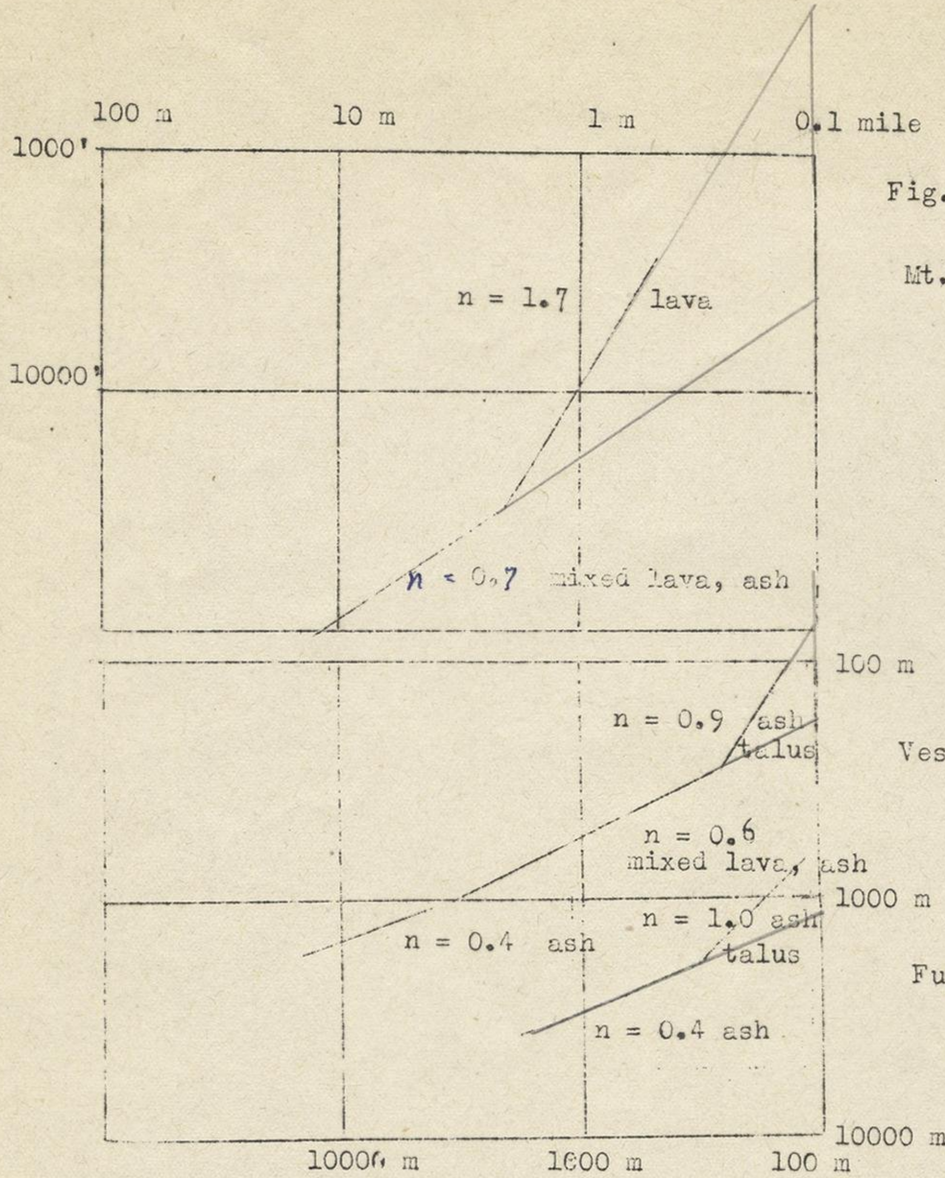


Fig. 35, p. 26, cont.

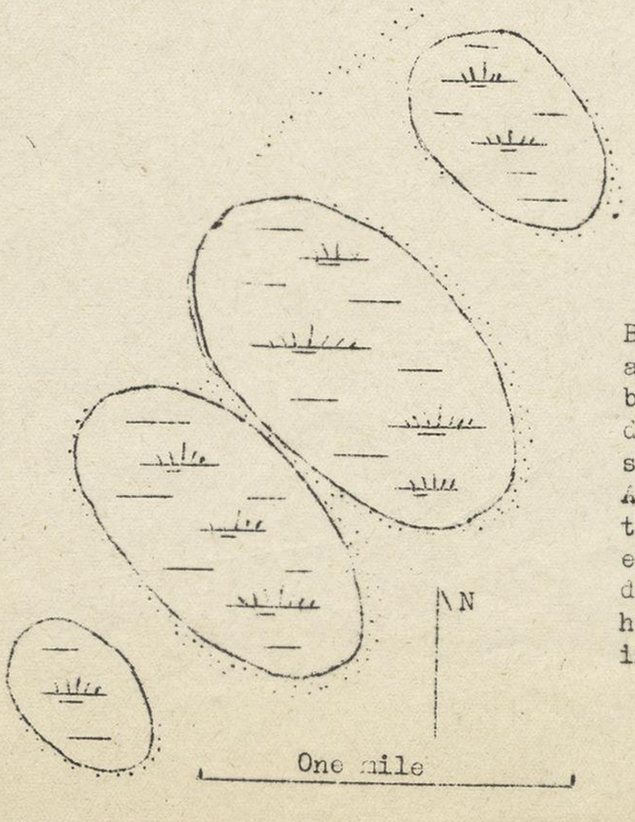


Fig. 46, p. 30

B ays 7 m. south of Mullins, S. C. after aerial photograph published by Johnson. Sandy ridges are difficult to ma p becaus e all surrounding soil is sandy. Are t he Bays (these are all transformed to swamp) scars of exploding meteorites or are they depressions of any origin which have been smoothed by wave action in lakes?

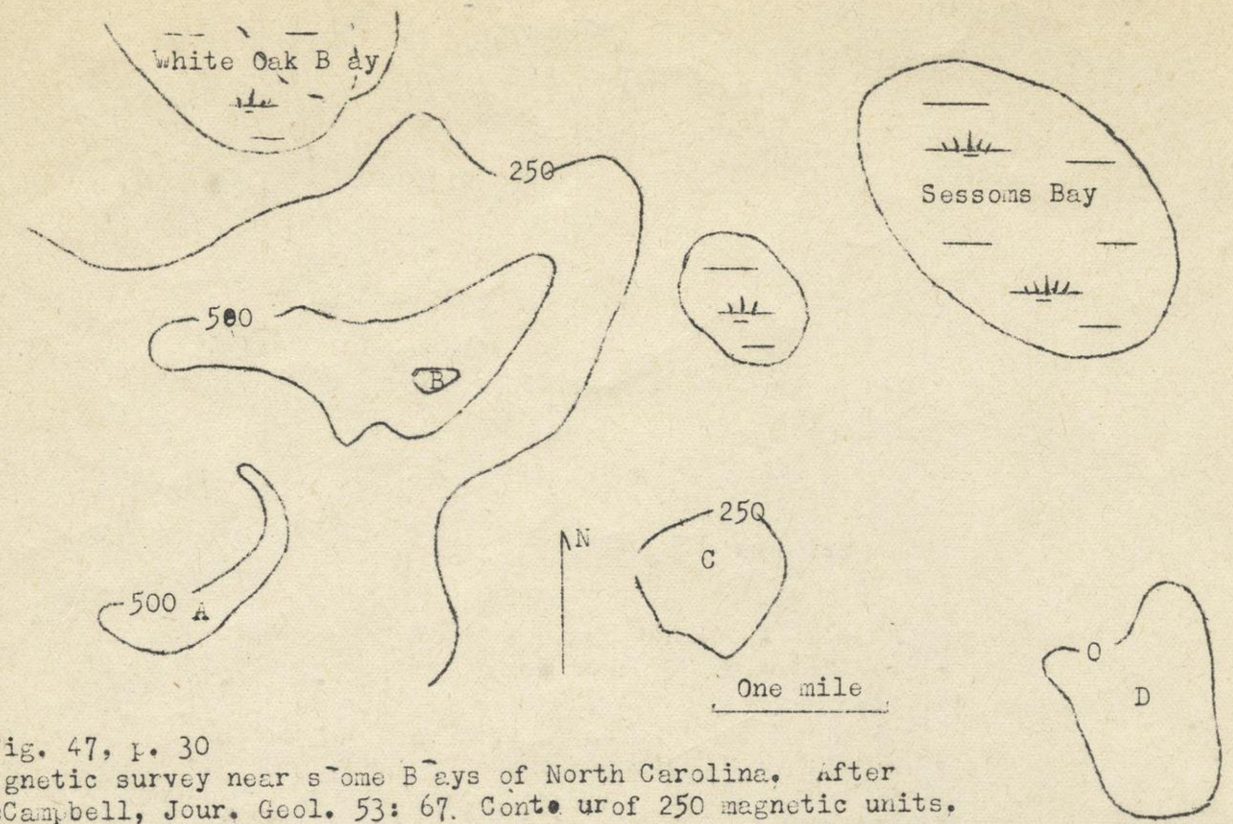


Fig. 47, p. 30
Magnetic survey near some Bays of North Carolina. After McCampbell, Jour. Geol. 53: 67. Contour of 250 magnetic units. High A is thought to be related to a Bay to N W, outside of map. High B is south of White Oak Bay, high C in the same direction from the unnamed bay. D is reported as a "high" but contouring of original map makes it a low. Is the magnetism related to buried framents of meteorite or to magnetic minerals either in the sediments of the Coastal Plain or in the buried hard rocks. It is very unlikely that it is due to limonite as suggested by Johnson. Such surveys should be extended over larger areas and checked by drilling on highs.

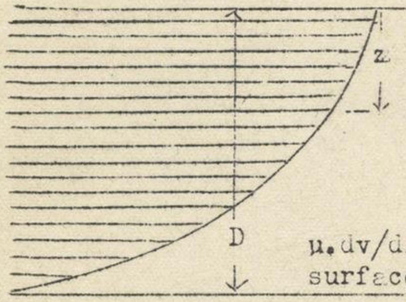


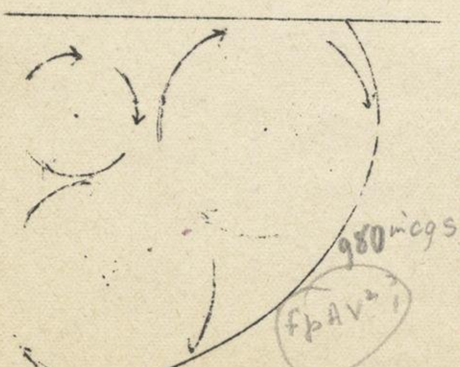
Fig. 48, p. 32
Laminar flow of water, one layer over the next below.
D = total depth S = slope (tangent of angle of slope)
V = velocity z = depth to any point above bottom.
dv = change in velocity for slight change in depth, dz
u = viscosity.

= force at that point, z

u.dv/dz = Z.S / u.dv = Z.dz.S by integration from 0 to D
surface $V = D^2 . S / 2u$ (stream of great width)

$\frac{u\,dv}{dz} = F$

$F = Z.S$



Turbulent flow of water which occurs as velocity and thickness exceed a certain value. Then bottom friction sets up eddies which absorb much of the energy in kinetic energy of rotation thus slowing down the stream.
A = area of cross section p = wetted perimneter
f = a coefficient V = velocity w = weight of unit volume of water S = slope A/p = R or hydraulic radius
Force downstream = A.w.S (in unit length)
Resistance = $f.p.w.V^2$ Equate these and solve for V^2 and substitute R for A/p $V^2 = R.S / f$ Chezy
Experiment shows that value of f depends partly on nature of bottom and partly on R, hence a better formula is Mannings: $V = \text{coeff.} R^{2/3}.S^{1/2}$

$V^2 : R^{4/3}S$

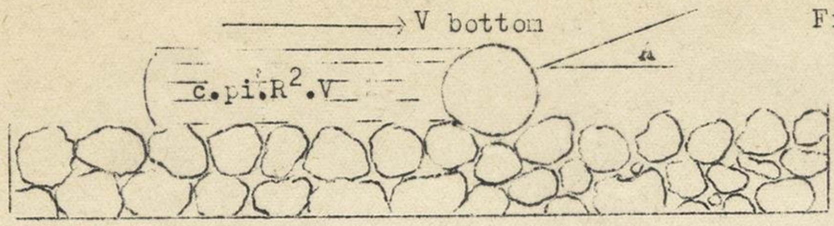


Fig. 49, p. 32
Force needed to start a sphere by elevating along a line at angle a. Weight under water= 4/3 pi.R^3.d.g c= coefficient R = radius of particle and column of water. d = density in air less 1 (density of water)

$4/3$ pi.R^3.d.g.sin a = c.pi. R^2.V^2. cos a (vertical comp.)= force of water   solving for $R = \frac{3\ c.V^2}{4\ \tan a.d.g}$

or diameter of particle is proportioned to square of bottom velocity.

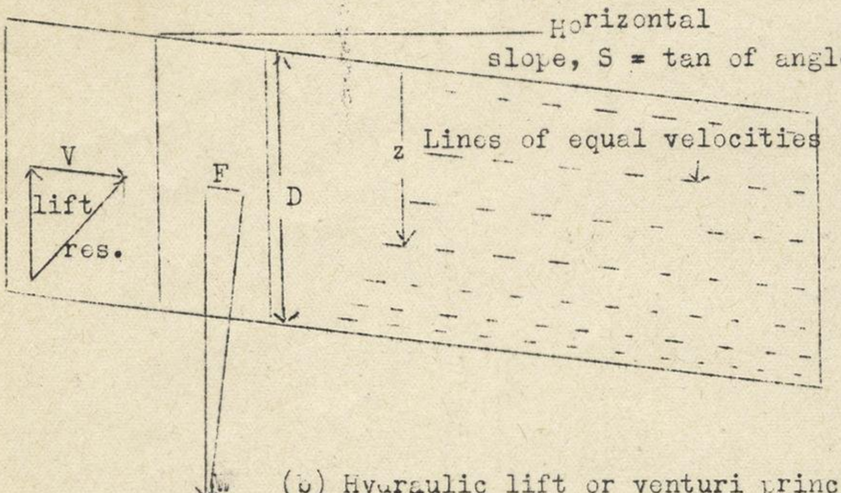


Fig. 50, p. 33
Different methods of expressing energy of a stream.
(a) Component of weight of a column of water of unit area= depth,(D).sin angle of slope. Since with low angles sin and tan are nearly the same this becomes D.S for weight of unit volume of water is unity (unless we use absolute units in which it is equal to g.)

(b) Hydraulic lift or venturi principle. When water in a pipe enters a smaller diameter its velocity is increased. By the principle of conservation of momentum this is done by reduction of pressure against the side wall. The lines of equal velocities shown above should exert an upward component of force thus raising material in the stream.

(c) Considering the same unit column of water as in (a) it is evident that its force downstream is opposed by the product of its kinetic energy times a coefficent. D. S = f. V^2 Now in a pipe carrying water with turbulent flow there is a loss of pressure(head) per unit of length which is equivalent to the descent of a stream in unit length or slope. Solving the above for S we find S = f.V^2/D This is the force per unit of depth or vertical gradient of kinetic energy and was used by Little as a measure of erosive force of running water.

(d) Some have tried to give a measure of intensity of turbulence by taking prisms with apeces at line of maximum velocity, sides normal to lines of equal velocities, and bases on bottom of stream of unit area. The intensity at any depth, z, is equal to force of prism above (by depth.slope formula) divided by rate of change of velocity with depth (dv/dz). Rsults are interesting but not convincing.

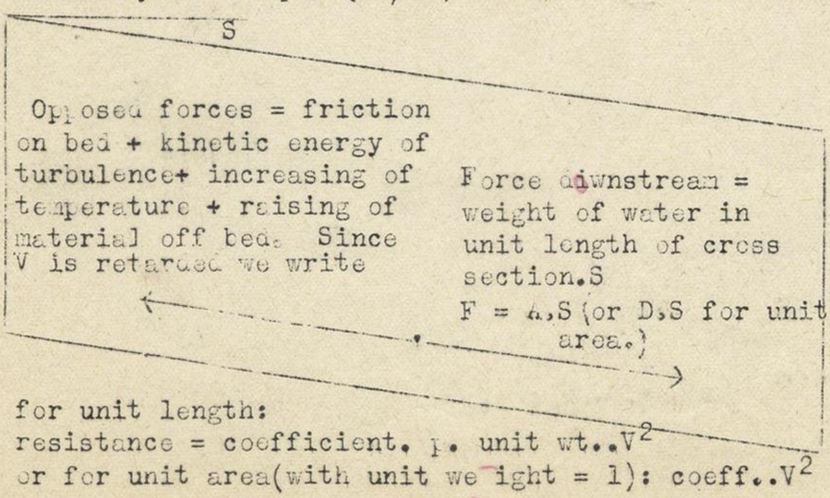


Fig. 53, p. 35
Opposed forces in a stream of water. As with a block sliding down an inclined plane it is possible to adjust the slope so that there is no acceleration.

Two layers: clean water on sand-water mixture (interface)

for unit length:
resistance = coefficient. p. unit wt..V^2
or for unit area(with unit weight = 1): coeff..V^2

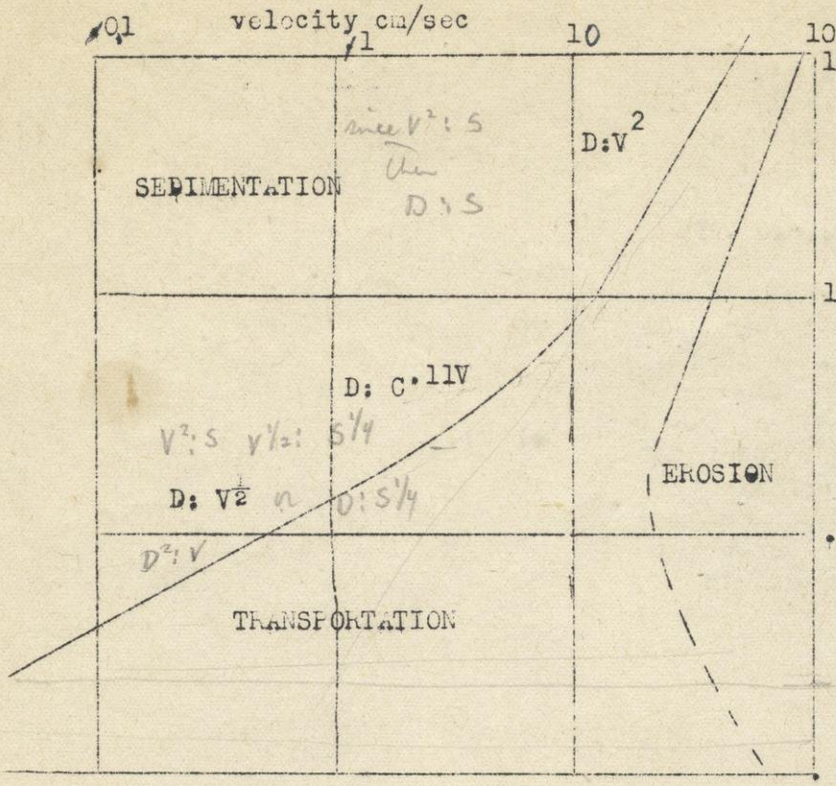


Fig. 51, p. 34
Relation of velocity of water to erosion, transportation and sedimentation of particles with different diameters, D. After Nevin, G. S. A. B., 57: 674
Note that for D 1 mm up the kinetic energy of the water is the major factor. For D less than about .2mm viscosity is dominant. There is a gradual change in the transition region. The curve for start of erosion is difficult to draw for so much depends upon the degree of packing of the small sizes of material.

mixed flow = $S^{7/10}$ $D: S^{14/10}$
or for small particles
$D: S^{1/4}$ or mixed $D: S^{7/40}$
(hw f: $3/4$ h or $\frac{33}{40}$ h)

$V: C \frac{R^{2/3} S^{1/2}}{D}$

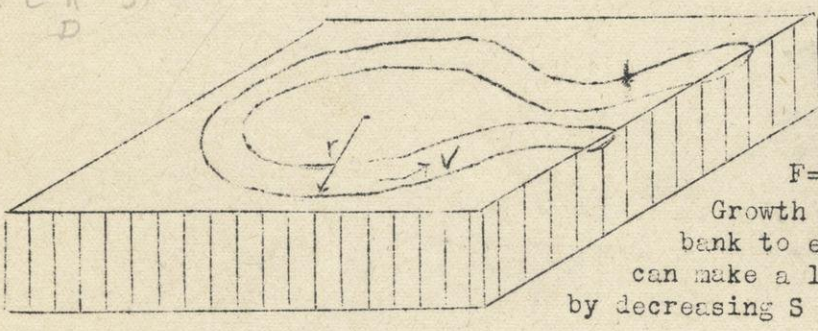


Fig. 55, p. 37
Rotational component of force in a curved or meandering stream.
Radius of curve = r Slope = S
$F = mV^2/r$ as $V: S^{\frac{1}{2}}$ then $F: m.S/r$
Growth ceases when F = resistance of bank to erosion. Only a large stream can make a large meander. Increase of length by decreasing S also serves to set a limit.

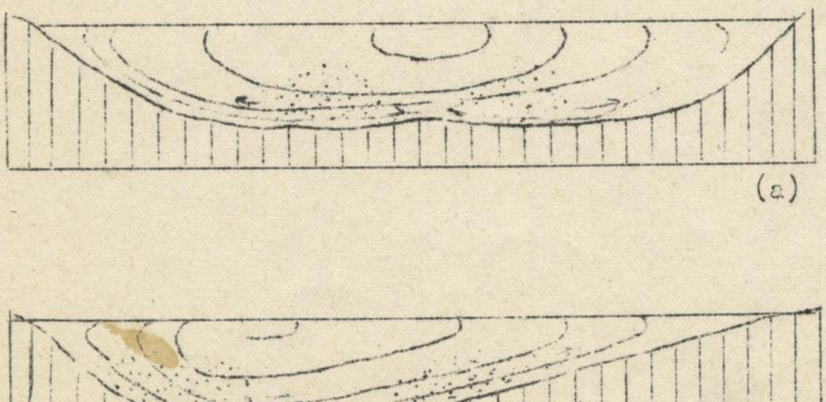


Fig. 56, p. 37
Distribution of velocity and inferred intensity of turbulence in (a) straight and (b) curved streams.
Intensity of turbulence is rate of transfer of energy. Hence sediments should move away from the dotted areas. This explains erosion of cut banks and transfer to insides of bends and may also explain start of sand bars in middle of a stream, prehaps the beginning of braiding where accumulation of deposits is rapid.

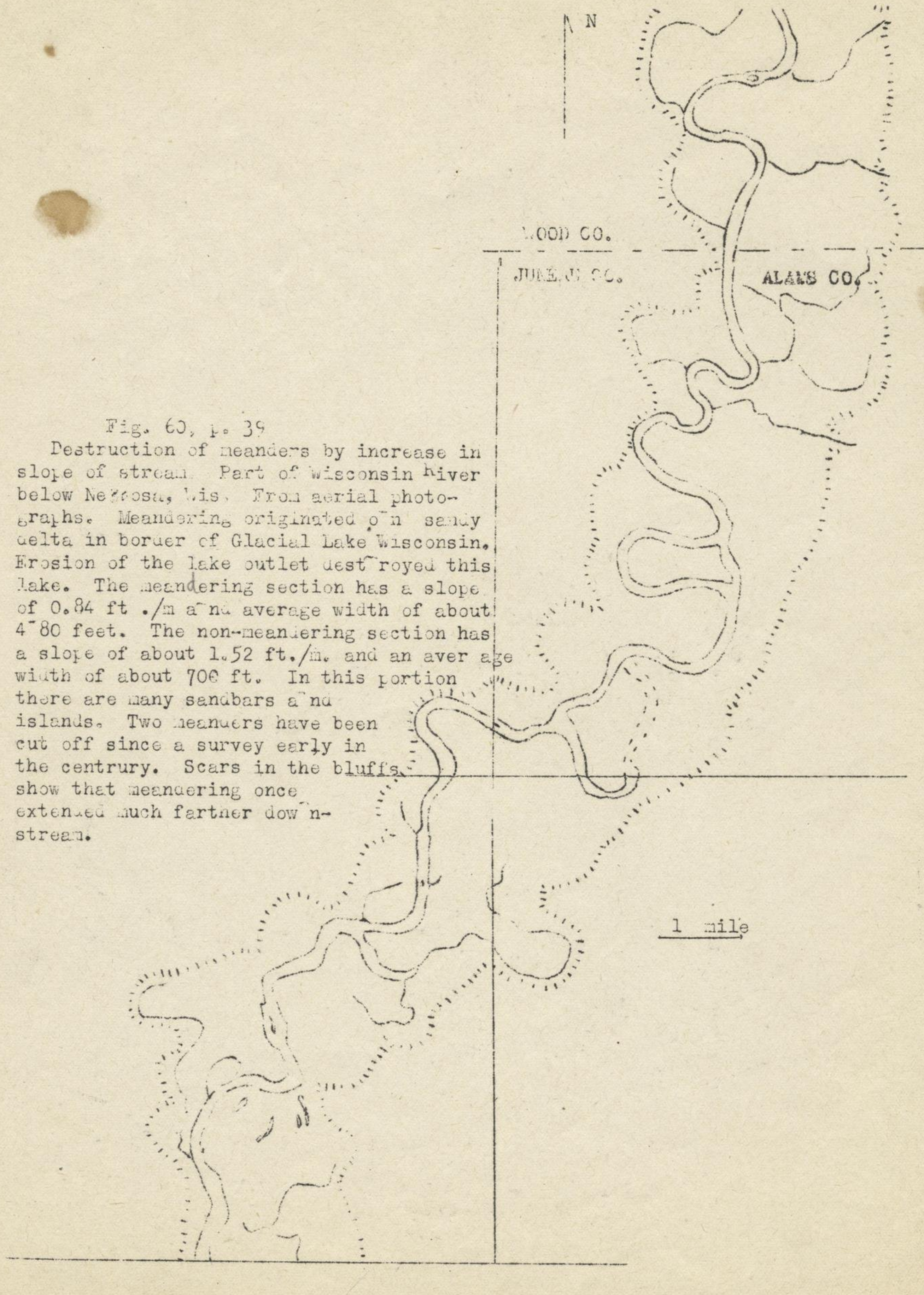


Fig. 60, p. 39

Destruction of meanders by increase in slope of stream. Part of Wisconsin River below Nekoosa, Wis. From aerial photographs. Meandering originated on sandy delta in border of Glacial Lake Wisconsin. Erosion of the lake outlet destroyed this lake. The meandering section has a slope of 0.84 ft./m and average width of about 480 feet. The non-meandering section has a slope of about 1.52 ft./m. and an average width of about 700 ft. In this portion there are many sandbars and islands. Two meanders have been cut off since a survey early in the centrury. Scars in the bluffs show that meandering once extended much farther downstream.

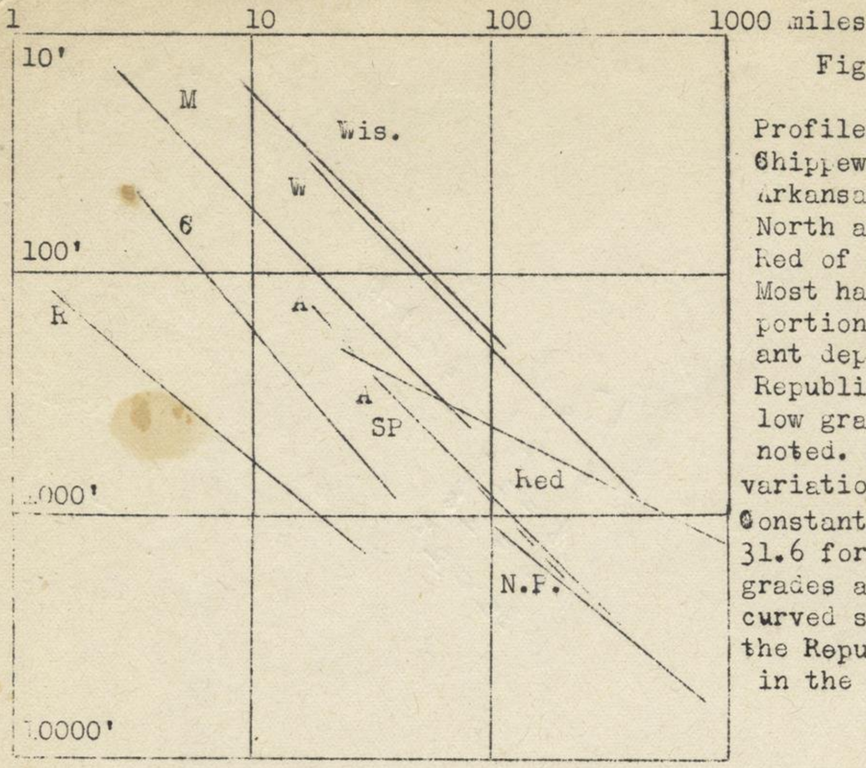
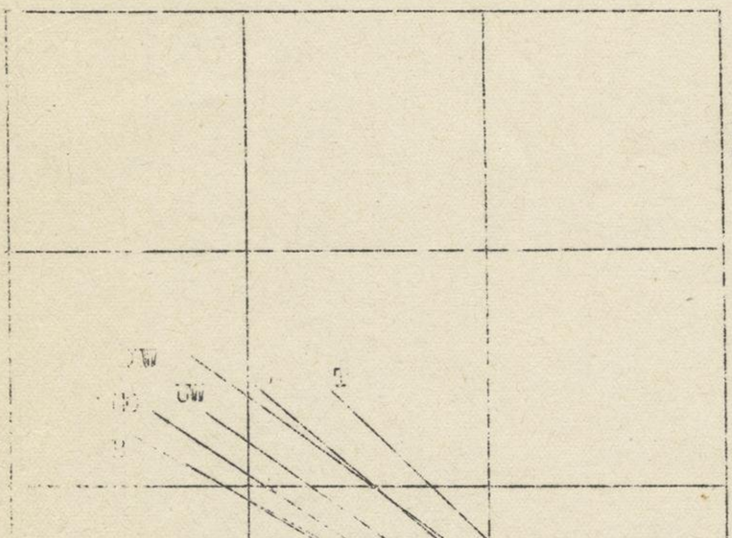


Fig. 61 , p . 40

Profiles of rivers on log plat. Chippewa, Mississippi below Keokuk, Arkansas in Great Plains, Republican, North and South Platte in Great Plains, Red of south, White (Arkansas), Wiscon sin Most have straight line grades in the portions platted, the only significant departures being North Platte .8, Republican .85, and Red .5. The low grade of the last has long been noted. Most rivers have too much variation in geology to permit study. Constants vary greatly, from 1.0 to 31.6 for streams with straight line grades and from 25.1 to 39.2 for the curved slopes. The highest is for the Republican River which lies all in the Great Plains.



Fig, 62, p. 40

Profiles of outwash terraces on log. plat. Streams include Black, Chippewa, Eau Claire of Marathon Co., Rock, Upper and Lower Wisconsin all in and near Wisconsin. For comparison the main terrace of the Lower Wisconsin is also shown. Exponents of outwash terraces vary from .55 to .85 and average 0.68 Constants vary from 6.6 to 31.6 apparently inversely to size of stream which is related to length of ice front which it drained. The terrace was due to erosion by drainage from a much longer ice front through two glacial lakes. It is almost a straight line slope like that of present river and has a constant of only 1.9

0 $1 = 2^0$

1 $4 = 2^2$

2 $8 = 2^4$

3 $16 = 2^6$

4 $32 = 2^8$

5

$= 2^{n-1}$