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Review

HYDROGEOMORPHOLOGY

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ABSTRACT

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A review of the current state of our knowledge of the mechanical effects caused by (liquid) water on the morphology of the Earth is presented. The mechanical interaction of surface water and groundwater with landscapes is described and the mechanical effects of the sea on coastal regions and on the ocean bottom are discussed.

FUNDAMENTALS

Scope of paper

The term "hydrogeomorphology" designates the study of landforms as caused by the action of water. By this definition, almost all of geomorphology is "hydro"-geomorphology, because water is the most important of the agents in the shaping of land forms. However, this paper is not intended to be a review of *all* of geomorphology; such a review has recently been provided by the writer in his book on theoretical geomorphology (Scheidegger, 1970). Rather, we shall confine ourselves here to the more violent effects of water upon landscapes. Furthermore, the effects of the solid form of water, i.e. of ice, upon the ground will not be considered here either.

Naturally, some of the background regarding the action of water upon the ground will have to be provided to make this review meaningful. However, in this we shall confine ourselves to the most recent developments only, inasmuch as for the earlier achievements the cited book provides all the information required.

The hydrological cycle

Any discussion of geomorphological effects of water has to start with the hydrological cycle. Starting from oceans and lakes, water evaporates, is carried aloft and finally condenses in the sky to form clouds. Eventually, it is precipitated back to the earth as rain, snow or dew. Some of the water then seeps into the soil and rock to proceed as groundwater, the other part forms surface water. Finally, the water returns again to the ocean.

The above description of the hydrological cycle is greatly simplified. Evaporation does not only occur from oceans and lakes, but also from plants (evapotranspiration), from rivers, and directly from the soil so that in the "large" cycle there are also small, partial cycles.

For the global hydrological balance one has (see De Wiest, 1965, p. 38):

precipitation in oceans	81.9 cm/annum
precipitation on land	67.4 cm/annum
evaporation from ocean	92.1 cm/annum
evaporation from land	42.0 cm/annum

Multiplying these values with the respective areas (oceans 70.8%, land 29.2% of the Earth's surface) does not yield a balance; thus there is a net runoff of water from the land to the sea; in land areas, it amounts on the average to about 25 cm/annum.

It is the runoff-water which causes geomorphic effects. This is the water that flows on and through the ground, thereby affecting the landscape. The part on the ground is termed surface water, the part in the ground is termed groundwater. The surface water is collected from drainage basins (catchment areas) into rivers and streams, susceptible to various amounts of flow. The groundwater can cause the fashioning of caves, can affect the stability of mountain sides and can trigger geodynamic phenomena.

Furthermore, the ocean (and other large bodies of surface water) can interact with the land on coasts and shore lines. This leads to rather special phenomena of great interest. Finally, morphological effects can also occur within large bodies of water on the land- (or rather, "sea"-) scape below its surface.

We shall discuss these phenomena in their turn below.

EFFECTS OF SURFACE WATER

River erosion

The geomorphological effects of hydrological phenomena are caused by the fact that they cause deviations from the equilibrium conditions. The geomorphological action of rivers in general can be split into bottom (channel) erosion and lateral erosion. The steady-state cases were summarized by the writer (Scheidegger, 1970).

Regarding bottom (channel) erosion we recall that, in the steady state, rivers cause a transportation of sediment along their bed. Over the course of the years, the river is at grade, ripples alternate with pools and differential transportation causes a regular gradation of pebbles along the profile. The various theories of "river bed processes" have been reviewed at length by the writer (Scheidegger 1970). The problem of riffles and dunes has been investigated more recently by Callander (1969) and by Engelund (1970) who showed that the instability of the turbulent flow can well explain these phenomena. Similarly a new study of the threshold of bedload transportation has been made by Müller et al. (1971).

Typical small-scale sedimentary structures have been studied recently particularly by Allen (1968) who wrote a book on current ripples. Other studies on the same subject have been reported by Kennedy (1969). Allen (1971b) has also investigated the provenance of longitudinal bed marks.

It is the nature of a geomorphologically significant hydrological event that it represents a deviation from the steady state: Generally one has to do with an increased flow. The general effect of such an increased flow is clear. For a short period of time, the carrying capacity of the river is greatly increased. This manifests itself in the increased erosion of large material in the headwater region, and in the deposition of such large material further downstream. In other words, the pebble-gradation curve is shifted downstream. The river profile is also affected ephemerally because of the general relation between bedload transport and channel shape (Wilcock, 1971).

The above statements are rather qualitative. Nevertheless, the general character of the predictions has been verified by Bradley et al. (1972) in an investigation of the effects of flood flows of the Knik River in Alaska. Coarse boulders have been found carried over a great distance. However, accurate investigations of the theoretical-mechanical aspects of the geomorphic devastations to be expected by river floods do not seem to have been carried out as of yet, expect that Bradley et al. (1972) have made some scaling experiments in an attempt to explain their observations.

What has been said above for flood channel erosion, also holds, mutatis mutandis, for lateral (bank) erosion. The equilibrium configuration of a river at grade is a sequence of meanders (Wundt, 1949; Langbein and Leopold, 1966; Thakur and Scheidegger, 1970). It stands to reason that floods will be the main agents of change in an established meander pattern: they cause the breakthrough to form an oxbow lake, the widening of loops, etc.

The writer is not awave of any studies, theoretical or experimental, of the mechanics of such violent changes. The available investigations deal with equilibrium conditions. Only Schumm and Khan (1972) made an experimental study of the effect of changes, albeit slow ones, in the external conditions (bedload, climate, etc.), but not of catastrophic changes, such as would be represented by the passage of a flood wave through and across the meanders.

The subject is therefore still a wide open one.

Water erosion on slopes

A heavy rainfall or snow-melt may not only cause floods in a river channel, but also exhibit a direct action upon slopes.

One can try to consider a slope simply as an impermeable (or partly permeable) inclined plane upon which a sheet of water develops which, then, flows downhill as a "sheet flood". This sheet flood, in turn, will act upon the underlying material. The dynamics of the buildup and decay of such a sheet flood due to a steady rainfall of finite duration was studied by Henderson and Wooding (1964). Accordingly, the dynamics of the water sheet is determined by a continuity equation and an equation of motion. The continuity equation is:

$$\partial q/\partial x + \partial h/\partial t = \nu_0 \tag{1}$$

where q is the flow or discharge rate; h the depth of the water; ν_0 the rainfall rate per unit area; x the downslope coordinate and t time.

Seepage into the plane also could be taken into account. The equation of motion is:

$$q = \alpha h^n \tag{2}$$

where there are two different cases:

$$\text{(a) laminar flow } \alpha = gS/(3\nu) \qquad n = 3 \tag{3a}$$

where g is the gravity acceleration, ν the kinematic viscosity and S the slope gradient;

$$\text{(b) turbulent flow } \alpha = C_1\, S^{1/2} \qquad n = 3/2 \tag{3b}$$

where C_1 is the Chézy coefficient.

If eq. 2 is inserted into eq. 1 one ends up with a partial differential equation for h (hyperbolic type) which has to be solved for the appropriate initial and boundary conditions; this is most conveniently done by the method of characteristics. It turns out that, during the buildup phase, a plateau parallel to the slope rises with the constant velocity ν_0, until a steady state given by the curve:

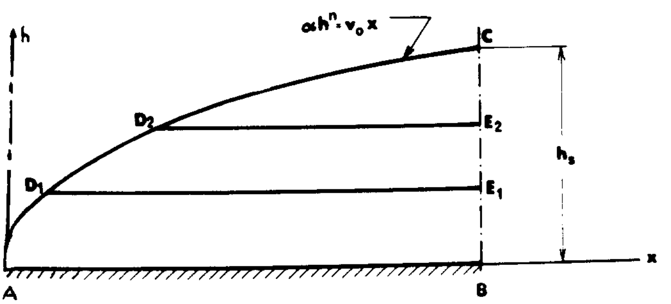

Fig. 1. Profile during buildup phase of a sheet flood. (After Henderson and Wooding, 1964.)

$$\alpha h^n = v_0 x \tag{4}$$

is reached. The situation is illustrated in Fig. 1, in which AB, AD_1E_1, AD_2E_2 ACB represent subsequent profiles. Referring to Fig. 1, the steady state is reached after a time t_s:

$$t_s = h_s/v_0 = L/\alpha h_s^{n-1} \tag{5}$$

where h_s is the depth at point B. The outflow at B is:

$$q = \alpha h^n = \begin{cases} \alpha v_0^n t^n & (t < t_s) \\ L v_0 & (t > t_s) \end{cases} \tag{6}$$

The decay phase (after cessation of the rain) can also be calculated, but the discussion of the case is more complicated.

The erosional activity of a sheet flood can be calculated according to one of the drag force equations for rivers (cf. Scheidegger, 1970). The forces of flowing water can be substantial, as evidenced by the devastations that can be caused by catastrophic floods in plains which come probably closest to the picture of a "sheet" flood considered above. Nevertheless, even the damage caused by such floods is generally to a large extent due to the fact that the water enters cellars and houses, not due to its streaming. Thus, the water depth h is of more significance than the velocity v.

Except for floods in plains, the cases encountered in nature are generally quite different from the theoretical picture of a "sheet flood", inasmuch as innumerable gulleys are immediately formed. This is most readily visible in the erosional features in badlands or heaps of mine tailings. Gerber and Scheidegger (1973) have analysed the characteristic features of such erosional gulleys and have shown that they are recognised by the fact that they have the

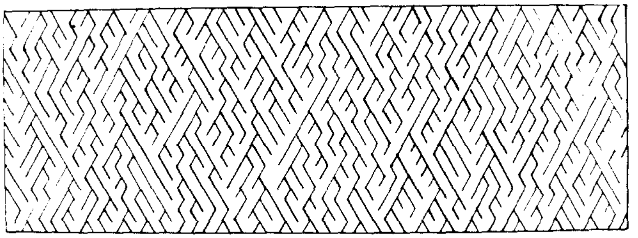

Fig. 2. Theoretical erosional gulley pattern. (After Gerber and Scheidegger, 1973.)

features of bifurcating topological trees. A theoretical gulley-pattern is shown in Fig. 2. The actual collapse of the features created by the gradual erosion of the gulleys (this by the usual drag-forces of the flowing water), i.e. of the "relics" between the gulleys, occurs through the eventual exceeding of the statically stable stresses; cases of such – sometimes very dangerous – collapses that occurred in a sand quarry have been described in the cited paper of Gerber and Scheidegger.

Tectonic effects of surface water

Surface water can cause tectonic effects which may be quite severe. It is well known that the filling up of artificially created reservoirs has caused swarms of earthquakes. A recent study of such cases has been given by Nikolayev (1972) and by Caloi (1970). It is in the periods of rapidly alternating decreases and rises of the water level that induced microseismicity and seismicity is observed. Thus, it is particularly the reservoir loading which causes local earthquakes (Rothé, 1969, 1970; Carder, 1970). Instances of this occurring have been observed in the United States (Hoover Dam), Spain (Canelos Dam), Greece (Cremasta Lake; Comniakis et al. 1968), India (Koyna Dam; Gubin, 1970), Pakistan (Adams, 1968), South Africa (Kariba Lake; Gough and Gough, 1970) and other countries. The induced earthquakes were sometimes found to reach nearly 6 on Richter's magnitude scale, which caused loss of human life and severe damage. Seismic activity is particularly likely to be induced if the water depth of the reservoir is 100 m or more. Not all reservoir fillups cause earthquakes; evidently the geotectonic conditions have to be such that stresses waiting for a triggering effect to release them have to have been already accumulated.

The above instances refer to the creation of artificial lakes. However, surface water can cause earthquake swarms also under natural conditions. It is again the load due to the weight of the water which causes these phenomena.

Thus, numerous earthquakes occur in the Mississippi Valley within a range of 200 miles of New Madrid (Mo.) which have been felt with an intensity of up to 7 units on the Mercalli scale since 1900, but have been experienced with 12 intensity scale units (total destruction) at New Madrid in 1811 and 1812. A careful study of the correlation of these earthquakes with the stage of the Mississippi River has been made by McGinnis (1964), who plotted the mean monthly stages against the observed bimonthly frequency of earthquakes. It is seen that there is a good correlation between earthquake frequency and the derivative of the river stage. The phenomenon can be explained by the changing load of the river. The difference in height between the highest and lowest river stages is somewhat more than 10 m, resulting in stress changes of up to $5 \cdot 10^5$ dynes/cm^2 at depth. It is quite conceivable that such stress changes will trigger earthquakes.

GEOMORPHIC EFFECTS OF GROUNDWATER

Principles of seepage

Groundwater is the water which does not flow on the surface of the earth, but rather flows within it. Like surface water, it can have various geomorphological effects.

The groundwater occupies the pore space of the soil and rocks, forming an aquifer. The aquifer may be confined or unconfined. The confined aquifer is bounded above and below by impermeable layers; an unconfined aquifer is bounded only at the bottom by an impermeable layer; above, the groundwater forms a free surface.

The flow of groundwater is seepage flow, governed by Darcy's law (Scheidegger, 1973). In differential form this law can be written as follows:

$$\boldsymbol{q} = -\frac{k}{\mu} \quad \operatorname{grad} p + \rho \boldsymbol{g} \tag{7}$$

where $\boldsymbol{q}$ is the seepage velocity vector (given as the volume seeping in unit time through a unit area); k the permeability (a constant characterizing the porous medium); μ the viscosity of the percolating water, p the pressure in the liquid in the pores; ρ the density of the liquid and $\boldsymbol{g}$ the gravity acceleration vector.

Darcy's law alone is not sufficient to make a flow problem determined; in addition, a continuity equation is needed:

$$-P\frac{\partial\rho}{\partial t} = \text{div}\,\rho\boldsymbol{q} \tag{8}$$

where P is the porosity (a fractional pure number characterizing the porous medium; it is equal to the average ratio of pore volume to bulk volume) and t time. Combined, eq. 7 and 8 yield:

$$P\frac{\partial\rho}{\partial t} = \text{div}\left[\rho\frac{k}{\mu}\left(\text{grad}\,p - \rho\boldsymbol{g}\right)\right] \tag{9}$$

which is a partial differential equation for ρ. If the connection between ρ and p (the rheological equation of the fluid):

$$\rho = \rho(p) \tag{10}$$

is added, a flow problem is determined, provided the applying initial and boundary conditions are known.

Any groundwater flow problem, at least in a first approximation, is described by a solution of the flow equations presented above. Under quite general conditions, eq. 9 can be approximated by a diffusivity equation, so that many groundwater flow problems can be solved by finding the appropriate solutions of a diffusivity equation.

The description of groundwater flow by Darcy's law may sometimes be in need of further refinements. The capillary pressure (the "suction" in the undersaturated zone in soil physics), nonlinear deviations from the linear flow law (in flow regions corresponding to a Reynolds number of more than 1) and molecular effects may have to be taken into account. For geomorphological applications this is, however, not very important.

Effect of pore water on slope stability

One of the main effects of pore water is on the mechanical resistance to collapse of a material. In a granular material, the yield-condition is given by the Mohr-Coulomb equation:

$$\tau_s = \sigma\tan\phi + c \tag{11}$$

where τ_s is the shearing limit; σ the normal ("overburden") stress; c a constant indicative of the cohesion of the material and ϕ another material constant, called the angle of internal friction. In a water-saturated assemblage of grains, the basic relation (11) is changed. The relation for the shearing resistance is (after Terzaghi and Peck, 1948; see also Bishop, 1959; Bishop and Bjerrum, 1960):

$$\tau_s = c + (\sigma - p_w) \tan \phi \tag{12}$$

Thus, all one has to do is to replace the normal stresses by the effective stress σ_e given by:

$$\sigma_e = \sigma - p_w \tag{13}$$

in the usual equations for slope stability. It is seen that the shearing resistance may become zero if p_w becomes large enough; in fact, eq. 12 breaks down for negative σ_e. Under such conditions, the porous medium undergoes "splitting" failure. Thus, slopes that are stable under dry conditions may become unstable when water-logged: It is a well-known fact that protracted periods of rain (by raising the pore pressure) may trigger landslides. A similar rise in pore pressure caused by the filling up of the artificial lake in the Vajont valley may have been a contributing cause of the catastrophic slide that occurred there on October 9, 1963.

The situation in rocks is similar as in soil masses. A review of this case has been given by Serafim (1969). One again has to take the effective pressure as defined in eq. 11 for the calculation of the mechanical response of the material. The flow in the joints has to be taken as a separate flow-system.

Consolidation

A further geomorphological effect of ground(pore)water occurs in the phenomenon of consolidation. The latter represents a two-fold process: the (porous) medium is undergoing finite deformations whereas the water contained in it is being displaced.

In consolidation studies the flow of the fluid per se is only of minor interest. Thus the flow velocity is eliminated altogether and one is left with equations describing the deformation of the mass as a whole. The standard way to do this (Terzaghi, 1951) is to assume that the effective pressure (cf. eq. 13) which under constant overburden pressure, is a linear function of the pore pressure, affects the porosity linearly. The pore water movement is then calculated according to Darcy's law which leads, as noted before, under quite widely applicable conditions (Scheidegger, 1973) to a diffusivity equation. Because of the assumed linear connection between pore pressure and porosity, one obtains a diffusivity equation for the description of the space-time dependence of the consolidation process under an imposed load.

In detail, the consolidation theory can be set up as follows. Let us consider an horizontal layer of thickness T of a consolidation-prone material. We assume with Terzaghi that the change of the ratio ϵ of void to grain volume is a linear function of the effective pressure p_{eff}:

$$\frac{\mathrm{d}\epsilon}{\mathrm{d}p_{\mathrm{eff}}} = a \tag{14}$$

Since:

$$p_{\mathrm{eff}} = p_{\mathrm{overburden}} - p_{\mathrm{liquid}} \tag{15}$$

one has equivalently:

$$-\frac{\mathrm{d}\epsilon}{\mathrm{d}p_{\mathrm{liquid}}} = +a = \frac{\mathrm{d}\epsilon}{\mathrm{d}p_{\mathrm{eff}}} \tag{16}$$

or:

$$\frac{\partial \epsilon}{\partial t} = -a\ \frac{\partial p_{\mathrm{liquid}}}{\partial t} = +a\ \frac{\partial p_{\mathrm{eff}}}{\partial t} \tag{17}$$

The ratio ϵ which is commonly used in consolidation theory, is connected to the otherwise more customary porosity P by the following well-known obvious relation:

$$\epsilon = \frac{P}{1-P} \tag{18}$$

A change of porosity causes a reduction of the thickness of a thin slice which can be calculated as follows. The bulk volume of the porous medium is the sum of the pore- and grain-volumes:

$$V_{\mathrm{bulk}} = V_{\mathrm{grain}} + V_{\mathrm{pore}} = (1-P)\ V_{\mathrm{bulk}} + PV_{\mathrm{bulk}} \tag{19}$$

Thus, if T be the thickness and A the area of the slice:

$$TA = V_{\mathrm{bulk}} = \frac{V_{\mathrm{grain}}}{1-P} \tag{20}$$

Assuming that the volume and the area stay constant during consolidation, we have:

$$\frac{\mathrm{d}T}{\mathrm{d}t}A = \frac{\mathrm{d}}{\mathrm{d}t}\frac{V_{\text{grain}}}{1-P} = -V_{\text{grain}}\frac{\partial P/\partial t}{(1-P)^2}$$

$$= -\frac{V_{\text{grain}}}{(1-P)}\frac{1}{1-P}\frac{\partial P}{\partial t} = -V_{\text{bulk}}\frac{1}{1-P}\frac{\partial P}{\partial t} \tag{21}$$

or, since $V_{\text{bulk}} = TA$:

$$\frac{1}{T}\frac{\partial T}{\partial t} = -\frac{1}{1-P}\frac{\partial P}{\partial t} \tag{22}$$

We note that (from eq. 18):

$$\frac{\partial P}{\partial t} = (1-P)^2\frac{\partial \epsilon}{\partial t} \tag{23}$$

so that eq. 22 can be written as:

$$\frac{1}{T}\frac{\partial T}{\partial t} = m\frac{\partial p_{\text{liquid}}}{\partial t} = -m\frac{\partial p_{\text{eff}}}{\partial t} \tag{24}$$

with:

$$m = a(1-P) = \frac{a}{1+\epsilon} \tag{25}$$

This quantity is called the coefficient of consolidation.

We now consider the following problem. At $t = 0$ the overburden pressure has been changed from an original equilibrium pressure p_0 by the additional amount p_1.

This causes a hydrostatic superpressure p_n in the pore fluid which during the consolidation will slowly disappear. At the beginning, it must be equal to p_1. Because of eq. 15 we have:

$$\frac{\partial p_{\text{eff}}}{\partial t} = -\frac{\partial p_n}{\partial t} \tag{26}$$

such that, with eq. 24:

$$\frac{1}{T}\frac{\partial T}{\partial t} = m\frac{\partial p_n}{\partial t} \tag{27}$$

The continuity condition, then, requires that this consolidation causes a flow of fluid upward with seepage velocity q so that:

$$\frac{\partial q}{\partial \zeta}\,\mathrm{d}\zeta = \frac{1}{T}\frac{\partial T}{\partial \zeta}\,\mathrm{d}\zeta = -m\frac{\partial p_n}{\partial t}\,\mathrm{d}\zeta \tag{28}$$

Furthermore, Darcy's law requires:

$$q = -\frac{k}{\rho g \mu}\,\frac{\partial p_n}{\partial \zeta} \tag{29}$$

Combining the last two equations yields:

$$\frac{k}{\rho g \mu m}\,\frac{\partial^2 p_\mu}{\partial \zeta^2} = \frac{\partial p_n}{\partial t} \tag{30}$$

as the fundamental equation of consolidation.

The various solutions for appropriate boundary conditions will yield the characteristics of consolidation processes.

These solutions can be interpreted as macroscopic creep equations: under the additional pressure p_{no}, the change of the thickness T of a layer due to consolidation has the time-dependence as calculated by Terzaghi and Fröhlich (1936).

Accordingly, if a load is suddenly imposed at $t = 0$ upon a consolidation-prone material, and if one denotes the ratio of instantaneous settlement $s(t)$ and final (complete) settlement $s_1 = s(\infty)$ by ψ:

$$\psi = \frac{s(t)}{s_1} \tag{31}$$

one has:

$$\psi = \mathrm{const}\cdot\sqrt{t} \tag{32}$$

for $\psi < 0.526$; and:

$$\psi = 1 - \mathrm{e}^{-\mathrm{const}\cdot t} \tag{33}$$

for $\psi > 0.526$. It is seen, thus, that the response of a settling mass is asymptotically that of a Kelvin-body.

The consolidation theory as discussed above assumes that there is a linear connection between pore pressure and porosity. More complicated models have been studied by Bjerrum (1967) and by Garlanger (1972) who considered cases of materials which exhibit creep under constant effective stress.

A specialized form of consolidation occurs if the process can be considered as very slow, i.e. as quasi-stationary. It may then be termed compaction, particularly if only the end product of the process (consolidated sediments) is of interest. Upon this basis, Marsal and Philipp (1970) set up a model according to which the compaction process of sediments occurs in three stages. In the first stage, fresh material is deposited in water leading to a loose assemblage of grains with high porosities (35–80%). In the second stage, the overburden builds up and the material begins to compact. In the third stage, the reduction of the porosity is caused by crushing of the individual grains.

All these stages may be represented heuristically by an empirical expression of compaction defined by:

$$\mathrm{d}P/\mathrm{d}\sigma = -\eta P \tag{34}$$

where σ is the overburden pressure and P the porosity. This empirical expression will represent η as a function of P and σ. A well-known relation of this type is:

$$\eta = \frac{\ln(P_0/P)}{\sigma - 1} \tag{35}$$

If this be inserted into eq. 15 and the corresponding integration be performed, one obtains porosity–depth curves for consolidated sediments. It also explains the subsidence observed if fluid is withdrawn from underground. A wealth of cases of this type was collected by Poland and Davis (1969). Similar cases were also collected by Prokopovich (1969, 1971) who was also able to make some quantitative statements about the course of the subsidence. Naturally, the rate of subsidence depends on the withdrawal rate of underground fluids. However, once the withdrawal has stopped, there is a time lag in the reaching of the ultimate subsidence. For this time lag, Prokopovich used an exponential decay law of the following type:

$$Z - y = A \exp(-Bt) \tag{36}$$

where y is the subsidence after the date when the withdrawal has stopped; t is time and Z the ultimate subsidence. A and B are constants. From a plot of the observational data and a fit of the curve corresponding to eq. 36 the constants A, B and therewith the ultimate subsidence Z to be expected, can be calculated.

An interesting occurrence is the inverse case to that discussed above: consolidation caused by wetting of dry material rather than by withdrawal of fluids from the ground (Lofgren, 1969). This "hydrocompaction" is presumably due to the lubrication and to the reduction in intergranular strength in loose grains caused by the addition of water. A proper mechanical model of the process, however, does not yet seem to exist.

Quick sand

When the pore pressure p_w becomes equal to the overburden pressure σ in a cohesionless ($c = 0$) material, an inspection of eq. 12 shows that the shearing resistance τ_s becomes zero. This is the condition in "quick sand". A sand bed may be solid at times, quick at others, depending on the instantaneous flow regime. A bed becoming quick in this fashion can lead to veritable geomorphic catastrophes.

Related to quick sands are "quick clays". However, clays are generally quite impermeable, although the porosity and therewith the groundwater content may be considerable. Thus, it is not Darcy-flow which governs the movement of water particles in clays, but rather osmotic diffusion. Therefore, clay can become "quick" owing to an alteration of the ionic composition (changing the osmotic potential) of the interstitial water. This effect, represented by the leaching out of salt from marine clays, seems to be the cause of many quickclay slides in Norway (cf. Rosenqvist, 1953).

Geodynamic effects of pore water

The basic failure condition for a granular material containing pore water (eq. 12) leads one to suspect that there ought to be a connection between pore water pressure and geodynamic effects: if a rise in pore water pressure can reduce the shearing resistance of the material to zero, catastrophic failures could be expected to be the result therefrom.

Indeed, changes in pore pressure (through the injection of wastes) seem to have been responsible for an earthquake swarm near Denver, Colorado (cf. Bardwell, 1970; Evans, 1970) and near Rangeley, Colorado (cf. Raleigh et al., 1971). However, no corresponding correlation between withdrawal of underground fluids and seismic activity was found by Sylvester et al. (1970).

Earthquakes are an expression of a faulting process. A major role of pore pressure in faulting was postulated by Hubbert and Rubey (1970) according to whom the thrust sheets and nappes in folded mountains are made mechanically possible by the pore pressure reaching similar values as the overburden pressure. Direct observations of fault movement in relation to the local rainfall in Italy have been reported by Caloi and Migani (1972). Accordingly, there is a definite correlation. Similarly, a study of fault stability and pore pressure has been made by Byerlee and Brace (1972) in an experimental set-up in which the formation of faults was induced by varying the pore pressure in actual rocks under realistic conditions.

Karsts and caves

Further geomorphological effects due to groundwater occur by the leaching out of solid material from rocks. This lies at the root of the well-known karst formations in limestone terrains which can lead to a collapse of underground caverns and to the appearance of such geomorphic features as dolines.

Much of the present knowledge on karsts and caves was recently summarized by the writer in a book (Scheidegger, 1970) to which the reader is referred for details. Let it just be noted that the chemistry of the leaching process appears to be established and that the stability of caves can be calculated.

Of the more recent literature, a survey of the hydrology of limestone terrains (cf. Joiner, 1969) may be mentioned as well as a study of the origins of dolines (cf. Morawetz, 1970). These studies, however, are entirely qualitative and do not contribute much to an understanding of the physics of processes involved.

SHORELINES AND COASTS

Equilibrium condition

The geomorphological aspect of a shoreline or coastline is the result of a steady-state process occurring in its vicinity. Basically, it is the near-shore circulation system which determines its configuration; much of the pertinent information has recently been summarized by the writer (Scheidegger, 1970). Accordingly, the basic physical principle of coastline formation lies in an equilibrium between the longshore material transport and the beach. If this dynamic equilibrium is changed in any way, drastic changes may take place on a coast (Davis, 1972). The connection between the variables involved can be investigated by process-response studies.

The nearshore circulation occurs in the form of cells which include rip currents at their boundaries and are connected with the cuspate structure of a coast. Recent studies on this question have been reported by Bowen (1969), Komar and Inman (1970) and Komar (1971). The general structure of the nearshore circulation cells has been investigated theoretically and experimentally in wave basins. Net longshore currents are created when the equilibrium balance in each cell is not complete (cf. Longuet-Higgins, 1970). Using suitable models with regard to frictional coefficients, etc., yields a connection between the longshore current and wave parameters. It is directly related to the variation of breaker height along the shore.

Sea-level changes

An alteration in the relative position of the sea level with regard to the land can cause severe geomorphological effects. As indicated, it is the relative position of the sea level which is important; such a change can be brought about by an absolute change in sea level height or by an absolute change in the level of the land. In this, it is particularly the case of a rise of the sea level (sinking of the land) which has the most severe effects.

World-wide, there seems to have been an overall rise in sea level during the past 17,000 years. This has been established by carbon-14 dating of organic materials whose provenance is from animals and plants which are known to have lived close to sea level. Data on this problem have been collected by Shepard (1963); they were later extended by Milliman and Emery (1968) who found a lowering of sea level from 35,000 to 17,000 years B.P., so that a low point may have been reached about 15,000 years B.P. This may have coincided with the greatest extent of glaciation during the last ice age when a great amount of water was locked on the land in the then existing ice shields. Fig. 3 shows a mean curve drawn for the data collected by Milliman and Emery.

For present-day hydrogeomorphology, the course of the sea-level curve in

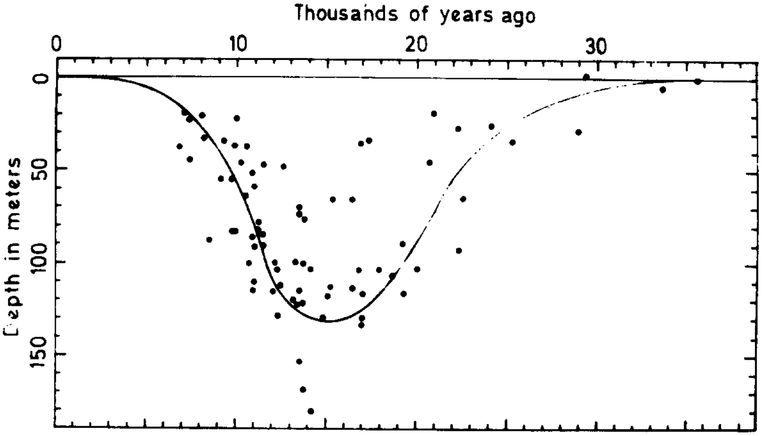

Fig. 3. Curve showing mean sea level for the past 35,000 years. (Drawn from data collected by Milliman and Emery, 1968.)

the last few thousand years or so is of importance. As is evident from the above, there has been a rise during that time; indications are that the rate of the latter has been 6 m in 6,000 years, i.e. 1 mm/year (Shepard, 1964). The above value was deduced for "stable" areas; it is a world-wide average that may not apply to specific areas. Thus, it is well known that large parts of Scandinavia have risen substantially during the past 12,000 years or so, owing to an isostatic readjustment to the ice melting off this region after the end of the latest glaciation. Thus, geodynamic effects in a particular locality may be the determining factor whether the sea advances or recedes at that place. The data on contemporary vertical movements on a world-wide scale have recently been collected by Gopwani and Scheidegger (1971) who have shown that these movements may also be (up or down) of the orders of millimeters per year; thus the net effect between sea-level rise and vertical movements may be of millimeters per year, up or down.

Additional effects may be caused by local compaction and consolidation in conformity with the regime of pore fluids (cf. section on "Consolidation"). The settling of the ground in certain areas and a corresponding invasion of the sea may have to be ascribed to such causes in a particular locality.

As noted above, it is the invasion of the sea which is of most concern to mankind. A famous case is the gradual flooding of the City of Venice in Italy (cf. Caputo et al., 1972) with catastrophic consequences. At that locality, the mean sea level seems to be rising at just about the world rate of approximately 1 mm/year (see Fig. 4). Similar rates have been found at other coastal points of Italy. Hence, there is probably little one can do to stop the process

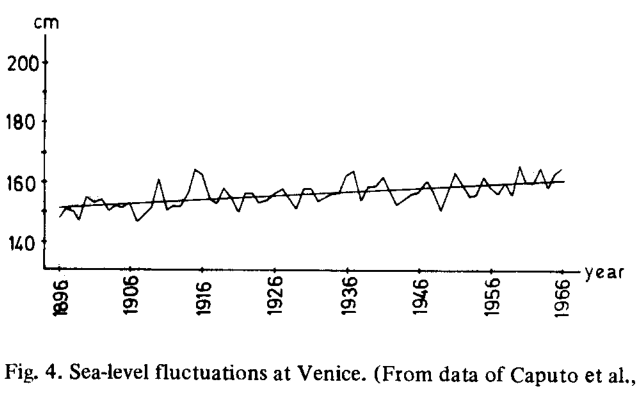

Fig. 4. Sea-level fluctuations at Venice. (From data of Caputo et al., 1972)

of land subsidence, inasmuch as it is the result of the general world-wide rise of sea level due to the melting of the glaciers.

In other places, the subsidence may be man-made, mostly due to the withdrawal of underground fluids. Cases of this type have been discussed before (see section on "Consolidation"). At Venice, the withdrawal of groundwater may in fact aggravate the situation.

Large-scale changes on coasts

The observed large-scale changes occurring on coasts are due, essentially, to a rise or lowering of the sea level. Of particular importance are the changes caused by a rise (advance) in sea level because they can cause severe damage. The particular alterations of the coast may be triggered by a particularly high flood, severe weather or such like, but without permanent changes in sea level an equilibrium would have been established long ago.

On a beach, the changes caused by a rising sea level are due to the displacement of the near-shore circulation system shoreward. Schwartz (1968) has described the mechanism. Accordingly, the material "eroded" from the upper reaches of the beach is deposited in the lower reaches and the near-shore zone in the water to restore the equilibrium that was present before the rise in sea level. The equilibrium profile is thus simply displaced upward (corresponding to the rise in sea level) and shoreward (to provide the material for the upward displacement of the near-shore zone).

On steep slopes, the mechanism of erosion is different, although here, too, the development of a new equilibrium situation conforming to the changed conditions is basic. A case of this type has been described by Hutchinson (1967) in connection with the degradation of cliffs in Essex and Kent. This degradation proceeds in the form of progressive land-slides.

Tsunamis

When a large (Richter magnitude > 6.5) shallow earthquake occurs (depth less than 50 km) whose epicenter is in the ocean, it usually causes vast water waves which can travel very far and cause disastrous effects on the coasts of the lands bordering that ocean. Such earthquake-caused water waves are called "tsunamis". Many instances of the sudden arrival of such waves have been reported, particularly from the rim of the Pacific Ocean (Adams, 1970), but also from other regions.

The problem of elucidating the physics of tsunamis has three aspects. First, there is the problem of the generation of a tsunami by an earthquake; secondly, the problem of the propagation of the disturbance over the vast distances observed; and thirdly, the problem of the action of the arriving wave upon the coast.

Regarding the generation of the tsunami, it is generally believed that it is effected by the surfacing of an earthquake-fault at the ocean bottom; in this, it is the dip-component of the fault which activates the tsunami. Physically, the problem is thus one of calculating the disturbance caused in an infinite, homogeneous and incompressible ocean of uniform depth H by an instantaneous displacement of a limited region of the sea floor represented by a step-function. The tools for solving this problem may be found in Stoker's (1957) book; the actual solution referring to the above problem was obtained by Ben-Menahem and Rosenman (1972). The result is a rather complicated expression for the far-field deep-water waves caused by the disturbance. Qualitatively, one obtains a series of more or less concentric wave-rings whose exact geometry is determined by the prevailing bottom-topography.

Next comes the propagation of the tsunami waves across the open sea. Van Dorn (1965) found that the tsunami waves propagate in mid-ocean with a (group) velocity of 234–237 m/sec, which corresponds approximately to the limiting formula for long water waves:

$$v = (gH)^{1/2}$$

for H = 5,500 m. Since the total energy of the tsunami is constant, the amplitude of the waves decreases with increasing distance from the source. The higher frequencies, however, propagate faster than the low ones, so that dispersion is observed.

The impingement of a tsunami wave onto a coast is basically governed by the same physical processes as the shoaling of any wave. Because this process is described by non-linear equations, a bore of great height may develop in a narrow channel. The exact behavior of the waves is determined by the bottom topography; strange (and as yet unexplained) resonance phenomena of oscillations in bays can cause great damage. Typically, there are 3 – 5 large oscillations, after which the motion dies down. Maximum waves may reach coastal runups corresponding to wave heights of some tens of meters above normal sea level. Such runups can cause large geomorphological effects.

HYDROGEOMORPHOLOGY OF THE OCEAN BOTTOM

General remarks

The morphological aspects of the ocean bottom are primarily caused by tectonic effects (mid-ocean ridges, continental slopes, guyots, etc.). The hydrological features of the ocean bottom, then, are superposed on these primary features. They are mainly caused by turbidity currents. Such currents are turbulent suspensions of solid materials in the water that may travel a very great distance.

The classic case of a turbidity current occurred on 18 November, 1929, when an earthquake triggered a turbidity flow; it was described by Heezen and Ewing (1952). Because of transatlantic-cable breaks, the exact progress of the flow could be traced. Other cases have been reported from the Western New Britain Trench (Krause et al., 1970). Again, the currents were triggered by earthquakes. They may, however, also be caused by the natural collapse of unstable submarine slopes. Interesting are particularly the high maximum velocities in such currents (55 knots in the Grand Banks event; 28 knots in the New Britain Trench event). The distances traveled are also great: 400 miles in the Grand Banks case, 160 miles in the New Britain Trench case. In the investigated currents, the bottom slope decreased with distance from the source; the velocity of the currents also decreased accordingly.

As noted, turbidity currents appear to be the main exogenic agents in the formation of the ocean bottom. They have been held responsible for the erosion of shelf-canyons and for the deposition of graded sediments.

Physical mechanism of turbidity currents

The possible physical mechanisms of turbidity currents were recently reviewed at length by the author (Scheidegger, 1970). Accordingly, there is no question that the suspension mechanism is caused by the prevailing turbulence There is still a problem, though, to explain the lack of quick dissipation of this turbulence into the surrounding medium. Some more recent studies have been made by Kuenen and Sengupta (1970) and by Van der Knaap and Eijpe (1968). These were experimental studies in which turbidity currents were generated in a tank. Komar (1970) made a semi-empirical analysis of the theoretical carrying capacity of turbidity currents and arrived at some correlation formulas. The problem of mixing at turbidity current heads has recently been studied more thoroughly by Allen (1971a) who was able to show that the head must be transversely "fingered out" (with clefts and lobes) to provide for a steady progress of the current. For a head of a turbidity current structured in this fashion he deduced various statistical relationships, using overall mass balance relations.

Turbidites

Turbidity currents cause characteristic features in the regions where they come to a stop and deposit their charges.

First, each turbidity current deposits a graded bed. The mechanism by which this occurs was studied by Potter and Scheidegger (1966) and by Scheidegger and Potter (1965). Flysch deposits (Middleton, 1970) also belong into this category. Most remarkable is the work of Allen (1971a) who was able to show that flute marks, tool marks, longitudinal ridges and furrows can be explained on the basis of his three-dimensional "fingering" theory of turbidity-current heads. It seems, thus, that many morphological features of the ocean bottom have found a physical explanation.

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