Erosion of Cohesionless Sediment by Groundwater Seepage

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Surface grains of cohesionless sediments eroded by emerging groundwater are worn upon by three forces: the time force of the continuous surface flow concentrated by dikeage seepage, the local seepage base, and gravity. The balance of the first moment determines the mode and rate of transport. Seepage flows are assumed to be a source "sapping zone" at the upwelling end and of the emerging flow, where erosion occurs by mass movement and the surface gradient is interrupted by the balance of the seepage and gravity moments. Most of the erosion occurs in the zone, and the resultant elevating trigger intermittent fall of overhanging slopes in a "undercutting slope" maintained at the stage of the slope to a "sapping" zone. In the "fossil" zone, the emerging seepage from the sapping zone the seepage force is small compared to the time force, and the water force by natural flow of groundwater. The overall rate of sapping erosion is the effect of the seepage transport into the sapping zone model the interaction between the sapping zone and the quantity and rate of transport. A simulation model simulating a groundwater flow model and sediment transport relationship closely replicates the observed evidence of sapping erosion in a two-dimensional tank filled with cohesionless sediment and subject to lateral groundwater flow.

INTRODUCTION

This paper examines the erosion of cohesionless sediments, by groundwater seepage and outflow, between experiments and theoretical modeling. In recent years geomorphologists have found that erosion by emerging groundwater is important in many landscapes. The earliest quantitative geomorphic studies, as those of Howard (1941, 1945), followed the conventional views that overland flow erodes slopes and channels. By the mid-1960s the recognition grew that overland flow is not universally important. Kirkby and Charley (1967) suggested that throughflow dominates hillslope hydrology where vegetation cover is present and soils are moderately thick and permeable. However, they felt that overland flow is required to erode or extend channels. DeVries (1976) recognized that seepage forces may enhance the erodibility of soils. Recently, Dune (1980, 1988) and Higgins (1982, 1984) have suggested that sapping erosion occurs in many natural landscapes. Groundwater contributes to slope erosion and channel development by several processes. The process most similar to overland flow is the outflow of pipe-like channels in cohesive soils or sediments having strong topographic relief and abundant near-surface fractures or cracks. Such hydraulically erosion is generally termed "piping" or "tunneling" (Dune, 1988) and is common in stable badlands and alluvial fans and in deserts. The pipes commonly collapse to extend the sudden dry network. Piping is enhanced by expansive clay minerals that break during rainstorms and develop destriching cracks between storms. The cohesion of the soil or shale keeps the pipes from collapsing. Intergranular flow and seepage forces that generally not important in the movement of such pipes, except possibly at their heads (Dune, 1988).

Hollis (1984) suggests that "groundwater sapping" be re-

trictive to erosional mechanisms related to emergence of groundwater, and he recommends that the term "gaseous" be used for "undercutting slope". [Sciocecmn, 1968, p. 691; Dune, 1988]. "Garrison's" sapping" (Miller, 1973, p. 4642) and "sapping sapping" (Small, 1961; Baker and Jackson, 1980, p. 600). In contrast, Soderlund rocks groundwater enhances weathering processes at the interface between the emerging groundwater and the surface, including leaching of intergranular cement and other chemical alterations, salt-feeding, frost-feeding, and physiographic chemical action of surface plants. Last and Mulm (1985) conclude that many of the weathered-valley systems in the Cascade Range are formed by groundwater seepage processes. Westerveld (1927) and Baker (1982) suggest that these weathering processes are due to the presence of weathered-rock valleys with stony, weathered badlands on Mesozoic rocks in the Cascade Range. Last and Mulm (1985) suggest that the term "gaseous" or "fossil" sapping" be used for the processes of outflowing groundwater reduces the effective particle weight, which facilitates entrainment and the development of badlands. Higgins (1982) describes the small drainage systems on some beaches that form during falling tides by emerging groundwater. Similar small networks have been observed by the present author on the sandbar islands "beaches" of the Colorado River in the Grand Canyon (subject to diurnal water level fluctuations due to releases from Glen Canyon Dam) and on sandy flood terraces along other rivers, and even in sand traps on golf courses. Dune (1980, 1988) argues that the "exclusion of" or "return flow" of shallow groundwater is widely important in extending and maintaining first-order channels. The three types of processes described above will be referred to in this paper as "normal sapping," "gaseous outflowing," and "gaseous-induced transport," respectively. All three processes

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occur in natural stream networks and slopes, but their relative importance is often uncertain due to indiscernible observation and poor characterization of processes. The theory and experiments discussed here pertain to seepage-induced transport and help to quantify the seepage processes and to explain the resultant fluid flow and slope morphology.

**Seepage-Induced Transport**

Seepage aids slope erosion and channel development directly by the destabilizing effect of seepage forces and indirectly by contributing to overland and channel flows. A few studies with partially contradictory results have examined the role of seepage forces in sediment entrainment. Soil engineers have long known that seepage forces reduce the effective weight of soil particles, and that quicksand results on a level surface when the upward hydraulic gradient equals the ratio of the buoyant unit weight of the soil to the unit weight of water. Taylor [1981] analyzed the stability of a saturated, infinite slope composed of uniform noncohesive sediment with flow parallel to the surface, finding that the maximum stable slope angle is approximately one half of the net angle of internal friction of the dry soil. Hacqel [1984] and Inore and Major [1986] extended the analysis to the case of inclined and outward seepage, but did not consider the attractive forces exercised by surface water.

Upward seepage 'flow channels' increases the lift on particles near the bed but reduces drag forces due to the outward discharge [Martin, 1970; Watters and Rao, 1971; Oldenburg and Brown, 1974]. The effects of seepage on an individual particle diminish rapidly as it is moved away from the bed. In general, these studies find that upward seepage enhances entrainment, but generally negligibly, even when seepage forces are close to causing an unstable (fusiloid) bed. Harrison and Clayson [1970] concluded from field evidence that emerging seepage into a sand-bed stream near the Sherman Glaucite, Alaska, greatly enhanced competence relative to a nearby section where the stream was recirculating into the bed. However, their laboratory experiments gave results more in line with the aforementioned studies, suggesting little effect of seepage.

The slope stability studies cited above neglect surface flow forces, while the channel flow studies used gradients that were well below the limit of bulk stability of the bed sediment. In a headwater channel being eroded by seepage-induced transport there is a range of conditions on proceeding downstream ranging from dry or damp mass wasting by undermining through a zone of seepage-induced slurry flow to transport in channels partially affected by seepage. These zones are discussed in the next section presenting an empirical database on seepage-induced transport that was collected in a series of experiments in a chamber filled with sand and subjected to groundwater flow and seeping erosion. The succeeding section introduces a theoretical model of sediment transport under conditions of emergent seepage coupled with surface flow. The experiments are distinctly tested to see the theoretical model through the use of a simulation approach.

**Experiments**

A series of experiments were conducted to observe and quantify the process of seepage-induced transport in cohesionless sediments to test the applicability of the theory presented below. An experimental chamber was constructed with dimensions 2.5 m in length, 0.6 m in height, and 1.1 m in width with Perspex sidewalls (Figure 1). One end of the chamber is sealed to form the fluid head reservoir, and an adjustable outflow scoop forms the opposite end. In the experiments reported here, the outflow was fixed at the level of the chamber base. A movable screen is placed near the upstream end to support the sediment while allowing hydraulic continuity with the fluid reservoir. On one sidewall an array of piezometer ports (15.2 cm interval and 5.1 cm vertical spacing) allows observation of hydraulic head distribution in the sediment. The other sidewall is used to monitor the temporal evolution of seeping erosion. The narrow chamber allows the system to be rotated analytically in two dimensions, in laminar groundwater-flow the sidewall effects are negligible, and in the fluvial outflow channel that forms downstream the wash depth ratio is always acceptably large. Most experiments have been conducted with a fixed reservoir head which is maintained by continuous inflow and a drain adjusted to the selected head.

Following the similarity criteria presented in the work by Inore [1986] the experiments in the chamber are a small but representative fluvial and groundwater flow system, so that the results of fluid or flow properties are not necessary. Accordingly, water was used as the fluid and quartz sand as the sediment. Two grades of well-sorted crushed quartzite sand were used in the experiments. Relevant physical information is presented in Table 1. The coarse sand is referred to as no. 3 Q Rock (no. 3 Q R), and the medium sand is no. 1 fine sand (no. 1 F-S).

In preparation for each experiment the dry sand was poured loosely into the chamber in a series of 1-10 cm layers until the

<table>
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<th>TABLE 1. Grain Size Distribution Parameters for Sands Used in Experiments</th>
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<td>Sand</td>
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<tr>
<td>No. 3 Q rock</td>
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<td>No. 1 Fine sand</td>
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Values are in phi units (-log d, d in millimeters). Mean size also shown in millimeters. Sands are from Pennsylvania Glass Sand Corporation, Berkeley Springs, West Virginia.
Fig. 2. Evolution of sapping erosion in experimental chamber for five experiments. Elapsed time is given in seconds.
Water head in reservoir shown at left.

Total thickness (53 cm in the reported experiments) was obtained. Loose sand was used in the experiments both because the density of packed sands is difficult to control and because the strong capillary cohesion of dense sand tends to create enormous overhangs as the sapping face retreats headward. Sand was allowed to spill through the outflow spout until the sloping face attained the dry angle of repose. Bulk densities were computed from the tank volume and the sand weight. Tap water was used as the fluid in the chamber, and temperatures were monitored to estimate fluid viscosity.

At the beginning of each experiment the water flow into the reservoir was started with the head fixed at a value just below that which initiates sapping erosion. After an equilibrium outflow became established, the head was raised to the desired value for the experiment and was fixed at that level. At intervals during the experiments several sets of data were collected. The sand surface was sketched on the sidewall with a marker pen. Outflow discharge was measured, and sediment samples were collected. Head readings were taken from the piezometer tubes attached to the array of ports.

The evolution of the experimental run is shown in Figure 2. The individual profiles are subject to several types of errors. Relief of up to 3 cm occurred across the tank, particularly near the zone of active sapping. The profiles were drawn as close as possible to the average elevation, but errors up to 1 cm probably occurred. The individual profiles converged downstream, so that the definition of the downstream portion of the profiles is poorer than that near the upstream end.

Finally, random errors probably occurred during transfer of information from the tank sidewall to graph paper.
Upon progressively dividing downstream along a typical erosional profile, several distinct zones are observed (Figure 1). The upper 10-45 cm of the sloping sand surface is composed of dry sand at its angle of repose. This surface episodically slumps by dry slumping as it is inundated by sapping erosion. Below the dry sand is a generally steeper slope, often overhanging, composed of damp sand of the capillary fringe. This zone was about 10 cm thick for the no. 3 Q-S and 20 cm thick for the no. 1 F-S. The damp sand slumps at the same times as the overlying dry sand, partially by flowage and partially by shear. The combined dry and damp sand forms are referred to as the "undermining zone."

Most of the hydraulic erosion is concentrated in the uppermost zone of seepage outflow, where there are steep surface gradients and rapid, upward supergeese. This "sapping zone" is about 2.5-2.5 cm in length, depending upon flow conditions and the grain size, with gradients of 17' to 17'. The surface of this zone is wet and smooth, although the flow rarely covers the grains. The appearance is similar to the dilatent reaction upon shattering ash. The zone is often ill-defined. Individual surface grains move downstream, but intermittent bulk flow failure predominates, occurring at intervals of tens of seconds to a few minutes, depending upon grain size and flow conditions. The failure area are a few millimeters thick and are a type of translocational or slurry flow (Parsin and Costa, 1987) as defined by the outward seepage; they are miniature examples of the seepage-induced slope failures observed by Ivanov-Morin and Major (1988). Supergeese flow failures in the sapping zone occur for each failure in the undermining zone.

Further downstream is the "fluvial zone," in which the water flow is generally several grain diameters thick and grains move individually as bed load. Channel gradients range from 5' to 7' at the outflow 10' to 13' at the transition to the sapping zone. In contrast to the sapping zone, low bed forms, generally oblique bars, are common. Whereas the height of the undermining zone decreases during each experiment and the length of the sapping zone remains nearly constant, the fluvial zone expands as the sapping face retreats (Figure 2). During high experimental discharges (20 mL/s) the flow extended across the 5.1 cm width of the tank, but as low discharges (< 2 mL/s) the flow occupied about 0.3 cm, increasing gradually from side to side. The combined fluvial and sapping zones are termed the "sorting channel." Due to the intermittent failure in the undermining zone and to a lesser extent in the sapping zone, the sediment and water outflow were nonlinearly variable, decreasing for a few minutes after a headwall failure as the sand dumped in the channel became saturated. The morphology of the undermining and sapping zones was highly transient, depending upon the length of time since the last major headwall failure. In particular, the sapping zone was generally broadened after head wall failures, but it would gradually re-establish.

Five experiments, three with the no. 3 Q-S and two with the no. 1 F-S, force the data base presented here. Each experiment was run with a constant head in the upstream reservoir (although load values varied between experiments). Experimental parameters are listed in Table 2. Temporal evolution of the experiments are shown in Figure 2. Koch et al. (1985) have statistically analyzed the experiments, relating the discharge of sediment and water to channel gradients and water and sediment properties and flow geometry. The following theory and simulation model provide a more fundamental analysis.

**Theory**

Figure 3 depicts the forces acting on a particle basking at the interface between the porous medium and the surface inundated produced by the emerging seepage assuming, to a first approximation, that the fluid forces act independently. The present analysis treats a cohesionless granular medium, although the force balance would be extended to include a cohesive force term. Lift forces due to surface flow are considered to be proportional to the force vector and are subject to the traction force term.

The forces acting on the interfacial particle are therefore the reactive force of the seepage flow F_s, the force of the adjacent seepage F_p, and the gravity force F_g. All forces are assumed to act through the centroid of the particle, although F_p would

![Fig. 3. Forces acting on an interfacial sediment grain.](image-url)
actually act above and F_v below the centroid. At the condition of incipient particle motion, the following balance of moments may be written [Howard and McLane, 1981]:

\[ P_{d} \cos \alpha + F_{v} \cos (\theta + \psi - \alpha) - F_{v} \sin (\alpha - \theta)/2 = 0 \]  

(1)

where \( l \) is the grain diameter, and the angles \( \alpha, \theta, \) and \( \psi \) are defined as Figure 3. The three forces can be expressed in terms of \( k_3 \)-moment and flow properties as follows:

**Tractive force.** The tractive force acting on a single interfacial bed particle is found by multiplying the projected area of the particle by the average bed shear stress

\[ F_{t} = C_{t} \frac{1}{2} \tau \]  

(2)

where \( C_{t} \) is a grain shape and exposure factor.

**Seepage force.** The drag force acting on a single grain in a laminar, linear seepage flow regime may be expressed as

\[ F_{e} = \rho \frac{1}{2} u \Delta u \]  

(3)

where \( \Delta \) is a tortuosity coefficient that takes into account the effects of neighboring particles, \( \mu \) is the dynamic viscosity of the fluid, and \( u \) is the local average seepage velocity. The coefficient \( \Delta \) reaches a limiting value of 3\( \pi \) for a single sphere in an infinite fluid, so that for an average particle

\[ \Delta = C_{p} 3\pi \]  

(4)

where \( C_{p} \) is a packing coefficient that expresses the effect of the geometry of the pore system upon local streamline configuration around the particle.

The seepage velocity may be expressed as

\[ u = \phi \eta = K \eta / \mu \]  

(5)

where \( \phi \) is the Darcian or superficial velocity, \( \eta \) is the porosity, \( K \) is the hydraulic conductivity, \( k \) is the intrinsic permeability, \( g \) is the gravitational acceleration, \( \rho \) is the fluid density, and \( i \) is the hydraulic head gradient. A general expression similar to the Hazen equation may be written for the intrinsic permeability

\[ k = C_{p} \eta \]  

(6)

where \( C_{p} \) is an empirical constant. Substituting (4), (5), and (6) into (3) yields the seepage force acting on a single particle

\[ F_{e} = C_{p} \frac{3 \pi \phi \rho \eta}{2} \]  

(7)

where \( C_{p} \) is the product of \( C_{t} \) and \( C_{p} \).

**Particle weight.** In a similar manner the buoyant weight of a particle may be expressed as

\[ F_{w} = \rho \frac{n}{2} \left( \rho - \rho_{d} \right) \]  

(8)

where \( n \) is the density of the sediment grain, and \( C_{p} \) is a shape coefficient.

The hydraulic gradient in the direction of seepage flow can be expressed as a unique function of the slope angle and the seepage exit angle [Howard and McLane, 1981]. Figure 4 shows the angular relationships for the case in which a seepage flow line or streamline exits the face of a slope. The total head difference across the unit surface length equals the elevation.

\[ L = \frac{\cos (\theta + \psi)}{\tau} \]  

(9)

**Equation (1) gives the moment balance at the initiation of motion, and it can be solved for the critical shear stress \( \tau \), by substitution of (2), (7), (8), and (9) and use of trigonometric identities to give

\[ \tau = \frac{C_{t} \rho}{\cos \alpha - C_{p} \phi \rho \eta} \]  

(10)

where

\[ C_{t} = 2C_{p} C_{1}/C_{1} \]  

(11)

\[ C_{p} = 12C_{1}/C_{2} \]  

(12)

The approximate value of the constants \( C_{t} \) and \( C_{p} \) can be determined by considering special cases that have been experimentally examined theoretically or empirically. The same of seepage parallel to a saturated infinite slope with no runoff has been described by Lambe and Whitenon [1962, p. 353]. For this case the maximum stable slope angle is given by

\[ \tan \theta = \frac{1}{2} \tau \]  

(13)

where \( \theta \) and \( \gamma \) are the buoyant and total specific bulk weights of the sediment, respectively. This case corresponds to (10) with \( \psi \), \( \alpha \), and \( \theta \) when compared to (13) gives

\[ C_{p} = \frac{1}{2} \tau \]  

(14)

where \( \tau \) is the buoyant unit weight of an individual sediment particle.

The special case of no seepage and flow over a nearly horizontal bed allows determination of \( C_{t} \). For this case the \( F_{w} \) term in (1) is zero and the second term on the right-hand side of (10) disappears. In sediment transport studies a dimensionless critical grain shear stress, or 'entrainment function,' is defined as

\[ \Psi = \frac{\tau}{\left( \rho - \rho_{d} \right) g} \]  

(15)
so that for the threshold of motion

\[ C_0 = \frac{1}{\kappa \tan \phi} \]  

(16)

For sand-sized particles in water \(1/3\)'s is about 0.04. The angle \(\phi\) is expressively equal to the angle of internal friction for the bed sediment.

As mentioned above, the shear and tractive forces do not necessarily act through the particle centroid. Thus angle corrections might be needed to the constants in (14) and (16).

If seepage outflow occurs on a slope the surface flow may be enough that the threshold of motion defined by (18) is exceeded and active erosion and transport occurs by the surface runoff. In the fluvial zone, grains move individually and transport of bed sediment is assumed to be expressed as a relationship between \(1/\phi\) and a dimensionless sediment transport rate, \(\Phi\).

\[ \Phi = \left( \frac{Q_p}{Q_s} \right)^{1/3} \]  

(17)

where \(Q_p\) is the specific sediment discharge expressed as weight of solids per unit time per unit channel width. Many sediment transport formulas can be expressed in the form

\[ \Phi = \frac{C}{\phi} \left( \frac{1}{H} \right)^{1/3} \left( \frac{1}{W} \right) \]  

(18)

where \(C\) and the exponent \(\alpha\) are constants. The following hydraulic equations can be used to estimate the transfer shear areas if the surface flow is fully rough turbulent

\[ q = \frac{v}{\kappa \tan \phi} \]  

(19)

where \(\kappa\) is the hydraulic radius and \(v\) is the depth-averaged velocity

\[ V = \frac{C}{R^{1/2}} \left( \frac{1}{\sin \phi} \right)^{1/2} \]  

(20)

where \(n\) is Manning's roughness coefficient, and \(C\) is a constant (unit for metric units)

\[ n = \frac{1}{R^{1/2}} \tan \phi \]  

(21)

The above is used in (21) instead of the usual tangent approximation because of the steep fluvial gradients. For dimensional analysis, and by analogy the sine is also used in (21).

In the sapping zone, grains move partly by individual grain motion but partly by interstream shallow stream running. Such mass wasting will commence when the slope angle exceeds a critical value \(\theta_c\), which can be estimated by solving (10) for \(\theta_c\) with \(\kappa = 0\). This yields a quadratic equation in \(\tan \phi\):

\[ \tan^2 \theta_c \left( 1 + C_1 + 1 \right) \tan \phi + \tan \phi - C_2 = 0 \]  

(22)

where \(C_2\) is

\[ C_2 = \frac{C_1 (\phi - \phi)}{\phi} \]  

(23)

In the absence of any established theoretical or empirical transport laws, the average long-term sediment delivery by seafloor-induced mass wasting as \(Q_s\) is assumed to depend upon the amount by which the actual slope angle \(\theta\) exceeds the critical value

\[ \theta_q = \theta - \theta_c - 1 \]  

where \(\theta_q\) is a constant. Also, \(\theta_s = 0\) if \(\theta_s = 0\). Thus the run out failures in the sapping zone are represented as a solution to the above equation in which mass wasting in each sapping zone slope angle exceeds the critical value for the prevailing seafloor conditions.

The boundary between the fluvial and sapping zone is defined by the condition that the total transport \(Q_s\) is additive

\[ Q_s = Q_p + Q_s \]  

Simulation Modeling

The interplay of the groundwater flow and the sediment erosion and transport processes complicates the testing theory and interpretation of experiments. The force on the front facies particles depends upon the surface gradients, the bed seepage forces, and the accumulated surface displacement force. The temporal and spatial interaction of these force determines the seepage rate and the surface morphology. However, the surface morphology in turn determines the flow field within the sediment and thus the pattern of surface displacement and seepage forces. Therefore, an analytical theoretical solution for seepage rate and surface morphology requires a coupled solution of the flow field with a model for the sapping processes. Also, analytical solutions to the flow field for arbitrary boundary conditions is impractical. Furthermore, the finite difference solutions of the flow field would be inadequate due to the temporal evolution of the flow field and surface morphology.

A simulation modeling approach allows for an iterative coupling of the groundwater flow and the seepage processes that can incorporate, if necessary, the sediment characteristics of the flow and the seepage processes. The model presented here couples a finite element groundwater flow model with a model of sapping processes based upon the theory presented above. The model is used to evaluate the temporal evolution of the experiments, and a comparison of the stresses and the observed sapping processes is used to validate the model and explain the observed sapping processes.

The model incorporates two major elements, the flow model and the sapping process model. These, as suggested by the example, are summarized below, followed by a discussion of how these elements interact in the temporal simulation.

Flow model. A major reason for conducting experiments is a narrow test time in that two-dimensional numerical solutions of groundwater flow are much more tractable than the three-dimensional modeling. The two popular numerical methods are finite difference method, which discretizes the flow domains by different methods. However, this method is not exactly represented by grid boundaries, such as the sapping face and channel, and the temporally moving boundary is difficult to treat. The finite element method used here employs high-order nodes arranged in the flow domain as the control points of individual elements that can simulate a wide variety of size and shape. Curved boundaries are easily matched. In addition, a method was developed that allows accuracy of order of magnitude of the grid nodes to response to changing boundary conditions. Details
of the flow model are presented elsewhere [McLane, 1984], but the major features are outlined below.

The flow model was designed to simulate two-dimensional cross-sectional steady-state flow in a saturated, anisotropic, heterogeneous soil, governed by the equation

\[
\frac{\partial h}{\partial x} + \frac{\partial (K_x u)}{\partial x} + \frac{\partial (K_y v)}{\partial y} + Q = 0
\]

(26)

where \(x\) and \(y\) are the horizontal and vertical directions, \(K_x\) and \(K_y\) are the horizontal and vertical hydraulic conductivities, \(h\) is the hydraulic head, and \(Q\) is a distributed recharge. In the present simulations, \(K_x\) and \(K_y\) are assumed to be independent of \(x\) and \(y\), and \(Q = 0\) for the experimental conditions. In addition, most simulations specified that \(K_x = K_y\). An examination of the data from the experiments revealed that recharge produced by the falling water table resulting from sapping erosion accounted for only 0.5-2.0% of the total discharge, so that transient effects on the flow are weak and a steady state approximation is justified.

A semi-analytic mesh generator was developed to set up the initial nodes and elements corresponding to the starting time of the simulation. The simulations began from initial conditions corresponding to an early stage of the sapping erosion in the experiments. The flow region is first broken into eight-node, four-sided regions (quadrilateral quadrilaterals). Triangular solution elements are automatically generated within the quadrilateral quadrilaterals by an interpolation procedure, and the nodes are numbered in an order that minimizes storage and computation requirements. Figure 5a shows a typical element net. Flow conditions on external boundaries are specified as either fixed head or fixed flux. Element edges in contact with the reservoir have fixed head \(h = H\), where \(H\) is the water depth in the reservoir. Edges along the fluvial and sapping zones have the boundary condition \(h = h_b\), where \(h_b\) is the elevation of the surface relative to the base of the tank.

The flow of the tank is a zero flux boundary (\(\partial h/\partial y = 0\)) and the free water surface in the sand has the double boundary conditions of fixed head \(h = 0\) and no flow \(\partial h/\partial y = 0\), where \(y\) is normal to the water surface. Near the downstream end of the fluvial zone the elements become very small and the seepage vectors are essentially horizontal. Consequently, a fixed head boundary vertical \(h = h_b\) terminates the flow simulation near the channel end (Figure 5a), and flow through that boundary comprises the seepage exiting through the last channel segment.

The finite element solution utilized Galerkin's method of weighted residuals and techniques for ensuring self-consistent solutions and consistent nodal gradients. An iterative approach solves for the free water surface. In particular, the last element side on the seepage face at its intersection with the free surface (at the head of the sapping zone) is established as a discharge (flux) boundary. The computed discharge is applied as a boundary condition for the next iteration, giving a self-consistent solution using only internally computed boundary conditions. Flow in the capillary fringe was not modeled and provided negligible source or sink to the saturated flow region. The adequacy of the flow model was verified by simulation of flow problems with known solutions. A typical solution of Darcian velocity vectors for boundary conditions during one of the experimental runs is shown in Figure 5b.

Sapping process model. The sapping process model estimates the downstream sediment discharge \(q\), along the elements between each surface node along the fluvial zone and the sapping zone from (25). The difference between bulk volumetric sediment outflow and inflow to each node gives the rate of erosion (or deposition) at the node. The flow simulation model provides the seepage exit angle \(\phi\) required by (10). Surface gradients \(\theta\) are determined from the current surface elevations. The flow model also provides the effluent discharge through each surface element which, when summed downstream gives the specific discharge \(q\). The procedures for calculation of fluvial and mass wasting sediment and transport are discussed below.

1. Fluvial sediment transport: (10)-(21) are the basis for calculation of fluvial sediment transport rates \(q_f\). The constants \(M\) and \(C_1\) in (10) were estimated by least squares regression between the logarithms of \(W/\Phi\) and \(\Phi\) estimated from using observed channel gradients at the lower end of the channel and observed sediment and water outflow discharges. The data included runs in addition to the five experiments reported here. At the lower end of the fluvial channel seepage forces are negligible compared to tractive forces (see discussion below), so that the term in (10) involving \(C_1\) is unimportant in estimating \(1/\Phi\). Table 2 lists the sediment and flow parameters used in the regression, and the resulting estimates of \(M\) and \(C_1\). A constant channel width of 2 cm was assumed. Figure 6 (constant width) shows the calculated values of \(1/\Phi\) and \(\Phi\) together with the regression line and the Meyer-Peter transport formula [Yalin, 1977]

\[
\Phi = \left[\frac{1}{\sqrt{\Psi}}\right]^{3.31}
\]

(27)

The discrepancy between the regression and the Meyer-Peter formula is primarily due to the assumption of a constant channel width. When the estimated channel width is allowed to vary as a power function of discharge from about 1 cm for the smallest flows to 5 cm for the largest, the regression curve and the Meyer-Peter formula are nearly coincident (variable width in Figure 6). The constant width regression was used in the simulations because no empirical data was collected on
channel width and the degree of explanation \( (R^2 = 0.73) \) was the same for both constant and variable width regressions. The unsteadiness of the sediment and water discharges resulting from slumps in the undermining zone is probably the primary source of scatter in Figure 6. Because the constants \( C_1 \) and \( M \) are estimated from measurements at the downstream end of the transport rates may be biased near the upstream end of the fluvial zone.

Equations (19)-(21) were also used to estimate the relative flow depth \( D/d \), the sediment Reynolds number \( R_s = d_k \rho u^2 \nu \), and the flow Reynolds number \( R_f = u D \nu \) for the variable width case (\( \nu \) is the kinematic viscosity), giving average values of about 2.9, 36, and 700, respectively. Manning's \( n \) becomes a strong function of \( R_s \) for shallow flows with \( R_s < 500 \) [Emmett, 1970; Yoon and Wenzel, 1971; Shew and Li, 1973]; and a few of the experimental cases are estimated to have \( R_s > 500 \). Also \( u D \) is a weak function of \( R_s \) for \( R_s > 70 \). Finally, sediment transport formulas may become dependent on \( R_f/u \) when that parameter is small [Tallis, 1977].

Some theoretical and experimental work has been done on extending sediment transport relationships to shallow flows [Komars, 1976; Jones and Simons, 1985], but the regression approach used here is sufficient for the present purpose of explanation through replication rather than extrapolative prediction.

2. Mass wasting transport: the mass wasting component of sediment transport \( q_m \) is estimated from (22) to (24). Each simulation model was run with \( C_m = 0.2 \) and some cases also with \( C_m = 0 \) (no mass wasting contribution). Both types of simulations are compared below with the experiments.

All surface nodes downstream from including the first emergence of seepage (channel nodes) have sediment transport modeled by (25). Of these, nodes in the fluvial zone have positive \( q_m \) and \( q_m = 0 \), whereas the sapping zone has \( q_m = q_m \) with one or two transitional nodes \( q_m \). However, runs with \( C_m = 0 \) have no sapping zone.

The erosion in the undermining zone is implicitly modeled as a continuous process. A virtual node is simulated between the upstream end of the uppermost sapping channel node and the lowest free surface node. This virtual node is assumed to move upstream along the free surface and to deliver sediment from the undermining zone at a rate equal to the sediment flux through the uppermost channel segment. The rate of migration of the virtual node is therefore inversely proportional to the length of the undermining zone face. When the virtual node reaches the lowest free surface node, that node is converted to a channel node (fixed head), and the process is repeated. This upstream propagation of boundary conditions is sustainable because the overall rate of erosion is controlled downstream by the transport capacity in the fluvial zone.

Simulation procedure. The simulation procedure consists of calculation of the groundwater flow pattern for the given topography followed by a series of iterations of sapping erosion which modifies the topography. The alternations of flow recalculation and erosion are continued throughout the simulations. A number of simplifying assumptions and optimizations were incorporated in the model to make the computations sufficiently conservative of computer storage and processor time to be useful. These model details are presented in the work by McLane [1984], and they include a longer time scale for recalculation of the flow set than for the erosion iterations and a scheme for deforming the finite element mesh as erosion proceeds to avoid the necessity for its reestablishment.

Comparison of Simulation Model and Experiments

The simulation model can be tested against the experiments through comparison of (1) the temporal histories of water and
sentiment discharge, the observed and replicated topographic profiles, and the relative change of base and profile shapes for runs with different grain sizes and reservoir heads. The observed topographic profile for an arbitrary stage of an experimental run was input as initial conditions to the simulation model and then replicated the future evolution of the topography and flow and sediment movement. Core considerations limited the simulation to a portion of the experimental run for the five experiments. Each simulation model used one fourth to one third of the total erosion occurring during the corresponding experiment. Three simulations were conducted for portions of the experiments with no. 3 Q-R, H = 33 cm, (H is the fixed reservoir head) to investigate the model over a wider range of the erosional history. In addition, the no. 3 Q-R, H = 30.5 cm case was chosen for a series of simulations investigating the model sensitivity to parameter variations. The model and experiments are compared below for several types of topography statistics.

**Flow histories.** The observed outflow discharge functions are functions of the upstream fixed head, the spatial and temporal pattern of permeability within the sand (including anisotropy and lithobrane), and the shape of the surface profile. The total flow is less sensitive to surface profile than to the others. The adequacy of the flow model was tested separately from the erosion model by using the flow model to estimate the outflow discharge for several times during the three experiments with no. 3 Q-R (Figure 7). The upstream head and observed topographic profile were input to the flow model and it was assumed that the flow was homogeneous and isotropic. The permeability for the sand was chosen so as to match the discharge for the initial profile for the experiment with H = 33 cm. The flow model correctly replicates the relative discharge for the three different heads as well as the overall trend of discharge increase as erosion progresses (Figure 7). However, the flow model leads to overestimate the discharge occurring late in the experiments with heads of 30.5 and 33 cm, with a maximum overestimation of about 20%. The most likely explanation for the widening disparity of erosion profile with time is that small bubbles of air formed in the pore as the experiment progressed (the laboratory was generally warmer than the tap water used in the experiments), decreasing the effective hydraulic conductivity. Such bubble formation was noted to occur during the experiments, but was not quantified. The fact that the overestimates are greatest for the longest experiments supports the bubble formation hypothesis (note that an equivalent amount of total erosion occurred during all of the experiments, so that a similar range of profiles occurred for each head level). The use of deuterated water in conjunction with a radioactive water system is recommended for future experiments. However, it is clear that the flow model performed adequately for use in the simulation model. As discussed earlier, the simulated channel shapes and relative erosion rates are not strongly sensitive to various assumptions of soil permeability.

**Channel profiles.** The channel profiles for all of the experiments show similar concave form and temporal evolution (Figure 2). Channel gradients gradually diminish as erosion progresses. Most of the erosant and sediment entrainment occurs at the head of the sapping zone. Since all of the eroded sediment is carried through the fluvial zone and since the rate of erosion in this zone is slight, the channel is primarily graded to transport sediments supplied from upstream. The channel is steeper the greater the upstream head for a given grain size, whereas for a constant head the gradient is steeper for finer sand. Because channel gradients are determined by the balance between tractive forces and sediment inertia, the steeper gradients in the finer sand for a given head indicate that the diminished groundwater flow through finer sediment more than offsets its greater entrainment.

The simulated and observed profiles are similar in form, up to their evolution during a run, and in the manner in which the form varies among runs with different grain size and hydraulic head. The model and the experiment can be compared with regard to the temporal evolution of the height of the upstream end of the sapping zone Hs. The solid curves in Figure 8 show Hs during the experiments plotted versus the location Xs of the upstream end of the sapping zone Superimposed on these plots are the equivalent Hs - Xs plots for the simulations run.
with $C_0 = 0$ and $C_1 = 0.2$. The ratio of $H_f$ to $X$ is the average gradient of the channel head. The observed and simulated elevations of course are equal for the earliest stage of each simulation (shortest channel length), since the observed elevations are the initial conditions for the simulations. The simulations with $C_0 = 0.2$ show a close match to the observed average gradients with the exception of overestimation during the early stage of erosion (channel head locations less than 0.65 m from the outflow). The simulations with $C_0 = 0$ generally underestimate the average gradient. This comparison is complicated by two factors. Although the simulations give an exact location and elevation for the channel head, in the experiments the sapping zone merges gradually with the undermining zone. Second, the intermittent slumping of the undermining zone in the experiments temporarily moves the channel head downstream by as much as 13 cm. The observed profiles were generally measured after the sapping zone had redveloped. Nevertheless, after major slumps the channel head sometimes moved downstream between measured profiles (Figure 2); these regressions have been smoothed out in Figure 8, but some irregularity remains.

The shape of the simulated and observed profiles are compared in Figure 9. Because the profiles of the simulation results could not be matched exactly in time with the observed profiles, a procedure was developed to allow comparison. The observed profiles were digitized and the gradients were estimated by interpolation for four relative locations along the channel, that is at 25, 50, 75, and 90% of the distance from the outflow to the head of the sapping zone. The corresponding gradients were recorded for the simulated channels at four times during the simulation (after 25, 50, 75, and 100% of the total elapsed time). Thus for the seven simulation runs this gives a total of 28 cases for comparison of the simulations and experiments for each of the four relative locations. The experimental channel gradients corresponding to the 28 simulation cases could be found directly by extrapolation of the observed gradients between profiles before and after the simulated profile. However, an indirect approach was used because of the previously mentioned uncertainties in recording the shape of the experimental profiles and the fluctuations in channel shape produced by the intermittent headwall slumps, both of which introduce nonrepeatable random variation into the experimental data. The random errors were reduced by use of the observed profiles for the five experiments to develop regression equations for the experimental gradients at the four relative locations using the relationship

$$\theta_p = K_f H_f^n X^d$$

(28)

where $\theta_p$ is the estimated gradient, $H$ is the hydraulic head, $d$ is the grain size, $X$ is the location of the channel head relative to the outflow, and $K_f$, $n$, and $c$ are coefficients estimated by least squares regression in the logarithmic transformation of
Simulations with $C_0 = 0$ underestimate observed gradients for all four relative locations (Figure 9). Simulations with $C_0 = 0.2$ are relatively unbiased at the 25, 50, and 75% locations but overestimate gradients near the channel head (90%). Overall, the simulations with both values of $C_0$ produce a channel shape that is more concave than was observed, although the simulations with $C_0 = 0.2$ come close to replicating the average gradients (Figure 9). Because of errors and possibly bias in measurement of the experimental profiles, the differences in measured versus simulated profile shape may

![Fig. 9. Comparison of observed and simulated channel gradients at four relative locations along channel. Line shows perfect agreement.](image)

**TABLE 3. Least Squares Fitted Parameters for Estimating Channel Gradients**

<table>
<thead>
<tr>
<th>Angle</th>
<th>$K_x$</th>
<th>$a$</th>
<th>$b$</th>
<th>$c$</th>
<th>$R^2$</th>
</tr>
</thead>
<tbody>
<tr>
<td>$\alpha_{30}$</td>
<td>0.067</td>
<td>-0.01</td>
<td>-0.36</td>
<td>1.82</td>
<td>0.92</td>
</tr>
<tr>
<td>$\alpha_{30}$</td>
<td>0.27</td>
<td>0.95</td>
<td>-0.40</td>
<td>1.54</td>
<td>0.79</td>
</tr>
<tr>
<td>$\alpha_{30}$</td>
<td>0.25</td>
<td>0.04</td>
<td>-0.32</td>
<td>1.66</td>
<td>0.61</td>
</tr>
<tr>
<td>$\alpha_{30}$</td>
<td>0.75</td>
<td>0.10</td>
<td>-0.28</td>
<td>1.28</td>
<td>0.48</td>
</tr>
</tbody>
</table>

**Simulated Angles, $C_0 = 0.0 (N = 64)$**

| $\alpha_{30}$ | 0.47  | -0.10 | -0.17 | 0.85 | 0.76 |
| $\alpha_{30}$ | 1.98  | -0.10 | -0.25 | 0.54 | 0.68 |
| $\alpha_{30}$ | 2.57  | -0.14 | -0.26 | 0.47 | 0.66 |
| $\alpha_{30}$ | 0.022 | 0.05 | -0.59 | 2.57 | 0.84 |

**Simulated Angles, $C_0 = 0.2 (N = 28)$**

| $\alpha_{30}$ | 0.040 | 0.15 | -0.54 | 2.36 | 0.85 |
| $\alpha_{30}$ | 0.13  | 0.06 | -0.59 | 1.99 | 0.86 |
| $\alpha_{30}$ | 0.52  | -0.28 | -0.27 | 0.99 | 0.93 |
| $\alpha_{30}$ | 0.41  | 0.13 | -0.23 | 0.72 | 0.85 |

Subscripts in first column give position along channel relative to end equal to 100%. $K_x$ is the proportionality constant in (12) for head $H$ in centimeters, grain size $d$ in millimeters, $X$ in centimeters, and $\theta$ in degrees; $C_0$ is constant in (22); $a$ is the exponent for grain size; $d$ is the exponent for location of channel head measured from mouth; $c$ is the exponent for head of water in reservoir; $N$ is the number of cases in regression; and $R^2$ is the percent of variance of gradient explained by fitted parameters.
possibly be an artifact. However, as will be discussed below, a smaller assumed value of the friction angle makes the simulated profile shapes close to the experimental profiles.

Erosion rates. Figure 10 shows the volumetric rate of sand erosion for the five experimental runs plotted versus the location of the channel head. Although erosion rates were measured by collecting sediment outflow during the experimental runs, the temporary reduction of the sediment outflow after slumping episodes and the short duration of sample collection introduced much random variation. Therefore the erosion rates portrayed in Figure 10 were measured by calculating the cross-sectional area of sand eroded between successive observed profiles. Erosion rates increase with increases in hydraulic head and decrease for finer grain sizes. The exception is the nearly equal erosion rates for the no. 1 F-S for the two different hydraulic heads, this is due to the lower permeability of the sand for the experiment with \( H = 41.9 \) cm.

All of the experiments show a similar temporal pattern of erosion rates. The highest rates occur during the initial stages, when the total channel and sapping zone length, was less than 50 cm. Rates reach a minimum at a channel length of about 75-100 cm, and for all but the no. 1 F-S at a high head (41.9 cm), the erosion rates increase toward the end of the experiment, but are well below the initial rates. The gradual increase in erosion rates during the later stages is due to decreasing path length of flow. The early peak in erosion rates is primarily due to two coacting factors. The initial gradients near the outflow (dry angle of repose) exceed the angle of bulk stability of the sediment with seepage until the sapping zone becomes completely established, producing high sediment concentrations (up to 70% by weight) during the initial stages. Furthermore, the seepage initially exists very close to the outflow so that seepage forces are high. As erosion progresses the outflow is distributed over longer sections of channel. The short interval that is required to develop peak erosion rates in the experiments is due to the time required to reach full flow volumes after initial establishment of the head. This time is shorter for the coarse sand than for the medium sand and also for lower hydraulic head.

Erosion rates during the experiments show little correlation with the discharge (compare Figures 7 and 10), particularly during the early stages. Both the erosion rates and the discharges tend to increase steadily near the end, but the erosion rates increase by a smaller percentage than the discharge. This is due to two factors. As erosion progresses, the sediment must be delivered further downstream, which requires a higher elevation of the channel head (Figure 2). Furthermore, the seepage emerges over a longer section of the sapping and fluvial zones, reducing seepage forces.

The simulations are reasonably successful at replicating both the relative erosion rates for different grain sizes and hydraulic heads and the observed variations of erosion rates.
TABLE 4. Effect of Parameter Varies on Simulation Results

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Percent of Variation</th>
<th>Changed Gradient</th>
<th>Erosion Rates</th>
<th>Length of Sapping Zone</th>
</tr>
</thead>
<tbody>
<tr>
<td>Variance</td>
<td>25% 75% 90% 100%</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>C0</td>
<td>30 12 18 20</td>
<td></td>
<td>0</td>
<td>8 28 80</td>
</tr>
<tr>
<td>B</td>
<td>40 20 10 5</td>
<td></td>
<td>2</td>
<td>4 1</td>
</tr>
<tr>
<td>T</td>
<td>60 40 20 10</td>
<td></td>
<td>-2</td>
<td>-2 2 -2</td>
</tr>
<tr>
<td>n</td>
<td>60 40 20 10</td>
<td></td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>M</td>
<td>50 30 10 0</td>
<td></td>
<td>-2</td>
<td>-2</td>
</tr>
<tr>
<td>M0</td>
<td>30 30 30 30</td>
<td></td>
<td>-2</td>
<td>-2</td>
</tr>
<tr>
<td>k0</td>
<td>30 30 30 30</td>
<td></td>
<td>-2</td>
<td>-2</td>
</tr>
<tr>
<td>K0</td>
<td>30 30 30 30</td>
<td></td>
<td>-2</td>
<td>-2</td>
</tr>
<tr>
<td>L0</td>
<td>30 30 30 30</td>
<td></td>
<td>-2</td>
<td>-2</td>
</tr>
</tbody>
</table>

The table gives columns for parameter that is varied for given simulations, and the second column gives the percent of the parameter variation from the nominal value of 0.25 for each simulation. C0 is the initial charge of channel gradient at location 25% of the distance from the channel exit to the sapping zone, measured at the point on the profile 75, 90, and 100% of the channel gradient to the sapping zone. Erosion rates are given for each simulation at the end of the simulation time. The results show that the parameter C0 has the most significant effect on the simulation results, followed by the parameter B. The other parameters have a smaller effect on the simulation results. The simulation results were compared to experimental data to validate the model.
(Figure 7) This resulted in a slight decrease in erosion rates for the end of the simulation but almost no change in channel profiles (Table 6).

The length of the sapping zone, along with shallow, episodic mass wasting predominates over bedload transport, varies with material and flow properties. The length is relatively unaffected by individual changes in Cv, d, s, M, or K but decreases rapidly as v increases or Cv/C decreases (Table 4). In comparing the simulations for different heads and the two sizes of sediment, the sapping zone increases in length as the head is increased and/or the grain size is decreased (Figure 11a). In the latter case, since the sapping zone length is relatively independent of both K and d, the longer sapping zone for the no. 1 F-S primarily results from the higher head required for sapping rates equivalent to the no. 3 O-R. The replicated length of the sapping zone also correlates strongly with the sediment concentration in the cut-off (Figure 11b). Similarly, the temporal change of sapping zone length during the simulation for no. 3 O-R at H = 33 cm correlates closely with changes in erosion rates and sediment concentration (compare Figures 11c and 10).

**Simulation Model: Conclusions**

The success of the simulation model in replicating the erosional history of the sapping experiments indicates that the theoretical model is an adequate description of the sapping process in non-cohesive sediments. The modeling with an assumed sapping zone of mass wasting at the upstream end (Cv = 0.1) is more advantageous than the assumption of purely fluvial erosion (Cv = 0) and is consonant with the observations of shallow, quasi-static flows in this zone. Figure 12 shows the magnitudes of bedload and seepage forces acting on an interfacial particle (relative to the gravity force) as a function of discharge from the outlet for the end of the simulation with no. 3 O-R, H = 30.5 cm. The other simulations exhibit similar behavior. Along most of the channel the shear force is nearly constant relative to gravity and is much greater than the seepage force. This means that the shear force could be neglected for most of the fluvial zone. The contribution of seepage to sediment entrainment is important only in the upper 18 cm of the channel (the actual length varies with grain size and head). The sapping zone, where the channel gradient is steeper than the critical value for slope stability, is about 5 cm long for this simulation (Figure 12). The erosion rate is determined by the capacity of the fluvial portion of the channel to deliver sediment to the outlet. That is, the erosion rate in the sapping zone is limited by the rate of removal of sediment delivered to the downstream end of this zone. This “transport-limited” behavior is characteristic of sapping erosion of non-cohesive sediments. If the sediment were somewhat cohesive, erosion rates might become “detachment-limited” and controlled by conditions in the sapping zone. The transport-limited behavior and the minor importance of seepage forces to fluvial transport along most of the channel means that the evolution of the
mapping experiments and the overall rate of erosion could be modeled more simply than was done here using equations of critical tractive force and sediment transport capacity that have been developed for more typical fluvial channels. However, modeling still requires coupling of the ground-water flow and fluvial transport components due to the dependency of downstream variation in discharge upon the steepness and sinuosity of the fluvial channel.

The experiments and theory presented here are consistent with the previously cited conclusions of Harrison and Clayton (1970), Morrow (1972), Waters and Rao (1971), and Overton and Bock (1974) in that seepage force is not an important direct contributor to entrainment in the fluvial zone with sand-sized channels. However, seepage is important in the narrow muck-wasting zone at the head of seeping channels. Similar but more localized muck wasting zones might occur in some shallow, coarse-bed channels at the downstream end of beds and riffles where subsurface discharge recharges. If gradients are sufficiently steep in such situations to cause seepage-induced transport, the resulting erosion might be a contributor to the bedform development, possibly aided by decrease in competence on the upstream-facing portions of the bedforms due to the mixing mechanism discussed by Harrison and Clayton (1970).

**Figure 12.** Variation of simulated flow and transport parameters as a function of position relative to the outflow for the case of the simulation for no. 3 (A) sink, d = 0.3 cm, (B) Ratio of shear to gravity moments (c) ratio of equal to equal channel gradients (see equation 32), and (D) seepage vs. angle relative to the horizontal (see Figure 9).

**Discussion and Conclusions**

The recent studies by Hupp (1962, 1964) and others have led to a recognition of an important role of ground-water sapping in natural geomorphologic environments. To assess the intensity and spatial occurrence of such processes and their effects on landform, the processes must be quantitatively characterized. The studies reported here are a step in that direction for a limited range of materials and flow conditions. These experimental and modeling efforts are currently being extended by several related studies. The two-dimensional experimental procedures are being supplemented through use of sediments of various size, finer, and more poorly sorted than those reported here.

A three-dimensional simulation of channels with dimensions of about 1.5 m x 1.5 m x 0.6 m has been constructed to examine the drainage networks resulting when flow is no longer constrained to two dimensions. These experiments (Howard, 1984) demonstrate that sapping in cohesive-sediment canals produces channels that erode headward and braid, as suggested by Durbet (1985). Experience with materials and bed ranges similar to those of the two-dimensional experiments are being conducted to examine the effects of converging and diverging flow on the sapping process. In addition, subaerial processes in the sediment have been introduced to simulate the effects of structural influences on sapping networks. Small percentages of cement are being introduced to the sediments to observe the effects of unconsolidated material. These threedimensional simulation models of sapping processes have also been programmed and is being tested against the tank experiments (Howard, 1984).

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References
