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EROSIONAL DEVELOPMENT OF VALLEY-BOTTOM GULLIES IN THE UPPER MIDWESTERN UNITED STATES

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INTRODUCTION

A gully is an incised drainage channel, is usually steep-sided, and transmits ephemeral flow, often with a steeply sloping and actively eroding head scarp. There are no widely agreed upon dimensions for distinguishing gullies from rills. Generally, a gully is an obstacle to farm machinery and is too deep to be obliterated by ordinary tillage; whereas a rill is of lesser depth and can be smoothed over by ordinary tillage. Gullies are classified on the basis of topographic location as valley-bottom, valley-side, or valley-head gullies (Brice, 1966). A valley-bottom gully becomes a valley-head gully as its head scarp migrates into the valley head. Valley-side gullies usually result from less concentrated flow coming from diverse directions.

Gullying is important to the geomorphologist as a mechanism for slope retreat. Depending upon the type of material under consideration, slope retreat progresses at a rate dependent upon the materials competence and the energy conditions of the local environment. The study of gullies is essential not only to the understanding of natural landscapes, but also as a practical means for controlling erosion and sedimentation. To the agriculturalist and the environmentalist, gullying is of concern because of its effects on land destruction and water pollution. In this chapter we discuss the dynamics and processes of gullying that occur on a short-term basis, with some implications for the long term.

Why certain landforms respond to external influences and result in rapid gullying is still somewhat puzzling. Is the erosion and sedimentation of a gully cyclic? If so, how does one identify or predict which physical parameters will trigger the onset of erosion? Vegetative cover, climate, and other variables change with time; the terrain merely responds to these alterations.

A landscape is an energy regime with a delicate balance between the form of the system and inflow and outflow of energy. The system is continuously importing energy, and an erosion threshold may be reached if the mode of energy utilization changes. The way in which this added potential energy is stored or utilized changes when a change of state threshold is reached. An example is a bank of dry loess standing vertically along a gully. It slowly absorbs water from surface infiltration or base wetting, increasing its potential energy, and, finally, reaching the threshold where the potential energy is transferred or converted into kinetic energy. The mass of soil slumps. Thus, the addition of potential energy triggers the entire system at the threshold point. The gully wall advances.

The bank that failed is now at the base of the gully wall, and, apparently, will remain there until it is disaggregated and entrained by runoff—or enveloped by runoff—with enough energy to transport the soil particles. This transport will continue until the supply diminishes or the energy level of the stream goes below the threshold required to do this work.

Gullying results from a complex array of many processes; any one of which may operate for limited duration or landscape position. Changes in the rates of gully development can be attributed not only to changes in external forces but also, in part, to thresholds within the geomorphic system. Schumm (1973) and Patton and Schumm (1975) recognized geomorphic thresholds in the process of gully erosion. They observed critical threshold valley slopes above which alluvium was entrenched. However, in small drainage basins (less than 10 km^2) in Colorado, variations in vegetation prevented the recognition of a critical threshold slope. In this chapter we trace gully development and examine the concept of geomorphic thresholds as applied to the processes of valley-bottom gullying in small, agricultural drainage basins in the loessial soils regions of western Iowa and Missouri.

THE STUDY AREA

The gully study area consists of four small watersheds, from 0.302 to 0.608 km^2 in size, in Pottawattamie County in western Iowa (Fig. 1). They are located about 26 km east of the Missouri River, near the small town of Treynor, within Soil Conservation Service Land Resource Area M-107 (Agric. Handb. 296, 1965), Iowa and Missouri deep loess hills. The topography of each watershed is shown in Figure 2. Adjoining watersheds 1 and 2 are about 5 km south of adjoining watersheds 3 and 4.

Pleistocene Geology

The study area has been subjected to two major glaciations, the Nebraskan and Kansan. Each resulted in emplacement of till and was followed by periods of soil formation and erosion. Glacial deposits and driftless areas were subsequently covered by two major loess sheets—Loveland loess of Illinoian age, and Wisconsinan loess. Wisconsinan loess deposition began about 25,000 years ago and continued to about 14,000 years ago. After loess deposition, erosion removed materials from hillsides; the resulting downstream deposition is herein described as loess-derived alluvium. This alluvial fill, defined by Daniels and Jordan (1966) as the DeForest Formation, occupies most drainageways (Fig. 3).

The Pleistocene history of the region was described by Daniels and Jordan (1966) and Ruhe et al. (1967). The Pleistocene and Recent deposits in watershed 3 were described in detail by Allen (1971) and are shown in Figure 4.

The Kansan till is not exposed at the land surface and is found at 4 to 9 m below the surface in the drainageways and is about 18 m thick. A Yarmouth paleosol about 3 to 5 m thick is present in the upper part of the till and is overlain by about 9 to 12 m of Loveland loess on the hilltops. A 3- to 5-m-thick Sangamon paleosol is blanketed on the ridges by about 8 to 12 m of Wisconsin-age loess.

The sequence of post-Wisconsinan alluvial fill in drainageways was established by Daniels and Jordan (1966) and Allen (1971). The oldest unit of the DeForest alluvial

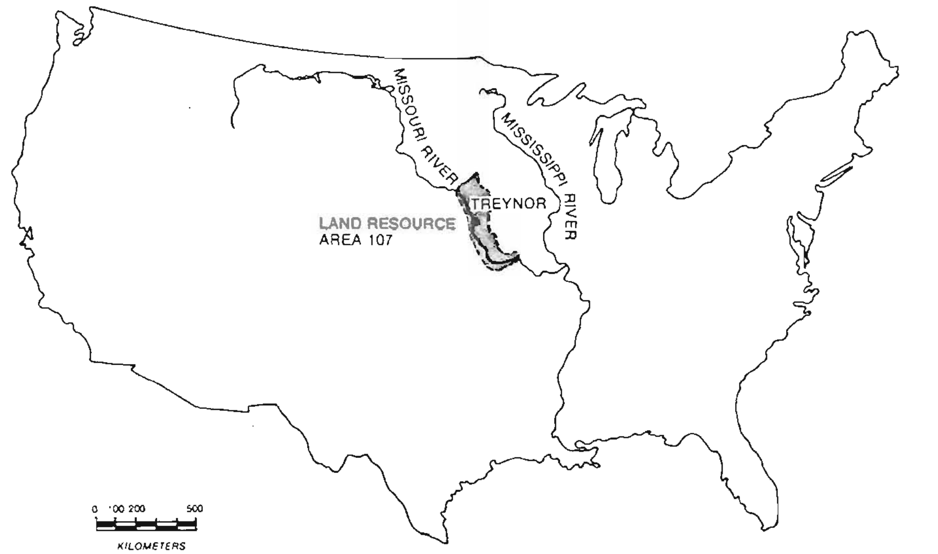


FIGURE 1 *Location of experimental watersheds within Land Resource Area 107 near Treynor, Iowa.*

formation, the Soetmelk member, lies over both Kansan till and the Yarmouth paleosol. Daniels and Jordan (1966) placed the end of this period of deposition about 11,000 years before present (YBP). Next is the Watkins member, which ranges from 11,120 ± 400 YBP at the base to 2020 ± 200 YBP at the top.

Above this is the Hatcher member dated at 2020 ± 200 YBP to 1800 ± 200 YBP. The next youngest fill is the Mullenix, 250 to 1800 years old with most of the accumulation

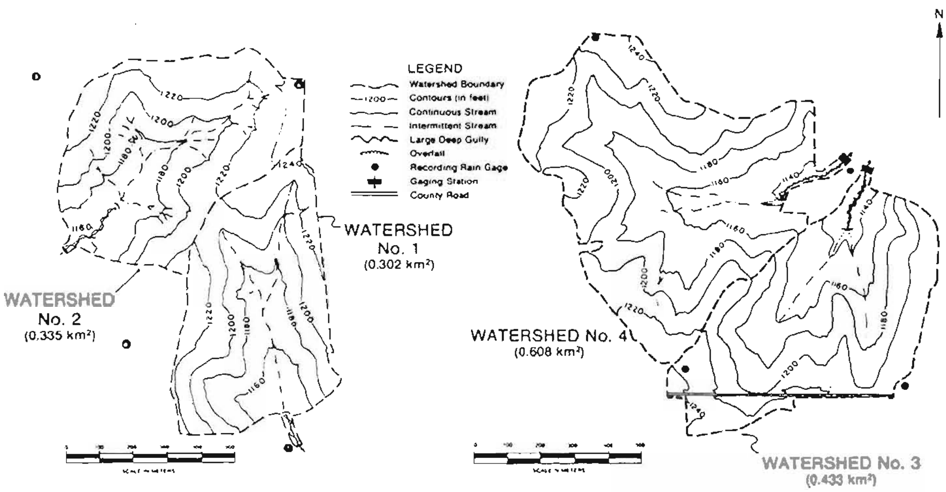
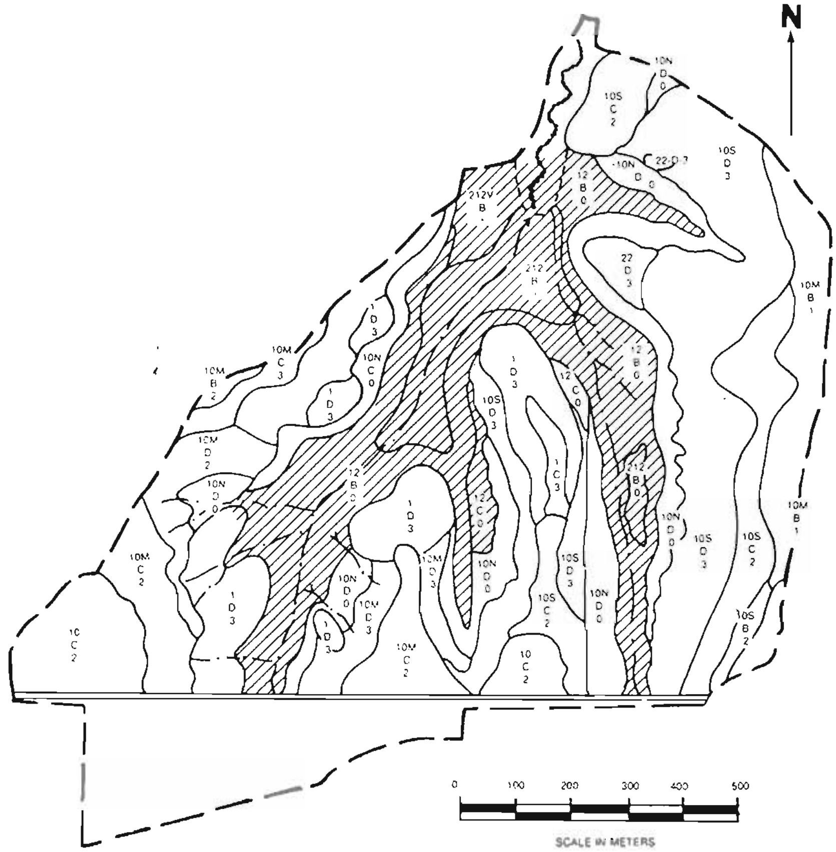


FIGURE 2 *Topographic map of Treynor, Iowa, watersheds.*

The idea of critical limits, boundary conditions,



Soil series and parent material
10 Monona silt loam - loess
12 Napier silt loam - silty alluvium
1 Ida silt loam - loess
212 Kennebec silt loam - silty alluvium
22 Dow silt loam - loess

Slope group
A 0-2 % slope
B 2-5 %
C 5-9 %
D 9-14 %

Erosion group
+ recent deposition of 17 to 44 cm
0 no apparent erosion
1 slightly eroded
2 moderately eroded
3 severely eroded

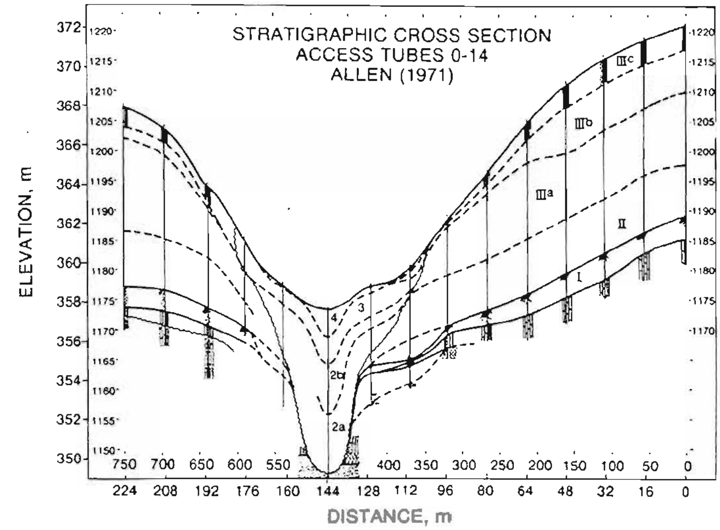
large deep gully
dark area — areas of silty alluvium
soil boundaries
watershed boundary

FIGURE 3 *Areas of natural loess and local alluvium within watershed 3, Treynor, Iowa.*

within 1100 years. The youngest fill is the Turton member dated at 95 to 250 years old. In addition, there is post-settlement alluvium that has accumulated since about 1850.

Vegetative Cover

The research watersheds have been managed so as to compare the effects of contoured corn, pasture grass, and level-terraced corn on hydrology and erosion. Table 1 lists



-1200	Mean sea level elevation (feet)
150	Distance from access tube 0
	Ground surface
	Recognized boundary
	Proposed boundary
	Observation site
4	Access tube number
	Unconformity

DeForest formation alluvium

2a	Watkins member
2b	Hatcher member
3	Mullenix member
4	Turton member

Wisconsinan loess

	Zone IIIc (Solum)
IIIb	noncalcareous increment of Zone III
IIIa	calcareous increment of Zone III
II	calcareous Zone II
I	"basal" Wisconsin loess
	basal Wisconsin paleosol
	Sangamon paleosol in Loveland loess
	late Sangamon paleosol
	Yarmouth silty clay
	Kansan till

FIGURE 4 *Stratigraphic cross section of watershed 3, Treynor, Iowa (Allen, 1971, Fig. 9, p. 78).*

the size, crop, and conservation practice of each of the four watersheds. Each watershed is in a single land use, and treatments are those commonly used in the area. The "nonconservation" watersheds are farmed on the contour with conventional tillage and planting. In 1972, watershed 3 was changed from bromegrass to continuous corn by planting directly into the grass sod. Thereafter, a mulch tillage system with minimum cultivation has been used. The effect of the sod cover has lasted for several years. Before fall 1972, watershed 4 was level-terraced with 92% of its area above terraces. The terraces had a storage capacity of about 5 cm of surface runoff. In 1972, the number of terraces was reduced by about one-half; the terrace system was made more parallel; and a complete pipe outlet system was installed to avoid excessive water pondage in the terrace channels.

TABLE 1 *Watershed Descriptions*

Watershed number	Area (km^2)	Year	Crop	Conservation practice
1	0.302	1964–77	Continuous corn	Field-contoured
2	0.335	1964–77	Continuous corn	Field-contoured
3	0.433	1964–72	Bromegrass	Rotation-grazed
		1972–77	Continuous corn	Mulch tillage
4	0.608	1964–72	Continuous corn	Level-terraced, conventional tillage
		1972–77	Continuous corn	Revised terraced system, mulch tillage

Climate and Hydrology

Climatic information before 1964 is available from records at Omaha, Nebraska, 38 km northwest of the watersheds. The mean annual temperature is 10.8°C with mean maximum and mean minimum of 16.0 and 5.2°C, respectively. The mean annual snow is 72 cm and the mean annual precipitation is 69.9 cm.

The 13-year mean precipitation (Table 2) over the four watersheds was 80.6 cm. Annual precipitation varied considerably from year to year but was similar for the four watersheds for each year. Certain years were characterized by low-intensity, long-duration rainfall, while high-intensity rainfall was characteristic of 1965 and 1967. For the study area, 1 storm every 10 years should have rainfall in excess of 10 cm/25 hr. The 1-year frequency for a storm duration of 24 hr is 7.1 cm.

TABLE 2 *Water and Sediment Yield Summary of Treynor, Iowa, Watersheds, 1964–76*

			Runoff (cm)			Sediment yield		
Year	Watershed number	Annual precip. (cm)	Base	Surface	Total	Sheet rill (kg/m^2)	Gully ($kg \times 10^3$)	Total (kg/m^2)
1964	1	90.45	4.88	11.58	16.46	5.60[a]	608[a]	7.62
	2	89.31	5.46	10.21	15.67	5.60[a]	300[a]	6.50
	3	85.06	5.99	1.07	7.06	0.07	58	0.20[b]
	4	88.39	14.38	2.01	16.38	0.16	9	0.18[b]
1965	1	115.19	9.04	26.97	36.02	9.86	1050	13.36
	2	112.65	7.54	27.13	34.67	8.16	599	9.95
	3	112.47	11.73	11.68	23.42	0.09[a]	78[a]	0.27
	4	113.97	26.82	6.38	33.20	0.20[a]	15[a]	0.22
1966	1	51.61	6.45	1.65	8.10	1.50	84	1.77
	2	52.15	6.10	2.24	8.33	1.93	161	2.40
	3	55.91	6.45	0.97	7.42	0.02[a]	9[a]	0.04
	4	55.58	15.01	0.48	15.49	0.13	13	0.16

TABLE 2 *(Continued)*

Year	Watershed number	Annual precip. (cm)	Runoff (cm)			Sediment yield		
			Base	Surface	Total	Sheet rill (kg/m^2)	Gully (kg × 10^3)	Total (kg/m^2)
1967	1	97.16	5.77	29.39	35.15	22.22	1320	26.56
	2	95.53	6.35	26.54	32.89	16.86	1250	20.53
	3	86.94	8.38	6.73	15.11	0.13	109	0.38
	4	87.76	18.49	1.85	20.35	0.65	−21[c]	0.61
1968	1	82.04	4.24	2.92	7.16	0.83	93	1.14
	2	82.55	4.62	2.87	7.49	0.92	40	1.03
	3	78.99	4.04	2.59	6.63	0.04	12	0.07
	4	81.74	10.74	0.30	11.05	0.07	2	0.07
1969	1	79.81	8.08	6.43	14.50	0.40	107	0.76
	2	80.11	7.54	5.97	13.51	0.22	50	0.38
	3	77.83	8.36	4.39	12.75	0.02	17	0.07
	4	77.98	15.52	0.69	16.21	0.02	−5	0.02
1970	1	80.04	5.61	5.44	11.05	2.65	161	3.14
	2	78.28	5.97	4.55	10.52	1.66	155	2.13
	3	73.30	5.56	0.94	6.50	<.02[d]	5	0.02
	4	73.13	10.13	0.33	10.46	0.02	<0.9[d]	0.02
1971	1	73.81	5.23	12.55	17.78	4.48	362	5.69
	2	74.09	6.65	9.75	16.41	2.98	219	3.63
	3	75.74	7.21	3.86	11.07	0.09[a]	27[a]	0.13
	4	76.40	14.02	1.73	15.75	0.34[a]	5[a]	0.36
1972	1	86.21	6.78	3.84	10.62	1.68	328	2.78
	2	86.46	7.65	3.89	11.53	1.77	109	2.08
	3	95.22	15.80	2.11	17.91	0.27	−29	0.20
	4	95.25	14.66	10.72	25.37	1.46	44	1.52
1973	1	105.94	20.78	6.63	27.41	0.22	91	0.54
	2	104.60	25.55	7.49	33.05	0.11	61	0.29
	3	103.20	37.03	2.72	39.75	0.02	−0.9	0.02
	4	102.39	30.51	8.48	38.99	0.22	45	0.31
1974	1	63.04	16.43	1.37	17.81	0.11	28	0.20
	2	62.15	21.77	1.42	23.19	0.07	34	0.18
	3	56.01	20.70	0.20	20.90	<0.02[a,d]	<0.9[a,d]	<0.02[d]
	4	53.80	18.90	0.61	19.51	<0.02[d]	5	0.02
1975	1	78.26	11.99	2.62	14.61	0.36	33	0.25
	2	78.66	19.79	2.11	21.89	0.18	34	0.07
	3	74.35	16.74	0.33	17.07	<0.02[a,d]	<0.09[a,d]	<0.02[d]
	4	73.76	16.00	3.05	19.05	0.04	5	0.04
1976	1	53.98	10.06	0.46	10.52	<0.02[a]	4	0.02
	2	53.92	12.32	0.38	12.73	<0.02[a]	2	<0.02
	3	60.71	12.62	0.97	13.59	0.25[a]	0.9	0.25
	4	64.14	10.44	1.83	12.27	0.16	12	0.18
Averages	1	81.36	8.89	8.59	17.48	3.86	328	4.91
13 years	2	80.80	10.59	8.05	18.62	3.09	231	3.79
1964–76	3	79.68	12.37	2.95	15.32	0.07	23	0.13
	4	80.34	16.59	2.95	19.53	0.27	10	0.29

[a]Division between sheet-rill and gully erosion estimated.
[b]Total and component erosion values estimated.
[c]Negative value indicates channel fill.
[d]Soil losses are less than 0.01 kg/m^2 from sheet-rill source or less than 0.4 kg from gully source.

Partitioning the total annual precipitation into its hydrologic components gives insights both into modification of vegetative cover and conservation treatments upon the erosion processes within each watershed and into the energy conditions of the local environment (Saxton and Spomer, 1968; Saxton et al., 1971; Saxton et al., 1974).

The annual runoff for each watershed and the base and surface flow components are given in Table 2. Paired, contoured-corn watersheds 1 and 2 had nearly the same precipitation and runoff; their total stream flow differed by less than 7%. The relative yearly amounts of base flow and surface runoff for watersheds 1 and 2 depended on antecedent moisture and rainfall intensity and duration. Base flow for watersheds 3 and 4 exceeded surface runoff by 300%, on the 13-year average.

Grass watershed 3 had the lowest total runoff of all four watersheds before 1972, for example, 38% less than that of watershed 1. However, 63% of its total was base flow. This lower water yield reflects the longer growing season and greater annual evapotranspiration of grass as compared with corn (Saxton et al., 1974). Surface runoff was less for grass than from corn, because infiltration rates were greater for grass. The dense grass prevented the soil surface from sealing, and the increased evapotranspiration from grass, especially from the early- and late-growing season, depleted soil moisture more than did with corn.

Total water yields from level-terraced corn watershed 4, before 1973, and contoured-corn watersheds 1 and 2 have been similar. There was a significant difference, however, in water source. Base flow accounted for most of the total stream flow from the level-terraced watershed. The large base-flow component from the terrace area was due to increased infiltration and percolation caused by ponding behind the terraces.

Gully Erosion

Gully erosion rates at the Treynor watersheds were measured, during periods of surface runoff, by sampling the sediment content of stream flow at two locations—immediately upstream and about 100 m downstream from each gully headcut (Fig. 5). With allowance for time of travel and minor added runoff between sampling locations, it was possible to monitor downstream sediment accretion due to gully scour throughout the storm.

The sediment yield from the outlet gully of each of the four Agricultural Research Service watersheds are summarized on an annual basis in Table 2. Large amounts of soil debris were transported. Gullies of watersheds 1 and 2, respectively, lost 3.3 and 2.3×10^5 kg of soil per year during the 13 years (1964–1976). Significantly smaller amounts were eroded from conservation watersheds 3 and 4.

Valley-bottom gullies within the four watersheds are characterized by near-vertical headwalls 2.7 to 4.3 m high and a total depth of 4.0 to 6.7 m. The sidewalls of the plunge pool area also have vertical walls but only after complete debris cleanout; normally, much debris is left at the base of the walls. Downstream from the headcut, the upper 0.5 to 1.0 m of the sidewall is vertical, and the lower sidewall has angles typically between 40 to 70°. Figure 6 gives the cross section of the gully approximately 14 m downstream from the headcut at watershed 1. The lower slopes, composed mainly of materials slumped from the sidewalls, are stable as long as runoff is insufficient to transport the debris from the base.

Gully voiding was sporadic. Although loosely correlated with storm runoff, the voiding rate varied according to the availability of soil debris which accumulated in the

FIGURE 5 *Aerial view of gully headcut at watershed 1 outlet drainageway, showing sampling footbridges and measuring weir.*

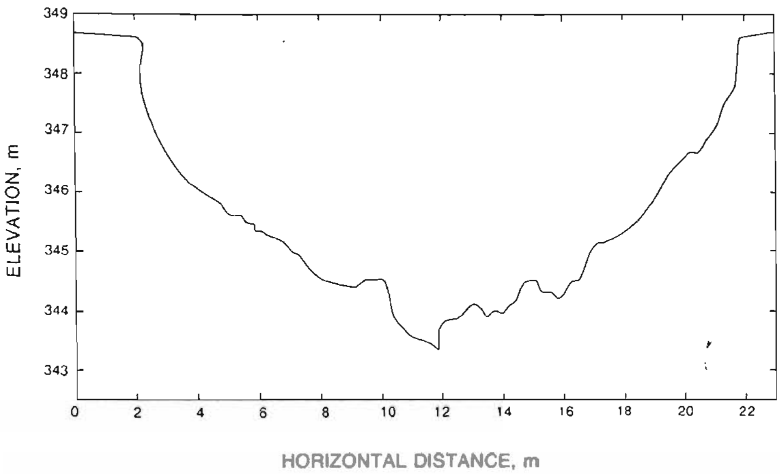


FIGURE 6 *Cross-profile of gully 14 m downstream from headcut at watershed 1, October 1973.*

channel from various weathering processes (Piest et al., 1975a). The 1967 gully erosion at watersheds 1 and 2 was 31 and 41%, respectively, of total 13-year gully erosion. At watershed 2, the single event of June 20, 1967, caused 21% of the 13-year gully erosion, principally by removing 1200 kg/m of soil, from each bank in the 213-m long reach of gully between upstream headcut and downstream measuring weir.

The gullies within watersheds 3 and 4 were actively eroding before the establishment of conservation practices. By 1964, these gullies had advanced about 210 to 250 m upslope. Before 1964, the four watersheds differed little in gully growth rate. The causes for present-day equilibrium in the conservation watersheds will be discussed in light of threshold processes.

INSTABILITY PROCESSES OF MASS WASTING

The dominant mode of gully growth within the thick loessial soils study area is by mass wasting of the walls, with most of the soil loss coming from the vicinity of the retreating headscarp. Little soil is eroded from the standing banks by tractive forces of flowing water.

No single mechanism can be used to describe the mass wasting process. Stability or instability is essentially a matter of whether or not driving forces exceed resistances and depend upon such factors as the amount of debris at the slope base, the slope angle, and moisture and soil conditions. The material begins to fail only when the forces involved become greater than the resistances. In mathematical terms, this can be expressed as

$$F_s = \frac{\tau_f}{\tau} \tag{1}$$

where F_s = factor of safety
τ_f = shear strength along some shear surface
τ = equilibrium shear stress along the same shear surface

A value of $F_s > 1$ implies stability, and a $F_s \leqslant 1$ implies instability. $F_s = 1$ is the threshold value at which the material is in a state of incipient movement.

Driving Forces

Changes in the driving forces require energy, and this energy is either associated with climate or gravity. Climate adds energy to the system in many ways through moisture and temperature changes. However, since gully wall slumping is mainly due to energy associated with water, other energy sources will not be discussed. Slope failure due to the addition of water can be attributed to (1) the weight of water added to the soil mass by surface infiltration and/or by a rise in the water table, and/or (2) increased seepage forces as water in the saturated zone exits through the bank as water table discharge or as drawdown of bank storage after passage of a flood wave through the gully, and/or (3) reduction in apparent cohesion of unsaturated soil, through reduction in capillary tension (or negative pore pressure).

Higher water table levels at the headcut are responsible, in part, for the geometry of the massive slumping near the headcut. The water table level along the gully wall, thus the seepage force at the base of the gully wall, decreases downstream from the headcut and causes little slumping of the walls. However, incipient bank failure is not directly related to the magnitude of the exit gradients near the headcut because headcut failure depends also on slope geometry, soil conditions, and runoff volume and duration.

Figure 7 shows the elevation of the phreatic surface in May 1977 at a cross section of (1) the nonincised upstream drainageway 3 m upstream from the gully plunge pool or headcut of watershed 1 and (2) about 1 m downstream from the vertical headcut. Seepage zones occur on the faces of the gully wall about 1.2 to 2.4 m below ground surface, depending on watershed and antecedent precipitation. These data, combined with water surface profiles in the gullies, indicate that rapid drawdown of water in the stream bed does not contribute to gully instability in the four watersheds; the height of flowing water at the gully head never reaches the level of the free water seepage surface in the gully banks. Stream stage in the gully is seldom greater than 0.6 m. For watershed 1, with a 5.5 to 6.0-m deep gully, the peak stage would normally be about 1.2 m below the seepage surface along the gully wall.

Resisting Forces

Forces that resist slope failure are due to the shear strength of the soil mass. Soil shear strength is usually expressed by Coulomb's equation:

$$\tau_f = c' + (\sigma_n - \mu) \tan \phi' \tag{2}$$

where c' = cohesion expressed in terms of effective stresses
ϕ' = friction angle expressed in terms of effective stresses
σ_n = total normal stress on the failure plane at the time of failure
μ = pore water pressure

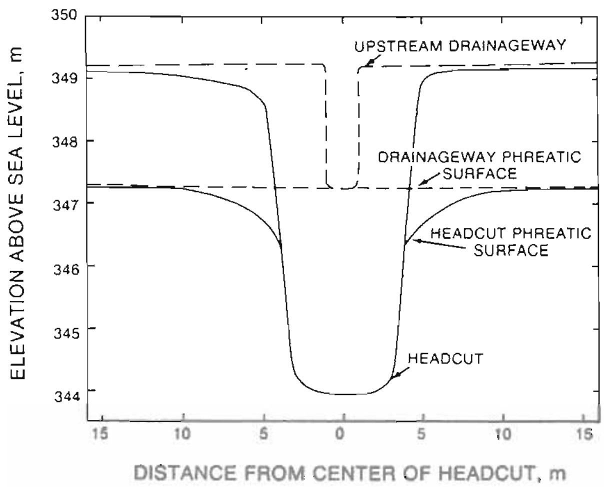


FIGURE 7 *Single-plane projection of cross sections of upstream drainageway and headcut and their phreatic surfaces.*

The valley-bottom gullies within each of the four watersheds are incised within silty alluvium. In general, the properties of this loess-derived silty alluvium are similar to the natural loess; however, specific differences and their influence on the mechanical behavior are not documented in the literature. Much of the literature deals specifically with the properties of loess.

The mechanical behavior of silty alluvium and loess depends largely on its internal structure and interparticle bonding. The structure of loess is typically a loose cluster arrangement of clay particles with a silt grain at the center (Rahman, 1973). Densities of undisturbed loess may be as low as 1.2 g/cm^3. Loess cohesion has been attributed not only to clay as a binder, but also to cementation by calcium carbonate. Strength of loessial deposits seems to be more sensitive to moisture changes than deposits of other origins; however, to our knowledge, no comparisons have been published. Turnbull (1968) presented a graph showing a radical change in shear strength with changing moisture content on undisturbed Mississippi loess soil.

The macrostructural features of loess and silty alluvium also contribute to its failure patterns. Loess cleaves on vertical planes; the cause of these planes of weakness is mere conjecture. In their review of this problem, Lohnes and Handy (1968) stated that vertical cleavage has been attributed to vertical root holes (Von Engeln, 1949), secondary carbonate filling vertical root holes (Schultz and Cleaves, 1955), or shrinkage cracks resulting from compaction (Strahler, 1963). Vertical faces in loess are quite stable under dry conditions; but when saturated, their stability is greatly decreased.

Undisturbed soil cores were taken within each DeFörest Formation unit (Fig. 4) with an 8.9-cm-diameter Shelby thin-walled sampler. The samples were taken in the drainageway at the same location (Access Tube 9) that Allen (1971) used in watershed 3, which is at the toeslope position with a 4.5% slope. Consolidated, drained direct shear tests were run on 2.00-cm-long, 6.37-cm-diameter hand-trimmed samples from each unit. The samples were saturated for 3 days before consolidation and shearing. The rate of shearing was 0.0005 cm/min. Sampling depths and shear strength parameters (eq. 2) are given in Table 3.

Table 3 *Direct Shear Strength Test Results of DeForest Formation Alluvium at Access Tube, AC 9 (Allen, 1971), Watershed 3, near Treynor, Iowa*

Stratigraphic unit	Horizon depth (cm)	Sample depth (cm)	Cohesion (g/cm^2)	Friction angle (deg)	Bulk density (g/cm^3)
Turton	0–152	80–100	138	15	1.24
Mullenix	152–274	190–210	70	20	1.34
Hatcher	274–503	400–420	124	16	1.43
Watkins	503–	740–760	77	23	1.39

Mode of Bank Failure

A full understanding of the processes of mass wasting of gully walls cannot be achieved without identifying the manner in which failure occurs. Three modes of failure have been identified in gully growth in silty alluvium in the loessial study area. The first is a deep-seated, circular arc failure. The second is slab failure. The third is a combination of base collapse or "popout," followed by slumping of overhanging material.

Deep-Seated Slides

Downstream from the near-vertical headcut, lower-angle slopes (possibly about 51° or lower, according to Lohnes and Handy, 1968) may fail along a circular arc or a logarithmic spiral (Fig. 8) under prolonged wet conditions or stream downcutting. Stability analyses of such failures can be conducted from conventional slope stability methods, such as the Swedish circle method (Fellenius, 1936), simplified Bishop method of slices (Bishop, 1955), or the Morgenstern and Price method (Morgenstern and Price, 1965). Under the simplified Bishop method of slices (Bailey and Christian, 1969) failure criteria, the slope heights necessary to cause instability (FS = 1) for 90°, 77°, and 51° slopes, are given in Table 4. For each calculation, the slope was assumed to be a uniform alluvial deposit with the shear strength parameters, c', and ϕ', as given in Table 3. Soil unit weights were calculated from the bulk densities in Table 3, assuming a soil particle density of 2.70 g/cm^3 and a degree of saturation of 100%. Pore water pressures along the slip surface were determined by assuming a flow-net system similar to that determined by Bradford and Piest (1977) for the Treynor study area. The phreatic surface in each case was taken to be 180 cm below the soil surface 15 m from the slope face. Slope angles of 77° and 51° were used due to Lohnes and Handy's (1968) discussion of slope-angle frequencies for loess in western Iowa. Their field observations had indicated that 77° slopes, however, usually fail on a plane, whereas 51° slopes can also fail along a circular arc.

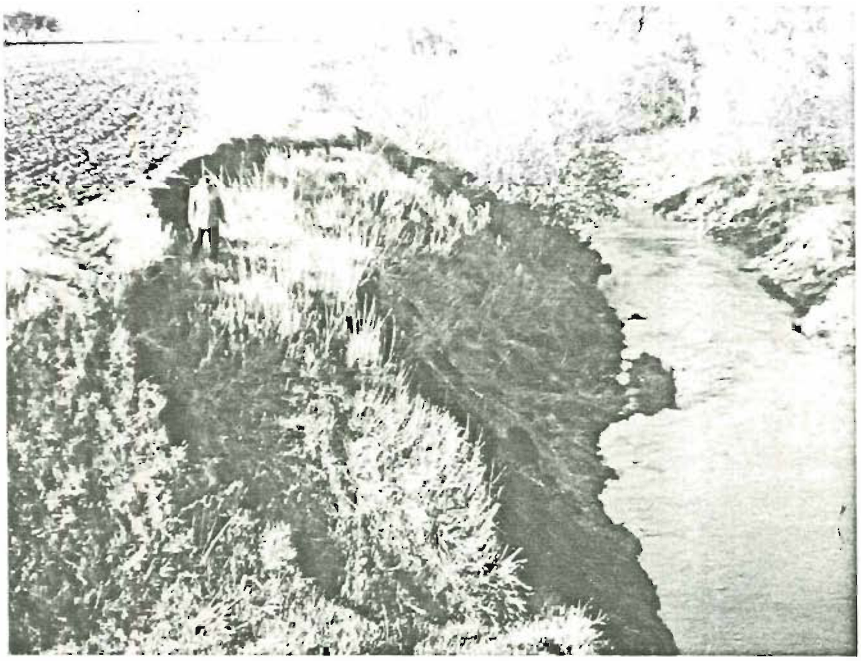


FIGURE 8 *Deep-seated circular slide along gully channel.*

TABLE 4 *Critical Slope Heights for DeForest Formation Alluvium*

Soil deposit	Slope angle (deg)	Calculated slope[a] height for FS = 1 (cm)	Maximum slope height[b] $Z = 0$ (cm)	$Z = Z_0$ (cm)
Turton	90	350	404	202
	77	386		
	51	579		
Mullenix	90	180	217	108
	77	235		
	51	350		
Hatcher	90	300	349	175
	77	342		
	51	495		
Watkins	90	210	248	124
	77	261		
	51	352		

[a]Calculated by the simplified Bishop method of slices (Bailey and Christian, 1969).
[b]Calculated from equations 3 and 4.

Slab Failure

Bradford and Piest (1977) have shown that a curved failure surface is inappropriate to describe the failure near the headcut in alluvium and that conventional slope analysis methods are not applicable. However, the failure surface can be approximated by two planes forming a trapezoidal mass of soil (Lohnes and Handy, 1968). Slab failure similar to that in a gully headcut in silty alluvium is seen at a road cut exposing Wisconsinan loess (Fig. 9). Intact slabs are seldom seen in the gully study area; upon falling, they break up because of the high moisture conditions. Vertical tension cracks form one of the planes, and the other plane is assumed to pass through the slope base at an inclination from the horizontal of $45° + \phi/2$ (Lohnes and Handy, 1968). Thus, the failure mode is similar to an unconfined compression test in that the slab fails when the weight of soil exceeds the compressive strength of the underlying soil. The vertical cracks caused by tensile stresses in the upper layers of the slope and the macrostructural features typical of loess seem to be responsible for this failure mode.

From Lohnes and Handy (1968), the maximum height H_c of the cut before slab failure is

$$H_c = \frac{4c}{\gamma[\cos\phi - 2\cos^2(45 + \phi/2)\tan\phi]} - z \tag{3}$$

where z is the depth of the tension crack and γ the unit weight of the soil mass. Maximum vertical slope heights were calculated (Table 4) for the case when $z = 0$ and

$$z = z_0 = \frac{2c}{\gamma}\tan\left[45° + \frac{\phi}{2}\right] \tag{4}$$

The slopes were assumed to be uniform, with zero tensile strength and other soil properties defined in Table 3.

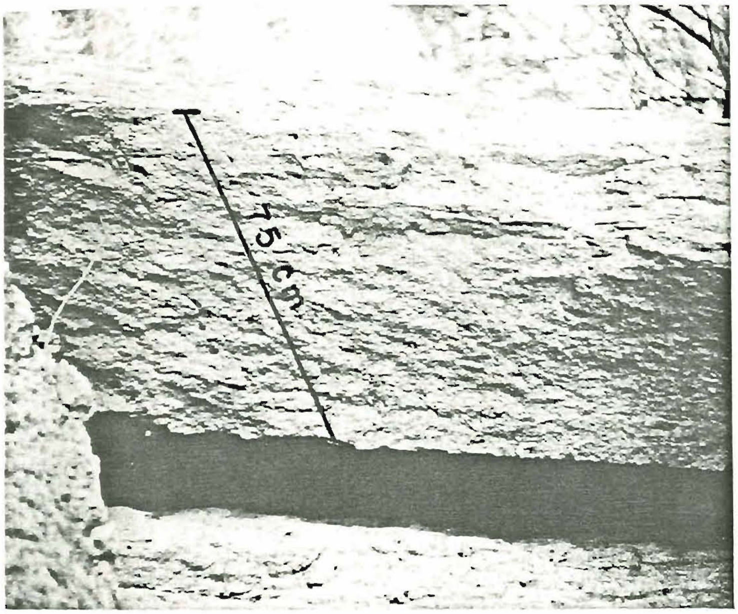


FIGURE 9 *Slab failure at roadcut near Missouri River.*

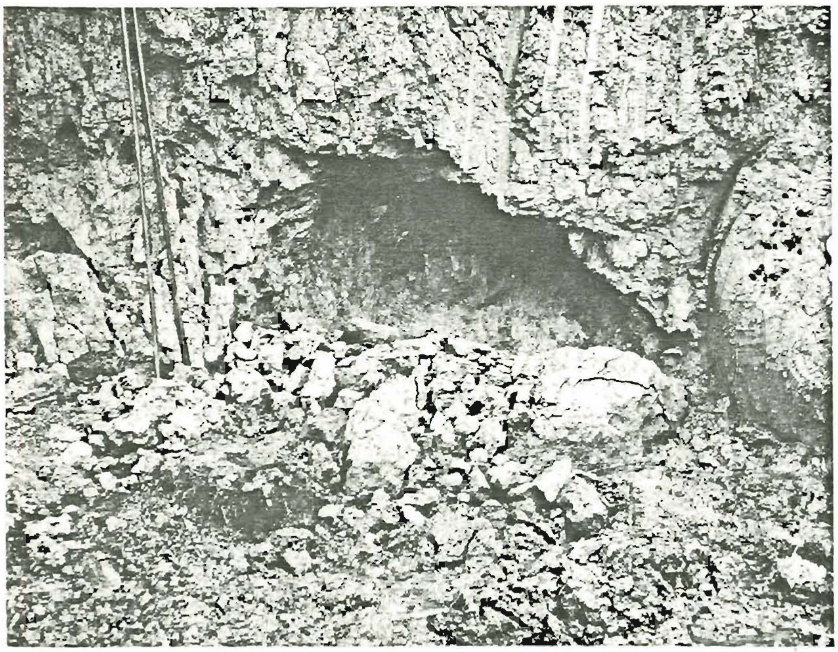


FIGURE 10 *Popout failure that develops at the base of the gully wall.*

Base Failure

Bradford et al. (1978) identified the typical failure sequence of gully headwalls in alluvium in the thick loessial area of western Iowa as (1) a popout or alcove failure near the toe of a near-vertical wall (Fig. 10), (2) columnar sloughing of the overhanging material, and, finally, (3) the transport of the eroded material downstream.

As described by Lutton (1969), both alcoves and popouts consist of pyramidal blocks freed from the wall by one fracture, inclined into the slope, and a second fracture below, inclined out of the slope. Popouts as compared with alcove failures are localized low on cut faces but above the base.

The initiating failure at the base of the wall is not thoroughly understood; however, three explanations can be offered.

1. The soil is, to a certain degree, collapsible (Handy, 1973). Handy defined collapsibility in loess as a state of underconsolidation related to apparent cohesive strengths of perennially unsaturated soils. Dudley (1970) stated that two prime requirements for collapse are a loose soil strength and a moisture content less than saturation. When water is added, the soil volume is reduced rapidly at some degree of saturation.

Reginatto and Ferrero (1973), however, distinguish between "truly collapsible" soils and "conditionally collapsible" soils. Conditionally collapsible soils are those able to support a certain level of stress upon saturation.

Collapsible loess occurs extensively in Iowa and in the study region (Handy, 1973) and collapsibility was proposed by Handy as a possible contributing factor to the development of valley-side and valley-head gullies within natural loess where no seepage water exists at the base of gully walls. Since the study area gullies are in alluvium and have a high zone of seepage above the base, collapsibility, as defined by Dudley (1970) and Handy (1973), could not be responsible for the base failure. To determine if the silty alluvium could meet the "conditionally collapsible" criterion, consolidation tests are run on undisturbed soil cores taken from each soil layer given in Table 3, and K_0 tests (Kane, 1973) were run on the most compressible alluvial member, the Mullenix unit.

K_0, the lateral stress ratio at rest, is the ratio of the total lateral stress σ_r to the total axial stress σ_n, where there has been no lateral strain. The K_0 tests were run on saturated samples in a standard triaxial cell on specimens with diameters of 7.0 cm and heights of 15 cm. The K_0 conditions were maintained by increasing the radial and axial stresses so that the volume change, as indicated by a burette, corresponded to the vertical deformation of the sample. The behavior of the soils was analyzed by the stress path method of representing the states of stress (Lambe and Whitman, 1969; Kane, 1973). In this method the state of stress is represented by a stress point on a p–q diagram, where $p = (\sigma_v + \sigma_H)/2$ and $q = (\sigma_v - \sigma_H)/2$. σ_v and σ_H represent the total vertical (normal) and horizontal (radial) stresses, respectively. An abrupt change occurred in the K_0 value upon collapse of the soil structure. For the Mullenix unit, K_0 was 0.420 before collapse and increased to 0.806 after collapse (Fig. 11).

Oedometer test results also indicated that a certain vertical pressure was required to collapse saturated samples. Little to no volume change occurred upon saturation at a vertical stress of 0.15 kg/cm^2.

The results of the K_0 and oedometer tests give further evidence that the collapse phenomenon due to wetting does not seem to be responsible for the base failure. However, a slow migration of soil particles from zones adjacent to the vertical wall and below

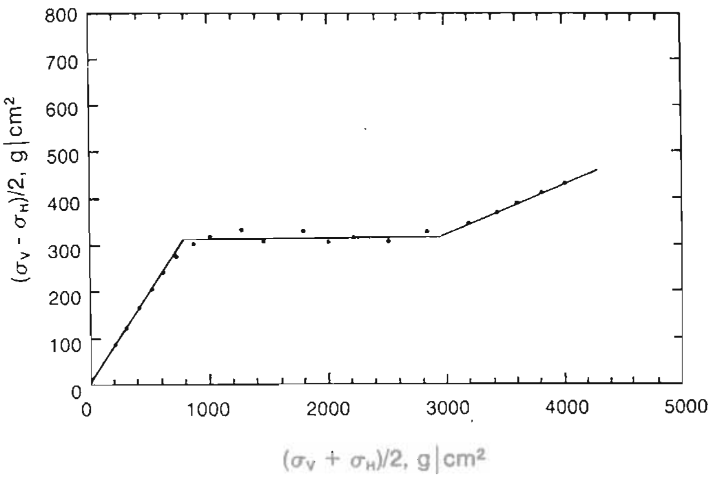


FIGURE 11 *Effective stress path for K_0 tests on the Mullenix member of the DeForest Formation Alluvium.*

the seepage level might trigger collapse. The particles would move along lines of concentration of seepage. The discussion by Brink and Kantey (1961) would support this conclusion. They stated that a marked characteristic of the collapse phenomenon is that it seems to be confined to slopes where the soils are well drained, and no recorded case has been found on flat plateaus or in depressions where the internal drainage is impeded. Visual observations of the exposed soil mass below the seepage zone and adjacent to the headcut walls indicated an orientation of drainage paths toward the walls.

2. Stress relief at the base of the slope causes bulging. The removal of debris from the slope base results in a change in the state of stress, with corresponding strains and displacements. For soils with a brittle structure in drained shear, slight deformations can result in a large decrease in the shearing resistance of the soil. This strength reduction may be due to the opening of cracks in the alluvium proceeding from the face into the wall. Since a zone of stress concentration occurs adjacent to the base of the slope, localized failure occurs there.

3. Reduced soil cohesion results from increased water content. Since undisturbed loess is a loose, open-structured soil composed of silt particles separated by clay coatings, and the silty alluvium of the DeForest Formation may be similar, as water contents increase, strength will decrease. Bradford et al. (1978) found that base failure can be initiated by a slight increase in the phreatic surface along a vertical wall of alluvium. Limit equilibrium slope stability analyses indicated that seepage forces exerted by the flowing groundwater had little effect on slope stability at the test site.

No satisfactory theoretical procedures have been developed for estimating conditions at time of base failure, because the mechanism of failure has not adequately been defined. Thus, even though mass wasting is responsible for gully growth, we are not able to predict response rates to external environmental changes because of the lack of quantitative description of processes. However, we do know that process thresholds occur in the development of gullies.

Thresholds of Instability

From the calculated maximum slope heights in Table 4, thresholds of instability can be defined within the processes of gully mass wasting. As a valley-floor gully advances up-valley, the scarp cuts into different materials, according to the geomorphology and stratigraphy of the drainage basin, and may gradually increase in height due to surface-erosion deposition. The additional gravity force caused by increased height and the differences in strength of the incised soil layers result in alteration of the stability of the gully scarp. For example, if we visualize the gully advance as shown in Figure 12, as the

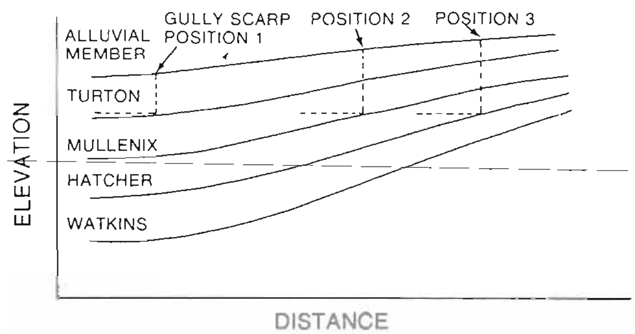


FIGURE 12 *Schematic representation of valley-floor gully advance.*

gully incises into the different alluvial soil units, the stability of the entire gully system is abruptly changed. Using the calculations in Table 4 and assuming the mode of failure to be slab-type, a vertical slope of Turton alluvium could reach a height of 404 cm before becoming unstable (position 1, Fig. 12). However, as the gully advances to position 2 and cuts into the Mullenix member, instability results because the Mullenix member can withstand only a 217-cm vertical wall. As the gully head moves from position 2 to position 3, stability is increased. Further gully growth up the drainageway into the Watkins unit again decreases stability.

This concept of abrupt changes in gully wall instability due to strength differences in the stratigraphic layers can be readily seen in the channel failures in the easily identifiable sandy alluvium in Mississippi (Grissinger, personal communication, 1977) and in gully failures in thick zones of sandy glacial outwash below loess in central Missouri. The gullies or channels seem to erode very little for many years; and then, in a matter of a few days, gullying can create huge voids in the land surface. This accelerated erosion is most often associated with eroding through to a weak layer and the alteration of the resistance forces of the system.

DEBRIS TRANSPORT

As previously stated, the production of soil debris by gully bank and head-scarp failure is a necessary, but not sufficient, condition for gully growth in this loessial region. In order to maintain gully erosion, runoff energies are required to entrain and transport soil debris. That is,

$$G = f(S, Q) \tag{5}$$

where G = gully growth rate
S = debris supply function
Q = runoff variable

Figure 13 (Piest et al., 1975a) shows gully sediment transport rates for May 25, 1972, at watershed 1, based on 30 streamflow samples. The supply of soil debris in the gully was exhausted (during a period near the runoff peak), and the transport of gully materials was essentially zero. Temporary gully cleanout for this storm occurred on the hydrograph

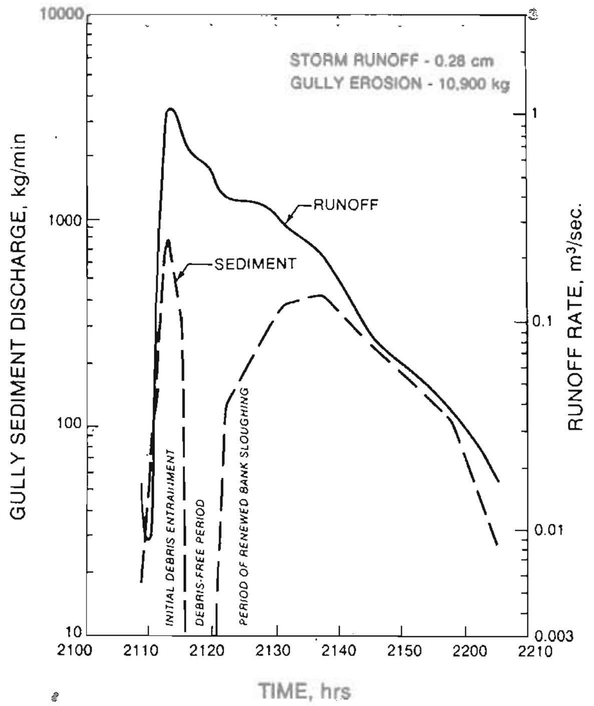


FIGURE 13 *Gully erosion rate, storm of May 25, 1965, at watershed 1.*

recession prior to a runoff rate of about 0.84 m^3/sec. Piest et al. (1975a, 1975b) reported the same processes in several other storms; that is, gully sediment discharge was maximum soon after runoff began but declined sharply during subsequent periods of high runoff. The large reduction in transport rate of eroded soil from the gully verifies that the tractive forces of runoff along the channel boundary do not play a major role in gully erosion but that soil debris from gully wall failure is the primary material removed from gullies of loessial regions.

If tractive force on the channel boundary had been the predominant eroding agent, either by direct shear on the channel boundary or by bank undercutting, a discharge of

The idea of critical limits, boundary conditions,

gully sediments would have been proportional to the stream velocity, and transport of gully sediments at significant runoff rates would not have been minimal. Piest et al. (1975a) measured runoff velocities exceeding 3m/sec in drainageways that showed little deterioration. If we assume that the soil debris accumulated in gullies is essentially co-

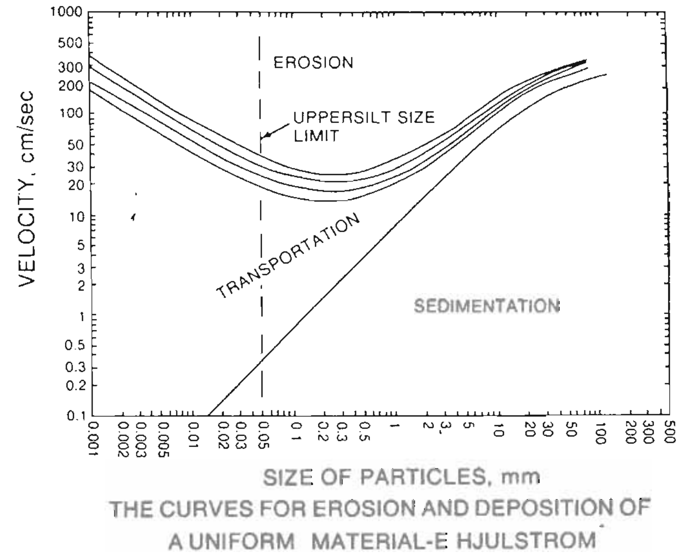


FIGURE 14 *Hjulstrom's (1935) critical drag curves, separating the flow regions where entrainment, deposition, and steady carriage occur.*

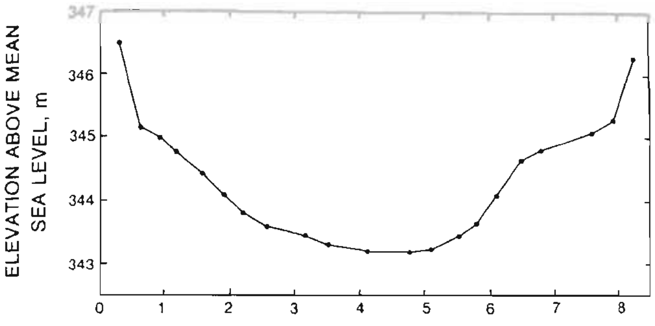
hesionless, Figure 14 (Hjulstrom, 1935) shows that nearly all discrete silt particles can be entrained and transported at velocities exceeding 0.61 cm/sec; indeed, most silt aggregates are transportable by the velocities commonly encountered.

Once the intermittent nature of debris supply and the transport effectiveness of runoff are known, a predictive model can be developed. Although the intricate input relations to equation 5 are poorly defined at present, the relative import can be postulated. For example, moderate storm runoff velocities are commonly found whenever flow depths exceed some minimum value (Fig. 15). The Manning formula was used to compute hydraulic radius and flow velocity for a typical gully cross section (Fig. 15):

$$V = \frac{K}{n} S^{1/2} R^{2/3} \tag{6}$$

where K = constant ($K = 1$, in SI units)
V = mean channel velocity, m/sec
n = roughness factor
S = hydraulic gradient
R = hydraulic radius, m (stream cross-sectional area/wetted perimeter)

Figure 15 summarizes hydraulic geometry and flow characteristics for several depths of flow and two approach slopes. Typically, the hydraulic radius, and therefore the



Flow depth (m)	Stream *cross-sectional* area (m^2)	*Wetted* perimeter (m)	*Hydraulic* radius (m)	n	Vs_1[1] (m/sec)	Vs_2[2] (m/sec)
0.30	0.16	2.68	0.02	0.1	0.48	0.59
0.61	1.52	3.96	0.38	0.1	0.75	0.91
0.91	2.71	4.69	0.58	0.1	0.98	1.20
1.22	4.04	5.55	0.73	0.1	1.14	1.40
1.52	5.53	6.40	0.86	0.1	1.28	1.57
1.83	7.21	7.38	0.98	0.1	1.39	1.70

[1]s_1 = a hydraulic gradient of 0.02
[2]s_2 = a hydraulic gradient of 0.03

FIGURE 15 *Hydraulic geometry—discharge relationships for typical cross section (top figure) for watershed 3 near Treynor, Iowa.*

velocity, increase rapidly with depth. Although it would be difficult to select any threshold runoff or velocity that would initiate erosion, Figure 16 attempts to define a threshold runoff volume that will sustain gully erosion. Cumulative storm runoff volume for

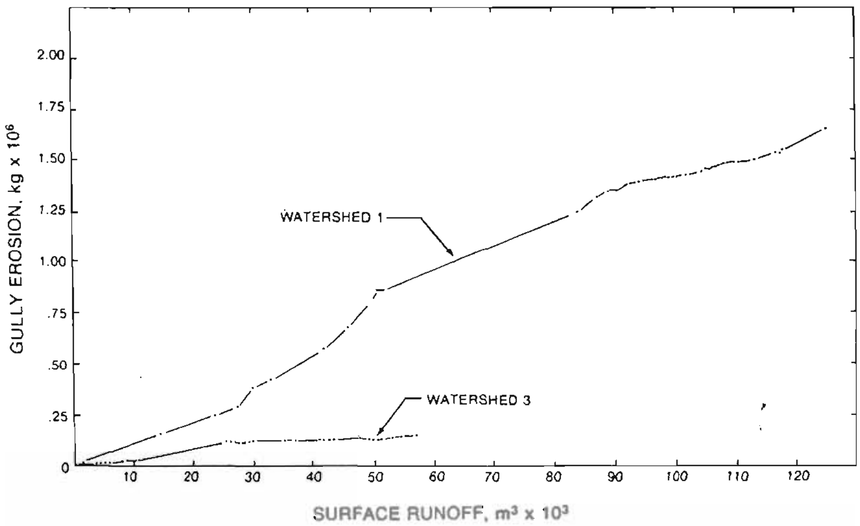


FIGURE 16 *Cumulative gully erosion and surface runoff at watersheds 1 and 3, Treynor, Iowa, for 1967–70.*

The idea of critical limits, boundary conditions,

1967–70 is plotted versus cumulative gully erosion for 59 storm events at watershed 1 and 44 events at watershed 3. Since watershed 3 was in a conservation management (grass), runoff for comparable storms was less than at rowcropped watershed 1, and 15 of the 44 rainfall events caused insignificant runoff.

If we assume that watershed 1/watershed 3 are representative nonconservation/conservation watersheds that drain through raw and active/healed and vegetated gully systems, then several conclusions can be made from Figure 16. Nearly any volume of runoff is adequate for removing gully debris from watershed 1, whereas the only appreciable gully erosion at watershed 3 required a storm runoff volume of 15,000 m^3 with a peak discharge of 6.1 m^3/sec. We can conclude that this large runoff exceeded the threshold necessary to initiate erosion in a vegetated gully.

Figure 17 describes several geomorphic stages or thresholds that pertain to a raw and active gully. At stage I, runoff is insufficient ($<Q_1$) to erode a gully, regardless of the supply of loose soil debris that is available for transport. Gully erosion under stage II circumstances is dependent upon debris supply and independent of runoff. Although the supply has been shown to vary within and between storm runoff events, the long-term supply can be considered essentially constant. Therefore, the erosion rate for a particular gully would define a horizontal line somewhere along plane II. Runoff values greater than Q_2 represent the class of catastrophic landform changes that cause an unlimited supply of

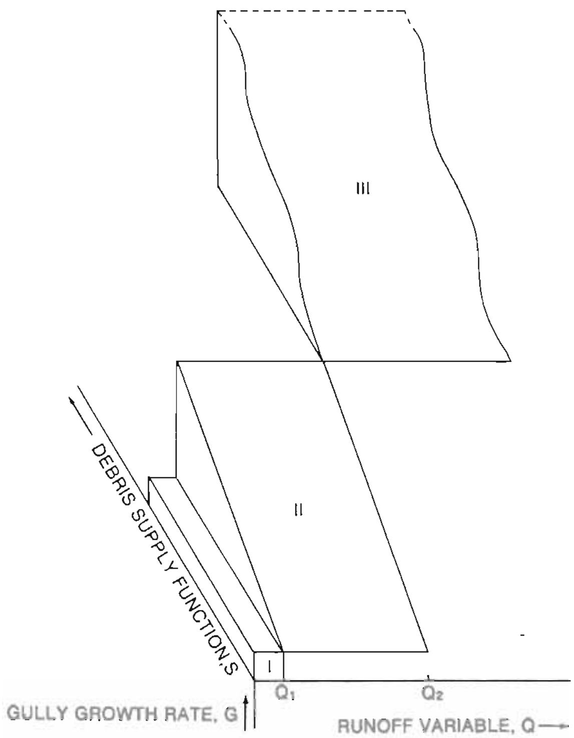


FIGURE 17 *Schematic representation of thresholds related to surface runoff and gully growth.*

debris to become available. This may have been caused by a "change of state," like the "washouts" near Fargo, North Dakota, in 1975 (Spomer et al., 1978) or by debris flows (Brown and Ritter, 1971).

PERSPECTIVES ON MODERN GULLYING AND GULLY CONTROL

Let us examine causes for the apparent rejuvenation of drainage systems, including gullying, in the context of Schumm's (1973) theories of geomorphic thresholds and the complex response of drainage systems. He cautioned against interpreting dynamic geomorphic phenomena on the basis of external causative variables when, in fact, some intrinsic erosion thresholds were exceeded. However, as Schumm implied, recognizing these erosion thresholds for areas of extensive agricultural activity can be extremely difficult.

The complexity of the relationship among climate, drainage area, and geomorphic characteristics may also be responsible for lack of establishing thresholds inherent to the landscape. We cannot be certain that the present cycle of gully erosion is due to land-use changes. Ruhe et al. (1967) recognized cycles of cut-and-fill (due to shifts in climate) in western Iowa during geologic time. We are uncertain about the influence of minor climatic shifts on gully formation (in the postsettlement period). A sensitive balance exists between the moisture regime (or hydrology) of the watershed and land surface instability. Tillage and cropping systems complicate the response of the valley floor to the processes causing gully erosion.

In this chapter we have attempted to relate gully erosion in agricultural watersheds to thresholds of mass wasting and debris transport. In the strictest interpretation of a geomorphic threshold, neither threshold is geomorphic. Neither threshold, using the words of Schumm, is inherent in the manner of landform change. That is not to say that true geomorphic thresholds are nonexistent; limiting the study to four watersheds would have, most likely, prevented such a finding.

The thresholds encountered in the processes of mass wasting of the gully walls are related to the stratigraphy of the land, and only indirectly related to the shape of the land surface. The shear strength samples represent only a limited location; however, the test results quantify what has been previously observed—that some soil layers are weaker than others. Not only is the strength per se of each successive stratigraphic layer critical to erosional processes, but also the relative permeabilities of each layer. Potentially unstable areas of the drainage basin can result from the concentration of flow lines, resulting in greater decreases in soil strength because of higher pore pressure gradients. Thresholds of the gully erosion processes are then reached.

The thresholds encountered in the debris transport process are, directly related to climate and vegetation. Prolonged wet periods increase the likelihood of gully bank failure, but only when runoff exceeds some critical value does the rate of gully development rapidly accelerate. This can be seen in the data of Table 2 and Figure 16. The years with above-normal rainfall and periods of high intensity result in increased sediment loss from gullies. Land use and vegetation overshadow climate's effect on present-day gullying. Referring again to Table 2, we see that mulch tillage or level terraces greatly reduce gully growth. This is due, in part, to the prevention of debris cleanout because of not exceeding some critical value of runoff. It is also due to the action of the water in the plunge pool

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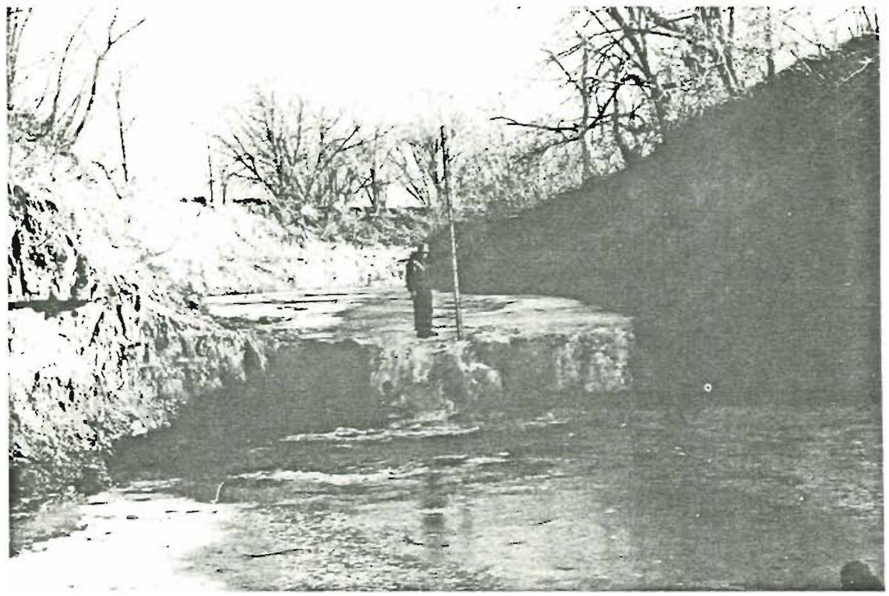

FIGURE 18 *Scarp 2 m high in channel reach of Keg Creek near Underwood, Iowa.*

area of the gully headscarp. Increased flow volumes and velocities over the headcut supply energy to the plunge pool region of the gully. This causes local scour in this part of the gully; slope height and steepness in the vicinity of the wall are increased. A reduction in energy at the headcut of watersheds 3 and 4 is the probable cause of the decrease in gully erosion since 1964, when the conservation practices were implemented.

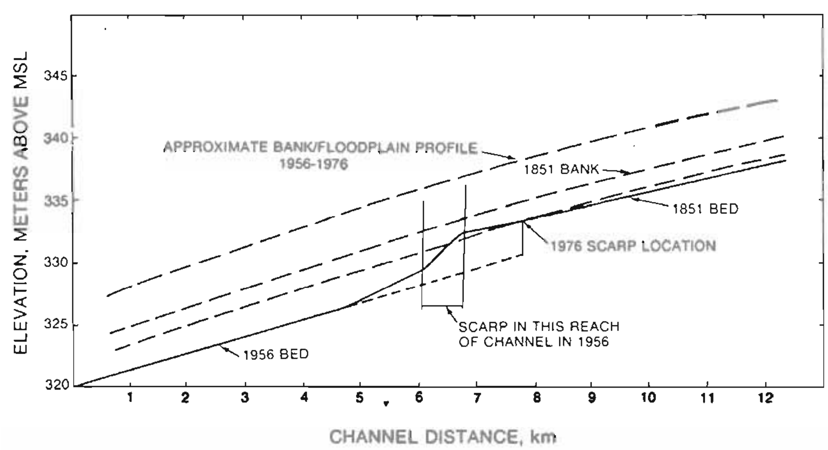

FIGURE 19 *Profile changes from 1851 to 1976 in an unstraightened channel reach of Keg Creek near Underwood, Iowa.*

Gullies that drain small upland fields are affected somewhat by the general degradation of downstream drainageways (valley trenches) to which they are tributary. Historic records throughout the loess hills region show a trend of base lowering that began after European settlement in mid-nineteenth century and has accelerated in the past half century. Piest et al. (1976) traced the development of the Tarkio River system, where some of the channel has degraded more than 6 m since dredging and straightening about 1920, with commensurate widening. Causes include changed base levels, coupled with channel alteration and drastically increased surface runoff brought about by agricultural land use.

Keg Creek, to which Treynor watersheds 1 and 2 are tributary, is typical of degrading streams that are causing upland dissection. Figure 18 shows a 2-m-high scarp migrating upstream in an unstraightened channel reach of Keg Creek near Underwood, Iowa. Figure 19 approximates the 1852 and the existing channel profile through the reach, as derived from the 1852 original land survey, the 1956 U.S. Geological Survey 1:24,000, 7 ½-min topographic map, and current measurements.

There has been an interaction of extrinsic/intrinsic stresses to accelerate gullying beyond any natural geomorphic cycle. Exceeded extrinsic thresholds result from anthropic land-use changes (that increased surface runoff) and overt channel alteration downstream. Intrinsic thresholds result from exceeding a particular soil or geomorphic parameter; this changes slopes and base levels of drainageways.

It is not possible to quantify the complex response of these drainage systems according to individual stimulus. We know, for example, that the higher flood runoff regime, generated by land-use changes, would alone have been sufficient to have altered (straightened) meanders and enlarged channels. Whatever the causes of deteriorating drainageways, however, the geomorphologist, engineer, and environmentalist should have one common goal—to understand the complex variables that are operating so that we can best manage this resource for the common good.

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