Landslides in the Vicinity of the Fort Randall Reservoir, South Dakota

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By CHRISTOPHER F. ERSKINE

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A study of some fundamental causes of landslides in the Pierre Shale along the Missouri River trench

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LANDSLIDES IN THE VICINITY OF THE FORT RANDALL RESERVOIR, SOUTH DAKOTA

By Christopher F. Erskine

ABSTRACT

The Fort Randall Reservoir, which first held water in 1952, is one of the several multipurpose reservoirs created by the U.S. Army Corps of Engineers along the Missouri River. It is in south-central South Dakota and extends from near the South Dakota-Nebraska boundary upstream about 130 miles to the northwest. The walls of the Missouri River valley and its tributaries consist predominantly of Upper Cretaceous Pierre Shale, a bentonitic claystone and shale that is subject to extensive landsliding. Alluvium underlies terraces and comprises the floodplain deposits in the Missouri River trench.

The impounded water of the reservoir had little effect on slope stability during the period of study. Emphasis was placed on investigating the nature and causes of landslides along the Missouri River trench above and below Fort Randall Dam.

The landslides along the trench are the following types: rockfalls, soilfalls, bedrock slumps, soil slumps, slow earthflows, mudflows, and slump-earthflows. The most common of these are slump-earthflows, earthflows, soilfalls (along the reservoir shore), and slumps, in that order.

All landslides are activated primarily by erosion and ground water, which may be considered as general causes or as trigger actions. General causes are long-term processes that decrease overall stability of a slope and increase its potential for landsliding. Trigger actions are rapid, often recurring processes that set off individual landslides in places where general causes have already reduced stability of the slopes. The main general cause of landslides is erosion. Ground water can act both as general cause and trigger action where oversteep, potentially unstable slopes exist. It can affect slope stability by increase of weight of slope material in a potential landslide, lubrication, hydrostatic pressure, and piping.

Analyses of the Pierre Shale indicate that it is mostly isolated silt-size grains in a matrix of clay-size particles that are predominantly montmorillonite. The silt grains do not interlock; hence, the strength of the shale depends almost entirely upon the montmorillonitic matrix.

Appraisal of slope stability in five representative areas, each including about 25 square miles, suggests that the Missouri River trench walls have been relatively stable during historic time. Most of the active and recently active landslides are related to erosion by tributary streams, whereas many of the old partially obscured landslide remnants, including most of the larger slides, shown an apparent topographic relation to the main river valley. Landslide activity probably was last at a maximum during the latest glacial advance.

A statistical comparison, made for individual members of the Pierre Shale, indicates a relation between montmorillonite in the clay-size component and the average slope angles and also a possible relation between slope angles and the relative amount of landslide terrain. There is no statistical relation, however, between montmorillonite and the relative amount of landslide terrain. The conclusions are that all members of the Pierre Shale along the Missouri River slide as a part of the normal erosion process, and the angle at which individual members develop a stable slope is a function of the montmorillonite content in their clay-size portions.

The apparent relation between landslide activity and ground-water conditions was explored in a ground-water observation program from 1954 through 1959. Twenty porous tube piezometers were installed at eight sites near the reservoir. Four of the sites are in bedrock and four in alluvium. The observations indicate that ground water in the shale bedrock moves primarily through fractures, and the direction and rate of ground-water movement depends on the orientation and size of the fractures.

Movements of control points established on five landslides—one earthflow, one slump, and three slump-earthflows—were measured over a 3-year period. The resulting data indicate that these three types of landslides have somewhat characteristic activity patterns. Earthflows have one short period of activity of a few days or weeks, after which they become stable. The movement of slumps is small relative to their size. Most of the movement probably occurs at one time, although minor movements as the slump approaches equilibrium may continue for several years. Slump-earthflows generally have a longer active life than either earthflows or slumps, the active life of a slump-earthflow being proportional to its size.

Possible correlations exist between (1) surplus precipitation available for ground-water storage, (2) landslide activity, and (3) ground-water conditions. Theoretical maximum potential evapotranspiration values were calculated from climatic data. The difference between actual precipitation and potential evapotranspiration is the precipitation-surplus or deficiency available to supplement ground-water supplies. The available information shows a distinct correlation between surplus precipitation and landslide activity. Long-term surplus precipitation values represented by a 12-month cumulative-precipitation-surplus curve were at a maximum in 1952 when landslide activity was at a maximum; long-term ground-water influxes, therefore, seem to be a general cause of unstable slopes. There are usually precipitation-surpluses during the winter and early spring, and most landslide activity occurs in the spring; short-term ground-water influxes thus seem to trigger landslides. The ground-water data from the piezometers are too limited to either prove or disprove that the precipitation-surplus actually affects ground-water conditions. The available data, however, are fully compatible with the precipitation-surplus–landslide-

activity relation. It seems likely that times of maximum landslide activity occur when infiltration of surplus precipitation appreciably augments the ground-water supplies. It may be possible to predict periods of appreciable landslide activity by appraising 12-month cumulative-precipitation-surplus curves.

INTRODUCTION

This report summarizes the results of landslide studies near the Fort Randall Reservoir in South Dakota (pl. 1). The reservoir occupies the Missouri River valley (Missouri River trench) and is formed behind the Fort Randall Dam, located 5 miles above the point where the Missouri River crosses the Nebraska-South Dakota State line. The reservoir extends upstream about 130 miles northwesterly nearly to Big Bend, a large meander curve in the river about 10 miles above Fort Thompson.

The walls of the trench in the segment containing the Fort Randall Reservoir are composed primarily of Pierre Shale, which here is virtually a bentonitic mudstone. Pierre Shale is very susceptible to landsliding when it is saturated or nearly saturated; ground-water conditions, therefore, have a close relation to slope stability in the shale. Filling of the reservoir (dam closure in 1952) has raised the free-water level in the trench and is undoubtedly changing the ground-water regimen in the area. The environs of the reservoir (pl. 1) are well suited to the study of the effect of changing ground-water conditions on slope stability.

PURPOSE AND SCOPE

This study was undertaken to learn more about landsliding, particularly reasons for the failure of slopes in bentonitic mudstone. The goal is not only to provide general qualitative information but also to point out specific problems for more detailed research. Factors affecting slope stability were studied throughout a large area, therefore, rather than exhaustive studies made of particular landslides.

The scope is restricted to three specific problems of landsliding. The first problem is to determine the relation between landslides and ground-water conditions. Because the Pierre Shale is relatively impermeable, its ground-water conditions will take many years to reach equilibrium with the free-water level in the reservoir. Little can now be said, therefore, about the effect of the reservoir on landsliding except where wave erosion and saturation are causing landslides along the reservoir shore. In this report emphasis is placed on ground-water effects independent of the reservoir and that result from seasonal precipitation variations as well as short- and long-range climatic changes. The second problem is to identify the factors that control times and rates of movement of individual and collective landslides. The third problem is to determine whether there is a relation between composition of the members of the Pierre Shale and their susceptibility to landsliding. The effects on slope stability of major compositional variations, such as quartzite versus clay shale, are obvious, but little is known about minor compositional differences between individual members of the Pierre Shale.

The landslide investigation consisted of analysis of the Pierre Shale, appraisal of selected study areas, ground-water investigations, and measurement of landslide movements; based on these studies, correlation was suggested between surplus precipitation available for ground-water storage, landslide activity, and ground-water conditions.

FIELDWORK

Most of the fieldwork was done during the summers of 1952–55 inclusive. During 1956 about 1 month was spent field checking and measuring landslide movements. U.S. Geological Survey personnel at Huron, S. Dak., made ground-water measurements during the winters of 1954 (Jan.–March) and 1954–55. They assumed full responsibility for the ground-water program in the fall of 1955 and continued to make piezometer measurements intermittently through 1959.

In the field, data were plotted on maps and on aerial photographs, and later they were transferred to base maps. The landslide movements were measured by periodically relocating control points on the landslides with a transit either by triangulation or by closed traverse. Porous tube piezometers were used to observe ground-water levels. Subsurface samples for analyses of the Pierre Shale were obtained with a power auger.

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The late Mrs. Helen Varnes accepted the author's duties when he resigned from the U.S. Geological Survey in 1957. Without her conscientious revision of the original draft and later work on the manuscript, the report could never have been completed. D. J. Varnes, C. G. Johnson, and R. D. Miller made significant contributions in further reviewing and revising the manuscript for publication.

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The Omaha District, U.S. Army Corps of Engineers, provided the project with maps, services, space for field offices, and storage facilities, and their personnel cooperated with project personnel in every way possible. A. H. Burling, J. A. Trantina, and L. B. Underwood, especially, gave the author many good suggestions at numerous discussions and field conferences.

Many other persons and agencies contributed to the project. Among them were the South Dakota State Geological Survey and the South Dakota State Highway Commission. The city officials of Chamberlain, S. Dak., allowed piezometers to be installed on city land. Landowners near Fort Randall Reservoir permitted landslide investigations on their land.

GEOGRAPHIC SETTING

CULTURE

The land around Fort Randall Reservoir is used entirely for farming and ranching. Generally, crops are raised on the relatively flat uplands, terraces, and flood plains, and the valley walls are used for grazing cattle and sheep.

The only permanent towns near the reservoir and their 1960 populations are Chamberlain, 2,598; Fort Thompson, 150; Oacoma, 312; and Pickstown, about 600. Pickstown, located at the Fort Randall damsite, was built by the U.S. Army Corps of Engineers.

The roads are mostly in the uplands somewhat remote from the reservoir. U.S. Highway 16 crosses the reservoir at Chamberlain, and U.S. Highway 18 crosses the Missouri River at the Fort Randall Dam. The uplands have a good network of State and county roads. There are few roads along the valley walls of the Missouri River and its tributaries; in most places the reservoir shore can be reached only on foot or by four-wheel-drive vehicles.

The Chicago, Milwaukee, St. Paul and Pacific Railroad line from Mitchell, S. Dak., to Rapid City, S. Dak., crosses the reservoir at Chamberlain. A spur track built by the Corps of Engineers to carry construction materials to the Fort Randall Dam runs from the Chicago, Milwaukee, St. Paul and Pacific Railroad line at Lake Andes to the damsite 7 miles to the south.

PHYSIOGRAPHY

The Fort Randall Reservoir lies entirely within the Missouri River trench (fig. 1), one of the 12 physical divisions of South Dakota of Flint (1955), but locally the landslide areas extend eastward into the Coteau du Missouri and westward into the Pierre hills.

Missouri River trench.—The Missouri River valley constitutes the Missouri River trench physical division (fig. 1). It is a broad southeast-trending trough, 300–650 feet deep, cut into the eastward-sloping Missouri Plateau. Between the Nebraska border and Big Bend (pl. 1), the trench is 1½–5 miles wide, and the floor is ¾–1½ miles wide. The walls of the trench, cut in easily eroded Pierre Shale, have a rugged "badlands" topography—alternating steep-sided gullies and spurs. Terrace remnants are common. Where large segments of old terraces remain, they protect the trench walls above from direct river erosion. As a result, in these places the walls have not been greatly affected by post-terrace valley cutting, and the topography is more subdued than in unprotected parts of the trench.

Pierre hills.—The Pierre hills physical division is characterized by mature topography of low rolling hills and a well-integrated drainage system. Downcutting by the Missouri River has rejuvenated the drainage system, and steep-walled youthful valleys now are encroaching on the rolling uplands.

Coteau du Missouri.—The Coteau du Missouri is a glaciated eastward extension of the Pierre hills. Most of the gently rolling surface is mantled by glacial drift that masks any preglacial drainage pattern. Compara-

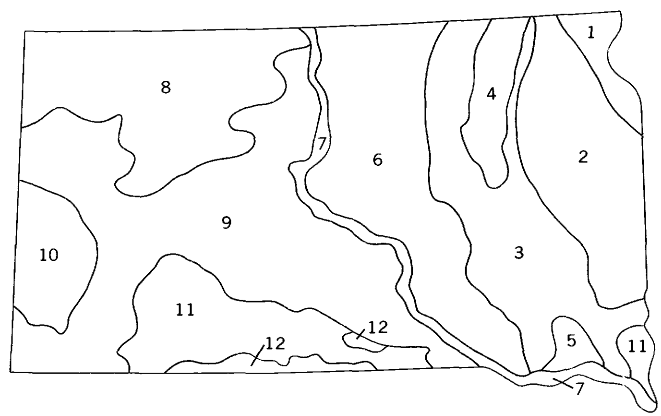


FIGURE 1.—Physical divisions of South Dakota. 1, Minnesota River-Red River lowland; 2, Coteau des Prairies; 3, James River lowland; 4, Lake Dakota plain; 5, James River highlands; 6, Coteau du Missouri; 7, Missouri River trench; 8, northern plateaus; 9, Pierre hills; 10, Black Hills; 11, southern plateaus; 12, Sand Hills. From (Flint, 1955, fig. 1.)

tively few youthful streams extend far beyond the limits of the Missouri River trench into the Coteau; consequently much of the area drains into intermittent ponds in numerous closed depressions.

CLIMATE

South Dakota has a dry subhumid climate with hot summers and cold winters, U.S. Department of Agriculture (1941) and Visher (1954). Within the reservoir area, precipitation and temperature decrease markedly northwestward. Average annual precipitation is about 23 inches at the dam and about 18 inches at Fort Thompson. More than 75 percent of the precipitation falls from early April through September, and 40–50 percent of the total precipitation falls in May, June, and July. Average January temperatures range from 20° F at Fort Randall Dam to 16° F at Fort Thompson; average July temperatures are about 75° F for the entire area. A temperature range of 130° F, from −25° F minimum to 105° F maximum, can be expected in a normal year. Winds, predominantly northwest in winter and southeast in summer, have an average velocity of 10–12 miles per hour.

VEGETATION

The natural plants of the uplands are drought-resistant grasses. Shrubs and trees grow in the valleys and gullies. Many good stands of trees, predominantly cottonwoods, are along the river bottomland not flooded by water in the reservoir. Trees growing away from permanent streams become increasingly scarce toward the northwest end of the reservoir area, reflecting the regional decrease in precipitation. Cultivated are corn, the main crop in the southeast end of the reservoir area, wheat, the main crop in the northwest end, and some oats, soybeans, and sorghum.

Vegetation commonly is lusher on slopes where landslides have occurred for the following reasons: the excess water that encourages landsliding also encourages vegetation; the closed depressions and fissures that result from landsliding trap water which would normally run off a smoother stable slope.

GEOLOGY

This report is concerned with the geologic formations of the Fort Randall Reservoir area relative to their involvement in landslides along valley walls of the Missouri River and its tributaries. The following geologic descriptions and the generalized stratigraphic section (table 1) are presented primarily as background information for discussion of the landslides. More detailed information on the regional geology is given by Petsch (1946), Flint (1955), Crandell (1958), Simpson (1960), and Schultz (1965).

UPPER CRETACEOUS SEDIMENTARY ROCKS

The Upper Cretaceous sedimentary rocks are virtually the only consolidated deposits in the area. They

TABLE 1.—*A generalized stratigraphic section of the Fort Randall Reservoir area*

Era	System	Series	Group	Unit		Thickness (ft)
Cenozoic	Quaternary			Alluvium, colluvium, and loess		Generally <10
				Loess		30±
				Alluvium	Slow draining	100±
					Fast draining	
				Till		Locally as much as 100
	Tertiary	Pliocene		Ogallala Formation		40±
Mesozoic	Cretaceous	Upper Cretaceous	Montana	Pierre Shale	Elk Butte Member	>180
					Mobridge Member	>100
					Virgin Creek Member	50–100
					Verendrye Member	90–170
					DeGrey Member	20–50
					Crow Creek Member	10±
					Gregory Member	30–90(?)
					Sharon Springs Member	35–55
			Colorado	Niobrara Formation		16–90 exposed (prereservoir)

include the most stable (Niobrara Formation) and the least stable (Pierre Shale) deposits in the area.

NIOBRARA FORMATION

The Niobrara Formation, a soft impure fossiliferous limestone, was widely exposed in the lower part of the walls of the Missouri River trench before the dam was built. Now, however, most of the formation in the downstream half of the area is permanently submerged.

The Niobrara crops out as nearly vertical bluffs bordering the river and its flood plain. Before the completion of the dam in 1952 these bluffs were from a few feet to more than 85 feet above the river level (Petsch, 1946, fig. 12). The exposed thickness above the river varied considerably, due to the uneven upper surface of the formation and also to the gentle open folding of the Cretaceous beds. Near Fort Randall Dam, the top of the Niobrara, at an altitude of 1,310 feet, was about 70 feet above river level (1,240-ft altitude). Upstream at the site of the former Wheeler Bridge, only 16 feet of Niobrara was exposed. Near Chamberlain, the Niobrara-Pierre contact is 1,420 feet in altitude, and about 90 feet of Niobrara was exposed above the Missouri River. The formation passes below river level near the lower end of Big Bend at an altitude of about 1,350 feet.

The Niobrara Formation, the oldest exposed bedrock in the area, is the uppermost formation of the Upper Cretaceous Colorado Group. The Niobrara rocks that crop out within the report area can be traced directly to the Niobrara type locality of Meek and Hayden (1861, p. 422–424) near the junction of Niobrara Creek and the Missouri River about 35 miles southeast of the Fort Randall Dam.

The Niobrara Formation is a dark-gray impure chalk containing many microscopic shells of Foraminifera and Ostracoda. Small light-colored shell fragments and clastic particles in a darker groundmass give the chalk a salt-and-pepper appearance when closely examined. Weathering changes the color from dark gray to pale orange. (All color terms conform with terminology used in the 'Rock-Color Chart" by Goddard and others (1948).)

The rock is massive and coherent, with a toughness and resilience that makes it difficult to fracture, although it is soft enough to be scratched by a fingernail. The nearly horizontal bedding is best shown by partings a few inches to 5 feet apart and by thin clay and gypsum layers along many of the bedding planes. Nearly vertical joints 1 inch to several feet apart are conspicuous in most exposures.

PIERRE SHALE

The Pierre Shale, overlying the Niobrara Formation, constitutes most of the valley walls of the Missouri River and its tributaries. The Pierre underlies a mature topography of low, gently rounded hills slashed by recently cut steep-sided gullies and valleys, and locally disrupted by landslides.

In this report the Pierre is divided into eight members following Crandell's classification (1958, p. 8–19), although individual members are not everywhere mappable, either because contacts are obscured or because the members are too similar lithologically. Because classification changes by Schultz (1965) after recorrelation of marl beds in the Gregory and Crow Creek Members were made after completion of the landslide studies, they have not been incorporated in this report.

The maximum thickness measured was 423 feet at the site of the former Wheeler Bridge 15 miles above Fort Randall Dam. Although a complete section was not compiled in the upper part of the area, the following figures for sections, including the lower six members, suggest that the whole formation thickens several hundred feet within the length of the reservoir. At Wheeler Bridge, the total thickness of the six members is 228 feet; in the vicinity of Chamberlain, it is at least 400 feet (interpreted from Petsch, 1952).

SHARON SPRINGS MEMBER

Before the filling of the reservoir the Sharon Springs Member of the Pierre Shale was exposed throughout the report area. The normal operating water level of the reservoir now covers this member south of the site of the former Wheeler Bridge. The Sharon Springs Member at Wheeler Bridge is 35 feet thick. The thickness in the upper part of the reservoir near Chamberlain is uncertain: Warren and Crandell (1952, p. 4) inferred 55 feet thickness from measurement of weathered shale chips in the section, but they saw no actual exposures of the shale more than about 15 feet thick.

The contact between the Pierre Shale and the underlying Niobrara Formation is sharp. The Sharon Springs Member commonly crops out in relatively steep slopes above the chalk, as the unweathered shale is stable and can stand in nearly vertical slopes. As weathering progresses, the shale breaks down into chips which tend to ravel down the slope face until a slope at the angle of repose, generally about 30°, is formed.

The Sharon Springs Member is olive to brownish-dark-gray bituminous shale. Fish scale fragments are widely disseminated through the member, and in many places the bituminous content is large enough to support combustion. Thin bentonite beds, generally less than 1 inch thick, are common, and secondary gypsum, either as disseminated selenite crystals or as intergrowths in the bentonite beds, is plentiful. A yellow powdery mineral, identified as a secondary hydrous sulfate of iron (Simpson, 1954, p. 63), in places coats the shale along parting and joint surfaces.

Partially weathered outcrops of the Sharon Springs Member are characterized by stacks of horizontal chips one-eighth inch or less thick and generally less than 2 inches in diameter. The rock is relatively hard (barely marked by a fingernail) and the only member of the Pierre Shale that does not readily weather to clay. Bedding is not apparent in fresh exposures, but horizontal fissility develop on weathered outcrops.

GREGORY MEMBER

The Gregory Member of the Pierre Shale underlies moderate grass-covered slopes throughout the area bordering the Fort Randall Reservoir. Locally, rapid erosion produces fairly steep slopes, although nearly all slopes more than 25 feet high are flattened by subsequent landsliding. The unit, which is separated from the underlying Sharon Springs by a sharp contact, is the lowest member of the Pierre Shale that will be completely exposed when the reservoir is at minimum pool level (1,310 ft).

Thickness of the Gregory Member increases from about 35 feet in the lower part of the reservoir (Gries, 1942, p. 31) to 50–90 feet near Chamberlain (Warren and Crandell, 1952, p. 5).

Near the dam, the Gregory corresponds lithologically to the lower beds of the type section (Crandell, 1958, p. 10) and is a dark-gray marl underlain by a basal silty layer and overlain by a dark-gray calcerous shale. In the upper reaches of the reservoir, the Gregory is a predominantly noncalcareous gray bentonitic claystone that closely resembles the upper part of the Gregory in the Pierre area (Crandell, 1958, p. 10), although a marl layer is locally present at or near the base.

The appearance of the outcrops varies with the lithology. The marl is a dense resistant rock that locally forms small ledges less than 10 feet high. The shale weathers rapidly; fresh shale alters to bentonitic clay in a few years, passing through a partially weathered stage of chips up to about 1 inch in diameter and one-eighth inch thick. The claystone weathers directly to a bentonitic clay without passing through a chip stage. Marl beds, bentonite beds, and occasional concretionary layers are the only depositional indications of bedding. Unweathered shale and claystone appear to be homogenous, but partially altered shale develops a fissility apparently parallel with the bedding.

CROW CREEK MEMBER

The Crow Creek Member consists of a blocky unbedded marl underlain by a few inches of laminated siltstone. The Crow Creek is remarkably persistent although the siltstone is only 10–15 inches thick and the marl is generally 7–10 feet thick. The marl bed has been recognized along the Missouri River trench for more than 250 miles from Yankton, 70 miles downstream from Fort Randall Dam, to the vicinity of Pierre, almost 200 miles upstream from the dam. The siltstone is present upstream from the former Wheeler Bridge site, but it has not been identified downstream.

Exposures of the Crow Creek Member become increasingly distinctive from the dam toward the upper part of the reservoir. In the lower part of the reservoir the member typically is covered by slope wash or soil and is visible only where recently eroded. In the upper part it stands out very noticeably as a light band in the otherwise dark Pierre Shale. The contact with the Gregory Member is sharp and disconformable, although no evidence of channeling has been observed.

Crandell (1952, 1958) made a detailed study of the Crow Creek Member, and the following brief lithologic description is largely a summary of his data. The unweathered marl is light gray; upon exposure it oxidizes to grayish orange. Secondary iron oxide in the siltstone commonly gives it a yellowish-brown appearance. The marl is soft and shows no indication of bedding, whereas the siltstone consistently shows bedding and breaks into thin slabs along bedding planes as it weathers. In some outcrops secondary iron oxide has cemented the siltstone to form ledges approximately 1 foot high; in other places it is cemented in varying degrees by calcium carbonate.

DEGREY MEMBER

The DeGrey Member of the Pierre Shale exposed in the valley walls bordering the reservoir is a massive dark-olive-gray bentonitic claystone containing numerous thin bentonite beds and, in the north half of the area, abundant iron-manganese concretions. A black band of residual concretions makes DeGrey outcrops a striking feature of the walls of the Missouri River trench around the upper part of the reservoir. The basal few feet of the DeGrey generally contains no concretions, and the contact between the Crow Creek and DeGrey Members is a relatively sharp transition from marl to noncalcareous claystone. The DeGrey of the Fort Randall area closely resembles the shale and bentonite facies described by Crandell (1958, p. 13–

14), but the thick basal siliceous shale facies described by Crandell apparently is not present in the report area.

The member ranges in thickness from 23 feet at the Wheeler Bridge site (Gries, 1942, p. 31) to about 50 feet near Chamberlain (Warren and Crandell, 1952, p. 7).

In a fresh cut the claystone appears well consolidated, but it weathers rapidly to gentle slopes blanketed by 4–8 inches of clay with a characteristic coarse, porous crumblike structure when dry. Wherever the more resistant bentonite beds and layers of concretions are eroded, a steplike topography develops. Continued weathering and erosion, however, break down these layers and leave the concretions scattered as a lag concentrate on the shale.

The DeGrey is very susceptible to landsliding, especially in the southern part of the area. Most exposures show at least minor displacements, most of which can be attributed to landsliding.

VERENDRYE MEMBER

The Verendrye Member is present throughout the report area. Most fresh outcrops are on active and recently active landslide blocks because exposed claystone weathers rapidly and is blanketed by residual clay soil. No complete sections were measured during the present investigation, but Petsch (1946, p. 37) showed 80 feet of thickness at the Wheeler Bridge site and about 170 feet at Crow Creek, 10 miles north of Chamberlain.

The contact between the Verendrye and the underlying DeGrey Member becomes less distinct southward along the reservoir. In the northern part of the reservoir, the contact can readily be recognized at the top of the iron-manganese concretion zone in the DeGrey. In the southern part of the reservoir, landslides and vegetation commonly obscure the contact. Moreover, the absence here of a conspicuous concretionary zone in the DeGrey results in a similar lithology that makes distinction between the two members difficult in the few good exposures available.

Lithologically, the Verendrye is a bentonitic olive-gray claystone very similar to parts of the DeGrey Member. In the Fort Randall area, the Verendrye has far fewer concretions and fewer discrete bentonite beds than the DeGrey. In the Pierre area to the northwest, however, Crandell (1958, p. 15) reported abundant concretions in the Verendrye and a gradational contact between it and the DeGrey.

The Verendrye Member breaks down to bentonitic clay after a few months of exposure. The material closely resembles the weathering products of the DeGrey except for the absence of lag concentrates. Throughout the area the Verendrye appears to be unstable and susceptible to landsliding.

VIRGIN CREEK MEMBER

The Virgin Creek Member, defined by Searight (1937, p. 35) and described by Crandell (1958, p. 15–16), is present above the Verendrye Member throughout the area. It consists of a noncalcareous bentonitic shale at the base and grades into noncalcareous bentonitic claystone at the top. In the upper half of the reservoir it is the highest member of the Pierre Shale extensively exposed in the walls of the Missouri River trench.

The Virgin Creek Member is about 50 feet thick near the dam and increases in thickness to the northwest. Poor outcrops make direct measurement difficult in the northern part of the reservoir, but comparison of the Chamberlain quadrangle geologic map (Petsch, 1952) and the topographic map suggests that the member is about 100 feet thick near Chamberlain.

Outcrops typical of the Virgin Creek Member are rare in the smooth, grass-covered slopes. Fresh outcrops are found only where stream erosion or landslides have recently exposed the member. Outcrops of the shale facies show a characteristic silvery sheen derived from the shale chips. In outcrops weathered to a bentonitic clay, it is difficult to distinguish the Virgin Creek Member from the underlying Verendrye Member, as apparently there is no sharp contact between them. The contact, which is inferred to be gradational over a vertical distance of 5–10 feet, is usually hidden by slump blocks derived from weathered Virgin Creek beds.

Both the shale and claystone are dark gray to grayish black where fresh and brownish gray where weathered. Concretionary layers and thin bentonite beds as much as 1 inch thick are common. The basal shale, with the exception of the Sharon Springs Member, is the most fissile material in the Pierre Shale. It weathers to small chips generally less than 1 inch in diameter and one-eighth-inch thick. After prolonged exposure the chips disintegrate to a bentonitic clay similar to the soils derived from the DeGrey and Verendrye Members.

The transition of shale to claystone is gradual. No obvious break separates the two facies, but the upper part of the member is less fissile and quickly weathers to bentonitic clay.

MOBRIDGE MEMBER

The Mobridge Member of the Pierre Shale consists of partly indurated marl and calcareous shale beds very similar to those described in the Pierre area by Crandell (1958, p. 16–17). In the south half of the Fort Randall area, the Mobridge crops out in the Missouri River trench walls bordering the reservoir; in the north half, it crops out only in the uplands away from the trench.

The Mobridge is about 100 feet thick near the Fort Randall Dam (Gries, 1942, p. 25). Although it thickens northwestward along the trench, no data are available on thicknesses in the northern part of the report area.

The lithology has considerable lateral variation. Along the Missouri River trench in South Dakota, the member grades from bentonitic marl (containing roughly 35 percent calcium carbonate) near Fort Randall Dam to calcareous bentonitic shale (containing less than 10 percent calcium carbonate) near the North Dakota border (Curtiss, 1950, p. 75). The color changes from medium dark gray in fresh exposures to yellowish gray and buff in completely weathered beds.

The rock appears massive and firm, although it is more friable than some shale members of the Pierre. Concretionary layers are common. As weathering progresses, partings and color bandings ½–1 inch thick develop parallel to the bedding. The contact between the calcareous Mobridge beds and the noncalcareous Virgin Creek beds is sharp, although the two members intertongue locally.

The Mobridge Member is more resistant to erosion than the underlying members. Although it typically forms grass-covered hillsides, the slopes are generally more pronounced and, in many places, the Mobridge crops out as fairly steep buff-colored slopes. The steeper slopes suggest that it is less susceptible to landsliding than are most members of the Pierre.

ELK BUTTE MEMBER

The youngest unit of the Pierre Shale is the Elk Butte Member, a locally calcareous bentonitic claystone and shale. Exposures are confined to the uplands and the upper trench walls bordering the south half of the Fort Randall Reservoir.

The original thickness of the Elk Butte Member in this area is unknown because the upper part was eroded and later overlain unconformably by upper Tertiary beds. At the Wheeler Bridge site, Gries (1942, p. 31) recorded nearly 100 feet of Elk Butte beds overlain by 80 feet of Fox Hills Formation. The Fox Hills as used by Gries is now considered to be part of the Elk Butte Member (Stevenson and Carlson, 1950), and therefore the total thickness is now estimated to be 180 feet in this location.

Exposures of the Elk Butte Member are similar to those of the underlying Mobridge, and the contact between the two members is transitional. The top of the calcareous beds is commonly considered to be the top of the Mobridge (Crandell, 1950, p. 2338; 1958, p. 18), but because some calcareous beds occur above the supposed base of the Elk Butte Member in the lower part of the reservoir, color is also used to separate the two members (Crandell, 1958, p. 18). The weathered Elk Butte material is generally darker than the Mobridge. The color ranges from olive gray where partially weathered to moderate brown where fully weathered; unweathered material is rarely seen. A distinctive feature noted by Crandell is the presence of yellowish-brown calcareous concretions throughout the Elk Butte. In the Fort Randall area concretions are commonly scattered throughout the member and are also concentrated locally in lenses and discontinuous beds.

The Elk Butte Member appears to be well consolidated. As it weathers it develops slight fissility, and partings are spaced about one-fourth inch apart parallel to the bedding. Continued weathering easily alters the material to bentonitic clay. Weathering also brings out a conspicuous color banding in outcrops of consolidated shale and claystone.

STRUCTURE

The exposed rocks of Cretaceous age are flat-lying undisturbed sediments. Some minor open folds undoubtedly exist, but they are too small to affect landsliding in the area. The Niobrara-Pierre contact has a vertical range of almost 150 feet within the length of the reservoir. These fluctuations in altitude may represent open folds.

Exposures of the Niobrara Formation and the Pierre Shale show some faulting. Most of the faults have no more than a few feet of displacement and rarely can be traced between outcrops. Subsurface damsite investigations by the Corps of Engineers have revealed intense local faulting of both formations.

Almost all the coherent outcrops show randomly oriented near-vertical joints, which presumably are present in all members of the Pierre Shale as well as in the Niobrara Formation.

TERTIARY AND QUATERNARY DEPOSITS

The Tertiary and Quaternary sediments are mostly unconsolidated terrestrial deposits. They include the Ogallala Formation of Pliocene age, glacial till, alluvium, colluvium, and loess. All these deposits form more stabe slopes than does the Pierre Shale. Only a few of the landslides along the reservoir occur wholly in the Tertiary and Quaternary deposits.

OGALLALA FORMATION

The Ogallala Formation of Pliocene age is a heterogeneous mixture of silt, sand, and fine to medium gravel locally cemented to form orthoquartzite. The Ogallala is restricted to the south half of the reservoir area. Uplands and buttes protected by resistant caps of orthoquartzite of the Ogallala rise above the general land surface.

The original thickness of the formation is unknown and the present thickness varies because of erosion. The average thickness in the area is about 40 feet. Many landslides in Pierre Shale have extended up into the Ogallala Formation, and masses of Ogallala material are found in landslide remnants on slopes below their true stratigraphic locations.

The orthoquartzite is grayish olive in fresh exposures and weathers to yellowish gray. Unconsolidated parts of the formation generally are pale shades of gray.

TILL

Unsorted glacial deposits blanket the uplands along much of the east side of the reservoir and are exposed in many places along the east wall of the Missouri trench. Till occurs chiefly in small isolated patches on the west side of the reservoir.

Because glacial deposits tend to fill in and smooth out irregularities in the preglacial topography, the thickness of these deposits varies greatly over short distances. Generally, the deposits are less than 50 feet thick although locally they may be at least twice as thick.

In the Fort Randall area till is primarily silt and clay wth some pebbles, cobbles, and boulders. Mechanical analyses of the till at Chamberlain show that about 70 percent is of silt and clay size (Warren and Crandell, 1952, p. 44). Most of the till is partially weathered to some shade of brown; fresh till is generally a dark gray.

ALLUVIUM

In this report the term "alluvium" is applied to all water-laid deposits of Quaternary age and includes glacial outwash as well as stream and flood-plain deposits. The general category has been subdivided into fast-draining alluvium and slow-draining alluvium. Fast-draining alluvium is sufficiently permeable to offer little resistance to the passage of water. Where it borders the shore of a lake or reservoir, there is no appreciable lag between fluctuations in the water level and resulting water table fluctuations within the alluvium. Slow-draining alluvium has relatively low permeability, and the movement of water is restricted. There is slower adjustment of the water table within slow-draining alluvium to changes of water level in an adjacent body of water.

Extended studies of alluvial deposits enabled the author to use field methods to set a relatively consistent boundary between fast- and slow-draining alluvium. Fast-draining alluvium was predominantly composed of particles of coarse-sand size (0.5–1.0 mm) or larger; slow-draining alluvium was predominantly composed of particles of medium-sand size (0.5–0.25 mm) or smaller.

The alluvium either covers nearly flat valley bottoms or is exposed in terraces. The valley bottom alluvium is widespread and deep at least along the Missouri River trench. Flint (1955, p. 147) gave a depth of 189 feet to the bedrock floor of the Missouri River trench at Fort Randall Dam. Because valley bottom alluvium is comparatively unimportant as a landslide medium, the author did not attempt to separate it into fast- and slow-draining groups. Most is permanently flooded by the reservoir or is too far up the tributaries to be affected. Some minor landsliding may be anticipated in the zone where periodic drawdown of the water level will expose banks of slow-draining alluvium.

The terrace alluvium is considerably more important for the purposes of this report. Alluvial terraces are common along the walls of the Missouri River trench and, in places, form the reservoir shore. Locally, erosion has produced a topographic reversal so that alluvial deposits cap uplands bordering the trench.

The thickness and composition of the terrace alluvium vary. In some exposures the alluvium is only a thin veneer over bedrock; in others it forms terraces as much as 80 feet high. Preponderance of the slow- or fast-draining alluvium varies from one outcrop to the next, but most terrace deposits are composed of fast-draining alluvium overlain by a few feet to perhaps 25 feet of slow-draining alluvium.

Many landslides resulting from the combined effects of wave erosion and changes in saturation have already developed in fast-draining and slow-draining terrace aluvium. Continued sliding may be expected until the slopes reach equilibrium.

ALLUVIUM, COLLUVIUM, AND LOESS

These deposits are composed of mixtures of alluvium, colluvium exclusive of landslide material, and loess. Acording to Stokes and Varnes (1955), colluvium is earth material moved or deposited mainly through the action of gravity. The grain size commonly is no coarser than fine sand, although locally there are thin lenses of pebbly material. The material behaves as slow-draining alluvium. These deposits are rarely more than 10 feet thick and commonly occur as relatively small areas in the bottom of small valleys.

LOESS

Loess, a porous wind-deposited sediment composed predominantly of silt with minor amounts of fine sand and clay, forms a patchy blanket along the Missouri River trench walls and on the bordering uplands. In most of the Fort Randall area it is a few inches to a few feet thick, although locally it is more than 30 feet thick.

When dry, loess has sufficient strength to stand in vertical bluffs. This characteristic apparently results from the bonding action of clay particles that adhere to the silt and sand grains (Holtz and Gibbs, 1951, p. 15). When the clay becomes wet, the bonding forces are reduced and the loess loses its strength. Loess bluffs saturated by the reservoir water soon collapse, and the fallen material is removed by wave erosion.

ORIGIN AND DEVELOPMENT OF MISSOURI RIVER TRENCH

A brief discussion of the origin and development of the Missouri River trench is pertinent because nearly all the landslides in the area are the result of the downcutting and widening of the trench.

History.—The Missouri River trench is a comparatively recent feature. Until mid-Pleistocene time the major drainage in central South Dakota was easterly (Flint, 1955, pl. 7). During the Illinoian Glaciation ice dammed the east-flowing streams so that they were forced to drain southeastward along the ice front (Warren, 1952). By the time the ice retreated, this melt-water channel, the incipient Missouri River trench, had become well established in the easily eroded Pierre Shale. Most of the downcutting in the trench probably occurred during the Sangamon Interglaciation (Warren, 1952, p. 1151). Successive glacial advances in Wisconsin time partly filled the trench with outwash and till. Although most of these deposits were removed by renewed downcutting after each ice advance, numerous terraces and terrace remnants along the trench remain from this period.

The Missouri River has reexcavated its valley in Holocene time and now has a gradient of about 1 foot per mile. Until the reservoir was filled most of the river's energy was spent cutting laterally into the valley walls.

Mechanics of erosion.—In the course of its development the Missouri River trench was excavated in the Pierre Shale primarily by mass wasting and stream erosion, and to a minor degree by slope wash. In this multiple effort, landsliding probably was at least as effective as stream cutting. Slope wash probably was most effective on slopes where there were few if any active landslides.

Erosion by landsliding is caused by and in many places perpetuated by stream action. Downward and lateral cutting by streams oversteepens and undercuts the valley walls. Blocks of material adjacent to the streams thereby become unstable and break loose, sliding down and outward until they reach a stable position. Movement of these blocks leaves unsupported oversteepened slopes above, and eventually other blocks will slide. By repetition of this process, the entire valley wall slowly slides toward the stream, which in turn makes further sliding inevitable by its removing the landslide debris when it reaches the valley bottom.

The upslope migration of a zone of active landslides is never as obvious in the field as it would seem from the preceding description, because in nature many factors other than removal of toe support affect slope stability. At numerous places along the walls of the Missouri River trench, nevertheless, a slide sequence grades upslope from old, barely discernible, stabilized landslide blocks along the riverbank, through more recently stabilized slide blocks, into areas of active landsliding near the uplands. In some places, moreover, new landslides are developing along the riverbank, starting a new cycle of landsliding.

Over a long period of time the cumulative effect of several landslide cycles may be considerable lateral and vertical movement, although individual blocks are not necessarily transported a great distance during a single cycle. Wave erosion along the west shore of the reservoir about 10 miles upstream from Fort Randall Dam has exposed an old landslide containing isolated blocks of Ogallala Formation more than 350 feet below the outcrop level of the formation in the nearby uplands.

LANDSLIDES

Because landslides are the result of variable combinations of material and movement, it was inevitable that several systems of clasifying landslides were devised. The Highway Research Board classification (Varnes, 1958) based on two main variables—type of material and type of movement—was selected as the most satisfactory for this investigation. Plate 2 is a graphic summary of this system, with notes added concerning the materials involved in the most common landslides in the Fort Randall Reservoir area.

The fundamental concepts of landslides and their behavior are stated concisely by Varnes (1958, p. 20), as follows:

> * * * the term "landslide" denotes downward and outward movement of slope-forming materials composed of natural rock,

soils, artificial fills, or combinations of these materials. The moving mass may proceed by any of three principal types of movement; falling, sliding, or flowing, or by their combinations. * * * Normal surficial creep is excluded. Also most types of movement due to freezing and thawing (solifluction), together with avalanches that are composed mostly of snow and ice, are not considered as landslides.

TYPES OF LANDSLIDES

Most landslides in the Fort Randall Reservoir area involve at least two types of movement and commonly more than one type of material. One type of movement predominates in most of the slides, and if more than one kind of material in involved, generally only one is responsible for the landsliding. In the slump-earthflow type of landslide, two kinds of movement occur, slump and flow, and a hyphenated compound term is used to describe them.

The various types of landslides along the reservoir are discussed in the order they are listed on the landslide classification chart (pl. 2), rather than according to their abundance or size in the Fort Randall area.

ROCKFALLS

Small rockfalls occur in the Niobrara Formation where the chalk beds are undercut by stream and wave erosion or are weakened by prolonged weathering. The blocks rarely are more than a few feet in maximum dimension and the process usually borders between raveling and rockfall.

SOILFALLS

Soilfalls—in till, slow-draining alluvium, the alluvium, colluvium and loess unit, and loess—(pl. 2), caused by stream and wave erosion, are very common where fine-grained unconsolidated deposits crop out along the reservoir (fig. 2) or along the tributary valleys. Generally the blocks are a few feet wide and no more than 30 feet long. The overall importance of soilfalls as a landslide process is usually overlooked because the individual blocks are relatively small. Probably more than half of the fine-grained unconsolidated material undergoing erosion along the reservoir shore is entering the water as soilfall blocks.

Although soilfalls are a common result of erosion, their mechanics are relatively simple. Soilfalls result from removal of support at the base of nearly vertical or vertical bluffs. Slopes must be so steep that unstable blocks tend to fall downward and outward instead of sliding along the surface of failure. Saturation, seepage pressure from ground water moving toward a free face, and unbalanced pore-water pressures may contribute to soilfalls, but they are never primary causes.

FIGURE 2—Soilfall and slumping along the left bank of Fort Randall Reservoir, Charles Mix County, S. Dak. This landslide consists of two distinct types of movement, soilfall at the water's edge and slumping upslope. The material is slow-draining alluvium capped by a few feet of loess and underlain a short distance below water level by Pierre Shale. Several small overhanging blocks of loess are visible in the center of the photograph. Photographed September 27, 1954.

BEDROCK SLUMPS AND BLOCK GLIDES

Slumps (rotational movement) and block glides (planar movement) (pl. 2; fig. 3) represent the end members of the series of landslide types caused by shear failure of coherent blocks. The relation between the two slide processes is shown in the schematic cross sections of figure 3. If slumps have cylindrical surfaces of shear failure, the shear surface appears in cross section as the arc of a circle with a radius of finite length, r (fig. 3*A*). If the curvature of the shear surface is reduced, the radius, r, will increase, gradually approaching infinity. The extreme case, where the radius, r′, is infinitely long and the shear surface is planar, is represented by the block glide (fig. 3*B*). Be-

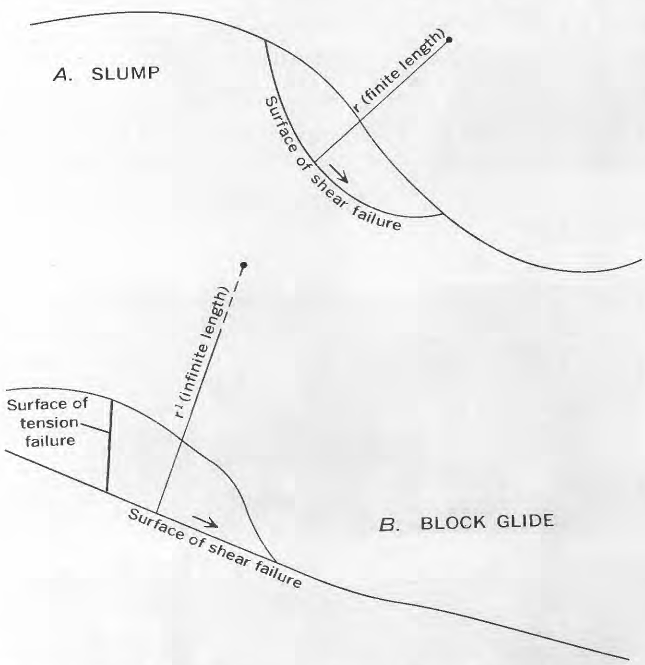


FIGURE 3. Relation between slump (*A*) and block glide (*B*), end members in the series of landslide types involving shear failure of coherent blocks.

cause of the similarity in causes and behavior, block glides in the Pierre Shale are considered slumps.

Slumping in the Pierre Shale is the most common type of landslide movement in the project area. In some parts of the Missouri River trench the trench walls are almost entirely slump blocks. These slides vary greatly in size. One slide, the Landing Creek slump-earthflow (fig. 4), has a scarp about 1,300 feet long at the head of the slump section; but some slides are no more than 30 feet across.

Although rotational movement predominates, most slumps seem to be composite, with both rotational and planar movement. Where the shale is relatively homogenous, the slump movement is rotational along a concave-upward shear surface. Abundant bentonite beds in the Pierre, however, represent zones of weakness. If a shear surface encounters a bentonite bed near the base of a slump block (fig. 5), the shear surface becomes planar, and movement is translated to glide. Sometimes there is contemporaneous glide movement along several parallel shear planes at the base of a slump. If a shear surface encounters bentonite beds at the back of a slump block above the base, it cuts through the beds and remains curved.

Most slumps start as movement of a single block (fig. 6, lower half of slump). If the block is large or if the shear surface is irregular, deformation caused by the movement breaks the single block into several smaller units. Where there is sufficient water, the toe of the slump, which is commonly highly fractured and most easily saturated, may become an earthflow (fig. 7). The original block continues to move, either as a single unit or in pieces, until it reaches a stable position. If the toe of a slump is removed by agency of nature or man, the landslide will not become stable until all or most of the slide material has been removed (fig. 8).

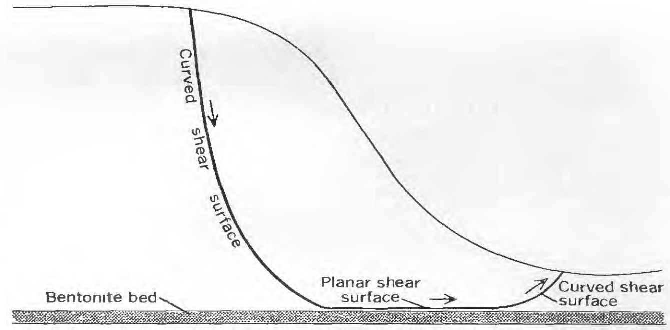


FIGURE 5.—Schematic cross section through a potential slump with a bentonite bed controlling development of the shear surface.

Slump movements become more complex as movement continues. The material in the unsupported scarp above the initial slump block generally cannot stand long without support on its downhill side, and soon a second block breaks loose. If the initial slump block continues to move, the second block follows, leaving a new unstable scarp at the head of the landslide. By repetition of this process, the slump area gradually migrates upslope until the individual blocks reach stable positions (figs. 7, 8).

SOIL SLUMPS

Soil slumps—alluvium, colluvium and loess unit, loess, and till?—(pl. 2) were rare before the filling of the reservoir but they have been fairly common along the shoreline since, and probably will continue to occur for many years. The lower part of most of the soil slumps is submerged, and many have been completely inundated by the rising water. Till adjacent to the uplands is involved in many slumps, but failure of the Pierre Shale underlying the till generally is responsible.

FIGURE 4.—Landing Creek slump-earthflow. The upper part of the landslide consists of slump blocks moving out on an old terrace remnant. Where gullies had eroded the terrace, the slump blocks received no support. and the material disintegrated to form earthflows moving down the gullies. SE¼NE¼ sec. 25, T. 100 N., R. 72 W., Gregory County, S. Dak. Photographed September 12, 1953.

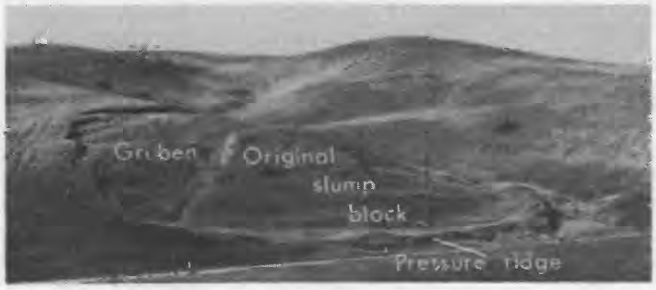


FIGURE 6.—Highway 16 slump. A slump southwest of U.S. Highway 16 about 1 mile south of the center of Chamberlain, S. Dak. The lower part of the slump, with a pressure ridge caused by forward and upward thrusting of the slump block at the toe, is inferred to be the original slump block. The upper part is a graben formed by collapse after the original slump block moved and left the upslope material unsupported. Photographed October 14, 1954.

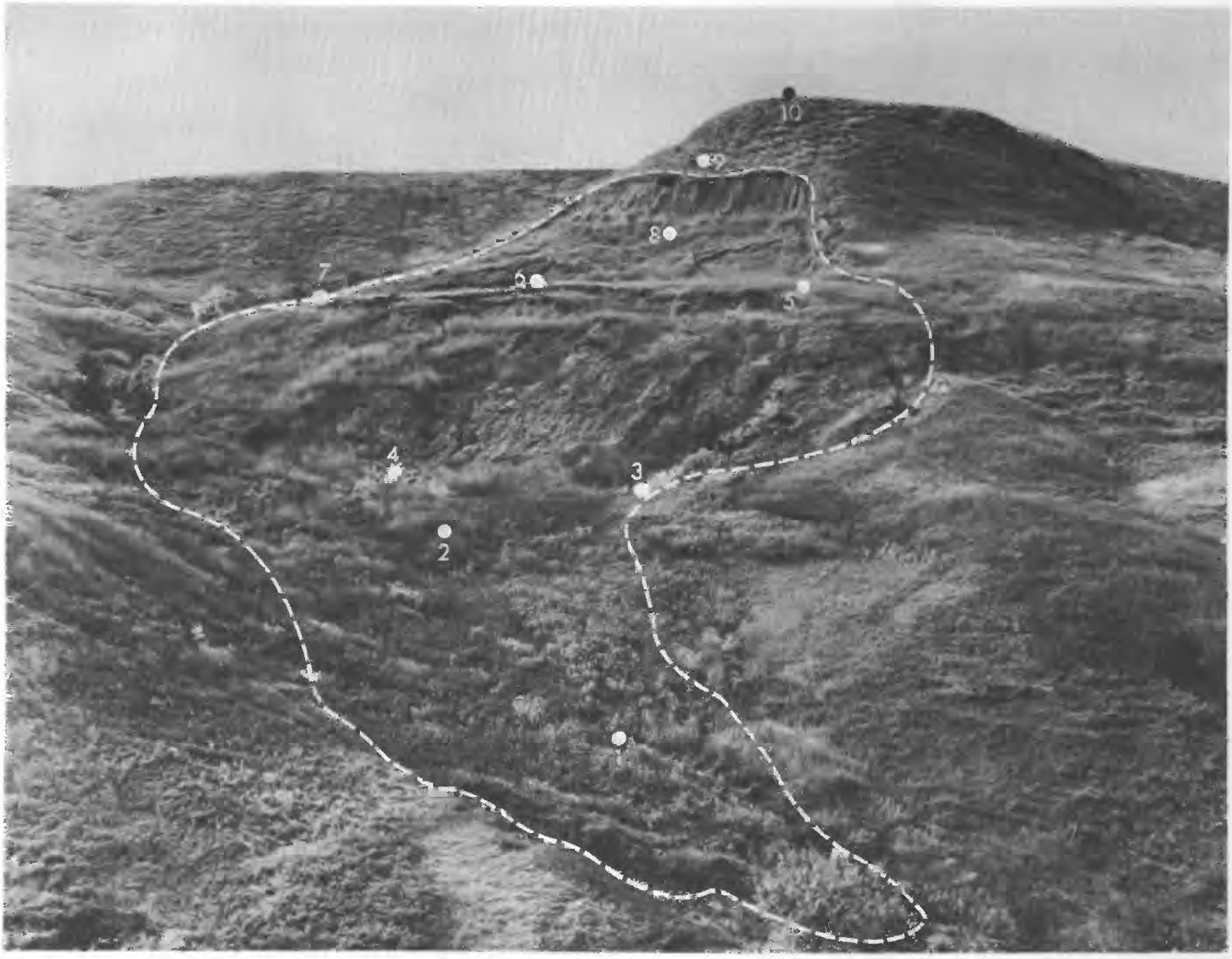


FIGURE 7.—Cable School stump-earthflow. NE¼NE¼ sec. 25, T. 103 N., R. 72 W., Brule County, S. Dak. The upper part of the slide (above points 5, 6, and 7) is numerous slump blocks. In the scarp between points 3 and 4 and points 5, 6, and 7, the slump blocks are disintegrating to chunks of weathered shale and clay. Water from springs near point 4 mixes with the disintegrated slump blocks and forms an earthflow (the lower part of the slide below points 3 and 4). Numbers refer to control stations established to measure movement. Photographed October 14, 1954.

Soil slumps now occurring along the reservoir shore range from those caused primarily by saturation to those caused by erosion. The slump shown in figure 9 has developed along the face of a terrace composed of gravel (fast-draining alluvium) blanketed by loess and colluvium. A small wave-cut bench about 8 feet above the water surface represents the reservoir water level before a drawdown beginning about 2 months before the photograph was taken. The wave-cut bench on the slide block is closely alined with wave-cut benches on each side of the slide area, a feature indicating that little or no movement has occurred since the drawdown. Also, the fact that the bench is as well developed on the landslide as it is on either side indicates that the slump occurred before wave erosion started. These facts indicate that neither minor wave erosion nor drawdown noticeably affected the slump block after it became stabilized. The slump movement apparently is related to saturation of the materials composing the

FIGURE 8.—Active slump along the Fort Randall Reservoir shore in Gregory County, S. Dak. Wave erosion at the toe of the slump is breaking up the slump blocks and removing the material. Photographed September 27, 1954.

FIGURE 9.—Slumping along the face of a gravel terrace blanketed by colluvium and loess. The limited amount of wave erosion implies that saturation, rather than oversteepening of the shore, was the primary cause of slumping. Fort Randall Reservoir shore, Charles Mix County, S. Dak. Photographed October 24, 1953.

slump block. Failure began after saturation destroyed the intergranular bonding effect of capillary forces in the gravel (see p. 18). Once the block began to settle, part of its weight was supported by water, the intergranular friction in the gravel was reduced by the buoyant effect of the water, and further settling occurred.

A soil slump caused primarily by wave erosion is shown in figure 10. This photograph, taken at the same time as figure 9, shows slumping in slow-draining alluvium. Here also, there has been a temporary drawdown of about 8 feet in the reservoir water level. There is no indication of a wave-cut bench, however, despite the fact that wave erosion is more active here than in near the area shown in figure 9. The slump shown in figure 10 was still moving at the time it was photographed, and the combination of landslide movement and wave erosion apparently obliterated all evidence of a wave-cut bench. Saturation followed by friction from draining water and unbalanced pore-water pressure probably decreased the stability of the slow-draining alluvium; however, the area should have become stable after the water level dropped. A logical explanation of the slump's continuing movement is that removal of support at the toe by continuing wave erosion was the main cause of slumping and was still preventing stability at the time the photograph was taken.

FIGURE 10.—Slumping in slow-draining alluvium along the right bank of the Fort Randall Reservoir, Gregory County, S. Dak. The area is exposed to wave erosion which, by removing support at the toe of the slump block, is probably the primary cause of slumping. Photographed October 24, 1953.

SLOW EARTHFLOWS

Slow earthflows—weathered Pierre Shale, till, and colluvium—(pl. 2) (Varnes, 1958, p. 38) are the most common type of slope failure in plastic unconsolidated materials along the reservoir. Most of the earthflows are small, less than 100 feet wide by 150 feet long. Where a larger area is involved, the upper part of the landslide generally is a well-developed slump, and the slide is classified as a slump-earthflow.

Earthflows form only on slopes that are composed of a mantle of plastic, unconsolidated material underlain by relatively coherent impermeable material. Two conditions are responsible for this restriction. Earthflows form only in material that is sufficiently porous to hold enough water to make it sufficiently plastic to flow as a mass or to be converted into a viscous fluid. Generally this qualification applies only to clay-rich colluvium or weathered shale and till. Also, earthflow material must be saturated, or nearly saturated, before it will flow. Where relatively impermeable material is overlain by more permeable material, water percolating downward from the surface will stop at the contact and build up a saturated zone in the mantle (fig. 11). Sections of the mantle will fail by flowing if the saturated zone is thick enough and the slope steep enough. Blocks of unsaturated material commonly are carried along on the surface of the earthflow.

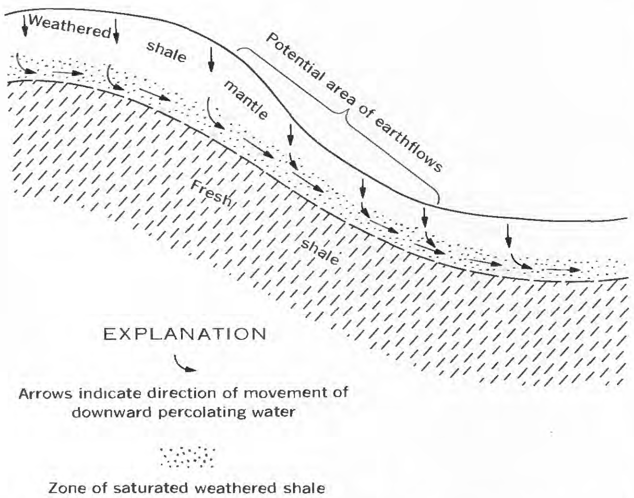


FIGURE 11.—Idealized diagram of conditions conducive to earthflows.

The three earthflows shown in figure 12 are typical of those along the Fort Randall Reservoir. They have formed in shale-rich gravelly colluvium and weathered Pierre Shale, all underlain by relatively competent shale. The small flow shown in the lower left corner of the photograph has virtually no slump movement. In the middle of the photograph is a flow in transition from slump-earthflow to earthflow. The upper half of the flow consists of many slump blocks so small that the overall behavior of the material more closely resembles flowage than movement of unit slump blocks. The third flow, on the right in the photograph, is the most common type. Although there are one or two large coherent slump blocks at the top, over three-quarters of the area involved is a true flow.

MUDFLOWS

Mudflows characterized by the rapid flow of a saturated mass of material are confined to the weathered Pierre Shale along the Missouri River trench. Though viscous, they are much more fluid than earthflows and, once started, they flow on very gentle slopes. Old flows closely resemble, and are locally incorporated in, alluvial fans at the mouths of gullies.

Most members of the Pierre Shale weather to bentonitic clays that when dry develop a very coarse, porous, crumblike structure. During heavy rainfalls this open structure rapidly absorbs large quantities of water before the wet bentonite expands and reduces the permeability. The saturated mass of weathered shale becomes a mudflow when added water has increased the weight and lessened the cohesion of the clay so much that it is no longer stable on a slope.

FIGURE 12.—Cable School earthflows, NW¼ sec. 30, T. 103 N., R. 71 W., Brule County, S. Dak. Three earthflows in gravelly colluvium and weathered Pierre Shale. The two larger flows have numerous small slump blocks in their upper portions. Note vehicle on top of ridge for scale. Photographed July 7, 1952.

CAUSES OF LANDSLIDING IN THE FORT RANDALL RESERVOIR AREA

The causes of landslides are in two general groups: (1) basic processes which slowly reduce the stability of a mass of material and (2) trigger processes which in themselves are of minor importance but which, when added to the stresses already operating in a potential slide area, are sufficient to set the mass in motion. The trigger actions are more obvious and often are erroneously considered the dominant causes of slides. A trigger process, however, can start a landslide only after an area has become unstable through the operation of other factors. Eliminating possible trigger actions, nevertheless, can prevent or at least postpone many landslides.

EROSION

Erosion, the main cause of landsliding along the reservoir, operates both as a basic cause of movement and as a trigger action. As a basic cause, it creates the topography that makes the downward and outward movement of landslides possible. Landsliding could not continue in structurally stable South Dakota if erosion did not create steep slopes. Erosion becomes a triggering process when it removes from the toe of a potential landslide material which otherwise would serve as a buttress (figs. 8, 10).

GROUND WATER

Ground water is the most versatile cause of landslides. It can affect slope stability by means of weight, lubrication, hydrostatic pressure, and piping. Some of these increase landslide stresses; others decrease sheer strength in potential slides. Ground-water conditions may be the trigger action or simply the general cause of instability.

WEIGHT

The effect of the weight of ground water on a potential landslide is not easily evaluated. Addition of water to earth materials increases their density because air in the voids is replaced by water; the greater density then produces some increase in the stresses within the earth materials. The present studies have not furnished sufficient data to determine if an increase in stress due to greater density will necessarily be accompanied by a decrease in stability.

LUBRICATION

Lubrication by water is one of the oldest explanations regarding the disastrous effects of ground water on the stability of slopes. Consequently, such lubrication is often acepted as the trigger action in most landslides that closely followed periods of heavy ground-water intake. The nature of the material to be lubricated and its reaction to water must be considered in order to analyze the lubricating effects of water in landslides along the Fort Randall Reservoir. Fresh Pierre Shale is a coherent aggregation of close-packed, interbonded particles of silt and clay size. Its present character is explained by the history of its deposition and compaction.

The sediment that comprises the Pierre Shale was deposited on the bottom of an inland sea as a mud composed of loosely bonded aggregates of poorly oriented particles that were relatively admixed with large quantities of free and adsorbed water. As the deposits thickened, compaction gradually broke down the poorly bonded, unstable aggregates. The platy clay-mineral grains probably developed the horizontal orientation that gives fissility to the shale, and much of the free water was squeezed out by the reduction in pore space. At this stage the material became a soft clay.

Compaction of the Pierre Shale continued as the overburden increased until much of the adsorbed water was driven from between the particles. The particles became closer packed, and molecular forces between particles formed a relatively tight bond. In subsequent periods of uplift, erosion removed much of the overburden, but molecular forces bonding the particles apparently remain strong enough to prevent re-expansion of the shale under most conditions.

The rapid weathering of exposed Pierre Shale to a poorly bonded clay indicates, however, that the adsorbed water layer was still sufficiently thick to prevent the molecular forces between particles from developing a bond between solids. When overburden pressures are removed, therefore, the attractive forces between the adsorbed water and available free water are greater than the molecular forces between the particles. Additional water is drawn into the adsorbed water layer, reducing the bonding forces between particles. The effects of weathering are aggravated by the presence in the Pierre Shale of montmorillonite, a clay capable of swelling by adsorbing water within its molecular structure (p. 25).

Behavior of unconsolidated material not directly derived from Pierre Shale depends on the size of the particles. In materials in which clay or fine silt is a binder, the area of intergranular contacts is high in proportion to the volume of solids. Molecular forces are important, therefore, and the material behaves much like Pierre Shale. In coarser grained material the area of intergranular contacts is small in proportion to the total volume of solids; the effects of molecular attraction are less and lubrication by water is a less important factor in stability.

The shear strength along the surface of contact between two solids is related to friction between the solids. If a liquid is placed so there is direct contact between the two solids, shear will occur in the liquid, which has much less shear strength than the solids. Reduction of friction by allowing shear to occur in some medium with very little shear strength is the basic principle of lubricants.

Terzaghi's theories (1950, p. 91) that a very thin film of water around a particle provides full lubricating effect and that most materials are always fully lubricated do not seem entirely valid with regard to the sediments along the Fort Randall Reservoir. Full lubrication implies shear within the lubricant; therefore, the shear strength of a material should equal that of the lubricant. Such conditions occur only locally and intermittently in any of the materials along the reservoir. In fact, full lubrication during movement is approached only in the more fluid types of flows.

Lubrication in Pierre Shale slumps, though probably never complete, is a major cause of movement. Ground-water investigations (see section "Ground-Water Investigation") indicate that the regional water table is high enough to saturate permanently most of the shear surface in large slumps. In smaller slumps the shear surface generally is saturated for at least 4 months of each year. Even in saturated shale, lubrication is not complete, however, because a combination of molecular forces and overburden pressures binds the particles (including adsorbed water layers) together at their contacts. The pressure of the saturating water normally is not great enough to penetrate these contacts and lubricate where friction is greatest. Although shale is not fully lubricated even in zones of permanent or semipermanent saturation, its lubrication is as complete as is possible for the given conditions.

Lubrication is more complete in flows than in slumps. An active flow is essentially an assemblage of solid particles, or aggregates of particles, in a matrix of water, and behaves like a viscous liquid. The shear strength of flow material, as a result, is much less than that of solid material in a dry state. In some mudflows it approaches that of water, indicating almost complete lubrication.

The difference in the degree of lubrication in Pierre Shale slumps and in flows is due to differences in materials and environment. Slumps occur in coherent materials that generally have sizable overburden pressures along the surface of failure. Flows occur in near-surface unconsolidated materials where coherence is relatively slight and overburden pressures are negligible.

Lubrication generally has a secondary effect in slumps in unconsolidated materials. If enough water is present, the failure is by flowage instead of by shear. When flow occurs in unstable material underlying coherent material, movement in the coherent block may resemble a slump.

The effect of saturation on apparent cohesion probably is a common trigger action for slumps in unconsolidated materials. Apparent cohesion is the cohesion developed by surface tension at air-water interfaces in materials than contain water but are not saturated (Terzaghi and Peck, 1948, p. 126). Saturation destroys air-water interfaces, and the resulting loss of apparent cohesion directly reduces stability of the material. Saturation also reduces the cohesion between the grains, and as a result more water can enter the adsorbed water layers as a lubricant.

The effects of lubrication on soil slumps (unconsolidated materials) are not as great as on slumps composed of Pierre Shale. The only exception is slumping in till. Being fairly compact material derived largely from Pierre Shale, till behaves like Pierre Shale.

The effects of lubrication cannot be classified on all landslides either as a general cause of decreased stability or as a trigger action. In general, however, lubrication by permanent or semipermanent saturation probably results in decreased stability. Lubrication by periodic or occasional inflow of ground water commonly is a trigger action.

HYDROSTATIC PRESSURE

The effects of hydrostatic pressures in ground water have increasingly been emphasized as landslide causes. Terzaghi (1950), one of the chief proponents, summarizes the engineering viewpoint on effects of hydrostatic pressures.

Many of the theoretical principles of hydrostatic pressure, though not directly applicable in nature, can be combined with empirical data to yield quantitative information. For example, the theoretical discussion of hydrostatic pressures that can develop in shale fractures cannot be used quantitatively in nature. Water pressure changes can actually be measured, nevertheless, and the application of theoretical reasoning to these data will determine the changes in forces acting on the shale along fractures. Thus, water levels in the standpipes of piezometers within stable shale have been known to rise more than a foot in less than a month. A 1-foot rise in water level represents 62.4 psf (pounds per square foot) increase in pressure. This increase in pressure acting on a surface 10 feet square creates a force of more than 3 tons.

Hydrostatic pressures are either uniform or variable. Uniform hydrostatic pressures create static ground-water conditions. Ideally, pressure can occur in unconfined water as a result of the weight of overlying ground water, in which case the pressure is directly proportional to the depth below the local water table; or else the pressure can occur in water confined to a permeable zone overlain and underlain by relatively impermeable materials. In the latter the pressure theoretically would have no direct relation to the overlying water table.

Uniform Hydrostatic Pressure

Independence between unconfined static ground water and the surrounding material is a basic concept of soil mechanics. Under static conditions the total pressure at any point in saturated material composed of incompressible grains is equal to the pressure exerted by the material (effective pressure) plus the pressure exerted by the contained water (neutral pressure). The two component pressures act independently of each other.

Hydrostatic pressure in confined static ground water behaves much like a hydraulic jack; pressures in the fluid are constant throughout any horizontal stratum in the system, and pressure exerted anywhere on the fluid system is transmitted through the entire system. A small increase in pressure over a large area thus can represent a major increase in force. A vertical rise in the hydraulic head of a small fracture connected with a horizontal permeable stratum can also create a tremendous increase in the total hydraulic force in the stratum.

Pierre Shale is the only material, with the possible exception of till, along the Fort Randall Reservoir in which pressure in confined static ground water may be important. The Pierre is a generally massive shale with very low permeability. Fractures are numerous, however, and investigation (see p. 50) shows that they carry most of the moving ground water. Ground water in these fractures reacts like confined water to pressure changes.

The effects of hydrostatic pressure in fractures within the Pierre Shale are demonstrated schematically in figure 13*A*, a cross section through a shale bank containing a horizontal permeable parting 40 feet below the top. It is assumed that the parting is sealed at the surface to permit static conditions. Another fracture, a steeply dipping joint, intersects the parting and crops out at the top of the bank. The block outlined by the two fractures and the surface of the shale represents a potential landslide.

When the fractures are filled with water, the pressure is proportional to the height of water. The water pressures throughout the parting shown in figure 13*A* are caused by the 40-foot head of water in the joint. Although water pressures exert force in all directions, the water pressures acting on the potential landslide block are the present concern. An essentially horizontal pressure at the back of the block pushes it laterally; an upward pressure along the base of the block reduces friction (see below). If the total forces from these combined pressures is great enough, the block becomes an active landslide.

Reducing friction by increasing hydrostatic pressures in confined ground water can renew movement in old landslides as well as initiate movement in new slides, the only difference being that in old landslides the surface of failure forms the permeable fracture (fig. 13*D*).

An analysis of vertical pressures on a small block of shale (block X, fig. 13*A*) illustrates the potential effect of hydrostatic pressures. When the fractures are dry, the vertical pressures on block X are a downward pressure from the weight of the overlying shale and equal upward pressure transmitted to the potential slide block from the shale below the parting (fig. 13*B*). If an average density of 155 lb per cu ft is assumed for the shale, the magnitude of each force acting on block X is about 6,200 psf.

When the fractures are filled with water (fig. 13*C*), the downward pressure on block X remains the same, but about 2,500 psf of upward pressure is now exerted by the water. Inasmuch as total upward pressure still equals 6,200 psf, the pressure transmitted through solid shale particle contacts now is only 3,700 psf. The reduction of the upward-acting solid-to-solid pressure to about 60 percent of its original amount correspondingly lowers friction along the base of the block and increases the chances of movement along the parting.

Variations in Hydrostatic Pressure

One important effect of lateral changes in hydrostatic pressure results from friction created by molecular attraction between the water and the solid particles between which it passes. This drag of moving water exerts a force on particles known as seepage pressure (Terzaghi and Peck, 1948, p. 54; Terzaghi, 1950, p. 99). Where the pressure gradient is steep, seepage pressures may become great enough to trigger landslides.

The effectiveness of seepage pressure depends on the particle size of the material on which it acts. In coarse gravel and other fast-draining materials, resistance to flow is slight, and generally large seepage pressures do

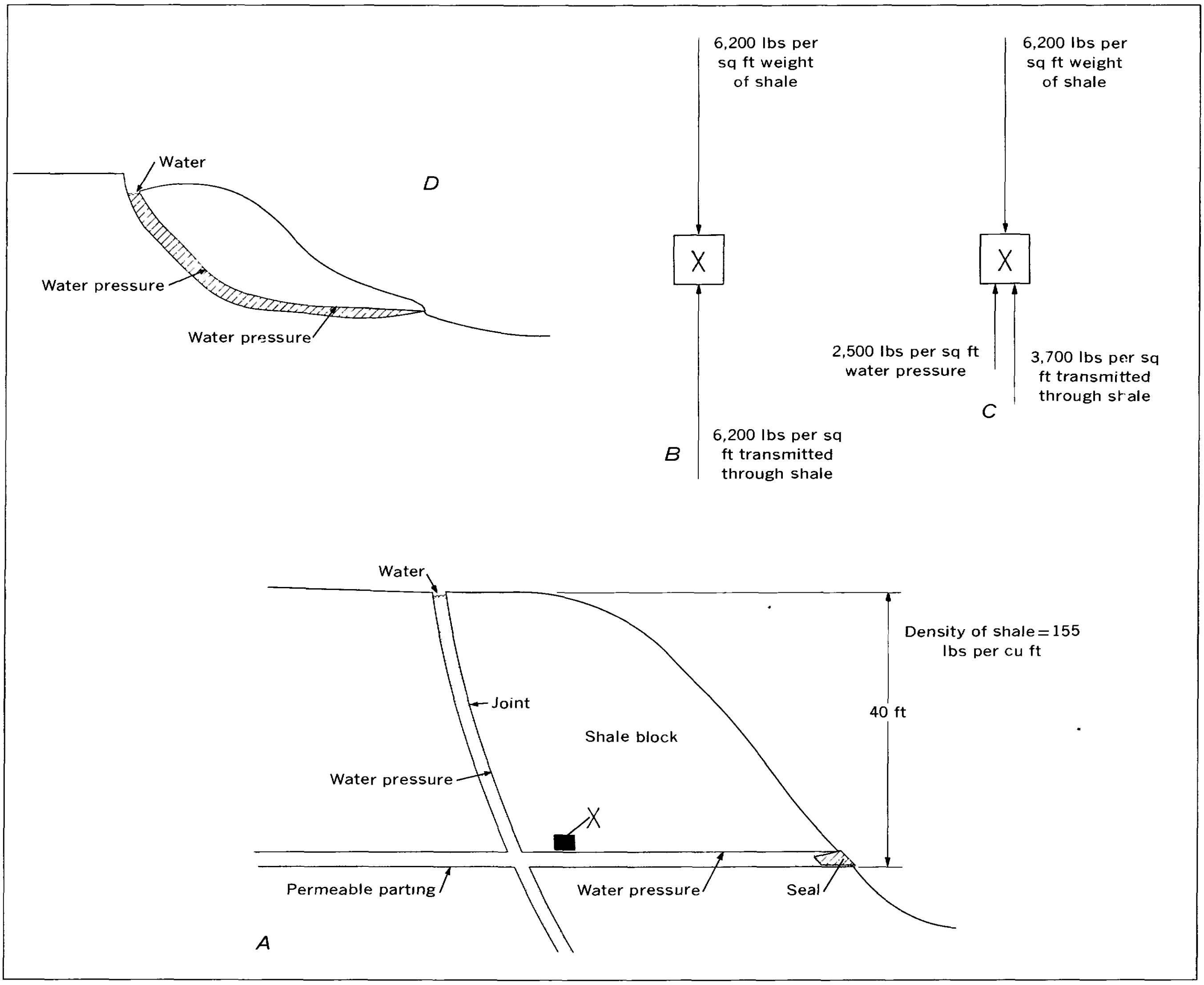


FIGURE 13.—Effect of hydrostatic water pressure in fractures within relatively impermeable shale. *A*, potential landslide block; *B*, pressure exerted on block *X* when fractures are dry; *C*, pressures exerted on block *X* when fractures are filled with water; *D*, old landslide block where the surface of failure is the permeable fracture.

not develop. Pressures caused by seepage in clays probably can be considered minor, also. In clay, the finest-grained of slow-draining materials, water pressure is virtually static. Slow-draining materials in the effective size range of silt and fine sand are most affected by seepage pressures because they are coarse enough to permit considerable movement of water, yet fine enough to develop large seepage pressures. Slope failures in which seepage pressures may be the trigger action occur in silt and fine sand along many reservoirs subject to rapid large drawdowns (Terzaghi, 1950, p. 99). Sudden drawdown of the free water level creates a steep pressure gradient in the ground water. The resulting movement in the ground water creates seepage pressures that may be large enough to trigger landslides.

The effects of saturation and drainage on the bonding forces of capillary water combined with the effects of seepage pressures are responsible for marked differences in the behavior of fast-draining alluvium and slow-draining alluvium.

Fast-draining alluvium becomes less stable as it is saturated, and it regains its stability as it drains. As water saturates fast-draining alluvium it quickly fills nearly all the pores and eliminates the air-water interfaces. As a result, the bonding effect of surface tension is destroyed, and the alluvium becomes less stable. As the alluvium drains, it soon regains the original condition in which capillary water bonds the particles. There is no appreciable friction between the draining water and the alluvium, and the water drains too rapidly to build up significant pore-water pressure.

Several factors cause slow-draining aluvium to be appreciably more stable when saturated than fast-draining alluvium. When the level is raised in an adjacent body of water, the low permeability of slow-draining alluvium resists entrance of the water. Instead of rapidly filling the pores, the water becomes a buttress against the alluvium, thus making the alluvium more stable. Moreover, air trapped by the slowly infiltrating water preserves some of the air-water interfaces and so the bonding effect of surface tension is not entirely destroyed. The adsorbed water layer on the surface of every particle (Terzaghi and Peck, 1948, p. 10–17) also increases the stability of slow-draining alluvium. Water in the adsorbed layer is less fluid than normal water and as a result has greater shear strength. The adsorbed water layer, consequently, has little effect in coarse-grained alluvium. In the finer materials of the slow-draining alluvium, however, the proportion of voids smaller than 0.2 micron wide becomes significant and the adsorbed water layers become important forces in helping the material resist shear.

Two forces tend to reduce the stability of slow-draining alluvium when it is draining. If the water level falls rapidly in a body of water in contact with slow-draining alluvium, the water in the alluvium cannot escape as rapidly as the free water level drops. Then the friction between the alluvial particles and the escaping water creates an internal stress which acts toward the free face of the alluvium. At the same time the pore-water pressure in the alluvium is no longer balanced by water pressure in the adjacent water body, and the unbalanced pressure creates a lateral stress toward the free face of the alluvium.

PIPING

Piping, internal erosion by moving ground water, is a minor process that gradually reduces the stability of slopes. The stability changes caused by piping are generally impossible to determine quantitatively. Internal erosion can occur by mechanical transportation or by solution. Mechanical transportation usually involves movement of particles of silt and fine sand small enough to be easily transported but not so small that they are held together appreciably by molecular forces. It is most common in poorly sorted gravels in which the coarser constituents can support the overlying material while the fines are washed away.

The effect of solution on stability in the Fort Randall area is unknown; the removal of material by solution does not leave large cavities, and appreciable surface deposits are not precipitated. The best evidence that solution ocurs in this area is the ground water itself. Searight and Meleen (1940) stated that there is hard water in more than three-quarters of the shallow wells in the counties along the reservoir. All these wells are less than 200 feet deep, and most are less than 50 feet deep. The water is relatively pure when it enters the ground; it must, therefore, dissolve minerals from the material through which it passes.

MISCELLANEOUS CAUSES OF LANDSLIDING

Earth tremors and loading on the heads of potential landslides can produce movement. Earth tides and atmospheric pressure changes are possible but generally unverified causes of landsliding. Although loading at the head of a potential landslide is rare under natural conditions, it occurs fairly commonly as a manmade phenomenon associated with various engineering projects. It may be either a general cause of instability or a trigger action. The slump pictured in figure 14 is a typical example of an earth movement produced by artificial overloading of an unstable area. Failure occurred after fill for a new highway was placed on a ridge of Pierre Shale. Abnormally lush vegetation in parts of the landslide area indicates that the shale was already wet and unstable; the added weight of the fill was the trigger that set the slope in motion.

In the Great Plains area the effects of earth tremors are probably very slight. Small or infrequent trigger actions, on the other hand, may be important complements to a larger trigger action. For example, the effects of saturation may not be enough to trigger a potential slide, but a very slight earth tremor occurring when material is saturated may produce movement.

RELATIVE IMPORTANCE OF THE LANDSLIDE CAUSES

Most landslide movement results from a combination of causes, but the various agents involved have been discussed separately so that their individual effects on slope stability could be appraised.

Table 2 summarizes landslide causes and presents estimates of their relative importance in the Fort Randall Reservoir area. Investigations along the reservoir were qualitative, and at best crudely quantitative. The interpretations presented in the table are intended to be a guide for future detailed investigations.

ANALYSES OF THE PIERRE SHALE

All the slope failures along the Fort Randall Reservoir, except soil slumps and soilfalls in alluvium and loess, involve either the Pierre Shale or material derived from it.

FIGURE 14.—Slump caused by highway fill being placed at the head of a potential landslide block. Relocation of South Dakota Highway 47, sec. 36, T. 103 N., R. 73 W., Lyman County, S. Dak. Photographed July 16, 1955.

A thorough investigation of the characteristics and behavior of the Pierre Shale is beyond the scope of the present study, and the data included here are only qualitative at best and are not complete. The information about the Crow Creek Member, which is thin and seems to have little effect on slope stability, was taken from Crandell's report (1952). The samples of the other shale members were collected from only one locality and, in most cases, one horizon per member; therefore they cannot show lateral and vertical variations although they give a good picture of the general nature of the Pierre.

Investigation of the shale included mechanical analyses, mineral analyses, and some clay studies. Mechanical analyses were made to determine possible relation between grain size and stability. Mineral analyses were made to evaluate the effects of mineral composition on stability. The clay studies were made in an attempt to correlate behavior of the clay minerals with slope stability in the shale.

MECHANICAL ANALYSES

The mechanical analyses were made by the hydrometer method virtually as prescribed by the American Society for Testing Materials (1950). Interpretation of particle size from the hydrometer readings was based on the empirical time particle-size relation used by the U.S. Bureau of Reclamation (1951).

The accuracy of mechanical analyses by the hydrometer method is limited by certain inherent factors. It is practically impossible to disaggregate completely fine-grained consolidated or partly consolidated sample material without also breaking some of the component particles. Although maximum possible disaggregation without excessive breakdown of the individual particles was attempted, most of the samples used included a small percentage of aggregates and some clay mineral particles broken during disaggregation. Some materials tend to flocculate even though deflocculating agents are added to the water-sediment mixture. Therefore, hydrometer analyses could not be made on the Mobridge and Virgin Creek Members of the Pierre Shale. Finally, errors of as much as 10 percent may result from the U.S. Bureau of Reclamation's empirical method of correlating time with the sizes of particles in suspension. This method does not ignore the effects of specific gravity, but there is no correction factor for variation in specific gravity of any particular sample from the mean of the samples from which the empirical curve was constructed.

Mechanical analyses of the Pierre Shale show the particle-size distribution in the shale and permit comparison of particle-size distribution between individual members. A cumulative curve of data from the analyses was plotted in figure 15 for five members analyzed: Sharon Springs, Gregory, DeGrey, Verendrye, Elk Butte. The data also were averaged to construct an average cumulative curve to approximate a composition curve for the entire formation in the Fort Randall area.

The curves are divided into three general size classes: greater than 0.074 mm (sand and coarser) (Truesdell and Varnes, 1950); less than 0.074 mm but greater than 2 microns (0.002 mm) (silt); and less than 2 microns

TABLE 2.—*Estimate of relative importance of agents involved in landsliding in the Fort Randall area*

Agent	Mode of operation	Relative effectiveness in Fort Randall area	Principal type of movement
I. Erosion	Creates slopes on which landslides start. Triggers landslides by oversteepening or undercutting slopes.	Probably the major general cause of instability. Also a major trigger action in shale or unconsolidated deposits.	Generally slumps in shale; some soilfalls in unconsolidated materials; rockfalls in Niobrara Formation.
II. Ground water		Various effects of ground water combine to make it the most common trigger agent. All landslides in the area, except those started by direct erosion, occur when ground-water conditions favor instability.	
A. Hydrostatic pressure.	Increase in hydrostatic pressure in joints and fractures, resulting from an influx of ground water, creates a large force acting outward and upward on a potential landslide block. A vertical rise in the hydraulic head of a small fracture connected with a horizontal permeable stratum can create a tremendous increase in the total hydraulic force.	Probably one of the main important trigger causes of landsliding.	Predominantly slumps in shale.
B. Seepage pressure.	As a result of a steep piezometric surface, moving ground water creates enough pressure on soil or rock particles to overcome frictional resistance of the material.	Probably not a major factor while the reservoir is filled, but it is a potentially effective agent in silt and fine sand whenever the reservoir is subject to rapid drawdowns.	Slumps and earthflows in weathered shale and unconsolidated materials.
C. Lubrication	(a) A sudden influx of ground water lubricates normally dry surface material and destroys its coherence.	Effective trigger action in flow type of movement.	Mudflows and earthflows in weathered shale and surficial materials.
	(b) Addition of ground water to a shear surface failure perpetuates movement already started by other processes.	Probably ineffective as a trigger agent in most slumps, but probably very effective in continued movement of active slides.	Predominantly slumps in shale.
D. Weight	(a) An influx of ground water into pores and fractures of previous or unconsolidated materials can produce an appreciable increase in weight.	Significant in flows but probably less significant than lubrication.	Earthflow and mudflows in weathered shale and (or) unconsolidated deposits.
	(b) Fractures in unweathered shale probably cannot hold enough water to have appreciable effect on the weight of a potential slide block.	Probably only very minor importance either as a general cause or as trigger action.	Slumps.
E. Saturation	Saturation of moist weathered shale and unconsolidated material in which surface tension has allowed abnormally steep slopes to develop, destroys their apparent coherence.	As a reservoir fills, saturation of steep sand and gravel banks may trigger landslides.	May be slumps or flows in weathered shale and surficial materials.
F. Piping	Subsurface removal of fine solids or soluble materials by movement of seepage water may weaken material thus eroded.	No present indication that piping is an effective force in this area.	Presumably produces slumps.
III. Loading head of potential slide.	Weight of added material increases shearing stresses in already unstable material to point of failure.	May be a general cause or a trigger action. Effective only locally.	Mostly slumps and slump-earthflows in fresh and altered shale.
IV. Tremors in earth's crust.	Vibrations are known to set potential slides in motion.	Very minor trigger action	Slumps or flows.

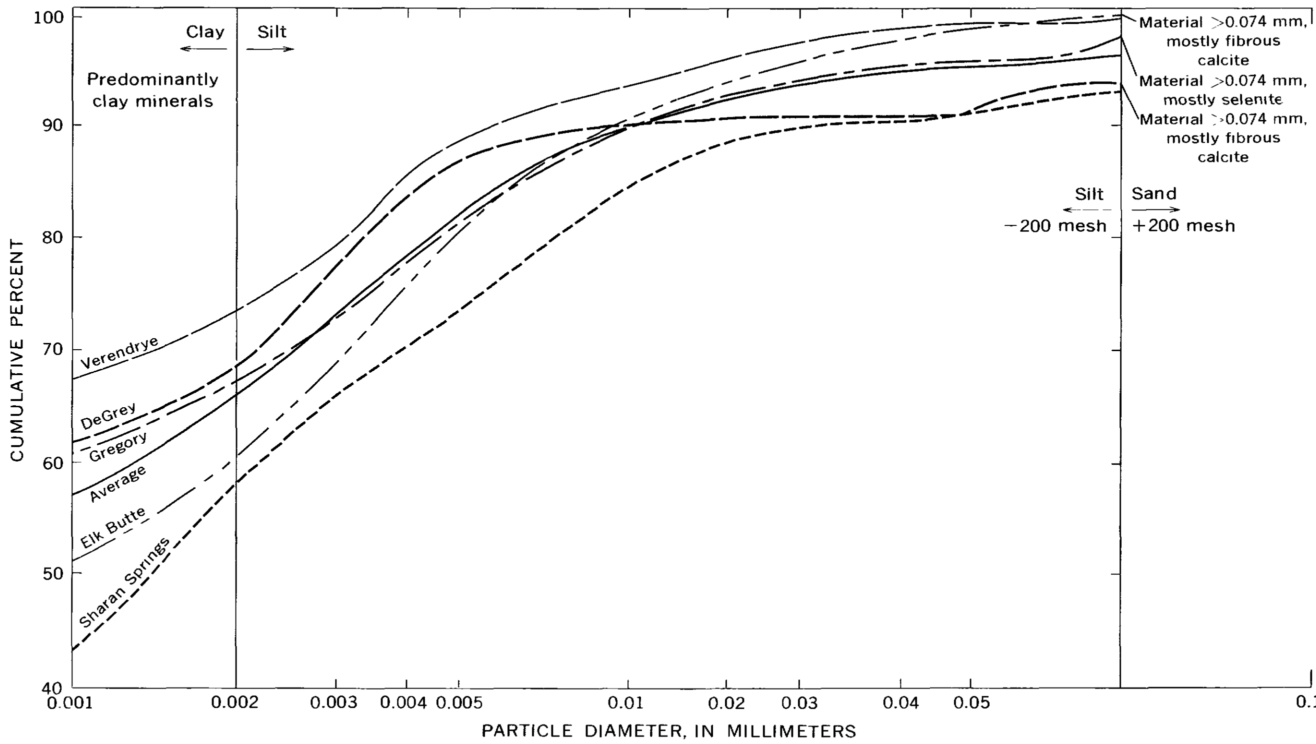


FIGURE 15.—Particle-size distribution in five members of the Pierre Shale.

(clay) (Grim, 1953, p. 1–2). The first division, 0.074 mm, is the dividing point between sieving and the more complicated methods of mechanical analysis, such as hydrometer analysis or elutriation. The division between silt and clay is placed between 2 and 5 microns in various size classifications (Truesdell and Varnes, 1950). The 2-micron division, as given by Grim (1953, p. 1–2), is selected for this report:

> Although there is no sharp universal boundary between the particle size of the clay minerals and nonclay minerals in argillaceous sediments, a large number of analyses have shown that there is a general tendency for the clay minerals to be concentrated in a size less than about 2 microns, or that naturally occurring larger clay-mineral particles break down easily to this size when the clay is slaked in water. Also such analyses have shown that the nonclay minerals usually are not present in particles much smaller than about 1 to 2 microns.

The cumulative curves show the predominantly fine grained character of the formation and similar particle-size distribution among five different members of the Pierre Shale. The average curve indicates that more than 95 percent of the particles are silt size and smaller. About 65 percent of the material has a particle diameter smaller than 2 microns. Particle-size distribution of the individual members shown in figure 15 is surprisingly similar to the average distribution. The material retained on the 200-mesh sieve ranges from less than 1 to about 8 percent of the individual samples. In several samples this material is mostly secondary minerals formed after deposition of the shale, and in all samples the +200-mesh material undoubtedly includes numerous aggregates of grains not separated during disaggregation of the samples. At the 2-micron size there is a range of about 16 percent. In the Sharon Springs Member about 58 percent of the material was finer than 2 microns; in the Verendrye Member about 74 percent was finer than 2 microns. Particle-size variation in the −2-micron sizes seems greater. At 1 micron, the smallest size measured, the range between members has increased to 24 percent, and presumably the curves continue to diverge in the smaller sizes.

The particle-size distribution is a significant factor because it seems to control mineral distribution (fig. 16), which in turn affects the strength of the material. The predominant mineral in the −2-micron size range is montmorillonite clay, but in the +2-micron size range the predominant mineral is quartz. The only exceptions, the Mobridge Member and the Crow Creek Member as described by Crandell (1952), contain large amounts of calcite. Shear strength of consolidated montmorillonite clay is small compared with the shear strength of consolidated quartz grains. The problem essentially is whether the +2-micron material, the −2-micron material, or the combination of both determines shear strength in the shale.

Relative effects of the two particle-size components on shear strength of the shale are controlled by structural arrangement of the grains. The effects of grain structure are considered first in theoretical homogeneous mixtures composed of varying ratios of two grain sizes: large relatively strong spherical grains, and minute relatively weak grains. The theoretical mixtures can then be compared with the Pierre Shale.

Two extreme types of particle or grain structure can develop in a homogeneous mixture of coarse grains and fine grains depending on the relative proportions of the two components. If the volume of the interstices between the coarse grains is the same as, or greater than, the volume of the fine grains, the coarse grains can develop a stable structural network. If the volume of fines is greater than the volume of the interstices, the fine grains can form a matrix in which coarse grains are isolated from each other. In the first case, shear strength of the mass is a function of the shear strength of the coarse grains and also of the friction between them. In the second case, shear strength of the mass is controlled by the shear strength of the fine-grained matrix.

The changes in shear strength from material in which the coarse grains predominate to that in which fine grains predominate are gradual. If the volume of fine grains only slightly exceeds the volume of the interstices between coarse grains, much of the shear strength of the coarse grains is retained. Even after the coarse grains are completely isolated from each other, they may effectively increase friction along shear surfaces.

The transition point at which the fine grains change from filler to matrix is controlled by the packing of the coarse grains. In systematically packed spheres of uniform size, the porosity ranges from 25.95 to 47.64 percent (Graton and Fraser, 1935, table 2). The transition point in the theoretical mixture thus should be about 26–48 percent fines, if the fines fill all the voids. In random packed spheres, if one asumes that a stable network of coarse grains exists, the transition point probably will fall in the same range as in the theoretical mixture.

Comparison of the +2-micron to the −2-micron particle- or grain-size mixture in the Pierre Shale with the theoretical mixture shows their general behavior to be similar.

The +2-micron fraction of the Pierre is predominantly quartz grains, which commonly are equidimensional if not spherical. Mica, one of the minor coarse-grained constituents, is platy and may increase porosity of the packed coarse grains. The fine grains are not without dimension as in the theoretical mixture, but their average size is small compared to the coarse grains. On the average cumulative curve (fig. 15) about 15 percent of the −2-micron particles occur in the 1- to 2-micron size range; about 85 percent of the fine-grained component has a diameter of less than half the smallest diameter in the coarse-grained component. The shale, moreover, consists of particle sizes undoubtedly mixed heterogeneously although they appear to be homogeneous when inspected through a binocular microscope.

These variations in the particle-size mixture of the shale should reduce the proportion of fines necessary to reach the transition point at which the coarse grains no longer form a continuous structural network. The closer packing should also counteract the tendency of the mica to increase porosity. The −2-micron particles in the shale cannot completely fill all voids between coarse grains, because the shale has some porosity; the volume of fine particles necessary to separate the coarse grains thus is less than the total volume of voids.

Comparison of the Pierre Shale and the theoretical mixture implies that −2-micron particles in the Pierre Shale become a matrix for the +2-micron grains at some proportion of fines less than 50 percent. The percentages of −2-micron particles in the five members of the Pierre Shale shown in figure 15 ranges from about 58 to 73. It seems valid, therefore, to assume that the −2-micron size fraction of the shale is a matrix for the coarser grains and that shear strength is determined by this matrix. Because the +2-micron fraction probably has little effect on the shear strength of the shale, it can be ignored in the shale behavior studies.

MINERAL COMPOSITION

Although the results therefrom are approximate, X-ray diffraction is the only feasible method of quantitative mineral identification for the dominantly fine grained Pierre Shale. Identifications for each member except the Crow Creek were made by Dorothy Carroll, J. C. Hathaway, C. J. Parker, and W. W. Brannock, U.S. Geological Survey, August 1955, using the following procedure:

> A portion of each sample was disaggregated and dispersed in distilled water with sodium tetraphosphate added as a dispersing agent. The silt (2-62 microns) and clay (less than 2 microns) fractions were separated by repeated centrifuging and decanting. Excess water was removed from the clay suspension by porcelain filter candles under vacuum, and the clay was Ca^{+2} saturated by passing the concentrated suspension through a Ca ion-exchange resin column. Oriented aggregates were prepared by pipetting portions of the concentrated suspensions on glass slides and allowing the water to evaporate at room temperature.

X-ray diffractometer patterns were made on each sample as follows:

Clay fraction

1. Oriented aggregate, untreated.
2. Oriented aggregate, ethylene glycol treated.
3. Oriented aggregate, heated to 400° C.
4. Oriented aggregate, heated to 500° C.
5. Randomly oriented powder.

Silt fraction

6. Randomly oriented powder.

Quantitative estimates are based on the intensity of the lines recorded by the X-ray diffractometer and are given as parts in 10. Inasmuch as many factors in addition to quantity of a mineral affect diffraction intensity, these estimates are not intended to give more than a very general indication of the relative amounts of the various minerals present.

Throughout the rest of this report, 74 microns is used for the upper limit of silt size, rather than 62 microns as given here. An average of less than 1 percent of the shale particles is in the 62- to 74-micron size range; therefore, the approximate quantitative mineral data should be equally valid for the larger silt-size limit.

Mineral compositions of the various shale members are shown graphically in figure 16; geographic locations of the samples are given in table 3. The Virgin Creek, Verendrye, and DeGrey Members are each represented by a set of samples, an auger sample from depths of 25 feet or less, and a surface grab sample. An additional auger sample was collected from a bentonite bed in the Sharon Springs Member to check possible variations between it and the enclosed shale. Sets of samples from two horizons were used for the Elk Butte and Mobridge Members. The grab samples were collected to determine the effects of near-surface weathering on mineral composition. Inasmuch as auger samples were from comparatively shallow depths, they are all presumed to be at least slightly weathered.

TABLE 3.—*Localities of shale samples shown in figure 16*

Sample Locality (fig. 16)	County	Sec.	T.N.	R.W.
1	Gregory	SE¼SW¼ 13	95	66
2	do	SW¼SE¼ 13	95	66
3	Lyman	SW¼SW¼ 21	104	72
4	do	SW¼SE¼ 22	104	72
5	do	NW¼SE¼ 22	104	72
6	do	SE¼SE¼ 22	104	72
7	do	SW¼SE¼ 14	104	72

The mineral composition of members of the Pierre Shale (fig. 16) is easily summarized. The clay-size material is predominantly montmorillonite, ranging from five parts in 10 in the Gregory Member to nine parts in 10 in the DeGrey, Verendrye, and Virgin Creek Members. Except in the Mobridge Member, the major mineral in the silt-size material is quartz, which is three to five parts in 10 of the silt-size category. In the Mobridge Member montmorillonite does not exceed four parts in 10 of the clay-size component, and quartz represents only one or two parts in 10 of the silt-size component. Calcite in the Mobridge Member, although only two to five parts in 10 in a single grain-size component, is an abundant mineral in both size components. Mineral composition of the Crow Creek Member is similar to that of the Mobridge Member, and calcium carbonate (presumably calcite) is the most abundant mineral (Crandell, 1952).

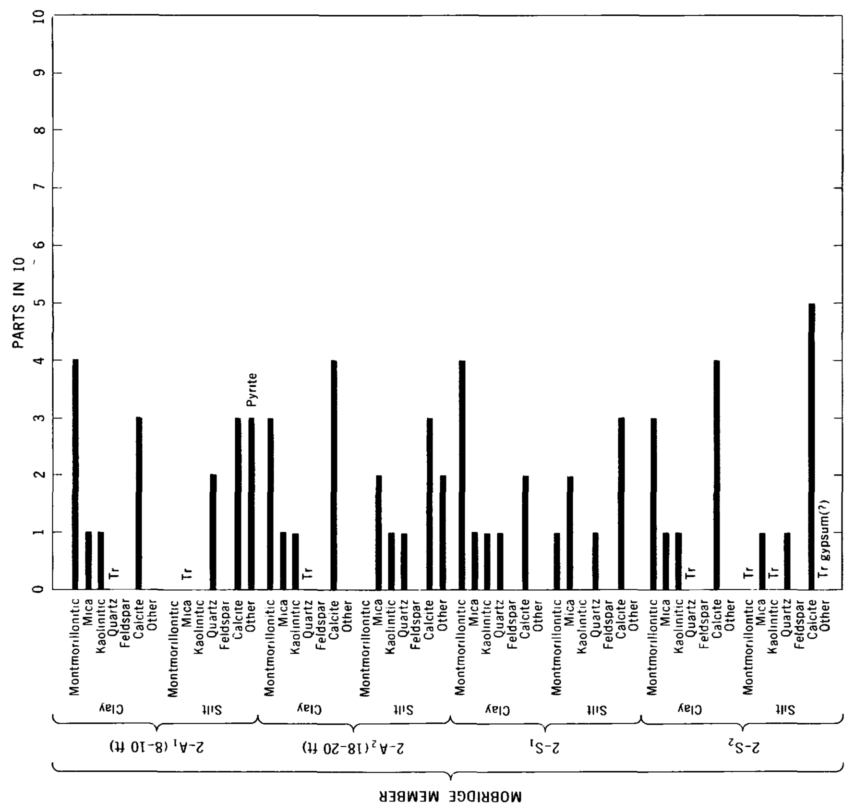
One of the most interesting features of the mineralogy of the Pierre Shale is the contrast in mineral composition between clay-size and silt-size material. Calcite in the Mobridge Member is the only nonclay mineral that exceeds the proportion of one part in 10 in the clay-size portions of the samples shown in figure 16. In the silt-size portions of the same samples no clay mineral exceeds two parts in 10, and in only four of the 19 samples does a clay mineral comprise more than one part in 10.

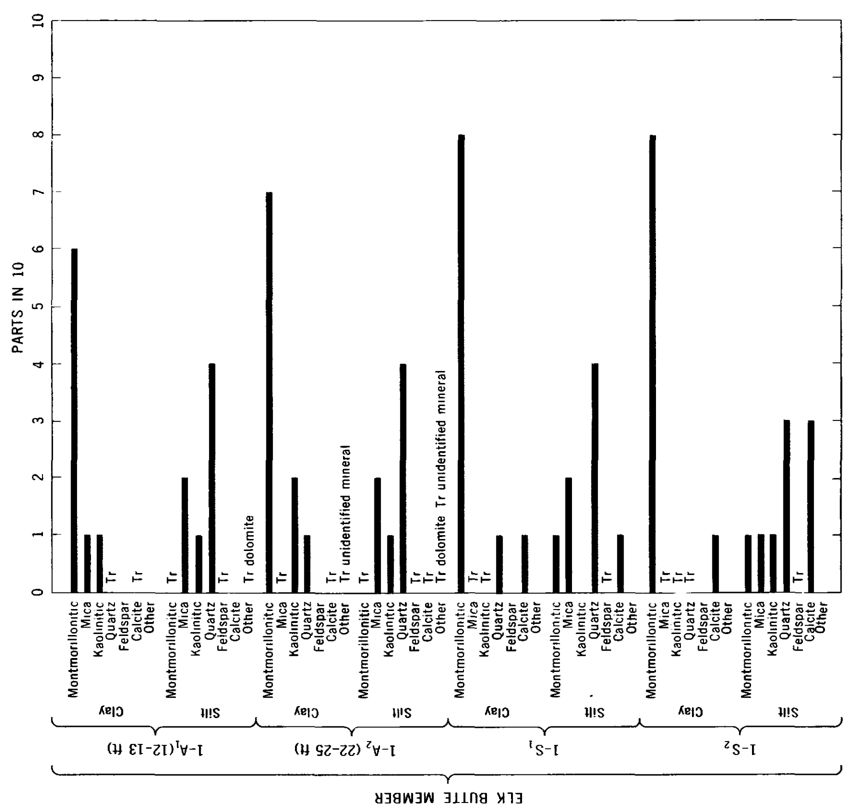
The data in figure 16 imply that near-surface weathering has no profound universal effect on the Pierre Shale. Apparently the effects of weathering are influenced more by local conditions than by any regional control.

The available data are insufficient to show the lateral and vertical ranges in composition and weathering effects. They suggest, however, that there is little variation within a given member. The two sets of samples from the Elk Butte and Mobridge Members show little variation over a vertical distance of 10 feet, but they cannot be considered representative of the total thicknes of each member. The Mobridge and Elk Butte samples show that minor variations (one part in 10) in mineral composition between surface and subsurface samples are not necessarily representative of the weathering conditions. Samples may show a small change in the amount present of a mineral between the surface and one subsurface horizon, but another horizon a few feet above or below the first horizon may show no difference in the amount present of the same mineral.

MONTMORILLONITE

Montmorillonite clay in the Pierre Shale apparently is responsible for the susceptibility of the shale to landslides. Interpretation of the data from mechanical analyses indicates that shear strength of the shale is determined by the shear strength of its clay-size portion, and the clay-size portion of most of the shale is dominated by montmorillonite. Slope stability of individual members, therefore, should depend on the pro-


PARTS IN 10
Pyrite
Tr gypsum(?)
Montmorillonitic
Mica
Kaolinitic
Quartz
Feldspar
Calcite
Other
Clay
Silt
2-A1 (8-10 ft)
2-A2 (18-20 ft)
2-S1
2-S2
MOBRIDGE MEMBER


PARTS IN 10
Tr dolomite
Tr unidentified mineral
Tr dolomite Tr unidentified mineral
Montmorillonitic
Mica
Kaolinitic
Quartz
Feldspar
Calcite
Other
Clay
Silt
1-A1 (12-13 ft)
1-A2 (22-25 ft)
1-S1
1-S2
ELK BUTTE MEMBER

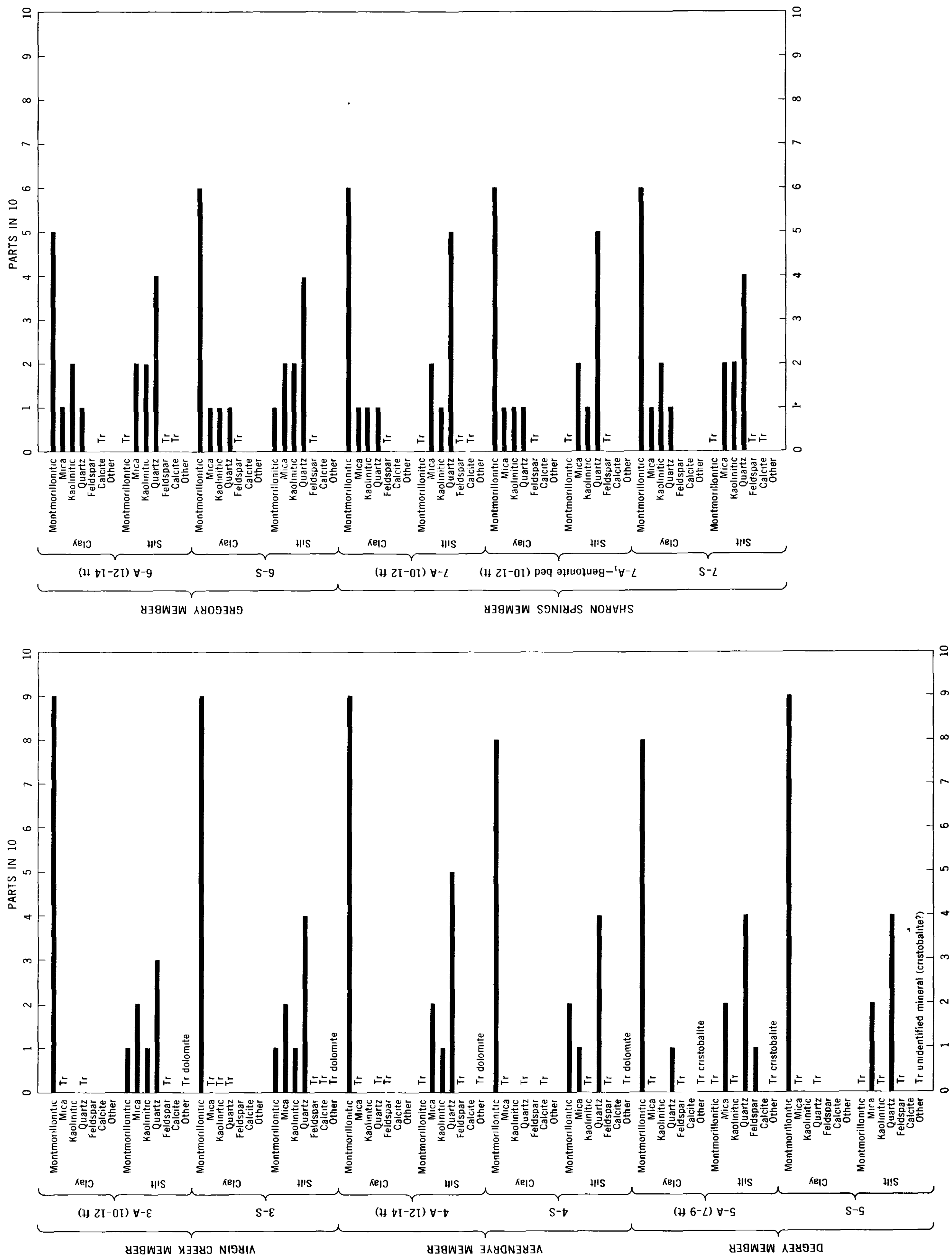


FIGURE 16.—Mineral composition of Pierre Shale. Numbers indicate sample localities. *A*, Auger sample; *S*, surface sample; *Tr*, trace amount. Sample localities described in table 3.

portion of montmorillonite in the clay-size components. In a general way this is true: the Mobridge Member, which contains noticeably less montmorillonite than the other members (fig. 16), commonly forms and maintains steeper slopes. Though susceptible to landsliding, it appears to be more stable than the other members. The following description of montmorillonite is adapted from Grim (1953).

The basic structure of montmorillonite is a series of silicate layers perpendicular to the *c*-axis. Unbalanced negative charges are on the surfaces of all silicate layers attracting various exchangeable cations, which are a necessary part of the montmorillonite structure.

The silicate layers also attract water molecules and other polar molecules to their interlayer surfaces. The thickness of the adsorbed water between silicate layers is controlled by the amount of water available and by the nature of the exchangeable cations held by the clay. The bond between adsorbed water and the silicate layers is not completely understood, but water molecules in direct contact with the silicate layer behave as a solid and apparently are held rigidly in an oriented position. Successive layers of oriented water attach themselves to the first molecular layer, but in each succeeding layer the orientation and degree of bonding are less and the water is more fluid.

If only a little water is adsorbed, the montmorillonite is held in a virtually solid state, and the clay has relatively high shear strength. If a large amount of water is adsorbed there is semifluid or fluid water held between the silicate layers and this water lubricates and allows movement between layers. The shear strength thus is reduced to almost nothing if sufficient water enters between the silicate layers.

As the addition of water reduces the shear strength of montmorillonite, it also reduces shear strength of the shale in almost direct proportion.

Exchange of cations at a given water content can also reduce shear strength in montmorillonite. As an example, replacement of cations that promote a thick layer of adsorbed water by cations that do not, can release some of the adsorbed water, which then becomes free water between silicate layers. If this free water is not expelled immediately by compaction of the clay, the shear strength of the clay is drastically reduced.

The stability of clay may also be reduced when cations that can adsorb only a thin layer of water are replaced by cations that can adsorb a thicker layer. In this case, if extra water becomes available to increase the thickness of the adsorbed water, the water in the center of the water layer is farther from the orienting effect of the silicate layer and therefore is not held as tightly. This water, as a result, is more fluid and is a better lubricant.

At present, the effect of cation exchange on the stability of Pierre Shale slopes cannot be measured. Studies—by Dorothy Carroll, J. C.Hathaway, C. J. Parker, and W. W. Brannock, U.S. Geological Survey, August 1955—of the exchangeable cations and ion exchange capacities in Pierre Shale samples indicated that near-surface weathering causes a general increase in exchange capacity, with calcium becoming the major exchangeable cation. These data, however, are from samples from horizons too shallow to have any major application to landslides. Data furnished by the Omaha District of the U.S. Army Corps of Engineers (D. K. Knight, oral commun.) show no significant contrast between exchangeable cations in clay from unweathered shale at Oahe damsite, about 100 miles upstream from the Fort Randall Reservoir, and exchangeable ions in the clay from shale samples along the Fort Randall Reservoir.

REPRESENTATIVE STUDY AREAS

The two major objectives of the Fort Randall Reservoir landslide investigations were (1) to determine the effect of the reservoir on slope stability along its shores, and (2) to determine the relation between different materials and the potential for landsliding. Both slope stability appraisals and geologic mapping are needed to achieve these objectives.

Study of the present slopes also is a basis for comparison of current to former landslide activity along the walls of the Missouri River trench. Correlation between slope stability and geology, supplemented by shale analyses and measurements of slope angles, indicates the susceptibility of individual geologic units to landsliding and makes possible estimates of the relation between geology, slope stability, and natural slope angles.

This approach to the study of landslides was tested in five areas representative of the reservoir environments.

METHOD OF SELECTION

There were two main factors to consider in selecting the areas for slope stability investigations. First, adequate large-scale base maps had to be available. Second, the areas had to be as representative of conditions along the entire reservoir as possible. At least one area near but outside the limits of the reservoir and its effects was needed as a control or reference area.

The areas selected are covered by sheets 23, 26, 30, 39, and 42 of the Missouri River Survey, Gavins Point near Yankton, S. Dak., to Stanton, N. Dak., topographic map series made in 1947 by the Omaha District, U.S. Army Corps of Engineers. These maps were available at scales of 1:12,000 and 1:24,000, and each covers an average of 25 square miles along the Missouri River trench. The maps have accurate photogrammetric planimetry of nearly the entire Missouri River trench. Topography is portrayed by contours at 10-foot intervals along the lower part of the trench from the river level to an altitude of 1,400 feet. For this study, contours at 20-foot intervals above 1,400 feet were completed, by the U.S. Geological Survey, on special manuscript sheets. The areas shown on map sheets 23, 26, 30, 39, and 42 are termed areas 1, 2, 3, 4, and 5, respectively.

Area 1, just below the Fort Randall Dam (pl. 1), is the reference area, which has an environment as similar as possible to that above the dam. Tributary valleys between the dam and area 1 should effectively isolate most of the control area from any near-surface ground-water changes caused by the reservoir.

A complete geologic map was made only in area 2 (pl. 3), which includes the entire Pierre Shale section as well as many old and new landslides. The area was chosen because it is an ideal locality for correlating slope stability and materials. In the other areas, only the geology that is to be permanently submerged was mapped as part of this project.

Areas 2 and 3 (pls. 1, 3), near the dam, were chosen to provide a record of slope stability changes in the part of the trench that undergoes greatest inundation. These areas, moreover, are sufficiently close to act as a check on each other. The general conditions controlling slope stability should have a similar effect in both areas, but any local factor presumably would apply in only one. Areas 4 and 5 (pl. 1), in the upper part of the reservoir, include parts of the trench that will have comparatively little inundation. These areas also are geographically close enough to provide checks on each other.

SCOPE OF INVESTIGATIONS

Investigations in the selected areas are in two categories: slope stability appraisals and geologic mapping.

SLOPE STABILITY APPRAISAL

Slope stability appraisals were made by examining the terrain in the field and plotting the landslides on vertical aerial photographs. This information was transferred to the base maps with the aid of a vertical sketchmaster.

The first purpose of these appraisals was to observe changes caused by formation of the Fort Randall Reservoir. The studies described in this report record conditions at the time the dam was built. Reappraisals can be made in the future to determine the slope stability changes occurring after the reservoir was filled and in operation. The second and more immediately important purpose of the slope stability appraisals is to supply data to help us better understand landslides and factors that control them.

The first phase of the appraisals was to set up a satisfactory group of slope stability conditions. There are no simple, absolute criteria, and therefore the classification developed was generalized. It includes the major classes of stability and is designed for reconnaissance-type observations. The classification, therefore, does not use core drilling to locate slide surfaces or firmly scheduled observation of control points to determine accurately the activity of individual landslides. The slope stability appraisal is only approximate, but examinations of the same area by more than one person indicate that the results in general are reproducible.

The slope stability classification is divided into two basic categories: (1) areas apparently always stable, and (2) areas that are or were subject to landsliding. The second category is further divided into a series of four classes of slide activity at the time of observation. The end points of this series are active landslides and stabilized landslides, with two gradations between —one applying to landslides that appear stable but show signs of very recent activity, and the other applying to landslides that were stable for a number of years but that have been recently reactivated.

STABLE GROUND

Stable ground includes all material that appears never to have been subject to landsliding. The only exceptions are thin patchy deposits of stable loess, alluvium, and colluvium that overlie but do not completely mask old landslide material. For such deposits the stability classification depends on the underlying material.

Stable material consists of two basic types: (1) material that was deposited independently of the Missouri River trench and (2) material deposited within the trench and its tributary valleys.

The stable material underlying the trench comprises the deposits into which the trench and its tributary valleys were cut. Inasmuch as these same deposits are also involved in most landslides, the distinction between stable material and very old stabilized landslides often is difficult. Material underlying uplands and gentle slopes (fig. 17) without steep scarps probably is stable. Evidence of stable ground is hard to find on steeper slopes, and undoubtedly some small stable areas are

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FIGURE 17.—Area of Pierre Shale slopes considered stable. View southwest into N½ sec. 4, T. 103 N., R. 72 W., Lyman County, S. Dak. (area 5 west of the Missouri River). Photographed August 24, 1956.

not recognized in large areas of old stabilized landslides.

Stable material deposited within the trench and its tributary valleys can generally be recognized easily. Where it is sufficiently thick and extensive to mask the underlying terrain, it forms relatively smooth surfaces, generally on flood plains or terrace remnants. Such deposits may give a false impression of stable conditions because they may cover older deposits with long complex histories of landslide activity. In some areas landslides have obviously been covered by stable natural fill material, but it is impossible to determine the conditions beneath the fill (fig. 18). Consequently, it is difficult to draw an accurate contact between stable ground and landslide areas where a stable fill feathers out against old landslide topography.

STABILIZED LANDSLIDES

Areas of stabilized landslides include landslide materials that show no evidence of recent activity. The surfaces of weakness are still present and the shearing strength is probably much less than in undisturbed material; therefore, changes in the factors controlling stability could result in renewed movement.

Surface indications of stabilized landslides range from slightly hummocky irregular topography to individual landslides with relatively fresh scarps and easily distinguished boundaries. The slump shown in figure 19 is typical of recently stabilized landslides that are still obvious entities only slightly modified by erosion and deposition. Although the scarp and the outlines of the slump block are still clearly discernible, deposition of loess and colluvium is beginning to obscure the limits of the block, and erosion is cutting gullies into the upthrust toe.

Identification of old slides is increasingly difficult as time and geologic processes continue to alter the slides. The characteristic older stabilized landslide terrain in the Pierre Shale is shown in figure 20. It is obvious from the photograph that any stable ground below the ridges cannot easily be distinguished from areas of stabilized landslides.

The hummocky topography shown in the left background of figure 20 is composed of old scarps and slump blocks that can still be recognized. Sufficient water from runoff and seepage has been trapped at the break in slope between some scarps and slump blocks to support growth of trees and shrubs. The backward rotation of the slump blocks has produced flatter slopes than the surrounding terrain so that they resemble terrace remnants. In most places, however, the two features can be easily distinguished because terrace remnants along a valley occur at about the same elevation, whereas the levels of old slump blocks rarely coincide.

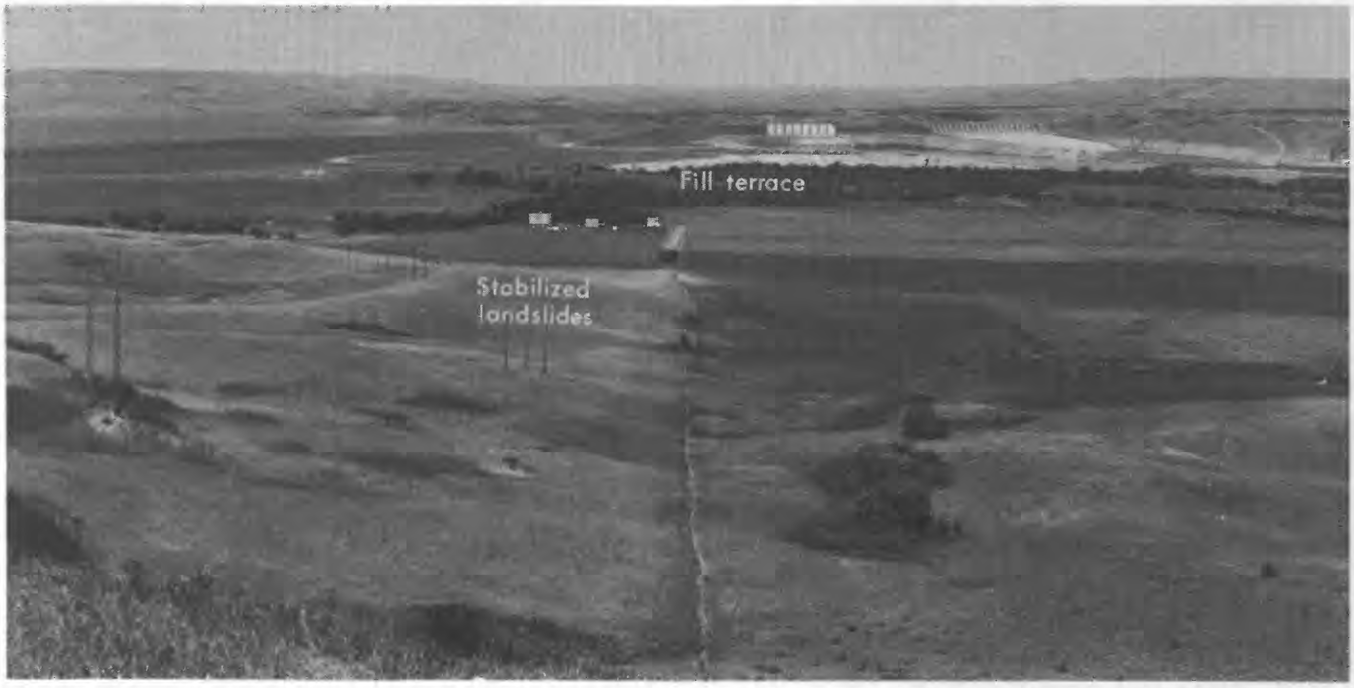

FIGURE 18.—A low-level stable fill terrace (center background) at the base of slopes composed of old stabilized landslides (foreground). The upper limits of the terrace seem to grade into the landslide area, which strongly suggests that the terrace fill was deposited on and masks old landslide topography. View north toward Fort Randall Dam from center of boundary between SW¼ and SE¼ sec. 20, T. 95 N., R. 65 W., Gregory County, S. Dak. (west side of area 1). Photographed August 16, 1956.

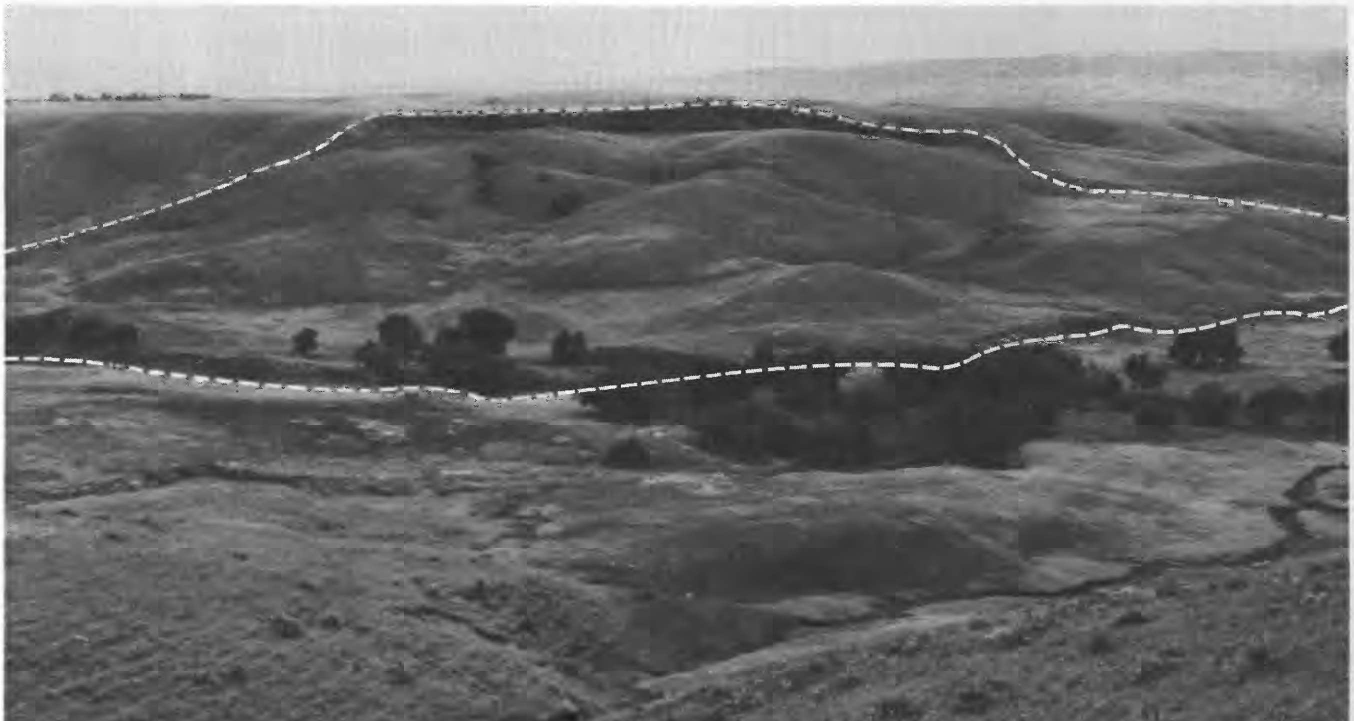

FIGURE 19.—Recently stabilized slump. This slump is still a distinct entity, but erosion and deposition locally obscure its limits. SW¼SE¼ sec. 6, T. 98 N., R. 69 W., Charles Mix County, S. Dak. Photographed August 19, 1956.

FIGURE 20.—Stabilized landslide terrain. All of the area, except some of the ridges along the skyline, probably slid at some time. The evidence of landsliding ranges from hummocky terrain to discrete slump blocks. E½ sec. 20, T. 97 N., R. 68 W., Gregory County, S. Dak. (area 3 west of the Missouri River). Photographed August 15, 1956.

Continual erosion and development of well-integrated drainage, coupled with deposition of loess and colluvium, modifies the landforms and reduces relief until the stabilized landslide area looks like the subdued hummocky slopes shown in the right-central part of figure 20. Surface evidence of landslides may be completely lost in the final stage, illustrated in figure 21 which shows loess deposition concealing a small slump in the Pierre Shale. This slump would never have been recognized if it had not been exposed by later erosion.

RECENTLY ACTIVE LANDSLIDES

Landslides showing recent movement are classified as recently active even though they were apparently inactive at the time of study. There is no certain way to judge whether they are actually stable or only temporarily quiet. Landslides in this category can become active without a major change in their stability conditions.

FIGURE 21.—Exposed slump along the Fort Randall Reservoir shore. The slump in the Pierre Shale was completely hidden by loess deposition. Erosion along the shore revealed the loess-shale relation. Charles Mix County, S. Dak. Photographed September 27, 1954.

Although the slump shown in figure 22 is recent, apparently it has not moved within the past few years. The slump block is surrounded by fresh-looking fractures, and at first glance it appears to be active, but cattle trails crossing the fractures without displacement and the edges of fractures not being sharp indicate that it is not active.

REACTIVATED LANDSLIDES

Reactivated landslides have started to move again after a period of stability. The term is not used for active landslides that include a part of one or more stabilized slides but is confined to old landslides moving again as their original entities. The significance of a reactivated landslide is that it implies redevelopment of the stress conditions that caused the original movement.

The slump-earthflow shown in figure 23 is a typical reactivated landslide. It apparently was a subsidiary part of a larger slump, part of which can be seen on the left of the reactivated area. The limits of the reactivated area follow a topographic break that outlines the original slump unit. Renewed movement of the subsidiary slump block has not affected the major stabilized landslide beside it. This reactivation may be complete within itself, or it may be a forewarning of more extensive reactivation.

ACTIVE LANDSLIDES

All landslides that show no signs of stability are classified as active. This negative criterion is necessary because actual movement was observed in very few

FIGURE 22.—A recently active slump. The fresh appearance of the fractures implies that this landslide moved recently, but close examination shows no present activity. The landslide may be permanently stable or temporarily quiet. SW¼SE¼ sec. 19, T. 103 N., R. 71 W., Brule County, S. Dak. (area 5 east of the Fort Randall Reservoir). Photographed August 23, 1956.

landslides along the Fort Randall Reservoir. Either landslides move too slowly for direct observation or else the movement is in periodic surges with only minor activity between. All unstable landslides are included in this category whether they are in formerly stable areas or whether they represent renewed activity in stabilized landslide areas.

Criteria suggestive of active sliding include fresh bare fractures and scarps, slickensides on exposed slide surfaces, fresh jagged blocks of landslide material, and dead or dying vegetation. Figure 24, a photograph of the head of an active landslide, shows all these features except slickensides.

Some landslides considered active on the basis of

FIGURE 23.—A slump-earthflow caused by reactivation of an old stabilized slump block. The slump-earthflow follows the outline of an old slump block apparently subsidiary to the stabilized landslide on the left side of the active area. NW¼NW¼ sec. 36, T. 95 N., R. 65 W., Charles Mix County, S. Dak. (east side of area 1). Photographed August 20, 1956.

FIGURE 24.—The head of an active landslide on the southwest bank of the Fort Randall Reservoir, about 14 miles upstream from the Fort Randall Dam, Gregory County, S. Dak. Photographed September 26, 1955.

appearance may be stable. Every landslide, after its final movement, passes through a stage when the signs of activity are still very fresh. By observation alone it is difficult to distinguish between a newly stabilized landslide and one that is stil active.

GEOLOGIC MAPPING

The geologic map of area 2 (pl. 3) is intended to show as accurately as possible the stable positions of the various geologic units in the Fort Randall reservoir area. All exposed and apparently stable geologic contacts were plotted on 1:20,000-scale vertical aerial photographs, then transferred to the map with the aid of a vertical sketchmaster.

Throughout much of the area, geologic contacts are hidden by loess, colluvium, and landslide deposits; however, sufficient reliable information was available to show that the Cretaceous beds are horizontal and that the thicknesses of the map units in the Pierre Shale are virtually constant throughout the area. Contacts in the Cretaceous units, therefore, are presumed to follow contours between reliable exposures. Contacts of the post-Cretaceous deposits are not necessarily horizontal, and their locations can only be approximated in areas of poor exposures.

The descriptions of the geologic units (p. 4) as they occur in the lower part of the reservoir apply to plate 3, with the few exceptions that follow.

Pierre Shale, Crow Creek Member.—The Crow Creek Member is not identified in area 2, but it may be present beneath either landslides or surficial deposits. Possible outcrops of the Crow Creek would be displaced downward by landslides and therefore would not be distinguishable from the marl facies of the underlying Gregory Member.

Pierre Shale, DeGrey and Verendrye Members.—The DeGrey and Verendrye Members have been mapped as a single unit in area 2. The lithologic similarity between the two members and the scarcity of stable outcrops make it impossible to draw any contact between them.

Pierre Shale, thicknesses of map units.—Thicknesses of the Pierre Shale map units, except for those units having unconformable contacts at the bottom and top of the section, are nearly constant throughout area 2. Thicknesses of the map units are as follows:

	Feet
Elk Butte Member	[1] ±140
Mobridge Member	100
Virgin Creek Member	50
DeGrey and Verendrye Members	100
Gregory Member	30
Sharon Springs Member	[2] +40

[1] Unconformity at top.
[2] Unconformity at base.

Slow-draining terrace alluvium.—Slow-draining terrace alluvium is distinguished from other slow-draining alluvium simply to aid interpretation of the geologic map. Physical characteristics of the two materials are similar, but the terrace alluvium was deposited when the altitude of the Missouri River was considerably higher than the present channel.

DESCRIPTIONS OF INDIVIDUAL AREAS

The five study areas have similar geologic settings. The Missouri River has eroded its valley, the Missouri River trench, through the flat-lying Pierre Shale and into the underlying Niobrara Formation. The valley walls in the lower part of the reservoir contain the entire Pierre Shale section and on some of the uplands the shale is capped by Ogallala Formation. In the upper part of the reservoir post-Cretaceous erosion has removed some of the upper members of the Pierre Shale from the vicinity of the river valley. Water, ice, and wind have by repeated cycles of erosion and deposition during Quaternary time modified the basic geologic setting.

AREA 1

Area 1 (pl. 1), immediately downstream from the Fort Randall Reservoir, extends from lat 42°59′ to 43°03′ N. and from long 98°27′ to 98°35′ W. The right (southwest) abutment of the Fort Randall Dam lies in the northwest corner and the Missouri River trench cuts diagonally southeastward through the central part of the map. All the area east of the Missouri River is in Charles Mix County, S. Dak. Approximately the southern one-fifth of the area on the west side of the river is in Boyd County, Nebr.; the rest is in Gregory County, S. Dak.

This study area provides a reference or control not affected by the reservoir in a geologic and topographic setting similar to the other study areas.

The base of the Missouri River trench is a steep-walled alluvium-filled valley about a mile wide incised into an earlier floor of the trench. The river occupies one-third to one-half of this inner valley at an altitude of about 1,240 feet. On the east side of the river the inner valley is walled by 50- to 100-foot-high bluffs. The older upper slopes of the trench wall rise steeply from the bluffs to a gently undulating surface 1,500–1,550 feet in altitude. Farther from the trench in the northeastern part of the area, this upland gives way to rolling hills as much as 1,800 feet in altitude.

West of the river, bluffs 80–100 feet high form the southern part of the inner valley wall. The height of these bluffs decreases to the northwest until they merge with the gentle slopes bordering the junction of Randall Creek and the Missouri River. Away from the inner trench walls, moderately sloping hills rise to an altitude of about 1,600 feet. The slopes then steepen abruptly and rise until they reach an upland that averages about 1,700 feet in altitude.

Geologic setting.—The entire pre-Quaternary sequence from the Niobrara Formation to the Ogallala Formation is exposed in the trench walls; however, erosion has removed the Ogallala Formation and the top of the Pierre Shale from all but the northern part of the area east of the Missouri River. Alluvium covers the flood plains of the river and its tributaries, and loess forms a patchy cover on the uplands. An ice sheet that advanced at least as far as the Missouri River after the beginning of downcutting of the river has left till deposits along the east wall of the trench. Deposits of fine and coarse (slow-draining and fast-draining, respectively) alluvium cap terrace remnants west of the river. In the northern part of the area, the terrace alluvium is thick enough to mask the Niobrara Formation and the lower members of the Pierre Shale. Southward it thins to isolated patches capping terrace remnants cut in the Cretaceous rocks.

Stability conditions.—Except for the limestone bluffs of the Niobrara Formation along the river, most of the steeper slopes in area 1 (pl. 1) have slid at some time or other. There are few exceptions to this steep-slope–landslide-terrain relation on the west side of the river. Some areas of relatively steep slopes east of the river, however, show no signs of landslides. An explanation for the seemingly anomalous stability of the slopes is that sliding has occurred, but that its surface evidence now is completely lost. The absence of landslides on some steep hills in the north part of area 1 may be due to deposits of till, which is more stable than Pierre Shale. Other slopes, such as those along Seven Mile Creek, may be stable because the adjustment of internal stresses in the underlying materials was able to keep pace with the changes wrought by erosion.

Almost all of the landslide terrain consists of stabilized landslides that long ago were reduced by erosion to patches of hummocky ground. Many of the individual landslides, one would infer from their remnants, must have been enormous.

Active, reactivated, and recently active landslides each represent a small proportion of area 1 (pl. 1). In most areas these landslides are randomly distributed, but they are concentrated in several localities, such as the W½ sec. 19 and the NW¼ sec. 31, T. 95 N., R. 64 W.

Two differences between the stabilized landslides and the active and recently active landslides indicate that the modern slopes are more stable than the old slopes: the presently active slides are considerably smaller than the old ones; current activity is not on the large and formerly unstable slopes of the Missouri River trench or major tributaries but in the small gullies of minor tributaries where streams have locally cut very steep banks or have removed the supporting toe of an old slide.

AREA 2

Area 2 (pl. 3), about 5 miles upstream from the Fort Randall Dam, extends from lat 43°03′ to 43°07′ N., and from long 98°39′ to 98°47′ W. Land east of the Missouri River is in Charles Mix County, S. Dak.; land west of the river is in Gregory County, S. Dak.

The Missouri River trench in area 2, as in area 1, has a steep-walled inner valley incised in an older trench bottom. This inner valley trends southeastward from the area's north boundary; near the center of the area it curves about 45° and heads due east. The inner valley increases in width from about 1 mile at the north boundary to about 1 1/3 miles at the east boundary. Before the construction of the dam, the Missouri River occupied only one-third to one-half of the inner valley, but water in the reservoir now completely fills it.

The terrain northeast of the river is rather subdued, although it contains some steep slopes. The northeast wall of the inner valley rises steeply from the valley bottom (altitude about 1,250 ft) or, locally, from narrow low terraces along the river to a broad terrace 75–130 feet high. The terrace is about three-fourths mile wide near the north edge of the area but broadens eastward to almost 1¼ miles. It slopes gently upward from 1,350 feet altitude along the river bluffs to an average of 1,480 feet altitude at its outer margin. In the northeast part the terrace grades into moderately sloping hills that rise to an upland 1,720–1,740 feet in altitude.

Southwest of the river (fig. 25) the terrain is much more rugged. At altitudes of 1,300–1,400 feet the steep walls of the inner valley merge with a belt of hills and ridges whose slopes are moderately gentle to moderately steep near the river but rise precipitously to the south and west to an upland 1,780–1,890 feet in altitude. The most striking feature of the slopes between the inner

FIGURE 25.—Terrain on the southwest side of area 2 showing hummocky terrain characteristic of old landslides. View northwest from SW¼SW¼ sec. 1, T. 95 N., R. 67 W., Gregory County, S. Dak. Photographed August 27, 1956.

valley and the upland is the erratic hummocky appearance characteristic of old landslide terrain.

Geologic setting.—The geologic setting of area 2 (pl. 3) differs only in detail from that of the other study areas.

The entire pre-Quaternary sequence from the Ogallala Formation to the Niobrara Formation is present in the southwest trench wall. Loess caps the uplands and occurs in small patches below the upland surface. There are also many pockets of loess, too small and discontinuous to be mappable. Slow-draining alluvium in the Missouri River flood plain forms the bottom of the trench and extends short distances up some of the larger tributary valleys. Farther up the valleys, and in many of the smaller valleys and depressions, the alluvium is mixed with colluvium and loess and is mapped as alluvium, colluvium, and loess.

Northeast of the river, the Ogallala Formation is missing and the middle part of the Pierre Shale is obscured by overlying Quaternary deposits. The Quaternary deposits in turn are more varied and more widespread than they are southwest of the river. Loess caps the Elk Butte Member of the Pierre Shale on the uplands. Loess also covers the broad terrace just above the inner valley of the Missouri River trench. In the eastern part of the terrace the loess is underlain succesively by slow-draining alluvium, fast-draining alluvium, and Pierre Shale. At the northwest end, the sequence below the loess descends through fast-draining alluvium to till and into more deposits of fast-draining alluvium before reaching the unconformable surface of the Pierre. Apparently there was a glacial advance between deposition of the upper and lower alluvium that was not recorded in the eastern part of this area.

The prereservoir Missouri River flood plain and a low terrace along the bottom of the inner valley are composed of a younger slow-draining alluvium that also floors the lower parts of some tributary valleys. In most of the small valleys and gullies the alluuvium is mixed with colluvium and loess. Alluvium, colluvium. and loess also blanket part of the inner-valley wall in sec. 32, T. 96 N., R. 66 W.

Stability conditions.—The slope stability conditions on opposite sides of the river are very different. Southwest of the river nearly all of the Pierre Shale, which makes up most of the steeper slopes, has been subject to landslides. Northeast of the river, landsliding apparently has been largely confined to the wall of the inner valley west of White Swan Bottom. The Pierre Shale exposed above the terrace contains few slides.

The difference in stability on opposite sides of the trench is caused by the difference in terrain. The slopes southwest of the river rise rather uniformly from the flood plain to the uplands; therefore their erosion and stability are controlled directly by the Missouri River. Periods of increased downcutting or sidecutting, such as have occurred several times in the recent history of the Missouri River, would have stimulated oversteepening of slopes and contributed to conditions favorable to landslides or landslide cycles at numerous points along the river.

Northeast of the river, only the steep inner wall of the trench is directly influenced by the river. The broad terrace that separates the uplands from the flood plain acted as a buffer that protects the uplands from direct erosion by renewed cutting by the river. The efficacy of this buffer is demonstrated by the fact that although some downcutting occurred recently in the small stream valleys above the terrace, erosion was relatively minor and only a few small landslides developed on the steepened slopes. In contrast, landslides are relatively common along sections of the inner valley wall.

Most landslides in area 2 are stabilized. Northeast of the river only three landslides are active. Southwest of the river there are more landslides that are active or recently active, but they are still only a fraction of the total landslide terrain.

Almost all the active and recently active landslides southwest of the river are on the higher parts of the older trench wall or along tributary valleys away from the inner valley of the Missouri River. These landslides, moreover, are concentrated in the eastern two-thirds of the area, where the inner valley walls are steepest. This localization probably is linked to active lateral erosion and downcutting by the Missouri River in the recent past. Erosion by the main river also accelerated downcutting along the major tributaries. The stable Niobrara Formation maintained steep, nearly vertical slopes at the base of the inner valley wall. Failure of the oversteepened slopes in the less stable Pierre Shale, however, began a cycle of landsliding that moved progressively up the trench walls and upstream along the tributaries. The first products of this cycle were removed or completely masked long ago. Later stages are present in the large areas of stabilized landslides and range from slides so altered that they can be identified only with difficulty to younger ones that are, as yet, only partly modified.

The current and recently active landslides belong to the waning stage of the cycle. The active slides on the main trench wall have reached the uplands. In the tributary valleys activity now is upstream at least a half mile from the valley mouths, and much of the landsliding occurs in the valleys of secondary tributaries.

AREA 3

Area 3 (pl. 1) is about 20 miles upstream from the Fort Randall Dam and includes the area between lat 43°10′ and 43°14′ N. and long 98°49′ and 98°57′ W. East of the Missouri River the land is in Charles Mix County, S. Dak.; west of the river it is in Gregory County, S. Dak.

The Missouri River divides area 3 so that about two-thirds of the land is west of the river. On the west side, the floor of the trench consists of a flood plain at about 1,270 feet altitude and low terrace remnants at various heights above the flood plain. The terraces grade upward into shale hills as much as 2,000 feet in altitude. The slopes in general are moderate (fig. 20) except where underlain by the relatively resistant Mobridge Member that forms a zone of steeper slopes at altitudes between 1,700 and 1,800 feet. The shale hills are part of a maturely dissected southward-draining upland that was established before the present course of the Missouri River was established. This upland surface is present on both sides of the river. In the southwest quarter of the area, the earlier drainage has been bisected and modified by Whetstone Creek.

The east side of the Missouri trench wall rises abruptly from the river in steep bluffs 80–180 feet above the prereservoir water level. Except for numerous small terrace remnants above the bluffs, the upper part of the trench wall continues to rise steeply to the southward-sloping upland. The dissected upland grades from maximum altitudes of 1,900 feet at the north edge of the area to about 1,450 feet where it merges with the upper part of a broad well-developed terrace near the south edge.

Geologic setting.—The geology in area 3 (pl. 1) is virtually the same as in the other areas. The entire pre-Quaternary section (lower part now below reservoir level) is present, although the Ogallala Formation occurs only in patches on the highest points in the northwest corner of the area. The flood-plain deposits of the Missouri River and Whetstone Creek are slow-draining alluvium, and both slow-draining and fast-draining alluvium underlie most of the terrace remnants. Ground moraine caps the uplands in the northeast corner of the area (Stevenson and Carlson, 1950). The patchy deposits of loess and of the alluvium, colluvium, and loess unit are widespread throughout the area.

Stability conditions.—Landsliding has occurred on most of the steeper slopes. Most of the landslides are now stable and the characteristic hummocky terrain prevails (fig. 20).

Nearly all slopes west of the reservoir show evidence of landsliding. The active or very recently active slides, except for some along Whetstone Creek, are near the uplands. Most occur on slopes that do not have any low-level terraces. The evidence implies that present landslide activity west of the reservoir is the waning phase of a cycle begun after the river had formed the terraces.

East of the reservoir are two general classes of slopes, west-draining and south-draining. The west-draining slopes adjacent to the reservoir resemble those across the river. They consist mostly of hummocky stabilized-landslide topography with a few active and recently active landslides. Much of the current landslide activity is near the uplands, but an appreciable amount is also on the lower slopes and along the river. Recent erosion along the river and the resulting renewed erosion along the tributaries appear to be responsible for the present landslide activity. Farther east, the south-draining slopes of the dissected upland show much less evidence of landsliding. As in area 2 (pl. 3), the terrace south of the upland has apparently acted as a buffer to protect the slopes above from recent direct erosion by the river; there is no evidence of sliding in terrace and preterrace time. In postterrace times, however, the streams crossing the terrace have done more downcutting and oversteepening of their walls than their counterparts in area 2. As a result, landslides, both stabilized and active, are more common above the terrace in area 3 than in area 2.

AREA 4

The limits of area 4 (pl. 1) are lat 43°33′ to 43°37′ N. and long 99°16′ to 99°24′ W. The area is about 55 miles upstream from the Fort Randall Dam. The part of the area west of the river is in Lyman County, S. Dak., and the part east of the river is in Brule County, S. Dak.

The Missouri River divides area 4 into two nearly equal parts. The river flows due south to about lat 43°35′ N., then swings gently southeastward to the south boundary of area 4. Flood plains and low terraces form a bottomland as much as three-fourths mile wide along the west side of the river. Moderately steep shale slopes (fig. 26) rise from the valley floor (about 1,320-ft altitude) to uplands ranging from about 1,750 feet to as much as 1,820 feet in altitude. In the northwest part of the area the upland is dissected by steep-walled valleys draining northwest

FIGURE 26.—Terrain west of the Missouri River in area 4. Typical steep slopes with hummocky stabilized landslide topography and some small active landslides along the valley bottoms. View north from SE¼SE¼ sec. 6, T. 101 N., R. 71 W., Lyman County, S. Dak. Photographed August 27, 1956.

away from the Missouri River. On the east side of the river erosion has removed or prevented development of an extensive flood plain. Except for isolated bits of flood plain, bluffs from 20 feet to more than 100 feet high rise directly from the river. In the southern third of the area the bluffs terminate in a clearly defined, gently sloping terrace as much as one-half mile wide. In the northern two-thirds of the area erosion has reduced the terrace to flat-topped spurs that cap the bluffs bordering the river. Behind the terrace, relatively steep slopes rise to an upland that is 1,700–1,800 feet in altitude.

Geologic setting.—The geology in area 4 (pl 1) differs from the areas downstream primarily in that the Mobridge Member of the Pierre Shale is the uppermost pre-Quaternary unit present. Much of the geologic information below was summarized from Baldwin and Baker (1952).

The flood-plain and low-terrace deposits of the river are mostly slow-draining alluvium. Except for exposures near the north and south edges of area 4, the top of the Niobrara Formation is concealed by the alluvium. Pierre Shale, from its base up through the lower part of the Mobridge Member, is exposed on the slopes rising from the bottomlands.

In the east half of area 4 the Niobrara Formation is well exposed north of lat 43°35′ N.; to the south it is found mostly in scattered exposures along small tributaries as far south as Elm Creek. The Pierre Shale, from the Sharon Springs through the Verendrye Members, overlies the Niobrara east of the river. Till, blanketed by loess, overlies the Verendrye Member and forms the upland surface. Alluvium comprises the terrace and the small flood plain at its base. Alluvium also floors the valley of Elm Creek above the northern limit of the terrace. Small deposits of alluvium also cap the terrace remnants to the north. The alluvium is mostly slow draining, but some fast-draining alluvium is also present.

Stability conditions.—Slides (fig. 26) have occurred on nearly all of the slopes in the Missouri River trench and in most of its tributary valleys. Landsliding is less prevalent along valley walls of the westward-flowing streams in the northwest corner of the area.

Active and recently active landslides are numerous on both sides of the river, but most of the slides are small and collectively they form a very small percentage of the total landslide area. Present landslide activity is related almost entirely to minor erosion and local oversteepening of slopes by the smaller tributary streams.

The extensive areas of now-stabilized landslides apparently were most active before the formation of the high terraces east of the river, because there is no evidence of landslides encroaching on the terraces. In postterrace time, moreover, the river has partially or completely removed the terraces north of Elm Creek, but extensive landsliding has not followed. This apparently anomalous erosion without landsliding may occur because the river is incised 40–50 feet into the Niobrara Formation, which has supported the weaker overlying shale here.

AREA 5

Area 5 (pl. 1), the northernmost study area, is about 70 miles upstream from Fort Randall Dam at the confluence of the White and Missouri River valleys (pl. 1), and is bounded by lat 43°42′ and 43°46′ N. and long 99°21′ and 99°29′ W. The land west of the Missouri River is in Lyman County, S. Dak.; the land east of the river is in Brule County, S. Dak.

The Missouri River, which flows from north to south in a gentle S-curve, divides the area nearly in half. The west half is further divided by the curving course of White River which isolates the southwest quarter from the rest of area 5.

The physiography in area 5 represents three stages of development. The oldest is the preglacial White River valley that crossed the area from west to east (Flint, 1955, pl. 7). Subsequently, this valley was modified by glaciation and the formation of the present valleys of the White and Missouri Rivers. The upland surface on both sides of the Missouri was originally part of the preglacial White River valley and is generally lower than the uplands in the other study areas. West of the Missouri River there are still remnants of several preglacial White River terraces. East of the river the former White River valley is filled with till and is identified only as a broad shallow sag in the present upland surface (Warren, 1952, fig. 2).

In the northern two-thirds of the area, vertical bluffs 70–100 feet high border the Missouri River (about 1,320-ft altitude). Above the bluffs, the walls of the trench rise steeply to a gently rolling upland that is 1,500–1,650 feet in altitude. Downstream, where the Missouri River makes a wide swing to the southwest, the steep trench wall gives way to gentler slopes, and a succession of terraces descend to a flood plain that broadens southward to the junction with the White River.

The floor of the White River valley is characterized by extensive flood plains, low flat islands, and low terraces. Supported by resistant outcrops of the Niobrara Formation, the valley walls generally rise above the floor as steep bluffs. The slopes above the Niobrara range from gentle in places not recently eroded to precipitous in stretches undergoing erosion by the river.

The east side of the Missouri River is bordered by a flood plain one-fourth to one-half mile wide surmounted by broad low terraces. The east wall of the Missouri trench, in the northern part of the area, is characterized by gentle multiterraced slopes. It steepens southward to precipitous bluffs as much as 100 feet high in turn surmounted by steep slopes. The upland east of the river ranges in altitude from approximately 1,620 to 1,700 feet.

Geologic setting.—Data on the geologic setting are derived from Baldwin and Baker (1952), Petsch (1952), Warren (1952), and Warren and Crandell (1952), supplemented by the author's observations.

The oldest exposed rock, the Niobrara Formation, has its upper limit at an altitude of about 1,380 feet. The lower five members (Sharon Springs to Verendrye) of the Pierre Shale overlie the Niobrara Formation and comprise most of the slopes. The Quaternary deposits, separated from the Cretaceous by a major unconformity, are alluvium and till with minor amounts of loess and of the alluvium, colluvium, and loess unit. Fast-draining alluvium caps most of the higher terraces of both pre- and post-Missouri River age. Fast-draining alluvium also crops out on the east side of the Missouri River trench wherever erosion has exposed the floor of the preglacial White River valley. The lower terraces and prereservoir flood plains along the contemporary drainage system are largely slow-draining alluvium. Till underlies the uplands east of the Missouri River. Loess occurs as a thin blanket overlying the till and as small isolated patches. The alluvium, colluvium, and loess unit forms small deposits in hollows and gullies.

Stability conditions.—Landslides make up a smaller proportion of area 5 (pl. 1) than of any other study areas. This can be attributed to the topography created by numerous terrace remnants left by both preglacial and postglacial rivers. These terrace remnants form steps in the valley walls and therefore areas of uninterrupted steep slopes are limited. The terraces, moreover, protect the slopes above them from direct river erosion.

In only two parts of area 5 has landsliding involved entire valley walls in the manner so common in the other representative areas. The first is the northern two-thirds of the west trench wall. Here a cycle of landsliding in the Pierre Shale apparently followed erosion of the supporting Niobrara Formation by the Missouri River. Most of the slides are now stable, which indicates that the Pierre Shale slopes have readjusted to prereservoir condition of the underlying Niobrara Formation.

The second part of the area is at the bend of the White River in the NE. cor. sec. 16, T. 103 N., R. 72 W. There the landsliding is less extensive than along the west trench wall, but it is more spectacular because large landslides are still active (fig 27). In this area also the landslides occurred after erosion of the Niobrara Formation removed support for the overlying shale.

Other landslides in the area are individual slides or relatively small areas of coalesced slides. Except for a few along the east trench wall, these slides are in the valleys of tributary streams.

Active and recently active landslides make up a very small part of the slides described above; moreover, more than half of the activity is concentrated in two areas. The first is at the bend in the White River in the NE. cor. sec. 16, T. 103 N., R. 72 W. The second area, which includes the S½ sec. 19, N½ sec. 30, T. 103 N., R. 71 W., east of the Missouri River and some land south of area 5, is informally called the Cable School landslide area for the Cable School, located about 1 mile to the east. The Cable School landslides (fig. 28) are along a branching tributary in the lowest part of the buried White River valley (Warren, 1952, fig. 2). The sliding occurred in shales saturated by seeps and springs fed from water-bearing gravels in the sediments of the preglacial valley.

INTERPRETATION OF DATA

The information derived from appraisal of slope stability and from geologic mapping is used for two purposes. The first is to present semiquantitative measures of prereservoir landslide distribution and activity, and to infer from these measures and from the relation between landslides and local topography that landslide activity was greater in the more distant past than in the time just before filling of the reservoir. The second purpose is to show the degree of correlation between slope stability, natural slope angles, and composition of geologic materials in part of the area included in area 2 (pl. 3).

PRERESERVOIR SLOPE STABILITY

The factors considered in judging the stability of prereservoir slopes are: the area in each landslide stability class relative to the total area of landslides, the relative sizes of individual landslide areas in each stability class, and the relation between landslide activity and local topography.

The area of each class of landslide stability in the five representative study areas was measured with a polar planimeter and the results by percentages of total landslide area are shown in table 4. Stabilized landslides are the predominant class in all areas. For present purposes the three other classes are considered current landslides in contrast with stabilized landslides.

FIGURE 27.—Bluffs along the left bank of the White River where river erosion is causing active landslides. The light-colored outcrops in the lower middle part of the slopes are Niobrara Formation. The Niobrara Formation is eroded by the river and leaves oversteepened shale slopes above. These shale slopes fail and the landslide debris masks the Niobrara Formation. NE¼ sec. 16, T. 103 N., R. 72 W., Lyman County, S. Dak. (west side of area 5). Photographed July 10, 1952.

FIGURE 28.—East part of the Cable School landslide area. View east-northeast from SW¼NW¼ sec. 30, T. 103 N., R. 71 W., Brule County, S. Dak. Photographed October 14, 1954.

TABLE 4.—*Percentage of total area of landslides in each class of landslide activity*

Study area	Stabilized landslides	Current landslides		
		Recently active	Reactivated	Active
1	98	<1	<1	1
2	92	2	3	3
3	97	<1	<1	1
4	99	<1	<1	<1
5	95	<1	—	5
Average	96	<1	<1	2

The stabilized landslide areas, large slides apparently are more common than small slides (figs. 20, 25, and 26). In current landslide areas, however, small slides (figs. 6, 12) are much more common than large landslides (figs. 4, 27). Large current landslide areas are numerous only in area 2 (pl. 3).

In the past, landslides have been prevalent on most of the steeper slopes, but many of these slopes now seem virtually stable. Wherever stabilized landslide areas are flanked by river terraces, there is no evidence of renewed sliding. New activity is confined to the valleys of streams recently entrenched into the terraces. Most new slides are directly related to erosion along the smaller tributary streams, and steepness of slope is a secondary factor in determining their location. Even in large areas of current landslide activity (pl. 3), there is very little landsliding along the precipitous inner trench wall.

The information summarized above indicates that in immediately preservior time the walls of this part of the Missouri River trench were more stable than at any time in their preceding history. Most evidence suggests that landslide activity was relatively small when the reservoir began to fill.

Two other factors probably have also contributed to the comparatively large landslide activity of the past and the small scale of recent slope failures. First, the change in climate that followed the vanishing ice sheets has resulted in a sharp decrease in the amount of water available to lubricate the potential slides. Precipitation must have been greater throughout the glacial periods. During the waning stages of glaciation, moreover, melt waters contributed appreciably to the ground-water supply in areas bordering the ice. The present climate must be very dry by comparison, with attendant diminution of ground-water supplies.

Second, during glacial and much of postglacial times, the Missouri was cutting through the easily eroded and unstable Pierre Shale in which slope failures are common. More recently, the river has entrenched itself in the Niobrara Formation which is harder and much less susceptible to failure. A decrease in landslide activity would predictably follow in most areas, as the slopes in the overlying Pierre reached equilibrium.

CORRELATION OF LANDSLIDES, SLOPE ANGLES, AND GEOLOGIC MATERIALS

The land southwest of the Missouri River in area 2 (pl. 3) is well suited to an investigation of possible correlations between landslides, slope angles, and geologic materials. It not only has more recently active landslides than any other study area but it also contains the entire Pierre Shale section, with the possible exception of the Crow Creek Member. The investigation was restricted to the pre-Quaternary rocks because these include the materials most prone to landsliding.

PREPARATION OF DATA

The information on slope stability conditions of the pre-Quaternary geologic units in area 2 (pl. 3) is given in percentages. First the areas of each unit and of each stability class in each unit were measured with a planimeter. Then the results were used to compute (1) the percentage of total pre-Quaternary exposures represented by each geologic unit, (2) the percentage distribution of each unit among the stability classes, and (3) the percentage of each stability class within individual units. The results are presented in table 5.

The Pierre Shale, as shown on table 5, comprises 97.4 percent of the pre-Quaternary exposures and includes all but 1–2 percent of each class of landslide. Field observations indicate, moreover, that landslides in the Ogallala and Niobrara Formations are usually the result of landslides in the shale. Inasmuch as virtually all the landslides are in or are controlled by the Pierre Shale, consideration of the correlation between landsliding, slope angles, and geologic materials is restricted to that formation.

Data for the slope angles were averaged from 10 cross sections; three of the cross sections and location of all 10 are shown on plate 3.

For every pre-Quaternary geologic unit exposed along each cross section, average slope angles were computed according to the following formula:

$$\tan\ (\text{average slope angle}) = \frac{\text{thickness of unit}}{\text{horizontal exposure of unit along cross section}}.$$

The angles for individual shale members in each cross section were then averaged with the corresponding angles along the other cross sections, and the results are taken as representative angles for each member of the Pierre Shale (table 6).

TABLE 6.—*Average slope angles, in degrees, of the members of the Pierre Shale along cross sections* A–A′ *to* J–J′

Cross section	Elk Butte	Mobridge	Virgin Creek	Verendrye and DeGrey	Gregory	Sharon Springs
A–A′	6	6	3	3	11	16
B–B′	18	18	8	3	15	6
C–C′	5	12	10	5	2	
D–D′	7	5	3	10	27	22
E–E′	17	7	7	5	5	2
F–F′	8	6	9	5	20	17
G–G′	6	10	4	6		
H–H′	10	13	5	3	6	
I–I′	17	12	8	3		
J–J′	17	20	7	7	3	5
Average	11	11	6	5	11	11

Assuming that only the proportional amount of montmorillonite in the clay-size component of each member affects the stability of that member, the author averaged and incorporated quantities for the montmorillonite content (fig. 16) in table 7. Although some of the samples (fig. 16) were collected as much as 60 miles from area 2 (pl. 3), their analyses are considered to be sufficiently accurate for this study because the mineral composition of the Pierre Shale apparently is relatively consistent throughout the Fort Randall area.

TABLE 5.—*Slope stability conditions of pre-Quaternary deposits in study area 2 southwest of the Missouri River*

Unit	Map symbol (pl. 3)	Percentage of total area of exposures of pre-Quaternary units	Percentage of unit						Percentage of each stability class in stratigraphic units				
			Stable	Stabilized	Recently active	Reactivated	Active	All landslide categories	Stable	Stabilized	Recently active	Reactivated	Active
Ogallala Formation	To	2.5	53	44	<1	2	<1	47	25	1	<1	2	1
Pierre Shale:													
Elk Butte Member	Kpc	22.9	12	79	2	4	3	88	56	21	14	34	24
Mobridge Member	Kpm	19.5	<1	90	4	3	3	100	1	20	31	20	22
Virgin Creek Member	Kpvc	16.1	0	92	3	3	2	100	0	17	19	16	11
Verendrye and DeGrey Members.	Kpvd	30.2	<1	93	3	2	2	100	2	32	33	24	23
Gregory Member	Kpg	6.0	5	89	<1	1	4	95	6	6	2	2	11
Sharon Springs Member	Kps	2.7	15	76	0	2	7	85	8	3	0	2	8
Total Pierre Shale		97.4	4	88	3	3	2	96	73	99	99	98	99
Niobrara Formation	Kn	.1	86	14	0	0	0	14	2	0	0	0	0

ANALYSIS

Although the data from area 2 (pl. 3) are quantitative and permit statistical analysis, the selection of a workable method for analysis was restricted by data being limited in both quantity and quality. Some information, such as the distinction between stable ground and landslide areas, is necessarily subjective. Many of the geologic contacts are either approximate or inferred. The slope angles are based on a limited number of cross sections, and the proportional values for montmorillonite clay are derived from scattered samples collected over the entire reservoir area.

Slope stability, slope angles, and montmorillonite clay content are interrelated, but geographic and stratigraphic locations are also important factors in slope stability of exposures. Exposures along a stream-cut bank, for example, are more susceptible to landslides than exposures of the same material on a gentle slope. Similarly, geologic units overlying stable strata are less prone to sliding than similar units overlying unstable strata. Finally, effects of ground water on slope stability are not taken into consideration.

Spearman's rank correlation coefficient was used to determine possible correlation between montmorillonite content, slope angles, and relative amount of landslide terrain of the members of the Pierre Shale. This coefficient indicates the relation between two classes of data ranked according to increasing or decreasing values. A method using ranks instead of actual numerical values minimizes minor discrepancies and also simplifies correlation of dissimilar classes of data.

Spearman's rank correlation coefficient (r_{rank}) is defined as:

$$r_{\text{rank}}=1-\frac{6\ \Sigma D^2}{N\ (N^2-1)}$$

in which,

D= difference in rank between paired item in two series,

ΣD^2=sum of values of D^2 for all paired items in two series,

N=the number of pairs of items in two paired series. (Croxton and Cowden, 1955, p. 478–480.)

Values for the correlation coefficient range from 1 for complete correlation, 0 for completely random distribution, to −1 for complete inverse correlation. Only two series can be correlated at one time.

The first step to determine rank correlation coefficients is to assign ranks to each item (table 7). Where two or more items are of equal value, the respective rank numbers are averaged between them. After the data are ranked, the rank correlation coefficient is calculated between each pair of series.

TABLE 7.—*Ranking of Pierre Shale data*

Member	Average montmorillonite content of clay-size fraction (parts in 10)	Rank	Average slope angle (degrees)	Rank	Relative amount of landslide terrain (percentage)	Rank
Elk Butte	7	3	11	4½	88	5
Mobridge	4	6	11	4½	100	2
Virgin Creek	9	1½	6	2	100	2
Verendrye and DeGrey	9	1½	5	1	100	2
Gregory	6	4½	11	4½	95	4
Sharon Springs	6	4½	11	4½	85	6

Montmorillonite-content–slope-angle rank correlation coefficient

Member of the Pierre Shale	Rank of montmorillonite content	Rank of slope angle	D	D^2
Elk Butte	3	4½	3/2	9/4
Mobridge	6	4½	3/2	9/4
Virgin Creek	1½	2	½	¼
Verendrye and DeGrey	1½	1	½	¼
Gregory	4½	4½	0	0
Sharon Springs	4½	4½	0	0
			$\Sigma D^2=5$	

$N=6$

$$r_{\text{rank}}=1-\frac{6\Sigma D^2}{N(N^2-1)}$$

$$=1-\frac{6(5)}{6(36)-1)}$$

$$=0.857$$

Montmorillonite-content–relative-amount-of-landslide-terrain rank correlation coefficient

Member of the Pierre Shale	Rank of montmorillonite content	Rank of relative amount of landslide terrain	D	D^2
Elk Butte	3	5	2	4
Mobridge	6	2	4	16
Virgin Creek	1½	2	½	¼
Verendrye and DeGrey	1½	2	½	¼
Gregory	4½	4	½	¼
Sharon Springs	4½	6	3/2	9/4
			$\Sigma D^2=23$	

$N=6$

$$r_{\text{rank}}=1-\frac{6\Sigma D^2}{N(N^2-1)}$$

$$=1-\frac{6(23)}{6(36-1)}$$

$$=0.343$$

Slope-angle–relative-amount-of-landslide-terrain rank correlation coefficient

Member of the Pierre Shale	Rank of slope angle	Rank of relative amount of landslide terrain	D	D^2
Elk Butte	4½	5	½	¼
Mobridge	4½	2	5/2	25/4
Virgin Creek	2	2	0	0
Verendrye and DeGrey	1	2	1	1
Gregory	4½	4	½	¼
Sharon Springs	4½	6	3/2	9/4
			$\Sigma D^2=10$	

$N=6$

$$r_{\text{rank}}=1-\frac{6\Sigma D^2}{N(N^2-1)}$$

$$=1-\frac{6(10)}{6(36-1)}$$

$$=0.714$$

The rank correlation coefficients indicate a large range in the degree of correlation between the three series. In any two series where N=6 there is only one chance in 20 that randomly selected independent variables will have a rank correlation coefficient as high as 0.829 (Olds, 1949, quoted from Dixon and Massey, 1951, table 17–6). It follows that the montmorillonite-content–slope-angle coefficient (0.857) has less than one chance in 20 of representing random correlation between variables. There is little doubt, therefore, that montmorillonite content and slope angles are related. The relation is inverse; that is, the higher the montmorillonite content, the lower the angle of slope. The montmorillonite-content – relative-amount-of-landslide-terrain coefficient (0.343), on the other hand, is too low to be significant. Between the two extremes is the coefficient for rank correlation between slope angles and relative amounts of landslide terrain (0.714). This value is not low enough to refute the possibility of correlation, but neither is it high enough to conclusively prove correlation.

INTERPRETATION

There is a correlation between montmorillonite content and slope angles if the conclusion is correct that shear strength of the Pierre Shale is an inverse function of montmorillonite content in the clay-size portion of the shale. The shear strength in turn largely determines the maximum natural slope angle of the shale. The maximum possible slope angles, therefore, are finally controlled by the montmorillonite content of the shale.

The line of reasoning in the preceding paragraph admittedly is much oversimplified. This generalization presents only the principle that there is a correlation between montmorillonite content and slope angle. Lack of complete quantitative correlation between the montmorillonite contents and slope angle data for the shale members (table 7) presumably is due to various modifying factors (p. 44). The data in table 7, and above show that the lack of correlation between the montmorillonite content and the relative amount of landslide terrain may be due largely to the apparently contradictory statitsics for the Mobridge Member. Although its average slope angle is among the highest and its montmorillonite content is the lowest in the whole formation, the Mobridge is classified as 100 percent landslides (table 7). This surprising lack of stability in the Mobridge probably arises because the underlying members are high both in montmorillonite content and in percentage of landslide terrain.

The only obvious correlation between slope angle and the relative amount of landsliding is that factors which permit the shale to maintain steep slopes also tend to reduce the amount of landsliding. The amount of landsliding, for example, seems to be controlled partly by stratigraphic and topographic location. The most stable members of the Pierre Shale (table 5) are the Sharon Springs and Gregory, which overlie the Niobrara Formation, and the Elk Butte, which is the uppermost member and comprises part of the stable upland surfaces (pl. 3). The relatively stable Niobrara Formation apparently increases stability of the overlying Sharon Springs and Gregory Members by forming a stable foundation for them. The fact that the Gregory, separated from the Niobrara by the Sharon Springs, has a higher percentage of landslides than the Sharon Springs shows that the effectiveness of support diminishes with distance from the foundation.

In contrast, the Mobridge, although its low montmorillonite content should make it more stable than the Sharon Springs and Gregory Members, is extremely susceptible to landsliding. The fact that it is underlain by the highly unstable Virgin Creek, Verendrye, and DeGrey Members undoubtedly tends to nullify its strength and to reduce its value as a potential support for the Elk Butte. The Elk Butte Member, in spite of its fairly high montmorillonite count, has fewer slides and steeper slopes, probably because there are no very thick overlying sediments creating stresses within it.

SUMMARY AND CONCLUSIONS

Investigations in the representative study areas have yielded several noteworthy results:

(1) There is a strong inference that before formation of the Fort Randall Reservoir the walls of the Missouri River trench were comparatively stable. Probably the last period of great landslide activity was in glacial times.

(2) Virtually all of the landslides occur in the Pierre Shale.

(3) All members of the Pierre Shale are subject to landslides. Relative stability of the members is reflected in the angles at which they form stable slopes and not by the relative number of landslides in each member.

(4) Natural slope angles in individual shale members, and therefore relative stability and strength of the members, are an inverse function of montmorillonite content in the clay-size fraction of the shale.

(5) Restricted landslide activity in some shale members seems due to stratigraphic and geographic locations of the members. Apparently it is not related to composition and behavior of the members themselves. Where external factors restrict landsliding, the shale

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may be stable on what otherwise would be abnormally steep slopes.

General conclusions are that the Fort Randall Reservoir should have little effect on slope stability beyond the immediate shoreline unless wave erosion is unexpectedly great or unless there is an unforeseen major rise in the ground-water table within the Pierre Shale. A major change from the present environment would be required to restore, or even approach, the conditions of the glacial times. Slope stability is directly related to the percentage of montmorillonite in the clay-size fraction of the Pierre Shale, although other factors may also influence slope stability.

GROUND-WATER INVESTIGATIONS

A continuing program was started in the winter of 1954 to investigate ground-water conditions at depths of less than 100 feet. The ground water data in this report are based on the first 3 years of ground-water observations.

PURPOSE

Ground-water observations in the Fort Randall Reservoir area were undertaken to correlate changes in ground-water conditions with landslide activity, and to better understand the behavior of ground water in the Pierre Shale and alluvium along the reservoir shores. Two general classes of factors control the ground-water conditions: (1) those that create permanent or semipermanent ground-water changes and (2) those of periodic or cyclic nature that occur either rather regularly or randomly.

The Fort Randall Reservoir itself is a factor in the first class. This permanent body of free water has formed a new base level at a higher altitude than the former Missouri River surface. The nearby water table consequently must reach a new, higher surface to redevelop a state of equilibrium. Such an adjustment is a long-range process controlled by the amount of water available and by the permeability of the ground. This report describes only the general effects of the reservoir water level on the table because the available records are insufficient for more detailed consideration.

The most obvious cyclic factor is seasonal climatic variations. Other factors are alternate periods of drought and above-normal precipitation and the effects of single, abnormally heavy, storms. Some of these recurring factors have relatively rapid effects on ground-water conditions and can be investigated qualitatively on available ground-water records.

METHOD OF INVESTIGATIONS

Porous tube piezometers, first described by Casagrande (1949), were used in the ground-water observation program. Casagrande's method of constructing and installing the piezometers was somewhat modified and simplified for the purposes of the investigation.

For several reasons, porous tube piezometers were considered suitable. Materials for their construction are inexpensive. Several piezometers can be installed independently at different depths in a single drill hole to provide ground-water data for several horizons without materially increasing installation costs. Because of their small diameter, they are sensitive to the small pressure changes caused by minor changes in water volume in relatively impermeable materials. They are nonmetallic and do not corrode in the mineral-rich waters of the Pierre Shale.

The porous tube piezometer is fundamentally an observation well with a well point consisting of a porous tube and a standpipe. Standpipes used on this project were made of ⅜-inch-inside-diameter plastic tubing (fig. 29). In the drill hole, the porous tube is

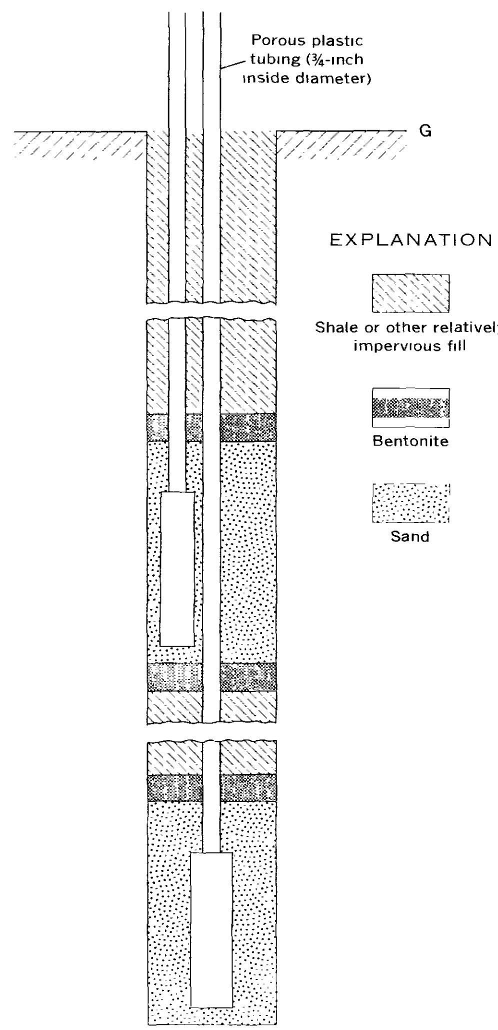


FIGURE 29.—Schematic diagram of piezometer installation. Plastic tubing extends 2–3 feet above ground surface.

packed in clean sand and sealed above and below with bentonite clay so that the water level in the standpipe reflects piezometric pressures at the horizon of the porous tube.

Installation of the piezometers along the reservoir was relatively simple. Holes for installation of the piezometers were drilled with either a power auger or a churn drill. Then the piezometers were lowered into the holes, which were backfilled. Where more than one piezometer was installed in a single hole, each additional one was emplaced when the hole had been backfilled to the desired depth.

PIEZOMETER LOCATIONS

Twenty piezometers were installed in the area at six sites along the reservoir shore and at two sites downstream from the dam. General locations of the piezometer sites are shown on plate 1; detailed descriptions of the locations are given in table 8. The two sites below the dam were selected as control installations to determine regional ground-water changes that may occur independently of the effects of the reservoir.

TABLE 8.—*Piezometer installations in South Dakota*

	County	Location	Altitude (ft)
A	Gregory	NE¼NE¼SE¼ sec. 20, T. 95 N., R. 65W.	1,360
B	do	SE¼NE¼NW¼ sec. 28, T. 97 N., R. 68 W.	1,444
C	Charles Mix	SE¼SE¼SE¼ sec. 28, T. 96 N., R. 66 W.	1,463
D	do	SW¼SW¼SW¼ sec. 31, T. 95 N., R. 64 W.	1,498
1	Brule	About 600 ft southwest of water plant intake building, Chamberlain.	1,381
2	do	About 450 ft northeast of intersection of 11th Ave. and Courtland St., Chamberlain.	1,390
3	do	SE¼SW¼SW¼ sec. 11, T. 105 N., R. 71 W.	1,414
4	do	SE¼NE¼NW¼ sec. 18, T. 103 N., R. 71 W.	1,450

The piezometer sites are grouped generally about the towns of Pickstown and Chamberlain to make them easily accessible for periodic measurements. The piezometers near Pickstown include the two control sites and two sites in the lower part of the reservoir where ground-water changes should be greatest. The piezometers near Chamberlain are near the head of the reservoir where the rise in free water level, and presumed corresponding rise in ground-water levels, should be small.

Sites 1 and 2 at the town of Chamberlain were chosen specifically to investigate seepage pressures in a terrace of fine surficial deposits similar to landslide-susceptible materials along the shores of Lake Roosevelt behind Grand Coulee Dam in Washington. The landslides along Lake Roosevelt have occurred most frequently after a rapid major drawdown of the reservoir water level, and it is likely that seepage pressures from water draining out of the saturated materials along the shore triggered these slope failures. In the event of similar major drawdowns in the Fort Randall Reservoir, the piezometers at Chamberlain will provide information about seepage pressures in the terrace.

Four of the piezometer installations—sites 4, B, C, and D—are in Pierre Shale; the other four—sites A, 1, 2, and 3—are in alluvium. This distribution was selected to determine ground-water trends in a typical variety of materials along the reservoir.

Piezometers in the shale were installed approximately 30, 60, and 90 feet below the surface so that the piezometers would be far enough apart to avoid ground-water migration from one horizon to another. An additional piezometer was installed at a depth of 12 feet at site B to record ground-water conditions near the surface. The piezometer numbers indicate site and depth; thus, piezometers B–12 and B–30 are at depths of 12 and 30 feet in the same hole at site B.

Distribution of piezometers in the surficial materials was more varied; depths ranged from 14 to 81 feet, and one to three piezometers were installed at each site.

RESULTS AND INTERPRETATIONS

Water levels for each piezometer and the drillers log for each site are shown graphically in plate 4. The reservoir water levels at both Pickstown and Chamberlain are also plotted for comparison with ground-water levels. The following results and interpretations refer to the data in plate 4.

Piezometric pressures were surprisingly constant during 1954–56; water levels varied little more than 1 foot in most of the piezometers, and they varied more than 4 feet in only three piezometers (B–92, D–92, and 1–60.

RELATION BETWEEN RESERVOIR WATER LEVEL AND GROUND-WATER LEVEL

As of November 1956 there was little indication of a relation between reservoir water level and ground-water level in the piezometers located in the shale. Only piezometer B–92 showed a gradual rise of pressure followed by a relatively stable water level.

The piezometers at sites 1 and 2 on the Chamberlain terrace are the only ones in the upper part of the reservoir that show any response to the rising water level. This is to be expected because the piezometers are near the reservoir in slow-draining alluvium, one of the relatively permeable materials along the reservoir. Piezometer 1–60 apparently shows a direct relation between reservoir water level and piezometric pressures in the terrace. The absence of the usual seasonal drop in water level in piezometer 1–29 during the summer of 1956 may be the result of ground-water changes caused by the rising reservoir water level.

During 1954-55, the water level in piezometer 2–24 was high in the winter and low during the summer months; in 1956, however, it rose steadily until in November it stood about 3 feet above the former normal level. This suggests that the water table at site 2 began in 1956 to respond to the raised water level of the reservoir. If this assumption is correct, there was a lag of about 1½ years from the time the reservoir water level began to rise until the water table about 750 feet inland from the shoreline also began to rise.

Comparison between reservoir and piezometer 1–60 water levels reveals an interesting relation. The water level in the piezometer was above the reservoir water level until about the end of September 1954, when, until May 1956, conditions were reversed, in that the reservoir water level was generally higher than the water level in the piezometer. The prereservoir hydraulic gradient obviously was toward the river, whereas from fall 1954 at least to spring 1956 water was moving from the reservoir into the terrace deposits. All the other piezometric data indicate that the normal hydraulic gradient along the Missouri River trench is toward the reservoir. The reversed hydraulic gradient that existed for at least 1½ years after the rise in the reservoir water level was, therefore, a locally anomalous condition that should exist only until ground-water conditions became adjusted to the new reservoir water level. Because ground-water adjustments in slow-draining alluvium were still incomplete less than 150 feet from the reservoir after 1½ years, one may assume that the overall readjustment along the reservoir may take at least 10 years.

RELATION BETWEEN CYCLIC PROCESSES AND GROUND-WATER LEVELS

Observations in 1954-56 indicated that cyclic processes had only minor effects on ground-water levels. Variations in water levels that could be correlated with seasonal climatic changes amounted to only a few feet in the shallowest piezometers and was even less in the deeper ones. There was no great change in annual precipitation during the 3 years that ground-water records were kept and, therefore, no corresponding change in ground-water conditions. Some small temporary rises of water levels in the shallower piezometers were probably the result of especially heavy local storms, but the precipitation records are not sufficiently detailed to make a definite correlation.

Most small variations in the piezometer water levels do not seem directly attributable to periodic or cyclic causes and have no obvious relation to permanent ground-water changes. Many are deviations in individual water-level readings that may be faulty measurements or temporary changes resulting from storms or other causes.

BEHAVIOR OF GROUND WATER

This section briefly outlines ground-water behavior at each piezometer installation and summarizes ground-water behavior in alluvium and in the Pierre Shale.

SITE A

The piezometers at site A are in slow-draining alluvium. Piezometers A–14 and A–30 measure pressures in a lean clay. The porous tube of piezometer A–81 is just below the alluvium-Niobrara Formation contact, but the sand filter around the porous tube extends upward about 8 feet into fine sand and clayey sand.

Piezometer A–14 is approximately at the water table. The system rarely contains much more than 2 feet of water, and frequently it is dry during part of the summer and early fall.

The water level in A–30 generally stands 8–10 feet above the base of the porous tube, or about 7–9 feet below the water table. Fluctuations in the water table are crudely reflected on a reduced scale by the water level in A–30; friction between the downward migrating water and the clay and silt particles reduces the hydrostatic pressures and dampens the effects of changes in the water table.

Piezometric pressures in piezometer A–81 apparently were smaller than in A–30 despite the greater depth. The water level in A–81 rarely rose to more than 6 feet above the base of the porous tube. The water-level fluctuations, moreover, show little correlation with water levels in the shallower piezometers except from December 1954 to February 1955. The water level in piezometer A–14 rose suddenly in December 1954, followed by a major drop in the water level in January; a similar rise and fall of the water level occurred in piezometer A–81 during January and February 1955. The failure of A–30 to show a similar response may be due either to unknown hydrologic factors or simply to clogging of the piezometer tube.

SITE B

The piezometers at site B are located in shale overlain by about 4 feet of colluvium. Piezometer B–12 is in the weathered part of the shale near the surface; the others are in unweathered shale.

Unusual water-level conditions were found in the piezometers at site B. Piezometer B–12 was dry almost half the time it was being measured, but 14 out of 24 measurements showed as much as 1.9 feet of water. Piezometer B–30, 18 feet below B–12, was dry at all measurements. Piezometer B–59, 29 feet deeper than B–30, consistently contained about 10 feet of water, and the minor water-level fluctuations roughly correlated with the fluctuations in B–12. The water level in B–59, moreover, was virtually at equilibrium when the first measurement was made in February 1954. The deepest piezometer, B–92, did not reach equilibrium until early in 1956 when the water level became stabilized at a height of about 34 feet above the piezometer.

The water levels in the piezometers at site B would be anomalous for normal ground-water conditions in material of uniform permeability, but are normal for the weathered and fractured conditions that apparently exist in the Pierre Shale. Although fresh shale is almost impermeable, weathered shale has greatly increased permeability. During intervals of rapid ground-water intake, therefore, the weathered zone can become saturated even though fresh shale below is dry. Piezometer B–12, located in weathered shale, records the presence of an intermittent perched water table, whereas piezometer B–30, in fresh shale but above the permanent water table, record nothing.

Although piezometer B–59 is in fresh unweathered shale, the comparatively rapid development of equilibrium and sizable monthly water-level fluctuations suggest that the water moves through fractures rather than through the shale itself. The presence of water in B–12 and B–59 and the absence of water in B–30 can be explained if B–59 intersects one of these fractures and B–30 does not. At times of heavy intake, some ground water would saturate the weathered shale and the remainder would move down through the fractures toward the permanent water table. An increase in piezometric pressures would develop throughout the entire fracture system and be reflected by piezometers intersecting that system.

Comparison of the water levels in piezometers B–59 and B–92 supports the theory that ground-water movement in the Pierre Shale is predominantly through fractures. The extra time for piezometer B–92 to reach equilibrium compared to the time for B–59 indicates that the material surrounding B–92 is little fractured and has a much lower permeability.

Field observations of the Pierre Shale further support the conclusion that fractures are the main avenues of ground-water movement. Seeps issuing from fractures in steep shale banks are common. In many shale banks that do not have active seeps, the fractures are coated with water-deposited secondary minerals. At Oahe Dam near Pierre, S. Dak., the Corps of Engineers also found that ground-water flow is related to the fractures in the Pierre Shale (A. H. Burling, project geologist, Oahe area, and L. B. Underwood, district geologist, Omaha District, oral commun.)

SITE C

The geologic section at site C has about 11 feet of lean clay (probably colluvium derived from the Pierre Shale) underlain by about 28 feet of weathered Pierre Shale, in turn underlain by fresh shale. The shallowest piezometer, C–29, is in the weathered shale and the two deeper ones are in fresh shale.

Water levels in the three piezometers at site C were relatively stable and all were within a range in elevation of about 10 feet. Piezometer C–29 seems to be located approximately at the water table; it never had more than 1.4 feet of water and frequently was dry. The depths of water in piezometers C–55 and C–88 were about 19 feet and 51 feet respectively. Altitudes of the water levels in these two piezometers never were as much as 3 feet apart, whereas there was commonly a difference of about 8 feet in the water levels in C–55 and C–29. The available data do not suggest a close correlation between water-level fluctuations in the three piezometers.

Such fluctuation of the water levels at site C may be expected if ground-water movement in the shale matches the movement postulated for water levels at site B. The water level in C–29 represents the water table, and the water levels in C–55 and C–88 represent piezometric pressures at depths of 55 and 88 feet below the surface. The rapid water-level equilibrium reached in both C–55 and C–88 indicates that they intersect relatively permeable fractures. The unrelated fluctuations of water levels in the two piezometers imply that the fractures do not directly interconnect.

SITE D

The three piezometers at site D are in unweathered Pierre Shale. Water in the shallowest piezometer, D–32, had an average depth of about 10 feet, which suggests that the water table was a few feet higher, or about 20 feet below the ground surface. Although the water level in the piezometer was relatively constant, there was a slow unexplained rise of about 2 feet in the 3-year period, 1954–56. Small fluctuations generally of less than a foot were superimposed on the gradual upward trend of the water level.

Although minor dissimilarities were noted, the ground-water conditions measured by piezometer D–62 apparently were closely related to the conditions at the horizon of D–32. With the exception of one measurement, the water level in D–62 was consistently about 0.5 foot below the water level in D–32.

The water level in piezometer D–92 indicates that it is subject to slightly different conditions than the shallower piezometers. After the piezometers were installed, the water level rose more in D–92 than in D–32 and D–62 and did not reach approximate equilibrium in D–92 until October 1954; from that time to November 1956 the upward movement of the water level was slightly faster than in the two shallower tubes. The depth of water in D–92 was about 65 feet in November 1956, and the water surface 4.5 feet below that of D–62. This contrasts with a difference of only 0.5 feet between water levels in piezometers D–32 and D–62.

Most of the ground-water conditions observed at site D suggest that here also ground-water movement is predominantly through fractures. Piezometers D–32 and D–62 apparently intersect relatively permeable fractures. Their levels adjust to water table fluctuations with only brief time lags, and the slight differences in water level in the two piezometers reflect the relative ease with which water moves along the fractures. Slow development of equilibrium in D–92 indicates less permeability than at the shallower horizons. The greater difference of water levels between D–92 and D–62 than between D–62 and D–32 may indicate that D–92 encountered a small relatively closed fracture or that it lies entirely in unfractured shale.

SITES 1 AND 2

The piezometers at sites 1 and 2 were installed to measure water pressures in the Chamberlain terrace. The material at both sites is slow-draining alluvium ranging in composition from clay to clayey sand.

Apparently the Chamberlain terrace has a permanent water table overlain by one or more perched water bodies. The wide range in permeability in the strata composing the Chamberlain terrace probably favors horizontal rather than vertical ground-water movement. The water level in piezometer 1–60 approximates the permanent water table near the reservoir shoreline. The water level in piezometer 2–24 probably represents the water table toward the back of the terrace. The water level in piezometer 1–29 is well above the level in 1–60 and undoubtedly is a local water table perched on the lean clay bed 33 feet below the surface of the terrace.

SITE 3

The single piezometer installed at site 3 measures water pressures in a 6-foot-thick bed of fast-draining alluvium underlain by the Niobrara Formation and overlain by about 39 feet of slow-draining alluvium. Measurements of one piezometer can give no data on vertical or horizontal movement, but they do show that the fast-draining alluvium has never been fully saturated and that at times it has had very little, if any, water in it. This suggests that the fast-draining alluvium may act as an underdrain for the slow-draining alluvium.

SITE 4

The hole at site 4 was drilled through about 8 feet of silty colluvium and about 80 feet of Pierre Shale into the Niobrara Formation. The two shallower piezometers, 4–32 and 4–57, measure water pressure in the shale. The deepest piezometer, 4–93, measures pressures in both the Pierre Shale and the Niobrara Formation near their contact.

The piezometer results at site 4 indicate the possibility of several separate and distinct ground-water levels. The independence of water levels in these piezometers is in direct contrast to the water levels in shale piezometer installations, each of which showed an interdependence between water pressures at different levels. The water in 4–32 had an average depth of 16 feet, greatest of the three, and showed a maximum variation in level of 1.5 feet. Piezometer 4–57, in contrast, had a maximum water depth of 1.8 feet and was dry periodically despite the apparently permanent ground-water body above it. The deepest piezometer, 4–93, had a remarkably stable water depth of 2 feet. Except for one probably erroneous measurement in January 1955, the level varied only 0.3 feet in 1954–57. Ground-water conditions at 4–93, therefore, show no correlation with conditions at the shallower piezometers. There is no correlation between water-level fluctuations nor is there a rise in water level toward a water table defined by one of the shallower piezometers.

Such varying water levels in the three piezometers seem to indicate that most ground water there moves horizontally through fractures in the shale. The vertical permeability apparently is slight, compared to the horizontal permeability.

SUMMARY

The following brief generalities of ground-water behavior at the eight piezometer installations are valuable only as working hypotheses inasmuch as they are based on data from only eight piezometer installations collected intermittently over a 3-year period.

Ground-water movement through alluvium is predominantly horizontal. Because most alluvium consists of layers of material with different permeabilities, ground water moves downward until it reaches a less permeable layer than the one through which it is moving. The less permeable layer deflects the ground water from predominantly vertical to predominantly horizontal movement. Water pressures at various levels in alluvium consequently may not show any direct vertical correlation.

Most ground-water movement in Pierre Shale apparently is along fractures. The water can move either vertically or horizontally at a rate depending on the degree of development of fractures in any particular direction. Where vertical fractures are well developed, there is a vertical component of movement that promotes vertical equilibrium in the water pressures. Where horizontal fractures predominate, the relative horizontal component of movement may be so great that vertical water pressure equilibrium cannot develop, and there will be no correlation between water pressures, or even rates of movement, at different levels. Behavior of the ground water in these circumstances is similar to that of ground water in stratified alluvium.

Piezometer measurements were continued until December 1959 by the U.S. Geological Survey, Huron, S. Dak. These are not plotted on plate 4, nor are they correlated with reservoir water levels and precipitation surpluses. In general, these later measurements add little to the preceding account except for the following minor corrections and additions:

Site A.—The difference in water level between A–14 and A–30 through the latter part of 1957 and through most of 1959 increased to about 15 feet, owing mostly to a steady drop in level of A–30.

Site B.—Conditions have remained practically the same as before except that B–30 had as much as 1.5 feet of water in the winter of 1958–59.

Site C.—Minor fluctuations in C–55 and C–88 were apparently unrelated. C–29 became plugged in early 1959, and other tubes became plugged later in the year.

Site D.—Water levels in D–32 and D–62 were almost the same from the spring of 1957 through 1959. They dropped gradually about 3 feet from July 1958 to October 1959 and rose about 1.7 feet in November and December 1959. These changes in the higher piezometers were reflected rather faithfully by similar changes at the same times in D–92, whose water level is 3–4 feet lower.

Sites 1 and 2.—Piezometer 1–29 remained dry or had less than 2 feet of water during 1957–59, similar to ground-water conditions in 1954–56. The water level in 1–60 rose to elevation 1,351.9 in July 1958, and dropped gradually to 1,344.9 by December. The water level in 2–24, at the back of the Chamberlain terrace, remained fairly constant between elevations 1,359.1 and 1,361.6 during 1957–59 except for a rise to 1,364.9 in July 1958.

Site 3.—Piezometer 3–46 showed little change in 1957–59. The water level ranged in elevation from 1,369.9 to 1,372.4 feet. Measurements were periodic because the tube was occasionally plugged.

Site 4.—During 1957–59 there was little change from previous readings. The records are discontinuous owing to plugging of the tubes or inaccessibility of the site during periods of poor road conditions.

LANDSLIDE MOVEMENTS

METHODS OF INVESTIGATIONS

An estimate of the time, rate, processes, and magnitude of movements was made for individual landslides and landslide complexes in five areas (pl. 1). A network of control points was established on and near each landslide. These points were occupied periodically and compared with previous readings to determine the amount and rate of movement.

Control points on the Highway 16 slump, Landing Creek slump-earthflow, and Paulson slump and on the Cable School earthflows were located by transit traverse. At least two reference points on each traverse were located on presumably stable ground. The main traverses were closed to check surveying errors, but short open traverses to isolated points were extended from the closed traverses.

The Cable School slump-earthflow is across a small valley from a stable ridge; so it was possible to triangulate control point locations from three stations established along the ridge. Triangulation had the advantages of being faster than a transit traverse and required only one man for the survey. Points on the Cable School slump-earthflow consequently were measured more frequently than those on the other landslides.

The control points were 1- by 2-inch stakes 2–4 feet long usually driven into the ground until 6 inches of the stake was exposed. At many points, however, more than 6 inches was exposed so the control point would be visible from adjacent stations. Aiming targets for triangulation were large tin-can covers nailed to the stakes. For the traverses, headless nails were set in the tops of the stakes for aiming points. A number of stakes were destroyed or damaged by cattle and were replaced or abandoned.

Horizontal control for each landslide was adjusted to an arbitrary grid system superimposed on a map of the landslide. The horizontal and vertical movements, computed from the triangulation and transit traverse data, were plotted vectorially on the individual landslides. Inasmuch as the two types of movement were similar in general behavior, only the horizontal movements are discussed in this report.

DESCRIPTIONS OF LANDSLIDES

Except in the geologic map of the Landing Creek slump-earthflow, the mapping represents the geology in each of the landslide areas before movement occurred. At the Landing Creek slump-earthflow, landslide material and other shallow colluvial deposits as well as bedrock geology were mapped to demonstrate the masking effects of the landslide materials. Mapping these also showed some of the larger stratigraphically displaced blocks of Pierre Shale that probably are remnants of old landslide blocks. Landslide and colluvial debris are not shown at the other four slide areas so that the causes of the landslides can be clearly separated from their effects.

HIGHWAY 16 SLUMP

The Highway 16 slump (fig. 6, pl. 5) is in NE¼ NE¼ sec. 22, T. 104 N., R. 71 W., on U.S. Highway 16 about 1 mile south of the center of Chamberlain. The landslide is on the nose of a Pierre Shale ridge at the contact between the DeGrey and Verendrye Members and is one of the few pure slump movements near the reservoir. Instead of disintegrating into an earthflow as is common, the base of the landslide has pushed upward and forward over stable ground as a single unit forming a pressure ridge along the toe.

The landslide is not located where exceptional quantities of ground water would be expected, but during 1952 and 1953 lush vegetation near the toe of the landslide indicated appreciable amounts of moisture in the ground. From 1954 to 1956 there was less evidence of moisture, although a closed depression in front of the landslide contained standing water each spring.

The slump (pl. 5) is a main, backward-rotated block in front of a complex graben area composed of several smaller blocks. Initial movement probably occurred in the large block and the two smaller ones adjacent to it on the south and southwest. As the mass moved downward and outward, the upper parts of the blocks collapsed into the open fracture behind them to form the two smallest blacks. Then the laterally unsupported edge of stable ground directly upslope from the fracture also collapsed into the opening.

Most of the movement in the Highway 16 slump apparently occurred in the early spring of 1952. When the slump was first observed in June 1952, the uprooted vegetation was still green and all the major fractures described above had already formed. The only noticeable change in the slump from June 1952 to September 1956 was minor sloughing along some of the scarps and weathering of some material exposed in scarps and fractures.

Control points were measured annually from 1953 through 1956. Six permanent points were established in 1953 and the rest in 1954. Control points 1 and 6 were used as stable reference points for movements from 1953 to 1954. Points Z_4, 6, and Z_3 were assumed to be stable from 1954 to 1956. The error in horizontal closure of the closed transit traverse was 0.74 foot in 1953 and not more than 0.19 foot in the following 3 years. The minor difference in locations for point 6 between 1953 and 1954 implies that the error in the 1953 measurements was in the locations of temporary points used to close the traverse from point 6 to point 1. None of the measurement vectors, consequently, are assumed to have errors larger than 0.2 foot.

The vectors plotted for each point show that there was some relatively minor movement between 1953 and 1956. Control point 3 at the top of the lowest slump block moved about 0.4 foot, the greatest for any of the points. The rates of movement of all the points on the landslide were about the same and the directions of movement were north and northeast downslope and toward the pressure ridge. Exceptions to this general direction were point 2 on the lowest slump block and point 1 below the slide which showed southeastward movements of about 0.3 foot between the 1954 and 1955 measurements. Inasmuch as no adequate reasons were found to explain a localized change in ground movement, it was assumed that these two stakes probably had been disturbed by cattle. The magnitude of movement at points 2 (excepting the presumably disturbed 1954–55 lateral displacement), 3, and 4 was greater between 1953 and 1954 than the magnitude of movement for later measurements. The points beyond the limits of the landslide, excepting point 1, had neither individual nor cumulative movements as much as 0.2 foot, which is within the limits of surveying error.

The apparent movements of points 5, Z_2, Z_5, and Z_6 are the only movements outside the landslide boundaries that may have some significance. Although small and within the limit of error in the survey their consistent downhill component strongly suggests actual movement that may be the result of any of three processes: (1) natural creep of the weathered shale on the steep slopes near the landslide, (2) slow sliding of the laterally unsupported material upslope from the landslide, and (3) plastic deformation of the unsupported material upslope from the landslide. However, information is too limited to assign a specific cause to these movements.

Before the first measurements in 1953, the total amount of movement in the Highway 16 slump and in individual blocks was probably less than 10 feet. The relatively short distance and the high cohesion of the shale enabled the toe of the landslide to remain fairly intact instead of disintegrating into an earthflow.

The trend from an initial major movement in 1952 through minor movements in 1953 and 1954 to very minor movements after 1954 implies that the landslide was becoming stabilized under the existing climatic conditions. If climatic conditions became more conducive to landsliding they might reactivate movement.

CABLE SCHOOL EARTHFLOWS

The Cable School earthflows (fig. 12 and pl. 5 (two larger flows)) in NW¼NW¼NW¼ sec. 30, T. 103 N., R. 71 W., Brule County (pl. 1), are part of a fairly extensive complex of active and recently active landslides in the Cable School area. The larger flow is 90 by 150 feet; the smaller one is about 100 by 110 feet. The greatest depth of disturbed material is less than 10 feet.

The two earthflows are on the side of a small ridge composed mostly of the Sharon Springs and Gregory Members of the Pierre Shale. The shale is overlain by a 15-foot-thick stratum of gravel (fast-draining alluvium) capped by 15–20 feet of till. Several feet of colluvium blankets all the earlier deposits. Although much of the colluvium moved normally by slope wash and creep, former landslides apparently also transported large quantities.

The two earthflows are similar in appearance. Below the prominent scarp at the head of each slide is a zone of slump block that is about half of the larger flow but is less than one-fourth of the smaller one. The remainder of each slide has moved as a true flow. The toes have bulbous appearance where flow material moved over the original surface. The smaller earthflow includes a subsidiary earthflow that behaved as a separate entity although it is physically connected to the main one.

A small seep was observed about 10 feet below the gravel-shale contact (pl. 5) on the larger earthflow when it was first examined in July 1952. The water was probably derived from local precipitation because a valley isolates the ridge from ground-water sources in the nearby uplands. The seep dried up later in the summer and did not reappear during the time of observation. Accumulation of efflorescent minerals at the seep indicated, nevertheless, some intermittent flow of water.

When the earthflows were first examined in July 1952, the still partially saturated materials, the freshly uprooted vegetation, and the active seep indicated that the flows occurred in spring or early summer of that year. From 1952 to 1956 the earthflows became drier and lost some of their fresh appearance but otherwise changed very little. The only indication of activity was the growth in 1952 and 1953 of a crack from the top of the larger slide (pl. 5) toward the smaller one; the crack showed no change after 1953.

Eight control points were established on and near the larger earthflow in August 1953, and three more were added in October 1954. Control points A, Y_1, and 2 were assumed to be stable, and all traverses were run from them. The maximum error in closure was 0.20 foot; movements, therefore, are assumed to be accurate to at least 0.2 foot. All movements of the control points from 1953 to 1956 were randomly oriented and neither single nor aggregate movement at any point was more than 0.2 foot. These measurements indicated that there was virtually no movement of the earthflow; the slight shifting of the control points undoubtedly reflects minor adjustments of the earthflow material as it became stabilized after the activity of 1952.

Observation of the two earthflows indicates that nearly all of their movement occurred at one time, or at least during a short interval of time. They now seem stabilized and give no indication of future movement.

CABLE SCHOOL SLUMP-EARTHFLOW

The Cable School slump-earthflow in SE¼NE¼NE¼ sec. 25, T. 103 N., R. 72 W., Brule County (fig. 7, pl. 5), was one of the most active landslides during the investigations along the Fort Randall Reservoir. The slide is about 400 by 700 feet and is divided nearly equally between slump and earthflow movements.

The slump-earthflow is on the southwest wall of the Cable School area directly across a small valley from the two Cable School earthflows. The geologic setting is similar: Sharon Springs and Gregory Members of the Pierre Shale are overlain by gravel (fast-draining alluvium) capped by till. Colluvium forms a thin blanket over most of the surface and obscures much of the underlying geology.

There is a zone of springs at and just below the gravel-Gregory Member contact. The ground-water source for the springs must be extensive because the flow was continuous, though small, during the 4 years of landslide observation. The water probably comes from gravels in the eastward extension of the pre-Missouri River White River valley.

The upper part of the Cable School slump-earthflow is a broad but relatively short area, about 400 by 250 feet, composed of numerous slump blocks. Differential settling of the slump blocks has produced a horst-and-graben structure both in the major blocks and in the subsidiary units. A steep scarp 20–30 feet high forms the upper limit of the slump. The lower limit is approximately at the till-gravel contact (pl. 5) near the middle of a second scarp, about halfway down the landslide. Below this point, the slump blocks break up and become part of the earthflow.

The earthflow part of the landslide narrows from almost 300 feet at the gravel-till contact to less than 50 feet at the toe about 400 feet away. The earthflow material above the line of springs, at the gravel-shale contact (fig. 6, pl. 4), behaves more like a debris slide than an earthflow because there is not enough moisture to permit flowage. The springs sap the base of the slump blocks, causing their partial disintegration and movement of the debris downslope as semiplastic masses. As the masses pass through the zone of springs, they become saturated and continue down the slope as an extremely viscous liquid.

The slide probably started as a predominantly slump-type movement with failure occurring in the saturated Pierre Shale immediately below the gravel. Aerial photographs taken in June 1949 show the landslide as a small slump-earthflow less than half the size it was in 1952. By July 1952 it had grown to the approximate size shown in plate 5. The absence of vegetation on the earthflow and freshness of the cracks and scarps showed that much of the movement probably occurred that spring. Although no other large slump blocks developed, saturation from the springs kept the earthflow fluid, and the entire landslide was active at least until August 1956.

Triangulation of 10 control points was started in August 1953. Eight points were located on the landslide; point 9 was established just above the upper scarp and point 10 on the ridgetop above the landslide (fig. 7, pl. 5). Measurements were made approximately monthly during the fall of 1953 and the summers of 1954 and 1955.

Control point 10, on stable ground, was farthest from the triangulation baseline, and because the angles of the other points were of the same general magnitude, it was assumed that the errors in measuring point 10 would be larger than those for any other point. Because the loci of measurements of point 10 all fall within a circle slightly less than 0.5 foot in diameter, the greatest error in any measurement is presumed to be less than 0.5 foot. The movements of point 9 greater than 0.5 foot are believed to represent the effects of creep on the steep slope just above the landslide.

Control points on the slump and flow parts of the landslide showed distinct differences in their rates of movement. On the slump blocks, point 5 moved only about 0.8 foot while point 6 moved about 2 feet. Movements of points 7 and 8 were too random to show any general direction of movement. On the earthflow, however, movements ranged from about 2.7 feet at point 1 to about 26 feet at point 2. Although point 3 along the edge of the flow showed only random movements over periods of a few months, it had a cumulative movement of about 2.7 feet over a period of 2 years. The next larger movement was at point 4 which moved about 14 feet, or five times as far as point 3.

The data show differences in the basic types of movements in the flow and slump parts of the landslide. On the earthflow, movement is dominantly linear with small lateral components. This behavior implies that the earthflow material is moving along a relatively well defined route. Movement of the slump part of the landslide does not show a similar dominant component. Instead, only points 5 and 6, which are at the top of the scarp directly above the earthflow, show any indication of a dominant linear movement. Apparently the individual slump blocks are shifting differentially in more or less random directions except for the blocks in transition from slump to earthflow.

An interesting conclusion derived from the control point measurements is that points on slides vary in rates of movement. Points 3, 4, and 6 decreased in rate of movement during the 2 years that measurements were made. For the first year, points 1 and 2 on the earthflow showed a corresponding decrease in rate of movement, but during the second year, movement at these points accelerated, and so they moved about three times as far during the summer of 1955 as during the summer of 1954. Because the rate of movement at the toe of the earthflow reflects changes that occur farther up the slide, the acceleration at the toe was probably caused by addition of materials to the earthflow from the slump part of the landslide before measurements were started in August 1953.

In summary, the initial failure at the Cable School slump-earthflow probably was by slumping of large unified blocks. As the material was displaced downslope, it became saturated by seepage from the springs at the gravel-shale contact and disintegrated into an earthflow. As the springs continue to sap the base of the slumps above them, more material sloughs off and is incorporated into the saturated earthflow. This process both replenishes the earthflow and removes support from the base of the slump blocks. The increased instability creates continual shift and probably gradual downslope movement of the individual slump blocks.

LANDING CREEK SLUMP-EARTHFLOW

The Landing Creek slump-earthflow (fig. 4, pl. 5) is the largest and most complex landslide on which movement was measured. The basic component is a large slump about 900 feet from top to bottom and 2,000 feet across. Several smaller slumps and earthflows are superimposed along the toe of the main slump.

The landslide is on a ridge that parallels the Fort Randall Reservoir in SE¼NE¼ sec. 25, T. 100, N., R. 72 W., Gregory County, S. Dak. The slide developed above an intensely dissected terrace remnant about 100 feet above the prereservoir Missouri River level, and moved forward onto the terrace. There is no direct connection, consequently, between river erosion and landslide activation here.

The ridge consists of the lower five members of the Pierre Shale and the underlying Niobrara Formation. In prereservoir time the Niobrara Formation formed bluffs about 35–40 feet above the river alluvium. About 35 feet of the Sharon Springs Member overlies the Niobrara Formation, and the Gregory-Sharon Springs contact is at an altitude of about 1,365 feet. The true thickness of the Sharon Springs and the overlying beds is obscured by ancient and recent landslides. A band of the Crow Creek Member about 10 feet thick crops out on several spurs of the terrace remnant at altitudes ranging from 1,375 to 1,420 feet. Above the Crow Creek, and comprising nearly all of the Landing Creek slump-earthflow material as well as the top of the ridge, are the undifferentiated DeGrey and Verendrye Members. Extraneous masses of the Virgin Creek and Mobridge Members, probably remnants of ancient landslides, are found with the DeGrey and Verendrye landslide material, but only the largest masses are indicated on the geologic map (pl. 5).

The Landing Creek slide can be understood by considering its components individually. The main slump is in itself a complex landslide. All of the recent movement has been restricted to the northwest two-thirds of the slump. The southeast third either is older or it stabilized sooner. Though considerably modified by erosion, the southeast third is a jumbled series of slump blocks below a 30- to 40-foot-high scarp. The scarp becomes increasingly fresh to the northwest and shows the greatest indications of recent activity in the northwest third, where it reaches a maximum height of about 70 feet. Immediately below the scarp, in the northwest part of the slump, is a graben about 900 feet long and 150–200 feet wide. The northeast margin of the graben is bounded by a second scarp as much as 20 feet high. The graben and the slide material bordering it have few fractures or other visible evidence of major activity, although the control points on them showed appreciable movements. Farther downslope, however, the ground surface is increasingly fractured and shows signs of major activity.

The small slumps and earthflows along the toe of the main slump make up about one-half its total length. Earthflows predominate; slump-type movement is dominant only in the northwest part of a subsidiary slide at the grid location of 1,100–1,400 feet north and 1,600–2,000 feet east (pl. 5). Two prominent earthflows extend down small valleys to the prereservoir Missouri River level. The largest heads in the north corner of the main slump and is about 750 feet long. It has a dumbbell shape, with the source area about 275 feet wide forming one bulge, and the toe about 225 feet wide forming the other. The ends are connected by a central section about 100 feet wide and 300 feet long which flows down a steep gully.

The Landing Creek slump-earthflow probably started as a large slump on the slope above the terrace remnants. Causes of the original movement can only be inferred, the most likely being that weathering processes over a long time weakened the material so much that it became unstable on the existing slopes. Ground-water and minor gully erosion probably then triggered movement in the weakened mass.

Direct erosion by the Missouri River had no effect on the Landing Creek slump-earthflow, terrace remnants acted as a buffer to protect the upper slopes. The terrace surface, however, was dissected by many small gullies in the general surface. As the slump moved onto the terrace, some material collapsed into these gullies and destroyed the unity of the original slump. Where the gullies were broad and shallow, the slump broke into fairly large blocks; where the gullies were deeper, the material disintegrated and became potential earthflows.

Excellent examples of the effect of the dissected terrace surface on the landslide are the earthflow in the north corner of the landslide and the slump just southwest of it (pl. 5). These subsidiary landslides developed over a large deep gully cut into the original terrace.

Ground water probably has played a greater part in perpetuating and enlarging the activity of the slide than in launching the original movement. The preslide drainage apparently was well integrated, and there was no unusually large supply of ground water available. The initial movement, however, created fractures and closed depressions that trapped most of the subsequent precipitation, thus appreciably increasing the amount of ground water available.

Little is known about the movement of the landslide before the control points were established in August 1953. A map of the area (U.S. Army Corps of Engineers, 1947, sheet 37), which was compiled photogrammetrically from aerial photographs taken in 1945, shows that the landslide was already extensive with at least one subsidiary earthflow that reached the river. Aerial photographs taken in 1949 show the landslide virtually the same size that it was from 1952 to 1956. The largest earthflow has a larger lobe than is shown on the earlier Corps of Engineers' map, and the smaller earthflow to the southeast has a small lobe that also extends into the river. Between 1949 and 1953, there were several small changes: both earthflow lobes in the river grew about one-fifth larger, a small earthflow developed along the toe at the southeast end of the slide, and the slump-earthflow northwest of it nearly doubled its size.

Thirty-one control points were installed on the northwest half of the slump-earthflow in August 1953, and they were measured annually through September 1956. During this time several control points were either covered by the rising reservoir or destroyed by landslide movements and cattle. Some of these points were abandoned; others were replaced.

The maximum error for transit traverse closure for 1953, 1955, and 1956 was 0.33 foot, and consequently all measurements for these years are considered accurate to at least 0.33 foot. In 1954 there was an error of closure of 1.37 feet, of which 1.36 feet is in the north component.

The measurements indicate that there was appreciable movement of all the points on the landslide during the 3-year period, and also that the range in amount of movement was very great. Point 4 had the least movement, about 2 feet, over the entire 3-year period. Measurements are available for only 2 years at points 3 and J, but during that time their movements were of the same order of magnitude as at point 4. The other extreme is point 16, which moved about 100 feet in 3 years.

The control points beyond the limits of the landslide remained fairly stable, with the exception of point 2. Points X, Y, and 1 at the head of the slide were assumed to be stable and were used as controls for orienting the annual traverses. The location of point 15, below the slide, was within the limits of error of measurement during the 3-year period. This fact indicates not only that point 15 was stable, but also that the assumed stability of points X, Y, and 1 was valid. Point 2 was placed on apparently stable ground above the top scarp in 1953, but during the following year the area became an active slump block.

The movement of the control points on the landslide revealed some interesting contrasts between the movement of the slump and the earthflows. All the control points on slump blocks were remarkably constant in their direction of movement, and lateral movements resulting from differential shifting of unit blocks were very small relative to the overall downslope component of the entire landslide. The subsidiary slump block above the largest earthflow showed some exception to this trend. In general, its movement was subparallel to and independent of the other parts of the main slump. In contrast to the unity of the main slump each point on the earthflows was influenced by conditions at its particular site instead of by the conditions over the entire flow.

This behavior is shown best by a comparison of the movements of points J, 14, and 14–B with movements of the other points on the earthflow. Point J, near the earthflow's edge, was at the side of the main current of the earthflow. Its movement, consequently, was much smaller than that of the remainder of the points. At the site of points 14 and 14–B, the direction of flow was deflected by local topography so that movement of these points was at about 45° to the other points of the earthflow.

The rate of movement on the main slump increased from the top to the base of the mass. The material on the lower half of the slump was not pushing a large weight ahead of it and consequently could move farther under the same conditions than the upper part. On the earthflow the magnitude of movements was not so dependent on the relative location of the control point. Point 16 moved farther during the period 1953 to 1954 than did point 17 a short distance below it. Point 14 moved almost as far as point 15–A from 1954 to 1955, and it moved farther than point 16 during the same time. The entire flow, moreover, moved rapidly than the slump moved.

Finally, control points everywhere on the slide had a fairly large movement the first year and smaller, nearly equal, movements during the succeeding years. Apparently the conditions affecting stability had a similar effect on all parts of the landslide despite its complex nature.

PAULSON SLUMP

The Paulson slump (pl. 5, fig. 30) is an old landslide that has been reactivated by the development of two subsidiary landslides on its lower slopes. It is about 500 by 500 feet and has about 90 feet of relief. Its shape is very irregular, largely because the lower slopes have been extensively eroded by an intermittent stream at their base.

The slump is on the southeast side of a Pierre Shale ridge in NE¼SW¼SW¼ sec. 6, T. 98 N., R. 69 W., Charles Mix County, between two intermittent streams tributary to the Missouri River. About 15 feet of the Gregory Member is exposed above the valley-bottom alluvium. The overlying Crow Creek Member is about

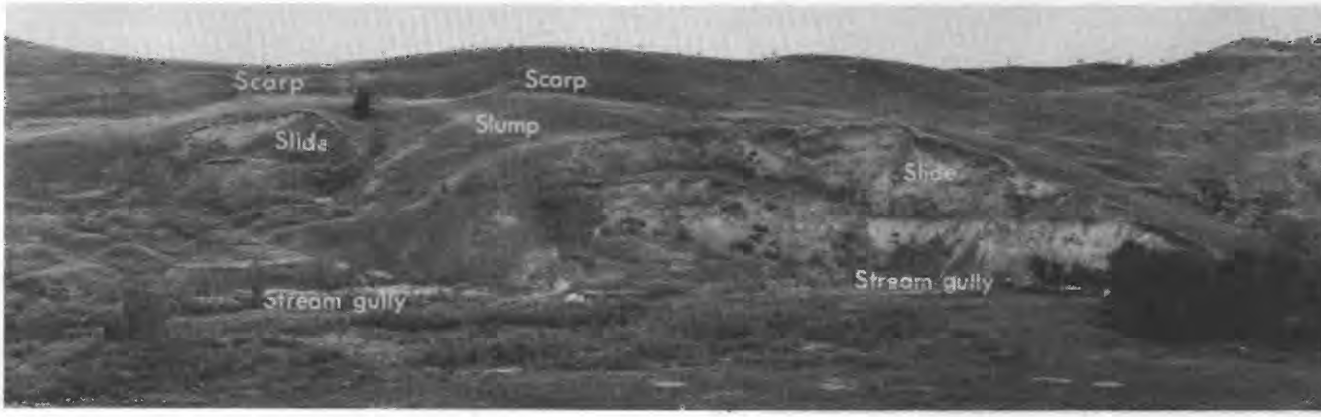


FIGURE 30.—The Paulson slump, a reactivated slump in Pierre Shale. The two active landslides are in the toe of the main slump, which includes the area between the two active slides and the rather flat surface just above them. Photographed August 19, 1956.

5 feet thick in this locality and is overlain by about 80 feet of undifferentiated DeGrey and Verendrye Members.

Interrupting the general slope of the ridge, the old slump formed a flatter area bordered by short steepened slopes that were the remnants of the original scarp. By 1956, reactivation had opened new fractures that closely outlined the top and sides of the original slide. There were also other fractures on the slump itself, especially between the scarps of the subsidiary slides.

The subsidiary landslides are slump-earthflows, although the earthflow parts are not very well developed on either slide. The slide in the southeast part of the old slump is about 225 feet long and 175 feet wide. Although the slide in the northeast part of the old slump was about 275 feet long and 200 feet wide, stream erosion had cut away its northeast corner until its total area is not much greater than that of the southeast slide. Both subsidiary slides have well-developed scarps. The northeast slide has a prominent scarp at its head and a second scarp on the margin of a large active slump block.

The causes of movement in the original slide cannot be determined. The low scarp remnants at the head of the block (fig. 30) and the relatively undisturbed shale exposed in the newer scarps indicate that the movement was small and that the landslide was predominantly a slump. After this movement, the slump became stable. Erosion subsequently modified it until the only remaining evidence of the landslide was the small scarp at the head and the flattened slope on the surface of the rotated slump block.

Aerial photographs show that the subsidiary slides were almost as well developed in 1949 as when the author first examined them in 1952, when they were active but the main slump was still stable. Renewed movement of the main Paulson slump started during 1952–53 and continued to increase as long as the slide was under observation.

Stream erosion probably was the major cause of the subsidiary landslide activity. The normally dry streambed may contain a swiftly flowing stream after heavy rains or during wet seasons. Shortly after a cloudburst in the area, the author saw a boulder more than 1 foot in maximum dimension lodged against a tree near the stream channel. A stream that can transport boulders of this size is capable of major erosion in the Pierre Shale. Although ground water may have contributed to the renewed instability of the old slump, during this study there was never evidence of appreciable amounts of ground water in the landslide.

The northeast subsidiary slide was definitely activated by stream erosion. The southeast slide was probably also caused by erosion, although it is farther from the stream chanel. After these subsidiary slides had removed an appreciable amount of material from the toe of the original slump, it became unstable and began to move again.

Measurements were started in 1952 and were made annually through 1956. Points A–1 to A–5 and B–1 to B–5 were installed in October 1952, and additional points to provide a closed traverse were installed in August 1953. The measurements in 1952 were taped slope distances between points; these measurements were not included with the movement vectors shown on plate 5, because they were approximate and did not indicate true directions of movement. Closed transit traverses were used for measurements from 1953 to 1956. The maximum error in closure was 0.14 foot, and consequently all movement vectors are assumed to be accurate within 0.14 foot.

From 1952 to 1956 the general trend on the Paulson slump was toward increasingly greater movement. Between 1952 and 1953 only points A–3, A–5, and B–5 moved appreciably, and the greatest movement, at A–5, was not more than 0.5 foot. Movement of all points on the slump between August 1953 and October 1954 indicated that the entire slump was then active. The amount of movement ranged from about 0.3 foot at points B–2, B–3, and B–4, to about 0.8 foot at point A–5. There was less movement from October 1954 to July 1955: points B–2, B–3, and B–4 each moved about 0.1 foot, which is within the possible limits of error, and point B_1–5 moved about 0.5 foot, the most movement of any point on the main slump. The greatest movements occurred during the last year of observation, July 1955 to September 1956. All the points on the northeast half of the slump were destroyed, either by movement or by cattle, but on the southwest half the movements ranged from about 0.6 foot at B–2 to about 1.8 feet at B_1–5. The appearance of new fractures and small scarps on the reactivated slump also showed that differential movement increased during this period.

The preceding observations indicate that renewed large-scale landsliding was developing in the Paulson area between 1952 and 1956. Renewed activity of the old slump began near the toe where the subsidiary slides had recently disturbed its equilibrium. Activity, renewed in one part of the slump, soon involved the entire area. Development of fractures throughout the slump block indicated that the mass was losing its unity and the lower half, at least, might disintegrate into an earthflow. The combination of a disintegrating slump unit and a potentially erosive stream at the toe could easily cause extensive landsliding.

SURPLUS PRECIPITATION, GROUND WATER, SLOPE STABILITY: CORRELATIONS

In this final section an attempt is made to correlate surplus precipitation available for ground-water storage, landslide activity, and ground water. Emphasis throughout the report has been on the relation between ground water and landslide activity, but little has been said about the possibility of correlations between (1) percentage of surplus precipitation available for storage as ground water, (2) the ground-water levels, and (3) slope stability. Such correlations can be valuable for prediction and interpretation of slope stability conditions.

PRECIPITATION AVAILABLE FOR STORAGE AS GROUND WATER

The amount of moisture potentially available to replenish and increase ground-water supplies is not easily determined. Precipitation can be recorded, but only a fraction of the moisture from precipitation ever becomes ground water. The rest is lost either to the streams and rivers as runoff or the atmosphere as evaporation and transpiration (evapotranspiration). Although runoff is an effective agent of erosion, it is less than 5 percent of the normal annual precipitation (Visher, 1954, p. 258). The average annual runoff at Pickstown thus amounts to less than 1 inch of precipitation per year. Inasmuch as runoff represents such a small part of total precipitation, only the relation between evapotranspiration and precipitation was used to determine the approximate amount of water available to affect ground-water supplies.

Evapotranspiration is a direct function of the climate and vegetation. It involves precipitation, temperature, hours of daylight, relative humidity, types of vegetation, and other variable factors. Adequate methods for directly measuring evapotranspiration have not yet been developed, because there is no direct way to satisfactorily measure the moisture that plants release to the atmosphere.

Thornthwaite (1948) has worked with basic climatic data to determine evapotranspiration without actual measurements of the rate of water loss from the ground to the atmosphere. He has developed empirical methods for computing "potential evapotranspiration," the amount of evapotranspiration that will occur under given climatic conditions if there is an unlimited source of water. Thornwaite cites evapotranspiration in a desert where water and vegetation are limited, and in a desert irrigation project where unlimited water is available and vegetation consequently is plentiful, as examples of natural and potential evapotranspiration.

Although the calculated potential evapotranspiration values are only approximate (Thornthwaite, 1948, p. 91), the basic concept is valuable in providing a comparative method of estimating the rate at which moisture can return to the atmosphere from the ground.

RELATION BETWEEN PRECIPITATION AND POTENTIAL EVAPOTRANSPIRATION AT PICKSTOWN

Pickstown is the only station along the Fort Randall Reservoir that has climatic data available (U.S. Weather Bureau, 1948-56) for investigation of the precipitation–potential evapotranspiration relation. Variations along the reservoir are probably fairly small, however, and Pickstown is assumed to be typical of the entire area.

The potential evapotranspiration values computed by Thornthwaite's method and the precipitation record at Pickstown for 1948 through 1956 are shown in a graph at the bottom of figure 31. Periods and amounts of precipitation surplus, and deficiency are shown by patterns; total annual values and surpluses or deficiencies are shown below the graphs. Curves for annual averages for the entire period of record are shown at the left side of figure 31.

The curves of annual averages show that both precipitation and potential evapotranspiration range from minima during the winter to maxima during the summer. The potential evapotranspiration has the greatest range, from almost no evapotranspiration during the winter months to more than 6 inches during July. The precipitation curve is bracketed by the evapotranspiration curve and shows that evapotranspiration is less than precipitation during the winter and greater during the summer. Precipitation in excess of evapotranspiration needs, therefore, is expected during the winter rather than during the summer.

The annual precipitation and potential evapotranspiration curves generally resemble the averaged annual 1948-56 precipitation curves. Annual precipitation ranges from 12.5 to 29.8 inches, but potential evapotranspiration ranges from 24.1 to 30.3 inches. Similarly, annual evapotranspiration curves vary less from their average than do the precipitation curves. Occasional irregularities show disparities between individual months, but variations from the 1948–56 curve for any year are relatively uniform and reflect a generally warm or cool year. The precipitation curves are more erratic. Although the trend is from a winter minimum to a summer maximum, records for individual months often appear to reverse this pattern.

The annual precipitation–potential evapotranspiration relations are similar, despite fluctuations from the average. Every winter and spring precipitation is greater than the amount of water used for evapotranspiration. The precipitation deficiency of summer, however, almost always exceeds the precipitation surplus of winter. Only 1 year, 1951, had a precipitation surplus (5.7 inches), and just 1 month, July, had an appreciable deficiency. Ground-water intake probably was continuous, therefore, for most of that year.

CUMULATIVE PRECIPITATION SURPLUS

Ground water can trigger landslide movement or condition a potential landslide mass for trigger action by another agent. The triggering effect occurs during or very soon after a period of excess precipitation. Ground water, to be a general cause of landsliding, requires an influx of ground water over an extended period of excess precipitation and involves a slow buildup of stresses and other factors that affect the stability of a mass.

A 12-month cumulative precipitation surplus (cps) curve (fig. 31, top) was constructed from monthly precipitation and potential evapotranspiration data in order to show periods of extended precipitation that could affect ground-water conditions. Each point on the curve indicates the total precipitation surplus or deficiency (negative surplus) for a 12-month period. The 12-month period was chosen because it seemed a reasonable time for ground water to permeate the shale, and cumulative values for 12 months automatically average the effects of normal seasonal variations in precipitation surplus.

The curve shows that a cumulative precipitation surplus for any 12-month period is rare; only 1 year of eight had a surplus. From June 1951 to May 1952 monthly surplus was about 5 inches, which helped provide a cumulative precipitation surplus that was 13 inches more than the average cps values for the remainder of the time of record. Also noteworthy during that 8-year time of record, the monthly precipitation showed an average deficiency of about 8 inches. June 1951 to May 1952 was therefore the only period in which there was enough precipitation to cause an appreciable increase in ground water.

CORRELATION BETWEEN SURPLUS PRECIPITATION AND SLOPE STABILITY

The hypothesis that there is a definite relation between surplus precipitation and slope stability is based on information in figure 31 and the assumption that most landslides are the result of general instability, supplemented by a trigger action that initiates movement. There may actually be several causes of instability, but only the apparent effects of ground water are considered in detail here.

Temporary increases in available ground water may raise the water table and also cause partial saturation of material above the water table. If such an influx of water continues for many months, the resulting changes in ground-water conditions may be sufficient to trigger landslides on some slopes. On most slopes, however, it will be a general cause of instability. Therefore, abnormally high peaks in the 12-month cumulative precipitation curve (for example from June 1951 to May 1952, fig. 31) may coincide with an appreciable increase in the amount of ground water and a corresponding decrease in general slope stability.

Periodically both general ground-water conditions,

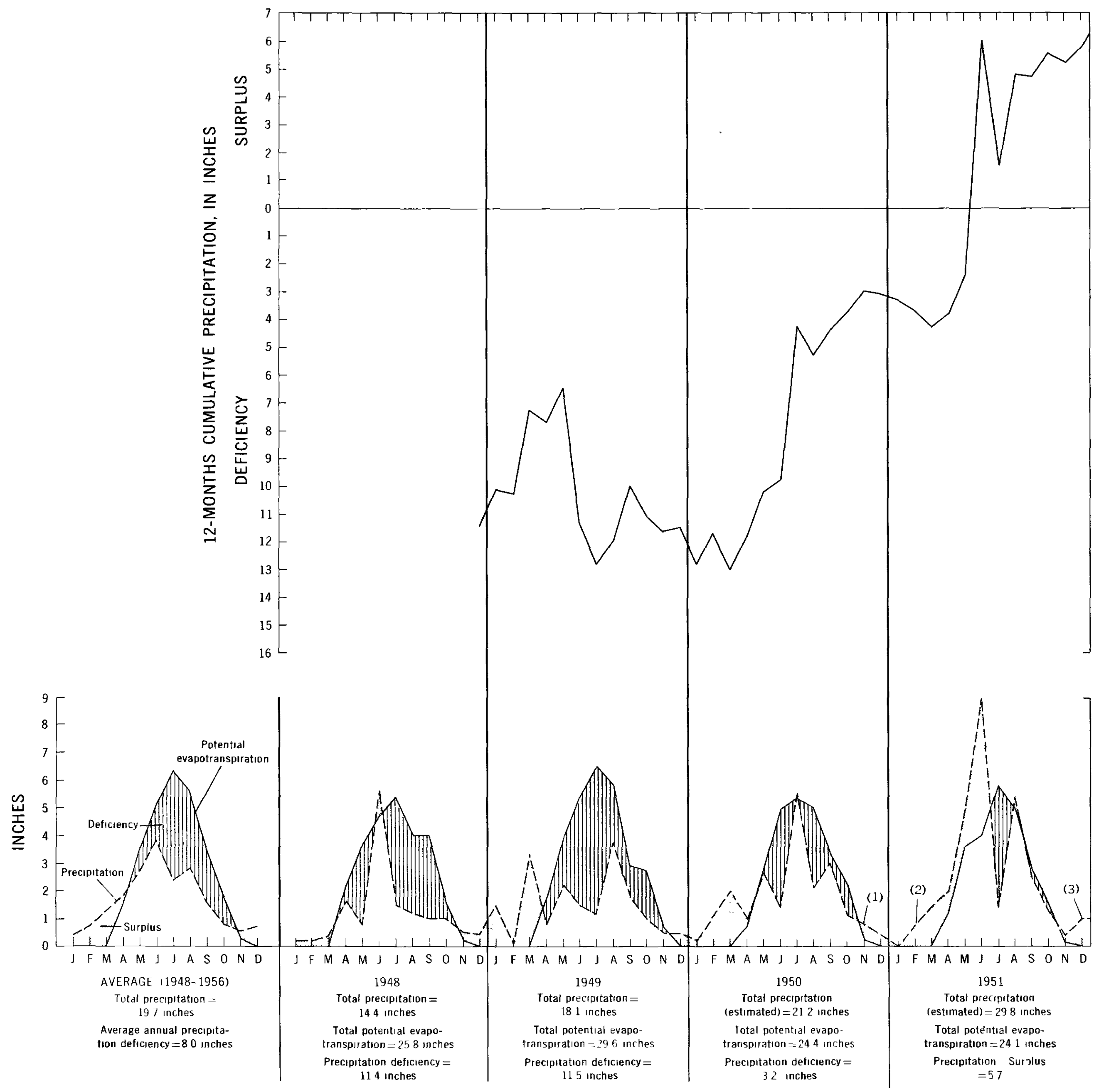
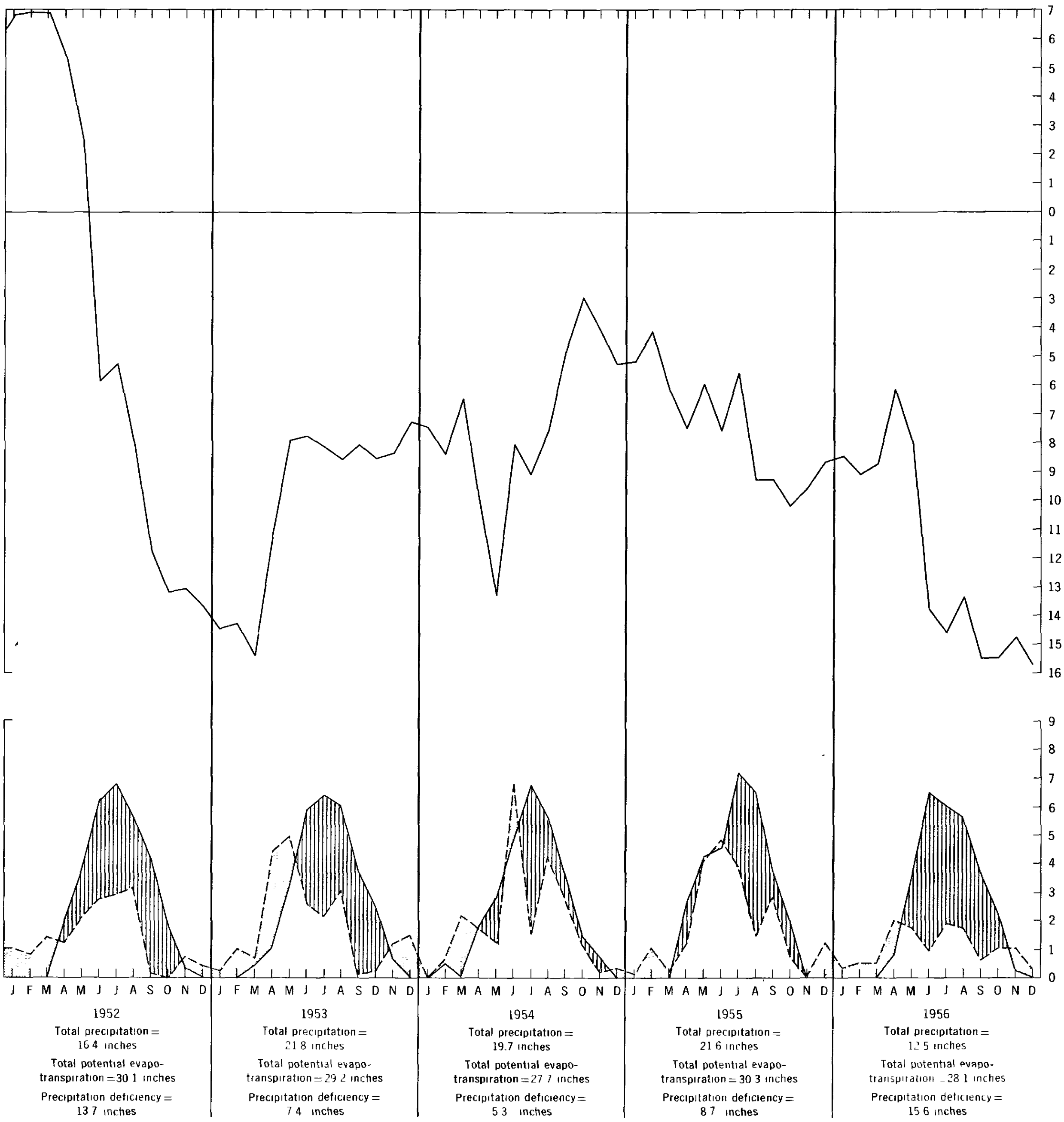


FIGURE 31.—Correlation between precipitation and potential evapotranspiration at Pickstown. Precipitation data from U.S. Weather at nearby stations. (2) No record at Pickstown; about average at nearby

evidenced by the 12-month cumulative precipitation surplus curve, and temporary ground-water conditions, shown by precipitation–potential evapotranspiration relations, should favor landslide activity, and exceptional amounts of activity may occur on both new and old landslides. Only one such period occurred from 1948, when the Pickstown climatic records were commenced, until 1956, and that was in the spring of 1952. The 12-month cumulative precipitation surplus curve was at a peak, and the precipitation and potential evapotranspiration curves showed a precipitation surplus.

During the first year of landslide examinations along the Fort Randall Reservoir (1952), an outstanding feature was the number of fresh apparently active landslides along the walls of the Missouri River trench and its tributaries from the streambeds tc the uplands.

The slides described in this report either originated in or were still active in 1956. The Highway 16 slump



Bureau climatological data (1948–56); all data given to nearest 0.1 inch. (1) No record at Pickstown; below average stations. (3) No record at Pickstown; well above average at nearby stations.

and the two Cable School earthflows all originated in 1952, and only minor subsequent movements occurred up to 1956. The Cable School slump-earthflow had major movement in 1952, probably greater than any later movement up to 1956. The Landing Creek slump-earthflow was active in 1952, but there are no data to compare the relative amount of movement that year with later movements. The Paulson slump alone showed increasing activity after 1952. It also is the only landslide in which the role of ground water apparently was secondary to another cause, stream erosion at the toe.

After the drop in the 12-month cumulative precipitation surplus curve, starting in May 1952 (fig. 31), there was a gradual increase in overall slope stability. The control points on all the landslides except the Paulson slump showed either negligible movements or a general decrease in rate of movement.

At Pierre, S. Dak., about 60 miles northwest of Chamberlain, Crandell (1958) noted a concentration of landslide activity in the spring of 1952. He stated (1958, p. 77): "Fewer than 10 large new landslides were observed in the Pierre area during the period 1948–51; in the spring of 1952, however, many new slumps and flows occurred in the area and some old slumps were reactuated." The increased landslide activity in the spring of 1952 is attributed by Crandell to an abnormally large amount of ground water.

Precipitation, potential evapotranspiration, and 12-month cumulative precipitation curves were prepared for the Pierre area to check the surplus precipitation-landslide activity hypothesis with Crandell's comments on landslide activity. The curves differed in detail from the ones at Pickstown, but the general trends were similar. The 12-month cumulative precipitation surplus curve at Pierre shows a deficiency from 1948 to the end of 1951. This period is followed by a surplus from December 1951 through May 1952, with a maximum of 4 inches in March. Then the curve again shows a deficiency from June 1952 through 1956. The precipitation and potential-evapotranspiration curves, moreover, show a more pronounced precipitation surplus at Pierre during the winter and spring of 1952 than there was at Pickstown. This agreement between observed conditions of greater than normal precipitation and landsliding at Pierre seems to validate further the hypothesis that precipitation surpluses can be correlated with periods of excessive landslide activity.

CORRELATION BETWEEN SURPLUS PRECIPITATION AND GROUND-WATER LEVELS

Available precipitation was compared with ground-water levels in the piezometers to check the presumed relation of precipitation surplus to ground-water conditions. A qualitative analysis was made for the piezometers near Pickstown that appeared to be in equilibrium with existing ground-water conditions. The average water level in each piezometer and the average monthly precipitation surplus (negative values included) were calculated first. Then the number of months in which above-average water levels and above-average precipitation surpluses, or below-average water levels and below-average precipitation surpluses, coincided was determined for each piezometer. The total number of times that above- or below-average values coincided was stated as a percentage of the total number of water-level measurements at each piezometer. The calculations were repeated using cumulative precipitation surplus values for periods increasing from 2 to 12 months by 1-month increments.

The results of the analysis were plotted graphically (fig. 32). Graphs of individual piezometers are divided into three groups on the basis of depth: (A) 12 to 30 feet, (B) 55 to 59 feet, and (C) 81 to 88 feet. Percentage values on the curves represent the degree of correlation relative piezometer water levels and relative cumulative precipitation surplus values. Complete correlation is 100 percent; random distribution is 50 percent; and complete inverse correlation is 0 percent.

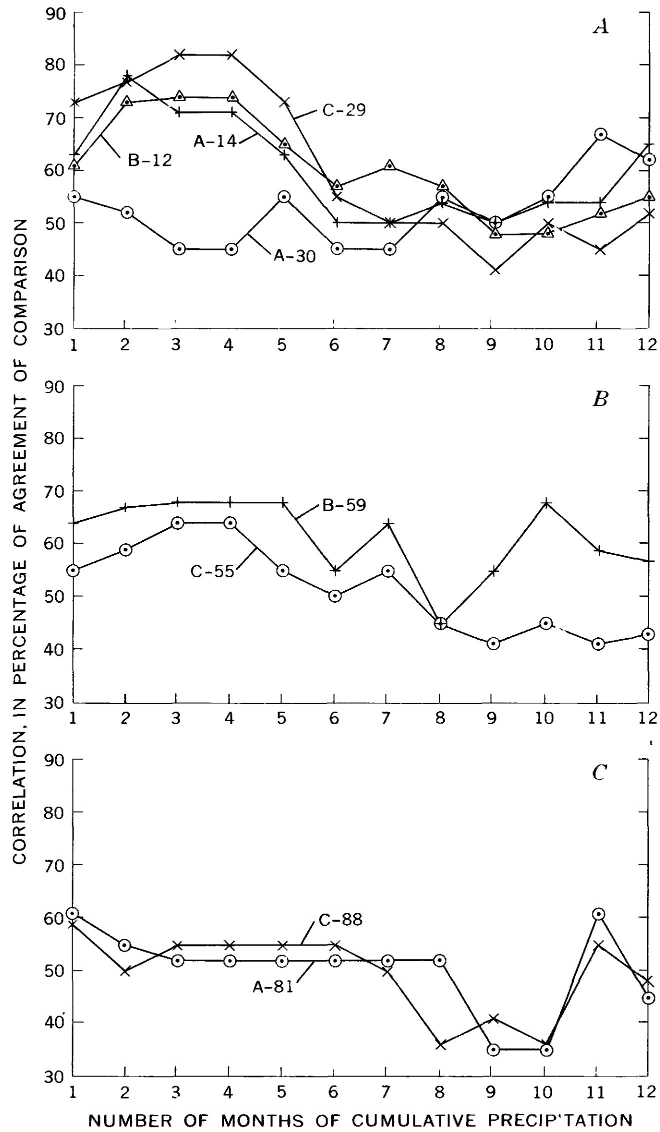


FIGURE 32.—Correlation between above- or below-average piezometer water levels and above- or below-average cumulative precipitation surpluses. *A*. Piezometer depths 12–30 feet; *B*. piezometer depths 55–59 feet; *C*. piezometer depths 81–88 feet. Number following letter designation indicates depth.

The curves show a relatively short-term relation between precipitation surplus and ground-water levels in the shallower piezometers (fig. 32, A, B), but effects are random in the deepest ones (fig. 32, C). Percentage values of the shallowest piezometers, except A–30, are above for the 2-, 3-, and 4-month periods of cumulative precipitation surplus. For the 55- to 59-foot piezometers there is a similar but smaller rise in the 2- to 5-month values. The inference is that short-term (a few months) precipitation surplus conditions have an appreciable effect on ground-water conditions down to depths of at least 30 feet. These effects decrease with depth; they become minor at about 60 feet and ineffective at greater depths.

The correlation between near-surface ground-water conditions and the short-term cumulative precipitation surplus values strengthens the assumption that ground water triggers landslides. The annual winter and spring precipitation-surpluses (fig. 31) give maximum values for 2- to 4-month cumulative precipitation surpluses in the spring, when landslide activity is greatest. The long-term cumulative precipitation surplus values (fig. 32)—in particular, the 12-month surpluses—show no correlation with piezometer water levels. Two factors may contribute to this apparent lack of correlation. The first is that the 12-month cumulative precipitation values were considered an indication of ground-water conditions in the Pierre Shale as a whole, whereas the piezometers used for ground water-precipitation surplus correlation probably reflect special conditions in the shale. Data only from piezometers in which water levels had reached equilibrium rapidly were used for the correlation because they gave the longest periods for comparison. The correlation data, as a result, probably are based on ground-water conditions in shale fractures, which would differ from conditions in massive shale. The second factor is the short period for which comparisons were made, inasmuch as the ground-water data used were only from February 1954 through November 1956. Although a correlation was found between landslide activity and a 12-month cumulative precipitation surplus after only one period of excess precipitation surplus might not be confirmed until at least two periods of abnormally high rainfall were observed.

CONCLUSION

The available data strongly suggest correlation between (1) surplus precipitation available to augment ground-water supplies, (2) landslide activity, and (3) ground-water levels. Both the short-term ground-water effects believed to trigger landslides and the long-term ground-water changes that have general effects on overall slope stability seem to indicate a correlation.

The best correlation seems between winter and spring precipitation surpluses and spring landslide activity triggered by ground water. The 2- to 4-month cumulative precipitation surpluses are highest in the spring after several months of surplus precipitation during the winter and early spring. Near-surface (less than about 60 ft deep) ground-water levels roughly reflect the cumulative precipitation surplus values, and they also tend to be above average in the spring, when most landslide activity occurs. It seems reasonable, therefore, to postulate that after 2 to 4 months of surplus precipitation, the surplus water that infiltrates the ground, along with ground water already present, can create sufficiently unstable conditions to promote landsliding.

The relation between available surplus precipitation, landslide activity, and long-range ground-water changes is less certain. There is an interesting coincidence between an extensive period of positive 12-month cumulative precipitation surplus values and abnormally active landsliding in the spring of 1952. Ground-water data are unavailable for that year; however, without positive evidence that ground-water conditions were unfavorable to earth movements, a logical conclusion is that a direct correlation exists between 12-month cumulative precipitation surplus values, ground-water conditions, and general slope stability conditions.

REFERENCES

American Society for Testing Materials, 1950. Standard method of mechanical analysis of soils, *in* A.S.T.M. procedures for testing soils: Am. Soc. Testing Materials, p. 44–55.

Baldwin, Brewster, and Baker, C. L., 1952, Areal geology of the Iona quadrangle: South Dakota Geol. Survey (map).

Casagrande, Arthur, 1949, A non-metallic piezometer for measuring pore pressures in clay, *in* Soil mechanics in the design and construction of the Logan Airport: Jour. Boston Soc. Civil Engineers, app., v. 36, no. 2, p. 214–221.

Crandell, D. R., 1950, Revision of Pierre shale of central South Dakota: Am. Assoc. Petroleum Geologists Bull., v. 34, no. 12, p. 2337–2346.

——— 1952, Origin of Crow Creek member of Pierre shale in central South Dakota: Am. Assoc. Petroleum Geologists Bull., v. 36, no. 9, p. 1754–1765.

———1958, Geology of the Pierre area, South Dakota: U.S. Geol. Survey Prof. Paper 307, 83 p.

Croxton, F. E., and Cowden, D. J., 1955, Applied general statistics [2d ed.]: New York, Prentice-Hall, Inc., 843 p.

Curtiss, R. E., 1950, Cement materials near Mobridge, South Dakota, *in* Report of the South Dakota State Cement Commission to the 32d session of the Legislative Assembly of the State of South Dakota: p. 63–81.

Dixon, W. J., and Massey, F. J., 1951, Introduction to statistical analysis: New York, McGraw-Hill Book Co., Inc., 370 p.

Flint, R. F., 1955, Pleistocene geology of eastern South Dakota: U.S. Geol. Survey Prof. Paper 262, 173 p.

Goddard, E. N., chm., and others, 1948, Rock-color chart: Natl. Research Council, 6 p. (repr., 1951, by Geol. Soc. America).

Graton, L. C., and Fraser, H. J., 1935, Systematic packing of spheres, with particular relation to porosity and permeability: Am. Assoc. Petroleum Geologists Bull., v. 43, no. 8, pt. 1, p. 785–909.

Gries, J. P., 1942 Economic possibilities of the Pierre shale: South Dakota Geol. Survey Rept. Inv. 43, 79 p

Grim, R. E., 1953 Clay mineralogy: New York, McGraw-Hill Book Co., Inc., 384 p.

Holtz, W. G., and Gibbs, H. J., 1951, Consolidation and related properties of loessial soils, *in* Symposium on consolidation testing of soils: Am. Soc. Testing Materials Spec. Tech. Pub. 126, p. 9–33.

Meek, F. B., and Hayden, F. V., 1861, Descriptions of new Lower Silurian (Primordial) Jurassic, Cretaceous, and Tertiary fossils, collected in Nebraska Territory * * *, with some remarks on the rocks from which they were obtained: Acad. Nat. Sci. Philadelphia Proc., p. 415–447.

Olds, E. G., 1949, The 5 percent significance levels for sums of squares of rank differences and a correction: Annals Math. Statistics, v. 20, p. 117–118.

Petsch, B. C., 1946, Geology of the Missouri Valley in South Dakota: South Dakota Geol. Survey Rept. Inv. 53, 78 p.

Petsch, B. C., and Fairbanks, Donald, 1952, Areal geology of the Chamberlain quadrangle: South Dakota Geol. Survey (map).

Schultz, L. G., 1965, Mineralogy and stratigraphy of the lower part of the Pierre Shale, South Dakota and Nebraska: U.S. Geol. Survey Prof. Paper 392–B, p. B1–B19.

Searight, W. V., 1937, Lithologic stratigraphy of the Pierre formation of the Missouri Valley in South Dakota: South Dakota Geol. Survey Rept. Inv. 27, 63 p.

Searight, W. V., and Meleen, E. E., 1940, Rural water supplies in South Dakota: 4 volumes, prepared by the Work Proj. Admin. as a report on the well survey conducted as Work Proj. Admin. Official Proj. 665–74–3–126: sponsored by the Ext. Service and the Expt. Sta., South Dakota State Coll., in coop. with the State Geol. Survey.

Simpson, H. E., 1954, Geology of an area about Yankton, South Dakota: U.S. Geol. Survey open-file report, 317 p.

——— 1960, Geology of the Yankton area, South Dakota and Nebraska: U.S. Geol. Survey Prof. Paper 328, 124 p. [1961].

Stevenson, R. E., and Carlson, L. A., 1950, Areal geology of the Bonesteel quadrangle: South Dakota Geol. Survey (map).

Stokes, W. L., and Varnes, D. J., 1955, Glossary of selected geologic terms with special reference to their use in engineering: Colorado Sci. Soc. Proc., v. 16, 165 p.

Terzaghi, K. C., 1950, Mechanism of landslides, *in* Paige, Sidney, chm., Application of geology to engineering practice: Geol. Soc. America, Berkey volume, p. 83–123.

Terzaghi, K. C., and Peck, R. B., 1948, Soil mechanics in engineering practice: New York, John Wiley & Sons, Inc., 566 p.

Thornthwaite, C. W., 1948, An approach toward a rational classification of climate: Geog. Rev., v. 38, p. 5–94.

Truesdell, P. E., and Varnes, D. J., 1950, Chart correlating various grain-size definitions of sedimentary materials: U.S. Geol. Survey.

U.S. Army Corps of Engineers, 1947, Missouri River survey, Gavins Point near Yankton, S. Dak., to Stanton, N. Dak.: U.S. Army Corps of Engineers, Omaha district, sheets 23, 26, 30, 37, 39, 42 of topographic map series.

U.S. Bureau of Reclamation, 1951, Mechanical analyses of earth materials, *in* Earth manual: U.S. Bur, Reclamation, tentative ed. (repr. with revisions Feb. 1952), p. 135–145.

U.S. Department of Agriculture, 1941, Climate of South Dakota, *in* Climate and man—1941 Yearbook of agriculture: U.S. Dept. Agriculture p. 1109–1118.

U.S. Weather Bureau, 1948–56, Climatological data, South Dakota section, *in* Climatological data for the United States by sections: U.S. Weather Bur. [volumes for the years 1947–55].

Varnes, D. J., 1958, Landslide types and processes, Chap. 3 *of* Eckel, E. B., ed., Landslides and engineering practice: Natl. Research Council, Highway Research Board Spec. Rept. 29, p. 20–47.

Visher, S. S., 1954, Climatic atlas of the United States: Harvard Univ. Press, 403 p.

Warren, C. R., 1952, Probable Illinoian age of part of the Missouri River, South Dakota: Geol. Soc. America Bull., v. 63, No. 11, p. 1143–1155.

Warren, C. R., and Crandell, D. R., 1952, Preliminary report on geology of part of the Chamberlain quadrangle, South Dakota: U.S. Geol. Survey open-file report. 45 p.

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