U.S. GEOLOGICAL SURVEY

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..es in the vicinity of the Fort Randall Reservoir

South Dakota

by Christopher F. Erskine

UNITED STATES DEPARTMENT OF THE INTERIOR

GEOLOGICAL SURVEY

[REPORTS - OPEN FILE]

# Landslides in the vicinity of the Fort Randall Reservoir, South Dakota

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This report is preliminary and has not been edited or reviewed for conformity with Geological Survey standards or nomenclature

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

by Christopher F. Erskine

Abstract

This report covers the first 4 years (1952 through 1956) of a project to investigate landslides in the vicinity of the Fort Randall Reservoir in South Dakota and to determine the effects of the new reservoir on landslide activity.



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The Fort Randall Reservoir, which first held water in 1952, is one of the several multipurpose reservoirs that the Corps of Engineers have created along the Missouri River. It is located in south-central South Dakota and extends from near the South Dakota-Nebraska boundary upstream to the northwest for about 130 miles. The walls of the Missouri River valley and its tributaries consist predominantly of Upper Cretaceous Pierre shale, a bentonitic mudstone that is subject to extensive landsliding. Alluvium that underlies terraces and comprises the flood-plain deposits is the next most important geologic unit in the Missouri River trench.

The reservoir had little effect on slope stability during this period and emphasis has been placed on investigation of the nature and causes of the landslides.

The landslides that occur along the reservoir are of 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.



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All of the landslide types are activated primarily by erosion and ground water. These agents may operate as general causes or as trigger actions. General causes are long-term processes that decrease the overall stability of a slope and prepare it for landsliding. Trigger actions are rapid, often recurring, processes that set off individual landslides in places where general causes have already reduced the slope stability. The most important general cause of landslides is erosion. Ground water can act both as a general cause and as a trigger action where oversteepened, potentially unstable slopes exist. It can affect slope stability in five ways: (1) increase of weight in a potential landslide, (2) lubrication, (3) hydrostatic pressure, (4) piping, and (5) chemical reactions.

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



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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 topographically related to erosion by tributary streams whereas many of the old, partially obscured, landslide remnants, including most of the larger slides, show an apparent topographic relationship to the main river valley. Landslide activity probably was last at a maximum during the latest glacial advance.

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



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The apparent relationship between landslide activity and ground-water conditions instigated a ground-water observation program that began in 1954. Twenty porous tube piezometers were installed at eight sites in the vicinity of 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 that the direction and rate of ground-water movement depends on the orientation and size of the fractures.



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Movements of control points on 5 landslides--1 earthflow, 1 slump, 3 slump-earthflows--were measured over a 3-year period. The data indicate that these three types of landslides have more or less characteristic activity patterns. Earthflows have one short period of activity, perhaps lasting a maximum 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. A rule of thumb is that the active life of a slump-earthflow is proportional to its size.



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The final section of the report discusses possible correlations 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. ~~and~~ the difference between actual precipitation and potential evapotranspiration is the precipitation surplus or deficiency that is available to supplement ground-water supplies. The available information showed a distinct correlation between surplus precipitation and landslide activity. Long-term surplus precipitation values as represented by a 12 months' 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. Precipitation surpluses usually exist 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 ~~is~~ are too limited to either prove or disprove that the precipitation surplus actually affects ground-water conditions. The available data, however, ~~is~~ are fully compatible with the precipitation surplus-landslide activity relationship. It seems very likely, as a result, that times of maximum landslide activity occur when infiltration of surplus precipitation from the surface appreciably augments the ground-water supplies. It may be possible, moreover, to predict periods of appreciable landslide activity by appraisal of 12 months' cumulative precipitation surplus curves.



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## Introduction

This report summarizes the first phase of a project to study landsliding in the vicinity of the Fort Randall Reservoir in South Dakota (fig. 1). The reservoir occupies the Missouri River valley,

Figure 1.--Near here.

often referred to as the Missouri River trench; Fort Randall Dam, 5 miles above the point where the Missouri River crosses the Nebraska-South Dakota State line, forms the lower limit. The reservoir extends upstream for about 130 miles northwesterly nearly to Big Bend, a large meadner 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 is essentially a bentonitic mudstone. Pierre shale is very susceptible to landsliding when it is saturated or nearly saturated; ground-water conditions, therefore, have a close connection with 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 (fig. 2) as a result, are well suited to the study of the effect of

Figure 2.--Near here.

changing ground-water conditions on slope stability.



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Figure 1.--Index map of the Fort Randall Reservoir landslide project.



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Figure 2.--Fort Randall Reservoir landslide studies.

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## Purpose and scope

The purpose of this study is to add to the basic knowledge of landsliding, particularly 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. The method is to study slope stability and factors affecting it throughout a large area rather than to make exhaustive studies of particular landslides.


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The scope is restricted to three specific problems of landsliding in order to place a practical limit on the investigations. The first problem is the relationship between landslides and ground-water conditions. Because the Pierre shale is relatively impermeable, ground-water conditions will take many years to reach equilibrium with the free-water level in the reservoir. Little can be said at this time, therefore, about the effect of the reservoir on landsliding except in the areas where wave erosion and saturation are causing landslides along the reservoir shore. The emphasis in this report is placed on those ground-water effects that are independent of the reservoir and that result from seasonal precipitation variations as well as short- and long-range climatic changes. The second problem involves the factors that control times and rates of movement both of individual landslides and of landslides collectively. The third problem to be considered is the relationship between composition of the Pierre shale ~~M~~embers and their susceptibility to landslides. The effects on slope stability of major compositional variations, such as quartzite versus shale, are quite obvious, but little is known about minor compositional differences such as occur between individual members of the Pierre shale.



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The landslide investigation was divided into four parts, each of which is essentially a separate entity and a fifth part that correlates data from several of the preceding four. The individual investigations are: (1) analysis of the Pierre shale, (2) appraisal of selected study areas, (3) ground-water investigations, and (4) measurement of landslide movements. The fifth phase of the investigation is correlation between surplus precipitation available for ground-water storage, landslide activity, and ground-water conditions.


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## Fieldwork

Most of the investigations were carried on during the summers of 1952-55 inclusive. The 1956 field season was limited to about 1 month spent field checking and measuring landslide movements. Personnel in the office of the Ground Water Branch, U.S. Geological Survey, at Huron, S. Dak., made the ground-water measurements during the winters of 1954 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.

Field data were plotted both on maps and on aerial photographs and later transferred to base maps in the office. The measurements of landslide movements were made by periodically relocating control points on the landslides. This was done with a transit either by triangulation or by closed traverse. Porous tube piezometers were used to observe ground-water levels. Subsurface samples for the Pierre Shale analysis were obtained with a power auger.


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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 her later continuing work on the manuscript the report could never have been completed.

The success of the ground-water observation program must be credited to personnel in several agencies. J. R. Jones, F. C. Koopman, and G. A. LaRocque, Jr., of the Huron, S. Dak., office of Ground Water Branch, U.S. Geological Survey, were advisers for the overall program. Information about porous tube piezometers was furnished by W. W. Daehn, E. E. Esmiol, and J. P. Gould of the Dams Branch, U.S. Bureau of Reclamation. W. H. Jackson of the U.S. Geological Survey designed the device used to measure water level in the piezometers. The Omaha District, U.S. Army Corps of Engineers, furnished a churn drill and crew to install the piezometers.

Personnel of the Omaha District, U.S. Army Corps of Engineers furnished the project with maps, services, space for field offices, and storage facilities and cooperated with the project in every way possible. Especial thanks must be given to A. H. Burling, J. A. Trantina, and L. B. Underwood, who gave the author many good suggestions in the course of numerous discussions and field conferences.

Many other people 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., kindly allowed the installation of piezometers on city land. The author also appreciates the cooperation of landowners in the vicinity of the Fort Randall Reservoir who permitted landslide investigations on their land.



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## Geographic setting

### Culture

The land around the 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.

There are only four permanent towns in the reservoir area: Chamberlain (1950 population: 1,912); Fort Thompson (1950 population: 250); and Oacoma (1950 population: 231). Pickstown (1960 population about 600), at the Fort Randall damsite, is a town built by the Corps of Engineers, U.S. Army.

The roads are mostly in the uplands along the reservoir. U.S. Highways 16 and 18 cross the reservoir at Chamberlain and at the Fort Randall Dam respectively. In addition to these two black-topped highways there is a good network of State and county graveled and graded roads on the uplands. There are few roads along the valley walls of the Missouri River and its tributaries, and in most places the reservoir shore can be reached only by jeep or on foot.

The Chicago, Milwaukee, St. Paul and Pacific Railroad line from Mitchell, S. Dak., to Rapid City, S. Dak., crosses the reservoir near 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 line at Lake Andes to the damsite 7 miles to the south.

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## Physiography

Flint (1955) divided South Dakota into 12 physical divisions (fig. 3). The Fort Randall Reservoir lies entirely within the Missouri

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Figure 3.--Near here.

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River trench, but locally the landslide areas extend eastward into the Coteau du Missouri and westward into the Pierre Hills.



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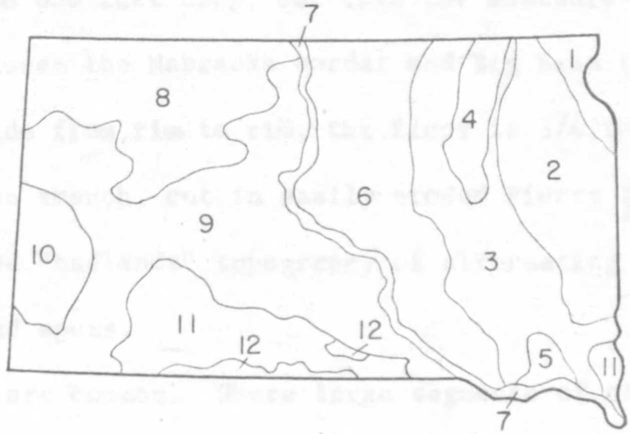


Figure 3. Physical divisions of South Dakota. 1, Minnesota River-Red River lowland; 2, Coteau des Prairies; 3, James River lowland; 4, Lake Dakota plain; 5, James River highlands; 6, Coteau du Missouri; 7, Missouri River trench; 8, northern plateaus; 9, Pierre hills; 10, Black Hills; 11, southern plateaus;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

## Missouri River trench

The present valley of the Missouri River constitutes the Missouri River trench physical division (fig. 3). It is a broad southeasterly trending trough, 300 to 650 feet deep, cut into the eastward-sloping Missouri Plateau. Between the Nebraska border and Big Bend (fig. 2) it is 1½ to 5 miles wide from rim to rim; the floor is 3/4 to 1½ miles wide. The walls of the trench, cut in easily eroded Pierre shale, typically have a rugged "badlands" topography of 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, the walls in these places have not been greatly affected by post-terrace valley cutting and are characterized by a more subdued topography than the unprotected portions of the trench.


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## Pierre Hills

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



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## 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. Comparatively, few youthful streams extend far beyond the limits of the Missouri River trench into the Coteau and much of the area drains into intermittent ponds in numerous closed depressions.



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## Climate

South Dakota has a dry subhumid climate characterized by hot summers and cold winters._/ Within the reservoir area precipitation and temperature decrease markedly in a northwesterly direction. Average annual precipitation is about 23 inches at the dam; at Fort Thompson it is only about 18 inches. More than three-fourths of the precipitation falls from early April through September, and 40 to 50 percent of the total precipitation is concentrated in May, June, and July. Average January temperatures grade from $20^{\circ}F$ at Fort Randall Dam to $16^{\circ}F$ at Fort Thompson; average July temperatures are about $75^{\circ}F$ for the entire area. A temperature range of $130^{\circ}F$ from $-25^{\circ}F$ minimum to $105^{\circ}F$ maximum can be expected in a normal year. Winds, predominantly northwest in winter and southeast in summer, have an average velocity of 10 to 12 miles per hour.

_/ All climatic data are from "Climate and man" (U.S. Department of Agriculture, 1941) and "Climatic atlas of the United States" (~~Vister~~ Visher, 1954).



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## Vegetation

The natural plants of the uplands are drought-resistant grasses. Shrubs and trees grow in the valleys and gullies where more moisture is available, and there are many good stands of trees, predominantly cottonwoods, along the part of the river bottomland that has not been flooded by the reservoir. Reflecting the decrease in precipitation trees growing away from permanent streams become increasingly scarce toward the northwest end of the reservoir area. Cultivated crops grade from predominantly corn in the southeastern end to predominantly wheat at the northwest end. Some oats are raised throughout the area and soy beans and sorghum are becoming important.

Vegetation commonly is more lush on slopes where landslides have occurred. This results from a twofold relationship. First, the excess water that encourages landsliding also encourages vegetation. Second, the closed depressions and fissures that commonly result from landsliding trap water which would normally run off of a smoother stable slope.


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## Geology

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



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| Era | System | Series | Group | Unit | | Thickness |
|---|---|---|---|---|---|---|
| CENOZOIC | QUATERNARY | (undifferentiated) | | Loess | | Up to 30± feet* |
| | | | | Alluvium-colluvium | | Generally less than 10 ft. |
| | | | | Alluvium | Slow-draining | Up to 100± feet |
| | | | | | Fast-draining | |
| | | | | Till | | Locally up 100 feet |
| | TERTIARY | Pliocene | | Ogallala formation | | 40± feet |
| MESOZOIC | CRETACEOUS | Upper Cretaceous | MONTANA | Pierre shale | Elk Butte Member | >180 feet |
| | | | | | Mobridge Member | >100 feet |
| | | | | | Virgin Creek Member | 50-100 feet |
| | | | | | Verendrye Member | 90-170 feet |
| | | | | | DeGrey Member | 20-50 feet |
| | | | | | Crow Creek Member | 10± feet |
| | | | | | Gregory Member | 30-90? feet |
| | | | | | Sharon Springs Member | 35-55 feet |
| | | | COLORADO | Niobrara formation | | 16 to 90 feet exposed* |

*~~Exposed thickness in~~ pre-reservoir ~~time~~

Table 1. A generalized stratigraphic section.

## Upper Cretaceous sedimentary rocks

The Upper Cretaceous sedimentary rocks are essentially the only consolidated deposits in the area. They include both the most stable (Niobrara Formation) and the least stable (Pierre shale) deposits in the area.



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## Niobrara Formation

The Niobrara Formation, a soft impure, fossiliferous limestone, was widely exposed in the lower portion 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 outcrops as nearly vertical bluffs bordering the river and its flood plain. Prior to the completion of the dam (1952) these bluffs ranged in height from a few feet to over 85 feet above the river level (Petsch, 1946, fig. 12). The exposed thickness above the river varied considerably due in part to the uneven surface of the formation and in part to the gentle open folding of the Cretaceous beds. In the vicinity of Fort Randall Dam, the top of the Niobrara, at an elevation of 1,310 feet, was about 70 feet above river level (1,240 feet elevation). Upstream at the site of the former Wheeler Bridge, only 16 feet of Niobrara were exposed. Near Chamberlain, the upper Niobrara contact was 1,420 feet in elevation, about 90 feet above the Missouri. The formation passes below river level near the lower end of Big Bend at an altitude of about 1,350 feet.


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The Niobrara Formation, the oldest exposed bedrock in the area, is the uppermost member of the Upper Cretaceous Colorado group. The Niobrara beds within the report area can be traced directly to the 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 ~~sheels~~ shells of Foraminifera and Ostracoda. Small light-colored shell fragments and clastic particles in a darker groundmass gives 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 in the Rock-Color Chart distributed by the National Research Council.

The rock is massive and coherent and has a toughness and resiliency that make it hard to fracture although it is soft enough to be scratched by a fingernail. The essentially 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. A system of near-vertical joints from 1 inch to several feet apart is well developed in most exposures.

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## 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 subdivisions, following Crandell's classification (1958, p. 8-19) although individual members are not everywhere mappable either because contacts are obscured or because they are too similar lithologically.

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 six members total 228 feet thick; in the vicinity of Chamberlain, they are at least 400 feet thick (interpreted from Petsch, 1952).



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## Sharon Springs Member

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

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



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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 ~~great~~ 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, 1952, p. 63), often 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 member is relatively hard (barely marked by a fingernail), and it is the only member of the Pierre shale that does not easily weather to clay. There is no apparent bedding in fresh exposures, but horizontal fissility ~~parallel to the bedding planes~~ develops on outcrops as the material weathers.



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## 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 more than 25 feet high are flattened by subsequent landsliding.

The unit, which is separated from the underlying Sharon Springs material 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 feet).

The thickness increases from about 35 feet in the lower part of the reservoir (Gries, 1942, p. 31) to at least 50 feet, perhaps as much as 90 feet, in the vicinity of 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 calcareous shale. In the upper reaches of the reservoir, the Gregory is a predominantly noncalcareous gray bentonitic claystone that closely resembles the upper 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 vary 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 shalè alters to bentonitic clay in a few years, passing through a partially weathered stage in which it consists 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 homogeneous, but partially altered shale develops a fissility apparently parallel with the bedding.



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## 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 to 15 inches thick and the marl is generally 7 to 10 feet thick. The marl bed has been recognized along the Missouri River trench for a distance of over 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 can be seen 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 lithologic description is largely a summary of his data. Unweathered marl is light gray; upon exposure it oxidizes to grayish orange. Secondary iron oxide deposited in the siltstone commonly gives it a yellowish-brown appearance.


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The marl is soft and ~~has~~ 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 so that it forms small ledges, generally about 1 foot high; in other places it is cemented to varying degrees by calcium carbonate.



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## De Grey member

The De Grey member 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 outcrops of the De Grey member one of the striking features of the walls of the Missouri River trench around the upper part of the reservoir. The basal few feet of the De Grey generally contains no concretions, and the contact between the Crow Creek and De Grey members is a relatively sharp transition from marl to noncalcareous claystone. The De Grey of the Fort Randall area closely resembles the shale and bentonite facies described by Crandell (1958, p. 13–14). Crandell's thick basal siliceous shale facies apparently is not developed in the report area.

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


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

The De Grey is very susceptible to landslides, especially in the southern part of the area. It is difficult to find exposures that do not show at least minor displacements, most of which can be attributed to landslides.



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## Verendrye Member

The Verendrye Member is present throughout the report area. Most fresh outcrops are found on landslide blocks because undisturbed claystone exposures weather rapidly to rounded slopes blanketed by residual clay soil.

No complete sections were measured by the author, but Petsch (1946, p. 37) shows the thickness at the Wheeler Bridge site as 80 feet and at Crow Creek, 10 miles north of Chamberlain, as about 170 feet.

The contact between the Verendrye and the underlying De Grey Member becomes less distinct southward along the reservoir. In the northern part of the reservoir, the contact can readily be placed at the top of the iron-manganese concretion zone in the De Grey. In the southern part of the reservoir, landslides and vegetation commonly obscure the contact. Moreover, the absence here of a well-developed concretionary zone in the De Grey results in a similarity of lithology that make separation of the two members unfeasible in the few good exposures available.


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(p. 39a follows)

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

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

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## Virgin Creek Member

The Virgin Creek Member, as defined by Searight (1937, p. 35) and described by Crandall (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 that is extensively exposed in the trench walls.

The Virgin Creek Member is about 50 feet thick in the vicinity of 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) with the topographic map of the same area suggests that the member is about 100 feet thick in the vicinity of Chamberlain.

Outcrops are rare in the smooth, grass-covered slopes typical of the Virgin Creek. Fresh outcrops are found only where stream erosion or landslides have recently exposed it. Outcrops of the shale facies show a characteristic silvery sheen from the shale chips. Where the outcrops have weathered to bentonitic clay the member is difficult to distinguish from the Verendrye Member. No sharp contact has been observed between the Virgin Creek and the Verendrye Members; apparently, the contact is gradational over a vertical distance of 5 to 10 feet. Moreover, the contact is commonly hidden by slump blocks derived from weathered Virgin Creek beds.



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Both the shale and claystone are dark gray to grayish black when fresh, changing to brownish gray where weathered. Concretionary layers and thin bentonite beds up to 1 inch thick are common.

The basal shale is, with the exception of the Sharon Springs Member, the most fissile material in the Pierre Shale. It weathers to form small chips that are usually less than 1 inch in maximum diameter and one-eighth inch thick. After prolonged exposure the chips disintegrate to a bentonitic clay similar to the De Grey and Verendrye Members.

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



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### Mobridge member

The Mobridge member consists of partly indurated marl and calcareous shale beds very similar to those described in the Pierre area by Crandall (1958, p. 16-17). In the southern half of the Fort Randall area, the Mobridge crops out in the trench walls bordering the reservoir; in the northern half, it is found 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 is known to thicken northwest along the trench, no data are available on thicknesses in the northern part of the report area.

The lithology shows considerable lateral variation. Along the Missouri River trench in South Dakota, it grades from bentonitic marl (containing roughly 35 percent calcium carbonate) in the vicinity of 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.



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The material 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 to 1 inch thick develop parallel to the bedding. The contact between the calcareous Mobridge beds and the noncalcareous Virgin Creek beds is sharp rather than gradational 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 landslides 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 southern half of the Fort Randall Reservoir.

The original thickness of the Elk Butte Member in this area is unknown because the upper limits have been eroded and then overlain unconformably by late 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" of Gries is now considered to be part of the Elk Butte Member (Stevenson and Carlson, 1950) so that the total thickness is now estimated at 180 feet in this location.

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Exposures of the Elk Butte Member are similar to those of the underlying Mobridge and the contact 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 some calcareous beds were observed above the supposed base of the Elk Butte Member in the lower part of the reservoir. Therefore color is also used to separate the two (Crandell, 1958, p. 18). Usually the weathered Elk Butte material is darker than the Mobridge. The color ranges from olive gray when partially weathered to moderate brown where fully weathered; unweathered material is rarely seen. Another distinctive feature also noted by Crandell is the presence throughout the Elk Butte of yellow-brown calcareous concretions. In the Fort Randall area concretions are commonly found scattered throughout the formation but 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 changes the material fairly easily to bentonitic clay. Weathering also brings out a prominent color banding in consolidated outcrops.



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## Structure

The exposed rocks of Cretaceous age are essentially flat-lying undisturbed sediments. There are undoubtedly some minor open folds but their magnitude is 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 U.S. Army Corps of Engineers have revealed intense local faulting of both formations.

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



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## Tertiary and Quaternary deposits

Although most of the Tertiary and Quaternary sediments are unconsolidated terrestrial deposits, They are much more stable than the Pierre shale. Only a very small percentage of the total number of landslides along the reservoir occur wholly in Tertiary and Quaternary deposits.



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## Ogallala Formation

The Ogallala Formation of Pliocene age is a heterogeneous mixture of silt, sand, and fine to medium gravel which is locally cemented to an orthoquartzite. The Ogallala is restricted to the south half of the 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 is variable because of erosion. The average thickness in the area is about 40 feet.

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



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## Till

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

Since glacial deposits tend to fill in and smooth out irregularities in the preglacial topography, the thickness of these deposits varies greatly in 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 consists primarily of silt and clay with 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.



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### Alluvium--fast draining; slow draining

In this report the term alluvium is applied to all water-laid Quaternary deposits of age and includes glacial outwash as well as stream and flood-plain deposits. The general category has been broken down into fast-draining alluvium (Qaf) and slow-draining alluvium (Qas). 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 a relatively low permeability and the movement of water is restricted. ~~These~~ There is considerable delay in adjustment of the water table within slow-draining alluvium to changes of water level in an adjacent body of water.



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Extended field studies of the behavior of alluvial deposits enabled the author to set a relatively consistent boundary between fast-draining and slow-draining alluvium. Material in which the components identifiable by field methods were predominantly coarse-sand size (0.5-1.0 mm) or larger behaved under changing conditions of saturation and drainage as fast-draining alluvium. Material in which the components appear to be predominantly medium-sand size (0.5-0.25 mm) or smaller reacted as slow-draining alluvium.

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 the two alluviums.

Fast-draining alluvium becomes less stable as it is saturated and redevelops its stability as it drains. As water saturates fast-draining alluvium it quickly fills essentially all the pores and destroys 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 redevelops 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.



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Several factors cause slow-draining alluvium 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 the 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 so that the bonding effect of surface tension is not entirely destroyed. The adsorbed water layer that occurs 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 has little effect in coarse-grained alluvium. In the finer materials of the slow-draining alluvium, however, the proportion of voids less than 0.2 microns wide becomes significant and the adsorbed water layers becomes an important force in helping the material resist shear.



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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. As a result, 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. As a result the unbalanced pressure creates a lateral stress toward the free face of the alluvium.

The alluvium either floors nearly flat valley bottoms or is exposed in terraces. The valley bottom alluvium is both widespread and, at least along the Missouri River trench, deep_/. No separation of slow- and fast-draining alluvium in the valley bottoms was attempted because the valley bottom alluvium is comparatively unimportant as a landslide medium. Most is either permanently flooded by the reservoir or is too far up the tributaries to be affected. Some minor landsliding can be anticipated in the zone where periodic drawdown of the water level will expose banks of slow-draining alluvium.

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_/ Flint (1955, p. 147) gives a depth of 189 feet to the bedrock floor of the Missouri River trench at Fort Randall.

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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 varies. In some exposures the alluvium is a thin veneer over bedrock; in others it forms terraces up to 80 feet high. Preponderance of the slow- or fast-draining alluvium varies from one outcrop to the next, but most 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 both fast-draining and slow-draining terrace alluvium. Continued sliding can be expected until the slopes reach equilibrium.



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## Alluvium-colluvium

Alluvium-colluvium is a general term for mixtures of alluvium, colluvium_/ exclusive of landslide material, and loess. The grain size is usually that of fine sand or finer, although occasionally there are thin lenses of pebbly material. The material behaves essentially as slow-draining alluvium.

Deposits of alluvium-colluvium are rarely more than 5 to 10 feet thick and commonly occur as relatively small areas in the bottom of small valleys.

_/ "Earth material that has moved or been deposited mainly through the action of gravity." (Stokes and Varnes, 1955.)



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## Loess

Loess is a porous wind-deposited sediment composed predominantly of silt with minor amounts of fine sand and clay. It 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 high 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 soon collapse and the fallen material is removed by wave erosion.



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## Origin and development of the Missouri River trench

A brief discussion of the origin and development of the Missouri River trench is pertinent because essentially all of the landslides in the area have resulted from 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 partially 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 re-excavated its valley in Recent time and now has a gradient of about 1 foot per mile. Until the filling of the reservoir most of the river's energy was spent cutting laterally into the valley walls.



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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 with slope wash contributing to a minor degree. In this joint action landsliding probably was at least as effective as stream cutting. Slope wash probably was important only on slopes where there are few, or no, active landslides.

Erosion by landsliding is instigated and, in many places, perpetuated by stream action. Downward and lateral cutting by streams oversteepen and undercut the valley walls. Blocks of material adjacent to the streams 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 slide. By repetition of this process the entire valley wall slowly slides toward the stream, which in turn makes further sliding inevitable by removing landslide debris after it reaches the valley bottom.



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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 there are many factors other than removal of toe support that affect slope stability. There are numerous places along the walls of the Missouri River trench, nevertheless, where a slide sequence grades upslope from old, barely discernible, stabilized landslide blocks along the river bank, 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 from several landslide cycles may cause considerable lateral and vertical movement, although individual blocks are not necessarily transported a great distance during a single cycle. As an example, 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 level at which the formation crops out in the nearby uplands.



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## Landslides

The phenomena included in the term landslides are generally compound features with many variables, and, inevitably, numerous systems of landslide classification have been devised. The Highway Research Board classification, (Varnes, 1958) based on two main variables--(1) type of material and (2) type of movement, has been selected as the most satisfactory for the purposes of ~~the~~ this investigation. Figure 5 is a graphic summary of this

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Figure 5.--Near here.

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system, with additions of notations concerning the materials involved in the most common types of 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."



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**Figure 5.--Classification of landslides. (Adopted from Varnes, 1958, pl. 1.)**



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## 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, where more than one kind of material is involved, generally only one is responsible for the landsliding. In some slides, the slump-earthflow type for example, two kinds of movement are well developed and a hyphenated compound term is used to describe them.

The various types of landslides along the reservoir are discussed in the order in which they occur on the landslide classification chart (fig. 5) rather than according to their abundance or size in the Fort Randall area.



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### Rockfalls (fig. 5-a) (Niobrara Formation)

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


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### Soilfalls (fig. 5-b) (Till, slow-draining alluvium, alluvium-colluvium and loess)

Soilfalls, caused by stream and wave erosion, are very common wherever fine-grained unconsolidated deposits outcrop along the reservoir (fig. 7) or along the tributary valleys. Generally the

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Figure 7.--Near here.

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blocks are a few feet wide and no more than 20 to 30 feet long. The overall importance of soilfalls as a landslide process is commonly overlooked because the individual blocks are relatively small. Probably more than one-half of the fine-grained unconsolidated material undergoing erosion along the reservoir shore is entering the water as soilfall blocks.

Although soilfalls are an important erosion process their mechanics are relatively simple. Soilfalls result from removal of support at the base of near-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 but they are never primary causes of soilfalls.



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Figure 7.--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 the water level by Pierre shale. Several small overhanging blocks of loess can be seen in the center of the photograph. Photographed September 27, 1954.

Bedrock slumps (including block glides) (fig. 5-c, -d, -e)

(Pierre shale)

Slumps (rotational movement) and block glides (planar movement) (fig. 8) represent the end points of a series of landslide types

Figure 8.--Near here.

caused by shear failure of coherent unit blocks. The relationship between the two slide processes is shown in the schematic cross sections of figure 8. Assuming that 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. 8a). 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. 8b). Because of the similarity in causes and behavior, slides in the Pierre shale that are essentially block glides have been included under slumps.



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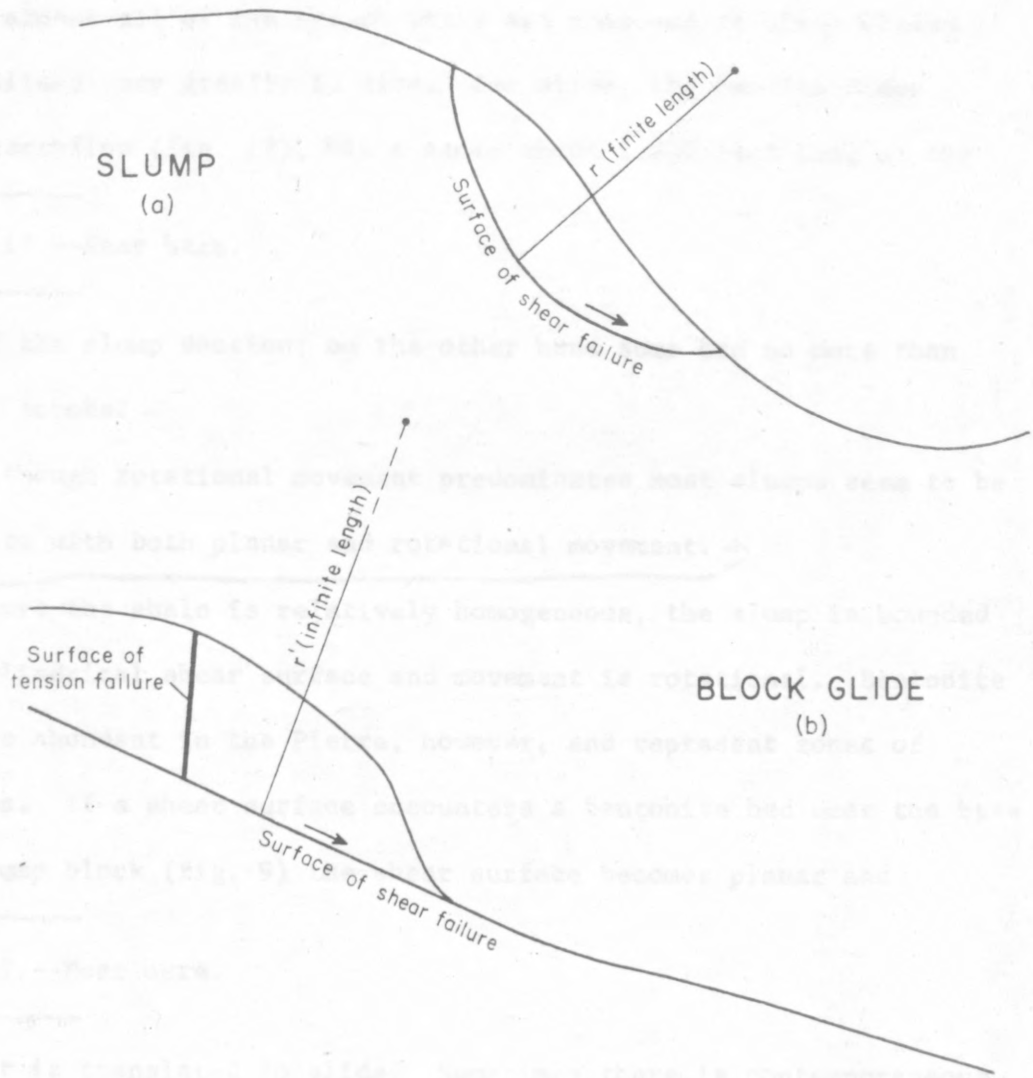


Figure 8. Schematic cross sections showing relationship between slumps (a) and block glides (b), end points in the series of landslide types involving shear failure of coherent blocks.

Slumping in the Pierre shale is the commonest type of landslide movement in the project area. In some portions of the Missouri River trench almost all of the trench walls are composed of slump blocks. These slides vary greatly in size. One slide, the Landing Creek slump-earthflow (fig. 17), has a scarp about 1,300 feet long at the

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Figure 17.--Near here.

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head of the slump section; on the other hand some are no more than 30 feet across.

Although rotational movement predominates most slumps seem to be composite with both planar and rotational movement.

Where the shale is relatively homogeneous, the slump is bounded by a cylindrical shear surface and movement is rotational. Bentonite beds are abundant in the Pierre, however, and represent zones of weakness. If a shear surface encounters a bentonite bed near the base of a slump block (fig. 9) the shear surface becomes planar and

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Figure 9.--Near here.

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movement is translated to glide. Sometimes there is contemporaneous glide movement along several parallel shear planes at the base of a slump. When the shear surface encounters bentonite beds at the back of the slump block above the base, it cuts through the beds and remains curved.



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Figure 17.--Landing Creek slump-earthflow. The upper part of the landslide consists of slump blocks moving out on an old terrace remanent. Where gullies eroded the terrace the slump blocks have had no support and the material has 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.



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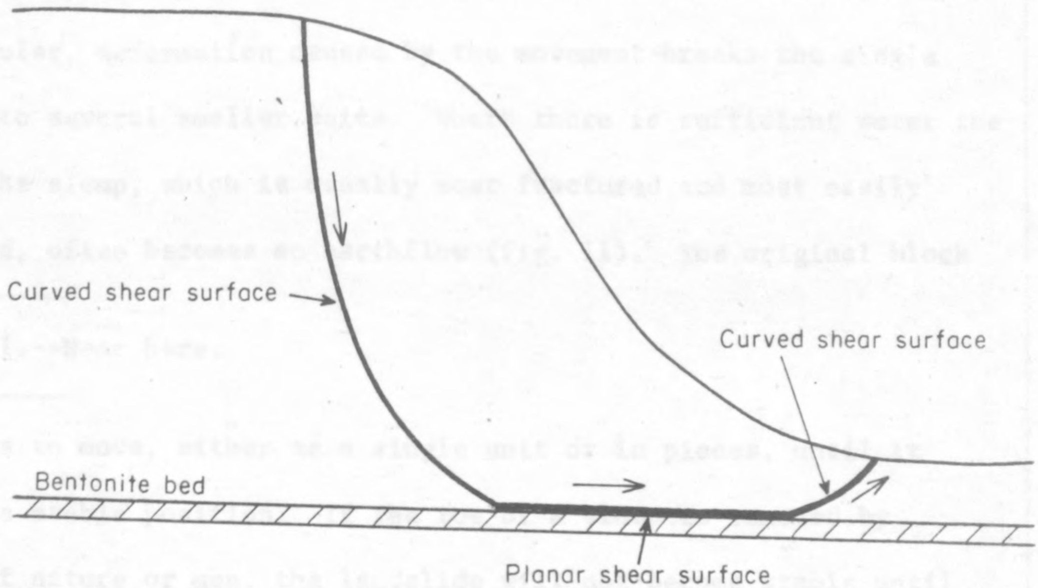


Figure 9. — Schematic cross-section through a potential slump with a bentonite bed controlling development of the shear surface.

Slumps usually start as movement of a single block (fig. 10, lower

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Figure 10.--Near here.

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half of slump). If the block is large, or if the surface of movement 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 usually most fractured and most easily saturated, often becomes an earthflow (fig. 11). The original block

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Figure 11.--Near here.

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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 carried away (fig. 12).

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Figure 12.--Near here.

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Figure 10.--Highway 16 landslide. A slump south of U.S. Highway 16 about one-half mile east 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.



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Figure 11.--Cable School slump-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) consists of 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.

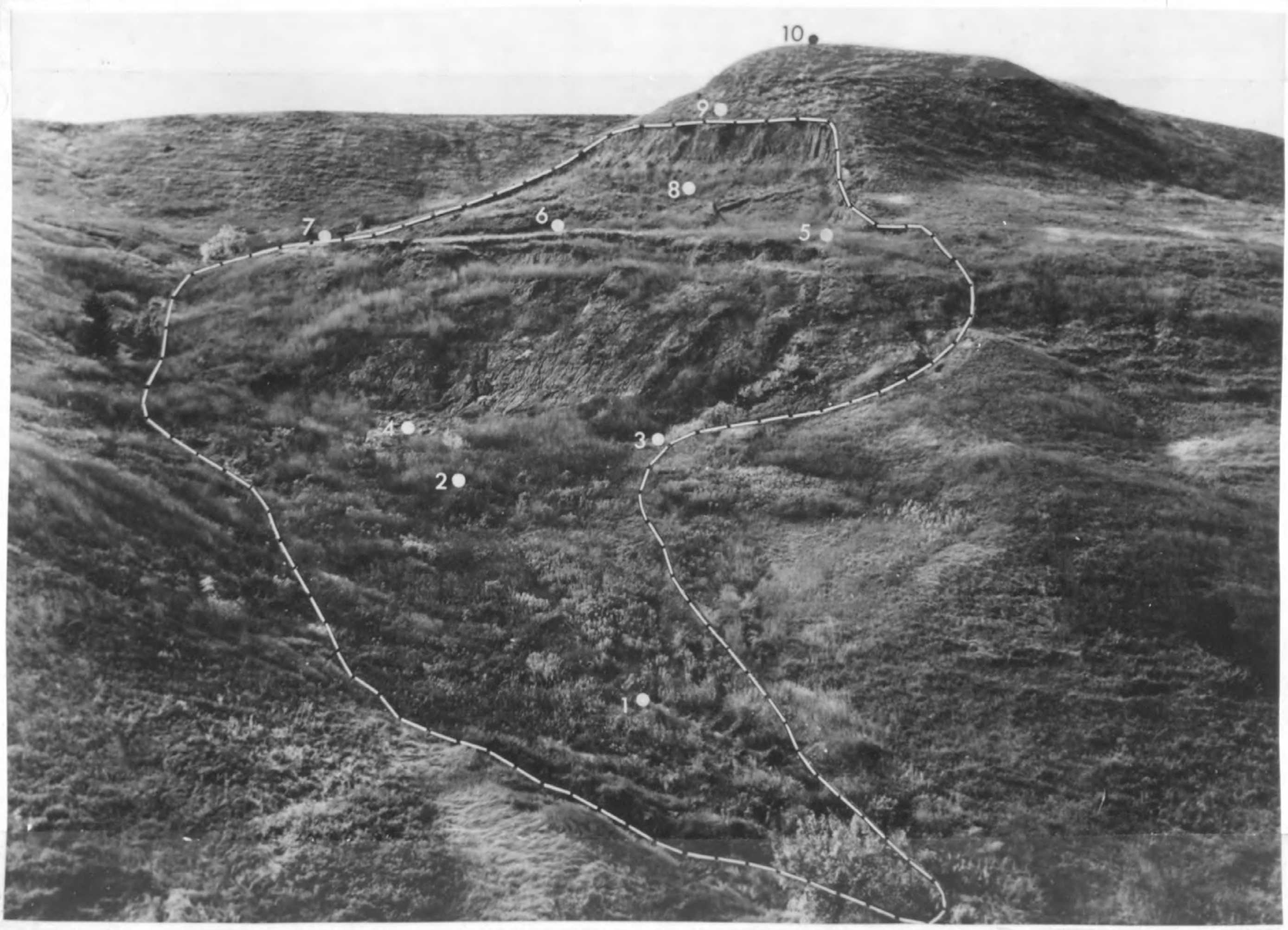

Figure 12.--Active slump along the Fort Randall Reservoir shore in Gregory County, S. Dak. Wave erosion at the toe of the slump is breaking up the slump blocks and removing the material. Photographed September 27, 1954.



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Slump movements become increasingly 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 a second block will break loose. If the initial slump block continues to move, the second block follows it 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 redevelop stable positions (figs. 11, 12).


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Soil slumps (fig. 5-h) (Alluvium, alluvium-colluvium, loess, till (?)

Soil slumps were rare before formation of the reservoir but they have been fairly common along the shoreline since that time and probably will continue to occur for many years. The lower part of most soil slumps is submerged, and many have been completely inundated by the rising reservoir level. Till adjacent to the uplands was, and still is, involved in many slumps, but failure of the Pierre shale underlying the till generally is responsible.



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Soil slumps now occurring along the reservoir range from those caused primarily by saturation to those resulting from wave erosion. In figure 13 a slump has developed along the face of a terrace composed

Figure 13.--Near here.

of gravel (fast-draining alluvium) blanketed by loess and colluvium. A small wave-cut bench about 8 feet above the water level represents the reservoir level before a drawdown that started 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 indicating that little or no movement has occurred since the drawdown. Also, the bench is as well developed on the landslide as it is on either side indicating that the slump occurred before wave erosion started. These facts indicate that neither minor wave erosion nor drawdown has noticeably affected the slump block after it became stabilized. The slump movement apparently is related to saturation of the materials composing the slump block. Failure began after saturation destroyed the intergranular bonding effect of capillary forces in the gravel (see p. ~~91~~ 51). Once the block began to settle, part of its weight was supported by water so that the intergranular friction in the gravel was reduced by the buoyant effect of the water. As a result the friction between the grains was also reduced and further settling made possible.



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Figure 13.--Slumping along the face of a gravel terrace blanketed by colluvium and loess. The limited amount of wave erosion implies that saturation was the primary cause of slumping. Fort Randall Reservoir shore, Charles Mix County, S. Dak. Photographed October 24, 1953.

The opposite extreme, a soil slump caused primarily by wave erosion, is shown in figure 14. This photograph, taken at the same time as

Figure 14.--Near here.

figure 13, shows slumping in slow-draining alluvium. Here, also, there has been a temporary drawdown of about 8 feet in the reservoir level. There is no indication of a wave-cut bench, however, despite the fact that wave erosion is more active here than in the vicinity of figure 13. The slump was still moving at the time of the photograph, and the combination of landslide movement and wave erosion apparently has obliterated all evidence of a wave-cut bench. Saturation followed by friction from draining water and unbalanced pore-water pressure ~~(see p. 101-103)~~ probably decreased the stability of the slow-draining alluvium; however, the area should have become stable after the water level dropped unless other forces are acting on it. The most logical explanation is that wave erosion has been the main cause of slumping and was still preventing stability at the time the photograph was taken.



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Figure 14.--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 has probably been the primary cause of slumping. Photographed October 24, 1953.



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Slow earthflows (fig. 5-p) (Weathered Pierre shale, till, colluvium)

"Slow earthflows" (Varnes, 1958, p. 38) are the commonest type of slope failure in plastic unconsolidated materials along the reservoir. Most are 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.



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Two conditions limit the development of earthflows to slopes composed of a thin mantle of plastic, unconsolidated material underlain by relatively coherent impermeable material. First, earthflows only develop in material that is sufficiently unconsolidated and plastic to flow as a mass and that also can hold enough water to convert it into a viscous fluid. Generally these qualifications are confined to clay-rich colluvium or weathered shale and till. Second, the earthflow material must be saturated, or nearly saturated, before it will flow. Where relatively impermeable material is covered by a mantle of more permeable material, water percolating downward from the surface will stop at the contact and build up a saturated zone in the mantle (fig. 16). Sections of the mantle will fail by flowing if

Figure 16.--Near here.

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.



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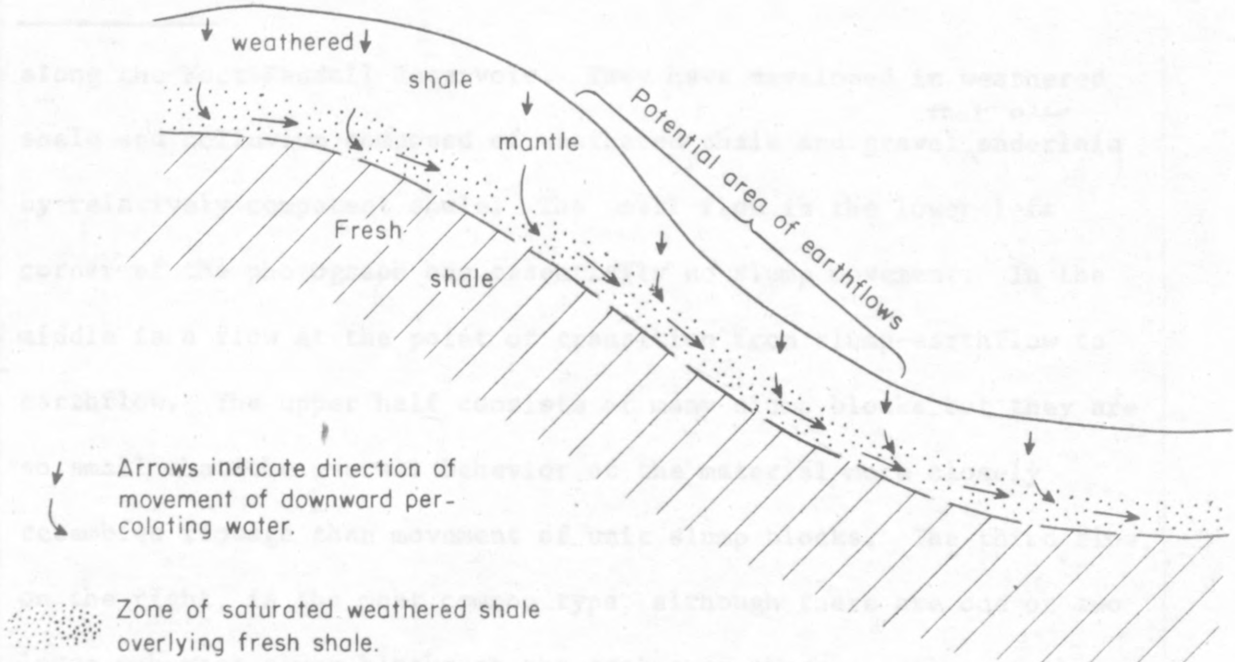


**Figure 16. Idealized diagram of conditions conducive to earthflows.**

The three earthflows in figure 15 are typical of those that occur

Figure 15.--Near here.

along the Fort Randall Reservoir. They have developed in weathered shale and colluvium composed of weathered shale and gravel that are underlain by relatively competent shale. The small flow in the lower left corner of the photograph has essentially no slump movement. In the middle is a flow at the point of transition from slump-earthflow to earthflow. The upper half consists of many slump blocks but they are 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, 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.



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Figure 15.--Cable School earthflows, NW¼ sec. 30, T. 103 N., R. ~~72~~ 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 jeep on top of ridge for scale. Photographed July 7, 1952.



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Mudflows (weathered Pierre Shale)

Mudflows, characterized by the rapid flow of a saturated mass of material, are confined to the weathered Pierre Shale along the Missouri River trench. Although 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 develop a very coarse, porous crumblike structure when they are dry. During sudden heavy rainfalls this open structure can rapidly absorb large quantities of water before the wet bentonite expands and reduces the permeability. The saturated mass of weathered shale becomes a mudflow when the added water has increased the weight and lessened the cohesion of the clay to the point where it is no longer stable on a given slope.



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Causes of landsliding in the Fort Randall Reservoir area

The causes of landslides may be classified in two generalized groups. The first includes those basic processes that slowly reduce the stability of a mass of material; the second is comprised of "trigger" processes that, in themselves, are only of minor importance but, added to the stresses already piled up in a potential slide area, are sufficient to set it in motion. The trigger actions are the most obvious and often are mistakenly considered to be the dominant causes of slides. A trigger process, however, can start a landslide only after an area has become a potential slide through the operation of other factors. Elimination of possible trigger actions, nevertheless, can prevent, or at least postpone, many landslides.



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## Erosion

Erosion, the most important 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 supply steep slopes. Erosion becomes a triggering process when it removes material, that otherwise would serve as a buttress, from the toe of a potential landslide (figs. 12, 14).



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## Ground water

Ground water is the most versatile cause of landslides. It can affect slope stability by means of: (1) weight, (2) lubrication, (3) hydrostatic pressure, and (4) piping. Some of these effects increase landslide stresses; others decrease shear strength in potential slides. In some places, ground-water conditions supply the trigger action; in others they are the general cause of instability.



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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-filled voids are replaced by water-filled voids; the greater density in turn produces some increase in the stresses within that material. The present studies have not furnished sufficient data to judge whether or not an increase in stress due to greater density will necessarily be accompanied by a decrease in stability.



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## Lubrication

Lubrication by water is one of the oldest explanations for the effects of ground water on the stability of slopes. Lubrication is often accepted 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 effects of water as a lubricant 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 character today can be explained by the history of its deposition and compaction.



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The shale was deposited on the bottom of an inland sea as a mud composed of loosely bonded aggregates of poorly oriented particles and 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 continued as the overburden increased until much of the adsorbed water was driven out from between the shale particles. The particles became closer packed and molecular forces between particles formed a relatively tight bond. In subsequent periods uplift and erosion have removed much of the overburden but these molecular forces apparently remain great enough to prevent re-expansion 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 have prevented the molecular forces between shale particles from developing a solid to solid bond. 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 and reduces the bonding forces between particles. The effects of weathering are further complicated by the presence of montmorillonite, a swelling clay (p. ~~118~~ 130).

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 is a less important factor in stability.


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The shear strength along the surface of contact of two solids is related to friction between the solids. If a liquid can be placed so that there will be no 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 princip~~al~~le of lubricants.

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



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Lubrication in Pierre shale slumps, although probably never complete, is an important cause of movement. Ground-water investigations (see section on Ground-water investigations, p. ~~252~~ 225) indicate that the regional water table is high enough to permanently saturate 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 cannot be complete, however, because a combination of molecular forces and overburden pressures hold the particles (including adsorbed water layers) together at their contacts. The pressure of the saturating water normally is not great enough to penetrate between these contacts and lubricate where friction is greatest. Although shale is not fully lubricated even in zones of permanent or semipermanent saturation, lubrication is nevertheless at a maximum for the given conditions.

Lubrication in flows is more complete 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 the shear strength of the solid material in a dry state. In some mudflows it approaches that of water, indicating almost complete lubrication.



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The difference in the degree of lubrication in Pierre Shale slumps and in flows results from the differences in materials and environment. Slumps occur in coherent materials, usually with sizable overburden pressures along the surface of failure. Flows occur in near-surface unconsolidated materials where there is relatively little coherence and overburden pressures are negligible.

Generally lubrication would be expected to have a secondary effect in unconsolidated slumps. If there is enough water present the failure should be by flowage instead of by shear. When flow occurs in unstable material underlying coherent material, the movement in the coherent block may resemble a slump.

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


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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. Till is fairly compact material derived largely from Pierre shale; it therefore behaves like Pierre shale.

It is not always possible to classify the effects of lubrication on landslides either as a general cause of decreased stability or as a trigger action. In general, however, lubrication due to permanent or semipermanent saturation probably results in generally decreased stability. Lubrication caused by periodic or occasional inflow of ground water commonly is a trigger action.



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Hydrostatic pressure

One development in the study of landslide causes is recognition of the effects of hydrostatic pressures in ground water. Soil mechanics engineers have increasingly emphasized the importance of hydrostatic pressure. Terzaghi, one of the chief proponents, summarizes the soil mechanics viewpoint on effects of hydrostatic pressures in the paper "Mechanism of landslides" (Terzaghi, 1950).

Many of the theoretical principles of hydrostatic pressure, although 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 lb./in.$^2$ increase in pressure. This increase in pressure acting on a surface 10 feet square in turn creates a force of more than 3 tons.


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The effects of hydrostatic pressures fall into two general categories. The first occurs in areas where uniform hydrostatic pressures create static ground-water conditions, and the second includes variations in hydrostatic pressures over a given area that produce dynamic ground-water conditions. Ideally ~~The~~ the 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 be in water that is locally confined to a permeable zone overlain and underlain by relatively impermeable materials. In theory, ~~The~~ the pressure then would have ~~has~~ no direct relation to the overlying water table.



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## Uniform hydrostatic pressure

Uniform hydrostatic pressures may or may not affect slope stability. Pressure in unconfined ground water has no effect on the material containing the ground water. Pressure in confined ground water, on the other hand, may exert large forces on the confining material.

Independence between unconfined static ground water and the surrounding material is one of the basic concepts of soil mechanics. Under static conditions the total pressure at any point in a 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), but 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 a 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.


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Pierre shale is the only formation, 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. ~~252~~ 225) showed 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 can be demonstrated schematically. Figure 19a is a

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Figure 19.--Near here.

---

cross section through a shale bank containing a horizontal parting 40 feet below the top. The parting is assumed to be sealed at the surface in order 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.



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Figure 19.--Effect of hydrostatic water pressure in fractures within relatively impermeable shale.



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When the fractures are filled with water the pressure is proportional to the height of water. In figure 19a the water pressures throughout the parting are caused by the 40-foot head of water in the joint. Although water pressures act in all directions, this discussion is concerned only with the water pressures acting on the potential landslide block. 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 p. 99)~~. If the total forces from these combined pressures is great enough, the block will become an active landslide.

The process of reducing friction by increasing hydrostatic pressures in confined ground water can cause renewed movement in old landslides as well as initiating movement in new slides. In old landslides the surface of failure forms the permeable fracture (fig. 19d). Otherwise the process is the same.



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An analysis of the vertical pressures on a small block of shale ("x", fig. 19a) illustrates the potential effect of hydrostatic pressures. When the fractures are dry the vertical pressures acting on "x" are a downward pressure caused by weight of the overlying shale and an equal upward pressure transmitted to the potential slide block from the shale below the parting (fig. 19b). If an average density of 155 lb./ft.$^3$ is assumed for the shale the magnitude of each force acting on "x" is about 6,200 lb./ft.$^2$.

When the fractures are filled with water (fig. 19c) the downward pressure on "x" remains the same but about 2,500 lb./ft.$^2$ of the upward pressure is now exerted by the water. Since the total upward pressure still equals 6,200 lb./ft.$^2$, the pressure transmitted through solid contacts must now be only 3,700 lb./ft.$^2$. The reduction of the upward acting solid-to-solid pressure to about 60 percent of its original value correspondingly lowers the friction along the base of the block and increases the chances of movement along the parting.



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Variations in hydrostatic pressure

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



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The importance of seepage pressure depends on the effective size of the material on which it acts. In coarse gravel and other fast-draining materials, resistance to flow is slight and in most cases large seepage pressures do not develop. Pressures caused by seepage in clays probably can be considered minor, also. Clay is the finest of the slow-draining materials and water pressure, as a result, is essentially in a static condition. Slow-draining materials in the effective size range of silt and fine sand are most affected by seepage pressures. Material in this range is 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 often occur in silts and fine sands along 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 the landslides.



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## Piping

Piping, the process of internal erosion by moving ground water, is one of the many minor processes that gradually reduce the stability of slopes. Quantitative determination of the stability changes caused by piping is impossible in most cases.

Internal erosion can occur by mechanical transportation or by solution. Mechanical transportation usually involves movement of particles of silt and fine sand size that are small enough to be easily transported but are 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 ~~unkwnon~~; material removed by solution neither leaves large cavities nor concentrates in appreciable surface deposits. The best evidence that solution does occur is the ground water itself. Searight and Meleen (1940) state that there is hard water in more than three-quarters of the shallow wells in the counties along the reservoir. All of these wells are less than 200 feet deep, and the majority are less than 50 feet deep. The ground water is relatively pure when it enters the ground; it must, therefore, dissolve minerals from the material through which it passes.



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Miscellaneous causes

Earth tremors and loading on the heads of potential landslides are known to produce movement. Earth tides and atmospheric pressure changes are possible but, as yet, generally unverified causes of landsliding. Although loading at the head of a potential landslide is rare under natural conditions, it is fairly common 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 20, is a typical example of an earth

Figure 20.--Near here.

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 indicate that the shale was already wet and unstable so that the added weight of the fill served as a "trigger" to set the slope in motion.

Figure 20.--Slump caused by placing a highway fill at the head of a potential landslide block. Relocation of South Dakota Highway 47, sec. 26, T. 103 N., R. 73 W., Lyman County, S. Dak. Photographed July 16, 1955.

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In the Great Plains area the effects of earth tremors are probably very slight. Small or infrequent trigger actions, on the other hand, can be important as a complement 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 at the time of saturation might be sufficient to produce movement.


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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. The following table attempts to summarize these causes and to estimate their relative importance in the vicinity of the Fort Randall Reservoir. Investigations along the reservoir were qualitative, and at best crudely quantitative. The interpretations presented in the table, therefore, are primarily intended as a guide for future detailed investigations.

## Analysis of the Pierre Shale

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

A thorough investigation of the characteristics and behavior of the Pierre Shale is beyond the scope of the present study. The data included here are only qualitative at best and are not complete. The information on 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 taken at only one locality per member and, in most cases, from only one horizon; therefore they cannot show the lateral and vertical variations although they give a good picture of the general nature of the Pierre.

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


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Mechanical analysis

The mechanical analyses were made by the hydrometer method essentially 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 relationship used by the U.S. Bureau of Reclamation (1952).



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



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Mechanical analyses of the Pierre shale show both the general trend of grain-size distribution in the shale and the similarities, or dissimilarities, of grain-size distribution between individual members. A cumulative curve has been plotted in figure 22 for

Figure 22.--Near here.

each member (Sharon Springs, Gregory, De Grey, Verendrye, Elk Butte) analyzed. The data also have been averaged to construct an average cumulative curve that should approximate a composite curve for the entire formation.

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 (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 has been placed between 2 microns and 5 microns in various size classifications (Truesdell and Varnes, 1950). The two microns division has been selected for this report following the line of reasoning used by Grim (1953, p. 1-2):



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p. 115a follows

Figure 22

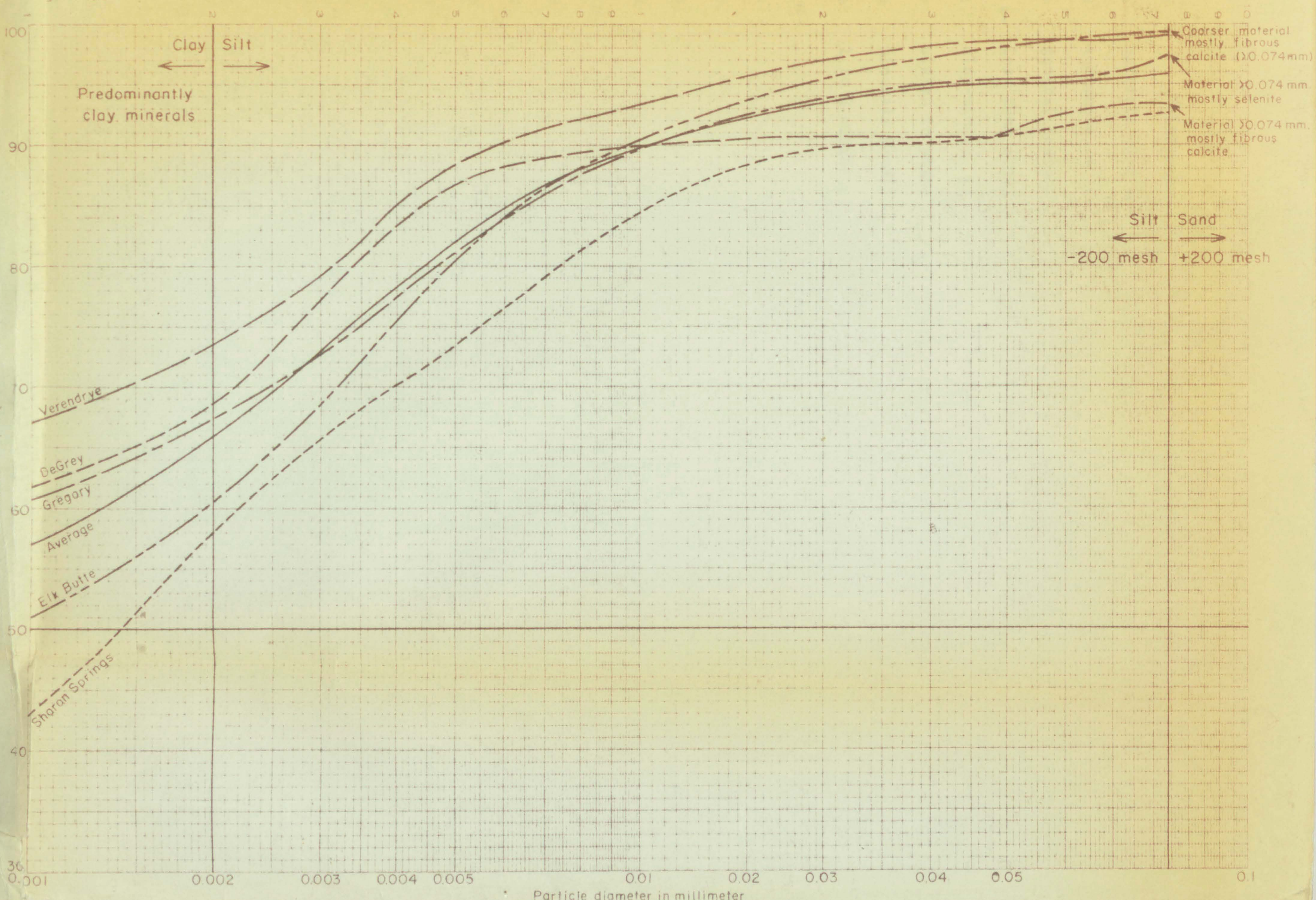


Figure 22. Particle size distribution in 5 members of the Pierre Shale.

"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."


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The outstanding features shown by the cumulative curves are the predominantly fine grained character of the formation and the similarity of grain-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 of less than 2 microns. Grain-size distribution of the individual members in figure 22 is surprisingly similar to the average distribution. The material retained on the number 200 sieve ranges from less than 1 percent to about 8 percent of the individual samples. In several samples this material consists mostly of secondary minerals formed after deposition of the shale, and in all samples the plus 200-mesh material undoubtedly includes numerous aggregates of grains that were not separated during disaggregation of the samples. At the 2 micron size there is a range of 17 percent. The Sharon Springs Member shows about 57 percent finer than 2 microns; the Verendrye Member about 74 percent finer. Grain-size variation in the -2 micron sizes seems to be 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 grain-size distribution is a significant factor because it seems to control mineral distribution (fig. 23), 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 predominate mineral is quartz. The only exceptions, the Mobridge Member and the Crow Creek Member as described by Crandell (1952), both 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 grain-size components on shear strength of the shale is controlled by structural arrangement of the grains. The effects of grain structure are best understood by examining them 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 grain structure can develop in a homogeneous mixture of coarse grains and fine grains depending on the relative proportions of the two components. (1) If the volume of the interstices between the coarse grains is the same or greater than the volume of the fine grains the coarse grains can develop a stable structural network. (2) 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. Shear strength in the second case 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 are most abundant are gradual. If the volume of fine grains only slightly exceeds the volume of the interstices between coarse grains much of the shear strength is retained. Even after the coarse grains are completely isolated from each other they may effectively increase friction along shear surfaces.



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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 percent to 47.64 percent (Graton and Fraser, 1935, table II). The transition point in the theoretical mixture thus should range from about 26 percent to about 48 percent fines if the fines fill all of the voids. In the case of random packing, assuming that a stable network of coarse grains exists, the transition point probably would fall in the same range.



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Comparison of the more than 2 micron to the less than 2 micron grain-size mixture in the Pierre Shale with the theoretical mixture shows that their general behavior should be similar in spite of the differences between the hypothetical and natural materials.

The plus 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 dimensionless as in the theoretical mixture, but their average size is small in comparison with the coarse grains. On the average cumulative curve (fig. 22) about 15 percent of the -2 micron grains occur in the 1 to 2 micron size range; about 85 percent of the fine-grained component has a diameter of less than one-half that of the smallest diameter in the coarse-grained component. The shale, moreover, consists of a range of grain sizes that undoubtedly are mixed heterogeneously although they appear homogeneous under a binocular microscope.



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These variations in the grain-size mixture of the shale should reduce the proportion of fine grains 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 grains in the shale cannot completely fill all voids between coarse grains because the shale has some porosity; the volume of fine grains necessary to separate the coarse grains thus is less than the total volume of voids.

Comparison between the Pierre shale and the theoretical mixture implies that -2 micron grains in the Pierre shale become a matrix for the +2 micron grains at some proportion of finer grains less than 50 percent. The percentages of -2 micron grains in the five members of the Pierre shale in figure 22 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 of the shale is determined by this matrix. The +2 micron fraction probably has little if any effect on the shear strength of the shale and can be neglected in shale behavior studies.



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Mineral composition

Although the results therefrom are only 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 the following procedure (Identifications by Dorothy Carroll, J. C. Hathaway, C. J. Parker, and W. W. Brannock, U.S. Geol. Survey, August 1955).



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" 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 $Ca^{++}$ 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."_/

_/ The author uses 74 microns for the upper limit of silt size, rather than of 62 as given above. An average of less than 1 percent of the shale grains lie in the 62 to 74 micron size range; therefore, the approximate quantitative mineral data should be equally valid for the larger silt-size limit.



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"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 ten. 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."



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Mineral compositions of the various shale members are shown graphically in figure 23; geographic locations of the samples are given

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Figure 23.--Near here.

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in appendix I. The Virgin Creek, Verendrye, and De Grey Members are each represented by a set of samples, an auger sample ("A") from depths of 25 feet or less, and a surface grab sample ("S"). An additional auger sample was taken from a bentonite bed in the Sharon Springs Member to check possible variations between it and the enclosing shale. Sets of samples from two horizons were used for the Elk Butte and Mobridge Members. The grab samples were taken to determine the effects of near-surface weathering on mineral composition. Since auger samples were from comparatively shallow depths, they all are presumed to be at least slightly weathered.


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Figure 23.--Pierre shale mineral composition.



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The mineral composition of the Pierre Shale members, as shown by figure 23, can be easily summarized. The clay-size material is predominantly montmorillonite with a range from 5 parts in 10 in the Gregory Member to 9 parts in 10 in the De Grey, Verendrye, and Virgin Creek Members. The most important mineral in the silt-size material is quartz, which makes up 3 to 5 parts in 10 of the silt-size category. In the Mobridge Member montmorillonite never exceeds 4 parts in 10 of the clay-size component and quartz comprises only 1 or 2 parts in 10 of the silt-size component. Calcite in the Mobridge Member, although only ranging from 2 to 5 parts in 10 in a single grain-size component, is an important 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 important mineral (Crandell, 1952).

One of the most interesting features 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 1 part in 10 in the clay size portions of the samples in figure 23. In the silt-size portions of the same samples no clay mineral exceeds 2 parts in 10 and in only 5 of the 19 samples does a clay mineral comprise more than 1 part in 10.

The data in figure 23 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.

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The available data are insufficient ~~ot~~ To give a clear picture of 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 that there is little variation over a vertical distance of 10 feet but they cannot be considered representative of the total thickness of each member. The Mobridge and Elk Butte samples show that minor variations (1 part in 10) in mineral composition between surface and subsurface samples are not necessarily representative of the weathering conditions. Samples at one horizon may show a small change in the content of some mineral between the subsurface and surface samples, but in another horizon a few feet above or below the first there may be no change in the content 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 indicate 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 proportion 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. 23), commonly forms and maintains steeper slopes. It, too, is susceptible to landsliding, but 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. There are always unbalanced negative charges on the surfaces of the silicate layers that attract various exchangeable cations, which are a necessary part of the montmorillonite structure.



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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 it is known that 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 the orientation and degree of bonding in each following layer is less and the water in each layer is more fluid.

If only a little water is adsorbed the montmorillonite will be held in an essentially solid state and the clay will have relatively high shear strength. If there is a large amount of adsorbed water there is semifluid or fluid water between the silicate layers that acts as a lubricant and allows movement between the layers. The shear strength thus will be reduced to almost nothing if sufficient water can enter between the silicate layers.

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



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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.

Replacement of cations that can absorb only a thin layer of water by cations that can absorb a thicker layer also may reduce stability of the clay. 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.


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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 indicate that near-surface weathering develops a general increase in exchange capacity with calcium becoming the most important exchangeable cation. These data, however, are from samples too shallow to have any major application to landslides. Data furnished by the Omaha District of the U.S. Army Corps of Engineers (Donald K. Knight, oral communication) 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 the shale samples along the Fort Randall Reservoir.


(p. 135 follows)
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## 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 shore, and (2) to determine the relationship between different materials and landsliding. Both slope stability appraisals and geologic maps are needed to achieve these goals.

Study of the present slopes also furnishes a basis for comparison of current landslide activity to former 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 it possible to estimate the relation between geology, slope stability and natural slope angles.

Since these investigations constitute a new approach to the study of landslides, it was decided that they should be tested in selected areas representative of the reservoir environment before any attempt was made to apply them over the entire area.



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Method of selection

Two main factors are involved in selecting the areas for slope stability investigations. First, adequate large-scale base maps must be available. Second, the areas must be as representative of conditions along the entire reservoir as possible. At least one area that is close to but outside the limits of the reservoir and its effects, is needed as a control or reference area.

The areas selected are covered by sheets 23, 26, 30, 39, 42 of the Missouri River Survey, Gavins Point near Yankton, S. Dak., to Stanton, N. Dak., series made in 1947 by the Omaha District Corps of Engineers, U.S. Army. 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 essentially the entire Missouri River trench. There also are contours at 10-foot intervals along the lower part of the trench from the river level to an elevation of 1,400 feet. Contours at a 20-foot interval were completed by Special Maps Branch, U.S. Geological Survey. For convenience of reference in this report, the areas shown on map sheets 23, 26, 30, 39, and 42 are termed areas 1, 2, 3, 4, and 5, respectively.


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Area 1 ~~Sheet 23~~, just below the Fort Randall Dam (fig. 32), is the reference area. This location was chosen so that the environment would be as similar as possible to that above the dam. Tributary valleys between the dam and area 1 ~~sheet 23~~ 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 in only the one area in order to save time and money. Area 2 ~~Map sheet 26~~ (fig. 34), which includes the entire Pierre Shale section and many old and new landslides, was chosen because it is an ideal locality for correlating slope stability and materials. In the other areas only the geology that will be permanently submerged was mapped as part of this project.



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Areas 2 and 3

Sheets 26 and 30 (figs. 34, 37), near the dam, were chosen to provide a record of slope stability changes in the part of the trench that undergoes greatest inundation. These ~~sheets~~ 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 particular local factor presumably would be active in only one. Areas 4 and 5 ~~Sheets 39 and 42~~ (figs. 39, 42), in the upper part of the reservoir, include portions of the trench that will have comparatively little inundation. These areas also are close enough to provide checks on each other.



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## Scope of investigations

Investigations in the selected areas fall into the two categories of 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 is to observe changes caused by formation of the Fort Randall Reservoir. The initial studies that are discussed in this report provide a record of conditions at the time the dam was built. Periodic reappraisals will be made in the future in order to determine the slope stability changes that occur after the reservoir fills.

The second and more immediately important purpose of the slope stability appraisal is to supply data that can help us better understand the landslides and the factors that control them.



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The first phase of the appraisal was to set up a satisfactory grouping of slope stability conditions. There are no simple, absolute criteria and the classification developed has been generalized in order to be usable. It includes the major classes of stability and is designed for essentially reconnaissance type observations. As a result it does not consider such techniques as core drilling to locate slide surfaces, or observations of control points to accurately determine the activity of individual landslides. The slope stability appraisal is approximate at best, but examinations of the same area by more than one individual indicate that the results in general are reproducible.

The classification is divided into two basic categories: (1) areas that seem to have always been stable, and (2) areas that now are, or in the past have been, subject to landslides. The second category is further divided into a series of four classes on the basis of activity at the time of observation. The end points of this series are active landslides and stabilized landslides. Between these end points are two gradations, one applying to landslides that appear to be stable but show signs of very recent activity, and the other applying to landslides that seem to have been stable for a number of years but that have been recently reactivated.



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## Stable ground

Stable ground includes all material that does not appear to have ever been subject to landsliding. The only exceptions to this general rule are thin patchy deposits of stable loess, alluvium, and colluvium that overlie but do not completely mask old landslide material. In this case the stability classification is controlled by the underlying material.

There are two basic types of stable material. The first is material that was deposited independently of the Missouri River trench. The second is material that has been deposited within the trench and its tributary valleys.

The stable material that was deposited independently of the trench comprises the material into which the trench and its tributary valleys have been cut. Since 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. 24) that do not contain steep scarps probably

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Figure 24.--Near here.

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is stable. Evidence of stable ground is hard to find on steeper slopes and undoubtedly some small stable areas are not recognized in large areas of old stabilized landslides.



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Figure 24.--Area of gentle Pierre Shale slopes considered to be stable. View southwest into N½ sec. 4, T. 103 N., R. 72 W., Lyman County, S. Dak. (~~map sheet 42~~ area 5 west of the Missouri River). Photographed August 24, 1956.



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Stable material deposited within the trench and its tributary valleys can generally be detected easily. Where it is thick enough and extensive enough 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 stability conditions because they may cover older deposits that have long complex histories of landslide activity. In some areas landslides have obviously been covered by stable fill material, but it is impossible to determine the conditions beneath the fill (fig. 25). Consequently, it is very

Figure 25.--Near here.

difficult to draw an accurate contact between stable ground and landslide areas where a stable fill feathers out against old landslide topography.



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## Stabilized landslides (~~limestone~~ Ls)

Areas of stabilized landslides include landslide materials that show no very 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 produce 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 in figure 27 is typical of recently stabilized landslides that are still

Figure 27.--Near here.

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 locally beginning to obscure the limits of the block and erosion is cutting gullies into the upthrust toe.



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Figure 27.--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 Aug~~.~~ust 19, 1956.



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Identification becomes increasingly difficult as time and geologic processes continue to alter the slides. Figure 26 is characteristic of

Figure 26.--Near here.

the older stabilized landslide terrain in the Pierre shale. All of the area in the photograph, except some of the ridges along the skyline, is believed to be old landslide blocks. It is obvious that any stable ground present below the ridges cannot easily be separated from areas of stabilized landslides.

The hummocky topography in the left-background of figure 26 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 of the 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 cases, however, the two features can be easily separated because terrace remnants along a valley occur at about the same elevation whereas the levels of old slump blocks rarely coincide.

Figure 26.--Stabilized-landslide terrain. Essentially all of the area, except some of the ridges along the skyline, is believed to have slid at some time. The evidences of landsliding range from hummocky terrain to discrete slump blocks. E½ sec. 20, T. 97 N., R. 68 W., Gregory County, S. Dak. (map sheet 30 area 3 west of the Missouri River). Photographed August 15, 1956.



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Continued erosion and development of well-integrated drainage, as well as deposition of loess and colluvium, further reduces the relief until the stabilized landslide area looks like the subdued hummocky slopes in the right-central part of figure 26. Surface evidence of landslides may be completely lost in the final stage. Loess deposition has completely concealed a small slump in the Pierre shale (fig. 28). This slump would never have been recognized if it had not

Figure 28.--Near here.

been exposed by later erosion.



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Figure 28.--Small concealed slump along the Fort Randall Reservoir shore. The slump in the Pierre shale was completely hidden by later loess deposition. Erosion along the shore revealed the loess-shale relationship. Charles Mix County, S. Dak. Photographed Sept~~.~~ember 27, 1954.



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## Recently active landslides

Landslides that show evidence of very recent movement are classified as recently active even though they are apparently inactive at the time of study. There is no sure way of judging whether they are actually stable or only temporarily quiescent. The landslides in this category are important because they can become active without requiring a major change in their stability conditions.

Although the slump shown in figure 29 is recent, apparently it has not moved within the past few years. There are fresh-looking fractures around the entire slump block and at first glance it appears active; but cattle trails cross the fractures without displacement and the edges of fractures are no longer sharp.

Figure 29.--Near here.



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Figure 29.--A recently active slump. The fresh appearance of the fractures implies that this landslide has moved recently but close examination shows no present activity. The landslide may be permanently stable, or it may be temporarily quiescent. SW¼SE¼ sec. 19, T. 103 N., R. 71 W., Brule County, S. Dak. (map sheet 42 east of the Fort Randall Reservoir). Photographed Aug. 23, 1956.



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## Reactivated landslides — (Lasa)

Reactivated landslides are those that 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 portions of several stabilized slides, but is confined to old landslides that are 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 in figure 30 is a typical reactivated

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Figure 30.--Near here.

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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.


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Figure 30.--A slump-earthflow caused by reactivation of an old stabilized slump block. The slump-earthflow follows the outline of an old slump block that apparently was 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.

area !

(east side of map-sheet-23). Photographed August 20, 1956.



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## Active landslides —(La)

All landslides that show no signs of stability are classified as active. This negative criterion is necessary because actual movement has been observed in very few landslides along the Fort Randall Reservoir. Either the 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 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 dying, or very recently dead, vegetation. Figure 31, a photograph of the head of an active landslide, shows all of

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Figure 31.--Near here.

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these features except slickensides.

Some landslides classified as active on the basis of appearance may be stable. Every landslide after its final movement must pass through a stage when the signs of activity are still very fresh, but it is impossible by observation alone to separate a newly stabilized landslide from one that is still active.


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Figure 31.--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 Sept ember 26, 1955.



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## Geologic mapping

The geologic mapping done of area 2 on map sheet 26 (fig. 34) is intended to

Figure 34.--Near here.

show as accurately as possible the stable positions of the various geologic units. All exposed and apparently stable geologic contacts were plotted on 1:20,000 scale vertical aerial photographs, then transferred to the map sheet 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 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 units (p. 26) as they occur in the lower part of the reservoir apply to figure 34 map sheet 26, with the few exceptions that follow.



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Figure 34.--Geology and slope stability appraisal of area 2 made during 1953-55. Prepared by C. F. Erskine, assisted by H. Ponder, J. H. Smith, and D. Kroenlein. Lines of sections A-A' to J-J' locate slope profiles shown in figure 45.



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Pierre shale, De Grey and Verendrye members

The De Grey and Verendrye members have been mapped as a single unit in map sheet 26 (fig. 34). The lithologic similarity between the two members and the scarcity of stable outcrops make it impossible to draw any contact between them.



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Pierre shale, Crow Creek member

The Crow Creek member has not been identified in the area. It is not necessarily missing from the Pierre shale section in map-sheet 26 (fig. 34) but it may be either covered by landslides or surficial deposits; moreover, outcrops of the Crow Creek, displaced downward by landslides, would not be distinguishable from the marl facies of the underlying Gregory member.

Slow-draining terrace alluvium

Slow-draining terrace alluvium is distinguished from other slow-draining alluvium only because the separation aids interpretation of the geologic map. Physical characteristics of the two materials are similar but the terrace alluvium was deposited when the elevation of the Missouri River was considerably higher than the present channel.

## Descriptions of individual areas

Each of the five study areas has a similar geologic setting. 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 Pierre Shale members from the vicinity of the river valley. Water, ice, and wind have by repeated cycles of erosion and deposition modified this basic geologic setting during Quaternary time.



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## Area 1

Area 1 (fig. 32), immediately downstream from the Fort Randall

Figure 32.--Near here.

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. The Missouri River trench cuts diagonally southeastward through the central part (fig. 32). All of the area east of the Missouri 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 was selected to furnish a reference or control in a geologic and topographic setting similar to the other study areas but not affected by the reservoir.



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Figure 32.--Slope stability appraisal of area 1 made during 1953-55. Appraisal made by C. F. Erskine assisted by H. Ponder and D. Kroenlein.


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The base of the Missouri River trench is a steep-walled alluvium-filled valley about a mile wide that is incised into an earlier floor of the trench. The river occupies one-third to one-half of this inner valley. 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 that is 1,500 to 1,550 feet in altitude (fig. 33). Farther from the trench in the northeastern part of the sheet, this upland gives way to rolling hills up to 1,800 feet in elevation.

West of the river bluffs 80 to 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. Inland from the inner trench walls, moderately sloping hills rise to an elevation of about 1,600 feet. The slopes then steepen abruptly and continue upward until they reach an upland that averages about 1,700 feet in altitude.



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Geologic setting

The entire pre-Quaternary sequence from the Niobrara formation to the Ogallala formation is present 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 river. Alluvium covers the flood plains of the Missouri 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 downcutting of the river had begun has left till deposits along the east wall of the trench. Deposits of fine and coarse (slow-draining and fast-draining) alluvium cap terrace remnants west of the river. In the northern part of the sheet, the terrace alluvium is thick enough to mask the Niobrara formation and the lower Pierre shale. Southward it thins to isolated patches capping terrace remnants cut in the Cretaceous rocks.



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Stability conditions

Except for the limestone bluffs of the Niobrara ~~limestone~~ Formation along the river, most of the steeper slopes in map ~~sheet~~ area 23 (fig. 32) have slid at some time or other. There are few exceptions to this steep-slope, landslide-terrain relationship on the west side of the river. Some areas of relatively steep slopes east of the river, however, show no signs of landslides. The simplest explanation for ~~there~~ their seemingly anomalous stability is that sliding has occurred, but that its surface evidence now is completely lost. The lack of landslides on some steep hills in the north part of ~~sheet~~ area 23 (fig. 32) may be due to deposits of till, which is more stable than Pierre shale. Other slopes, such as those along Seven Mile Creek, may owe their stability to the fact that 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 long since reduced by erosion to patches of hummocky ground (foreground, fig. 33). Many of the individual landslides,

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Figure 33.--Near here.

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judging from their remnants, must have been quite large.



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Figure 33.--Terrain in ~~map-sheet~~ area 1 23. View eastward across the Missouri River trench in the northern part of ~~map-sheet 23~~ area 1. Photographed from approximately the center of the boundary between SW¼ and SE¼ sec. 20, T. 95 N., R. 65 W., Gregory County, S. Dak. Photographed Aug~~.~~ust 16, 1956.

The categories of active, reactivated, and recently active landslides each represent such a small proportion of sheet 23 (fig. 32) that it is easier to group them together as "active and recently active landslides." In general these landslides have a random distribution, although several localities show a concentration of recent landslide activity. Typical of these localities are the W½ sec. 19 and the NW¼ sec. 31, T. 95 N., R. 64 W.

Two ~~The~~ differences between the stabilized landslides and the active and recent landslides indicate that the slopes are more stable today than in the past. First, the presently active slides are very much smaller than the old ones. Second, 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.



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Area 2

(fig. 34), ~~is~~ about 5 miles upstream from the Fort Randall Dam, extends from lat $43^{\circ}03'$ to $43^{\circ}07'$ N., and from long $98^{\circ}39'$ to $98^{\circ}47'$ W. Land east of the Missouri River is in Charles Mix County, S. Dak.; that west of the river is in Gregory County, S. Dak.

As in area 1 ~~sheet 23~~ (fig. 32) the Missouri River trench in area 2 (fig. 34) seems to have 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 it curves about $45^{\circ}$ and heads due east. The inner valley increases in width from about 1 mile at the northern boundary to about 1-1/3 miles at the eastern boundary. Before the construction of the dam, the Missouri River occupied only one-third to one-half of the inner valley, but the reservoir now completely fills it.



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The terrain northeast of the river is rather subdued,

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although there are some steep slopes. The northeast wall of the inner valley rises steeply from the valley bottom or, locally, from narrow low terraces along the river, to a broad well-developed terrace 75 to 130 feet high. The terrace is about three-fourths mile wide near the northern edge of the ~~sheet~~ area but broadens eastward to almost 1¼ miles. It ranges from 1,350 feet elevation above the river to an average of 1,480 feet elevation at its inner margin. In the northeast part ~~of the sheet~~ the terrace passes into moderately sloping hills that rise to an upland 1,720 to 1,740 feet in elevation.

Southwest of the river (fig. 36) the terrain is much more rugged.

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Figure 36.--Near here.

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At elevations of 1,300 to 1,400 feet the steep walls of the inner valley merge with a belt of hills and ridges. Their slopes are moderately gentle to moderately steep near the river but rise precipitously to the south and west to an upland 1,780 to 1,890 feet in elevation. The most striking feature of the slopes between the inner valley and the upland is the erratic hummocky appearance characteristic of old landslide terrain.



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area 2

Figure 36.--Terrain on the southwest side of ~~map sheet 26~~. View northwest from SW¼SW¼ sec. 1, T. 95 N., R. 67 W., Gregory County, S. Dak. Photographed Aug~~,~~ust 27, 1956.



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## Geologic setting

The geologic setting of area 2 sheet 26 differs only in detail from 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 is present 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.



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Northeast of the river, the Ogallala formation is missing and the central 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 successively 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 the deposition of the upper and lower alluvium that was not recorded in the eastern part of the sheet.



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The ~~present~~ (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 alluvium ~~(Qac)~~ is mixed with colluvium and loess. Alluvium-colluvium also blankets part of the inner-valley wall in sec. 32, T. 96 N., R. 66W.



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Stability conditions

There is a strong contrast between the slope stability conditions on opposite sides of the river. Southwest of the river practically all of the Pierre shale, which makes up most of the steeper slopes, has slid. 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 can be explained by the difference in terrain. Since the slopes southwest of the river rise more or less uniformly from the flood plain to the uplands, their erosion and stability would be controlled directly by the Missouri River. Periods of increased downcutting or sidecutting, such as are known to have occurred several times in the recent history of the Missouri, would have stimulated oversteepening of slopes and contributed to conditions favorable to landslides or even landslide cycles at numerous points.



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Northeast of the river, only the steep inner wall of the trench would be directly influenced by the Missouri. The broad terrace that separates the uplands from the flood plain has acted as a buffer that drastically modifies any effects that might have been produced on the uplands by renewed cutting by the river.

The efficacy of this buffer is demonstrated by the fact that although some recent downcutting has occurred in the small stream valleys above the terrace, erosion has been relatively minor and only a few small landslides have developed on the steepened slopes. In contrast, landslides are relatively abundant along sections of the inner valley wall.

By far most landslides in area 2 are stabilized. Northeast of the river only three landslides are active. Active and recently active landslides are more abundant southwest of the river, but they still make up a very small fraction of the total landslide terrain.



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Almost all of the active and recently active landslides southwest of the river are located 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, the part where the inner valley walls are steepest. This localization probably can be linked to a period of active lateral erosion and downcutting by the Missouri River in the recent past. Erosion by the main river would also have accelerated downcutting along the major tributaries. The stable Niobrara formation was able to maintain 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 have long since been removed or completely masked. 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 is occurring in the valleys of secondary tributaries.


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## Area 3

Area 3 (fig. 37) is about 20 miles upstream from the

Figure 37.--Near here.

Fort Randall Dam and includes the area between lat $43^{\circ}10'$ and $43^{\circ}14'$ N. and long $98^{\circ}49'$ and $98^{\circ}57'$ W. The land east of the Missouri River is in Charles Mix County, S. Dak.; west of the river the area is in Gregory County, S. Dak.

The Missouri River divides area 3 (fig. 37) so that about two-thirds of the land is west of the river. On this side, the base of the trench consists of a flood plain nearly at river level and low terrace remnants at various elevations 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. 26) except where underlain by the more resistant Mobridge member that forms a zone of steeper slopes at elevations between 1,700 and 1,800 feet (fig. 38).

Figure 38.--Near here.

The shale hills are part of a maturely dissected southward-draining upland that had been established prior to the present course of the Missouri. This surface is well-developed on both sides of the river. In the southwest quarter of the area, the earlier drainage has been bissected and modified by Whetstone Creek.

Figure 37.--Slope stability appraisal of area 3 made during 1953-55. Appraisal made by C. F. Erskine assisted by H. Ponder and D. Kroenlein.



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Figure 38.--View of the upper Pierre shale slopes on the west side of the Missouri River in area 3. The steeper slopes in the background are formed on the Mobridge member, which crops out as a buff-colored band. View north-northwest from NW¼ sec. 28, T. 97 N., R. 68 W., Gregory County, S. Dak. Photographed August 15, 1956.



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The east side of the Missouri trench wall rises abruptly from the river in steep bluffs 80 to 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 elevations 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.



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## Geologic setting

The geology in area 3 (fig. 37) is essentially 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 sheet. 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 terraces and terrace remnants. Ground moraine caps the uplands in the northeast corner of the area (Stevenson and Carlson, 1950). The patchy deposits of loess and alluvium-colluvium are widespread throughout the area.


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## Stability conditions

Landsliding has occurred on most of the steeper slopes. Most of the landslides are now stable and the characteristic hummocky terrain prevails (figs. 26 and 38).

Essentially all slopes west of the reservoir show evidence of landsliding. The active or very recently active slides, except for some along Whetstone Creek, are located near the uplands. Most occur on slopes that do not have any well developed 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.



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East of the reservoir there are two general classes of slopes. The west-facing slopes adjacent to the reservoir resemble those across the river. They consist mostly of hummocky stabilized-landslide topography with a small percentage of active and recently active landslides. Much of the current landslide activity is located near the uplands, but there also is an appreciable amount of landsliding 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 dissected south-draining upland shows much less evidence of landsliding. As in area 2 (fig. 34), the terrace south of the upland has apparently acted as a buffer to protect the slopes above from recent direct erosion by the river and no evidence remains of sliding in terrace and pre-terrace time. In post-terrace times, however, the streams crossing the terrace have done more downcutting and oversteepening of their walls, than their counterparts in area 2 (fig. 34). As a result, landslides, both stabilized and active, are more common above the terrace in area 3 (fig. 37) than in area 2 (fig. 34).



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area 4

The limits of area 4 (fig. 39) are lat 43°33' to 43°37' N.

Figure 39.--Near here.

and long 99°16' to 99°24' W. The area is about 55 miles upstream from the Fort Randall Dam. Land west of the river is part of Lyman County, S. Dak., and east of the river the land is part of Brule County, S. Dak.



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Figure 39.--Slope stability appraisal of area 4 made during 1953-55. Appraisal made by C. F. Erskine assisted by H. Ponder.



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The Missouri River divides area 4 into nearly equal parts. The river flows due south to about lat 43°35' N. then swings gently southeastward to the south boundary. 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. 40) rise from

Figure 40.--Near here.

the valley floor 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 away from the Missouri. 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 well-developed terrace up to one-half mile wide. In the northern two-thirds of the area the Missouri 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 ranging from 1,700 to 1,800 feet in elevation.



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Figure 40.--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 ~~W.~~ N, R. 71 W., Lyman County, S. Dak. Photographed Aug.ust 27, 1956.



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(190 follows)

## Geologic setting

Geology in area 4 (fig. 39) differs from the areas downstream primarily in the fact that the Mobridge member of the Pierre shale is the uppermost pre-Quaternary unit present (fig. 39). Much of the geologic information used 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 the 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~~ern~~ half of ~~sheet 39~~ area 4, the Niobrara formation is well exposed north of lat $43^{\circ}35'$ 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, overlies the Niobrara. The Verendrye is the highest member of the Pierre shale exposed 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. It also floors the valley of Elm Creek above the northern limit of the terrace. Small deposits of alluvium also cap the terraces remnants to the north. Most of the alluvium is slow draining but some fast-draining alluvium is also present.



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## Stability conditions

Slides (figs. 40, 41) have occurred on essentially all of the slopes in the Missouri River trench and in most of its tributary valleys. Landsliding is less prevalent only along valley walls of the westward flowing streams in the northwest corner of the area.

Active and very recently active landslides are numerous on both sides of the river but most of the individual slides are small and form only 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 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 post-terrace time, moreover, the river has partially or completely removed the terraces north of Elm Creek but extensive landsliding has not followed. This seemingly anomalous condition of erosion without landsliding may be due to the fact that here the river is incised 40 to 50 feet into the Niobrara formation which has supported the weaker overlying shale.


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## Area 5

Area 5 (fig. 42), the northernmost study area, is about

Figure 42.--Near here.

70 miles upstream from Fort Randall Dam at the confluence of the White and Missouri river valleys (fig. 2), and is bounded by lat $43^{\circ}42'$ and $43^{\circ}46'$ N., and long $99^{\circ}21'$ and $99^{\circ}29'$ W. That part West of the Missouri River is in Lyman County, S. Dak.; that part east thereof 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 effectively isolates the southwest quarter from the rest of the area.



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Figure 42.--Slope stability appraisal of area 5 made during 1953-55. Appraisal made by C. F. Erskine assisted by H. Ponder.



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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 surface 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 old White River terraces. East of the Missouri the former White River valley has been filled with till and can be 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 to 100 feet high border the Missouri River. Above the bluffs, the walls of the trench rise steeply to a gently rolling upland that is 1,500 to 1,650 feet in elevation. Downstream, where the Missouri 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.


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The floor of the White River valley is characterized by a well-developed flood plain 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 currently attacked 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 sheet, is characterized by gentle multiterraced slopes. It steepens southward to precipitous bluffs up to 100 feet high that are in turn surmounted by steep slopes. The upland east of the river ranges in elevation from 1,620 feet to as much 1,700 feet.



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## Geologic setting

Data on the geologic setting ~~is~~ 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 elevation 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, consist predominantly of alluvium and till with minor amounts of loess and alluvium-colluvium. 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 ~~flow~~ floor of the preglacial White River valley. The lower terraces and prereservoir flood plains along the contemporary drainage system are composed largely of 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. Alluvium-colluvium forms small deposits in hollows and gullies.



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## Stability conditions

Landslides comprise a smaller proportion of area 5 (fig. 42) than of any other study areas. This can be attributed to the numerous terrace remnants left by both preglacial and postglacial rivers. These terrace remnants form steps in the valley walls so that areas of uninterrupted steep slopes are limited. The terraces, moreover, protect the ~~insulate~~ slopes above them from direct river erosion.



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There are only two areas in area 5 where landsliding has involved entire valley walls in the manner that is so common in the other representative areas. The first area 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 indicating that the Pierre shale slopes have readjusted to prereservoir conditions in the underlying Niobrara formation.

The second area is at the bend of the White River in the NE corner ~~of~~ sec. 16, T. 103 N., R. 72 W. The landsliding is less extensive than along the west trench wall, but it is more spectacular because there still are large active landslides (fig. 43).

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Figure 43.--Near here.

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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.



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Figure 43.--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.


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Active and very recently active landslides comprise 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. (see p. 218). 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 referred to as the Cable School landslide area after the Cable School located about 1 mile to the east. The Cable School landslides (fig. 44) are along a branching tributary

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Figure 44.--Near here.

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carved 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.

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Figure 44.--~~Panorama of~~ 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 Oct. 14, 1954.



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## 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 for prereservoir landslide distribution and activity, and to infer from these measures and from the relationship between landslides and local topography that landslide activity was much greater in the more distant past than in the time just prior to filling of the reservoir. The second is to show, by means of some rough statistics, the degree of correlation between slope stability, natural slope angles, and composition of geologic materials in part of the area included in figure 34.



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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 quadrangles was measured with a polar planimeter. Table 3 shows the results by percentages of total landslide area. Stabilized landslides far outweight all other classes in all areas. For the present discussion the three other classes are grouped as "current" landslides in contrast with "stabilized" landslides.



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| Map sheet | Ls | Current landslides | | |
|---|---|---|---|---|
| | | Las | Lsa | La |
| 23 | 98% | <1% | <1% | 1% |
| 26 | 92% | 2% | 3% | 3% |
| 30 | 97% | <1% | <1% | 1% |
| 39 | 99% | <1% | <1% | <1% |
| 42 | 95% | <1% | — | 5% |
| Average | 96% | <1% | <1% | 2% |

**Table ~~2~~ 3. Percentage of total area of landslides that occurs in each class of landslide activity.** Ls, stabilized ~~landslides~~; Las, recently active; Lsa, reactivated; La, active.

A comparison of the relative sizes of individual landlsides indicates that, in stab~~lizied~~ilized landslide areas, large slides apparently were more common than small slides (figs. 26, 36, 38 (below the Mobridge member), 40 and 41). The reverse seems to be true for current landslide activity. There are a few large landslides (fig. 17, 43), but there are many more small ones (figs. 10, 15). Large areas of current landslide activity are numerous only in area 2 (fig. 34).

A study of landslide activity as related to topography shows that, in the past, landslides have been prevalent on most of the steeper slopes. Many of these slopes, however, ~~no~~ now appear to be essentially 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.

The topographic distribution of current landslides reveals that most new slides are directly related to erosion along the smaller tributary streams with steepness of slope only a secondary factor in determining their location. Even in large areas of current landslide activity (fig. 34) there is very little landsliding along the precipitous inner trench wall.



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The information summarized above indicates that, in immediately prereservoir time, the walls of this portion of the Missouri River trench were more stable than at anytime in their preceding history. A preponderance of the evidence suggests that landslide activity was relatively minor at the time the reservoir began to fill.

Two other factors probably have also contributed to the comparatively large landslide activity of the past and the minor 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 reduce slope stability. ~~lubricate the slides.~~ Precipitation must have been higher 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 quite dry by comparison with attendant diminution of ground-water supplies.

Second, during the glacial and much of the postglacial times, the Missouri was cutting through the easily eroded and unstable Pierre shale in which slope failures are common. more recently, ~~past,~~ 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.


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Correlation of landslides, slope angles, and geologic materials

The land southwest of the Missouri River in area 2 (fig. 34) is well suited to an investigation of possible correlation 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 beds since these include the materials most prone to landsliding.



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## Preparation of data

The information on slope stability conditions of the pre-Quaternary geologic units in area 2 (Fig. [illegible]) is given as percentages. First the area of each stability class in every unit was 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 percent of each stability class within individual units. The results are presented in table 4.

Examination of table 4 shows that the Pierre shale comprises 97.4 percent of the pre-Quaternary exposures and includes all but 1 or 2 percent of each class of landslides. Field observations indicate, moreover, that landslides in the Ogallala and Niobrara formations are, with few exceptions, the result of landslides in the shale. Since essentially all of the landslides take place in, or are controlled by, the Pierre shale, the discussion of the correlation between landsliding, slope angles, and geologic materials is restricted to that formation.

| Name of unit | | Symbol | % of total area of exposures of pre-Quaternary formations | % of unit | | | | | | ~~% total of its own kind~~ % of each stability class in stratigraphic units. | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | | | | Stable | Ls | Las | Lsa | La | All landslide categories | Stable | Ls | Las | Lsa | La |
| Ogallala formation | | To | 2.5 | 53 | 44 | <1 | 2 | <1 | 47 | 25 | 1 | <1 | 2 | 1 |
| Pierre shale | Elk Butte | Kpe | 22.9 | 12 | 79 | 2 | 4 | 3 | 88 | 56 | 21 | 14 | 34 | 24 |
| | Mobridge | Kpm | 19.5 | <1 | 90 | 4 | 3 | 3 | 100 | 1 | 20 | 31 | 20 | 22 |
| | Virgin Creek | Kpvc | 16.1 | 0 | 92 | 3 | 3 | 2 | 100 | 0 | 17 | 19 | 16 | 11 |
| | DeGrey-Verendrye | Kpdv | 30.2 | <1 | 93 | 3 | 2 | 2 | 100 | 2 | 32 | 33 | 24 | 23 |
| | Gregory | Kpg | 6.0 | 5 | 89 | <1 | 1 | 4 | 95 | 6 | 6 | 2 | 2 | 11 |
| | Sharon Springs | Kps | 2.7 | 15 | 76 | 0 | 2 | 7 | 85 | 8 | 3 | 0 | 2 | 8 |
| | entire formation | | 97.4 | 4 | 88 | 3 | 3 | 2 | 96 | 73 | 99 | 99 | 98 | 99 |
| Niobrara formation | | Kn | 0.1 | 86 | 14 | 0 | 0 | 0 | 14 | 2 | 0 | 0 | 0 | 0 |

Table ~~3~~ 4. Slope stability conditions of pre-Quaternary deposits in map sheet 26 southwest of the Missouri River.

Data for the slope angles were averaged from 10 cross sections (fig. 45). Locations of the cross sections are plotted on figure 34.

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Figure 45.--Near here.

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Average slope angles were computed for every pre-Quaternary geologic unit exposed along each cross section 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 5).

Following the assumption made in the section on analysis of the Pierre Shale (p. 130) that only the proportional amount of montmorillonite in the clay-size component of each member affects the stability of that member, the montmorillonite content (fig. 23) was averaged and incorporated in table 6. Although some of the samples (fig. 23) were taken as much as 60 miles from area 2 (fig. 34), the mineral composition of the Pierre Shale apparently is relatively consistent throughout the Fort Randall area.

Figure 45.--Slopes along 10 cross sections in area 2. See figure 34 for location of lines of sections.



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| Cross section | Kpe | Kpm | Kpvc | Kpdv | Kpg | Kps |
|---|---|---|---|---|---|---|
| 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° |

Table ~~4~~ 5. — . Average slope angles of the Pierre shale members along cross sections A-A' to J-J'.

## Analysis

Although the data from area 2 (fig. 34) are quantitative and permit statistical analysis, the selection of a workable method was restricted by the fact that the data are limited both in quantity and quality. Some information, such as the distinction~~s~~ between stable ground and landslide areas, ~~are~~ 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 samples scattered over the entire reservoir area.

Slope stability, slope angles, and montmorillonite clay content, moreover, are not interrelated to the exclusion of all other factors. Geographic and stratigraphic locations undoubtedly have a major effect. 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 are unaccounted for.



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The method selected to investigate possible correlations between the data was the determination of Spearman's rank correlation coefficient. This coefficient indicates the chance for a relationship between two classes of data that are ranked according to increasing or decreasing values. The use of a method that deals with ranks instead of actual numerical values minimizes the effects of minor discrepancies and also simplifies correlation of dissimilar classes of data.

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

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

in which,

$D$ = difference in rank between paired items in two series

$\sum D^2$ = 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, through 0 for completely random distribution, to -1 for complete inverse correlation. Only two series can be correlated at one time.


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The first step for determination of the rank correlation coefficients is to assign ranks to each item (table 6). Where two or more items are of equal value the respective rank numbers are averaged between them. After the data ~~is~~ are ranked the rank correlation coefficient is calculated between each pair of series.



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| the Pierre shale member of | Average montmorillonite content of clay-size fraction (parts in 10) | Rank | Average slope angle (degrees) | Rank | Relative amount of landslide terrain (percent) | Rank |
|---|---|---|---|---|---|---|
| Elk Butte (Kpe) | 7 | 3 | 11 | 4½ | 88 | 5 |
| Mobridge (Kpm) | 4 | 6 | 11 | 4½ | 100 | 2 |
| Virgin Creek (Kpvc) | 9 | 1½ | 6 | 2 | 100 | 2 |
| Verendrye-DeGrey (Kpdv) | 9 | 1½ | 5 | 1 | 100 | 2 |
| Gregory (Kpg) | 6 | 4½ | 11 | 4½ | 95 | 4 |
| Sharon Springs (Kps) | 6 | 4½ | 11 | 4½ | 85 | 6 |

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Table ~~5~~ 6. — Ranking of Pierre shale data.

Montmorillonite content-slope angle rank correlation coefficient

| Pierre shale member of | Rank of montmorillonite content | Rank of slope angle | D | $D^2$ |
|---|---|---|---|---|
| Kpe | 3 | $4\frac{1}{2}$ | 3/2 | 9/4 |
| Kpm | 6 | $4\frac{1}{2}$ | 3/2 | 9/4 |
| Kpvc | $1\frac{1}{2}$ | 2 | $\frac{1}{2}$ | $\frac{1}{4}$ |
| Kpdv | $1\frac{1}{2}$ | 1 | $\frac{1}{2}$ | $\frac{1}{4}$ |
| Kpg | $4\frac{1}{2}$ | $4\frac{1}{2}$ | 0 | 0 |
| Kps | $4\frac{1}{2}$ | $4\frac{1}{2}$ | 0 | 0 |

$$\sum D^2 = 5$$

$N = 6$

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

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

$$= 0.857$$

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

| Pierre shale, member | Rank of montmorillonite content | Rank of relative amount of landslide terrain | D | $D^2$ |
|---|---|---|---|---|
| Kpe | 3 | 5 | 2 | 4 |
| Kpm | 6 | 2 | 4 | 16 |
| Kpvc | 1½ | 2 | ½ | ¼ |
| Kpdv | 1½ | 2 | ½ | ¼ |
| Kpg | 4½ | 4 | ½ | ¼ |
| Kps | 4½ | 6 | 3/2 | 9/4 |

$\Sigma D^2 = 23$

$N = 6$

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

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

$$= 0.343$$



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Slope angle - relative amount of landslide terrain rank correlation coefficient

| Pierre shale member of | Rank of slope angle | Rank of relative amount of landslide terrain | D | $D^2$ |
|---|---|---|---|---|
| Kps | $4\frac{1}{2}$ | 5 | $\frac{1}{2}$ | $\frac{1}{4}$ |
| Kpm | $4\frac{1}{2}$ | 2 | 5/2 | 25/4 |
| Kpvc | 2 | 2 | 0 | 0 |
| Kpdv | 1 | 2 | 1 | 1 |
| Kpg | $4\frac{1}{2}$ | 4 | $\frac{1}{2}$ | $\frac{1}{4}$ |
| Kps | $4\frac{1}{2}$ | 6 | 3/2 | 9/4 |
| | | | | $\Sigma D^2 = 10$ |

$N = 6$

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

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

$$= 0.714$$



p. 219a follows

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 1 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 1 chance in 20 of representing random correlation between independent 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 have any significance. 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.


(p. 220 follows)
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Interpretation

Correlation between montmorillonite content and slope angles is to be expected if the conclusion (p. 130)--that shear strength of the Pierre shale is an inverse function of montmorillonite content in the clay-size portion of the shale--is correct. The shear strength in turn largely determines the maximum natural slope angle of the shale. The maximum possible slope angles, therefore, would be finally controlled by the montmorillonite content of the shale.

The line of reasoning in the preceding paragraph admittedly is much oversimplified. It intends only to present the general principle of montmorillonite content-slope angle correlation. Lack of complete quantitative correlation between the montmorillonite contents and slope angle data for the shale members (table 6) presumably is due to various modifying factors (p. 213). Examination of the data in table 6 and on p. 218 shows that the lack of correlation between the montmorillonite content and the relative amount of landslide terrain may be due largely to the apparently contradictory statistics for the Mobridge member. Table 6 shows that, 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. this rather surprising lack of stability in the Mobridge is probably due to the fact that the underlying members are high both in montmorillonite content and in percentage of landslide terrain.



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The only obvious correlation between slope angle and the relative amount of landsliding is that factors which permit the shale to maintain steep slopes will also tend to reduce the amount of landsliding. The amount of landsliding, for example, seems to be controlled at least in part by stratigraphic and topographic location. The most stable members of the Pierre shale (table 4) are the Sharon Springs and Gregory, which overlie the Niobrara formation, and the Elk Butte member, which is the uppermost member and comprises part of the stable upland surfaces (fig. 34). The relatively stable Niobrara formation apparently increases stability of the overlying Sharon Springs and Gregory members by forming a ~~stable~~ firm foundation for them. The fact that the Gregory, separated from the Niobrara by the Sharon Springs, has a higher percentage of landslides shows that the effectiveness of the support diminishes with the distance from the foundation.


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In contrast, the Mobridge, although its low montmorillonite content should ~~enable~~ allow it to stand up better than the Sharon Springs and Gregory members, is extremely liable to landsliding. The fact that it is underlain by the highly unstable Virgin Creek and Verendrye.— De Grey members, undoubtly tends to nullify its strength and reduce its value as a potential support for the Elk Butte. The Elk Butte member in spite of its fairly high montmorillonite count, probably has fewer slides and steeper slopes because there is no great thickness of overlying sediments creating stresses within it. If the rank correlation coefficient between slope angles and relative amounts of landslides (0.714) results from some degree of correlation between the two series of ranks, the correlation must be because steeper slopes can develop where external factors restrict landsliding.



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## 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 related to glacial times.

(2) The Pierre shale is responsible for virtually all of the landslides.

(3) All of the Pierre shale members are subject to landslides. Relative stability of the members is reflected by the angles on which they form stable slopes and not by the relative amount of landslides in the members.

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

(5) Restricted landslide activity in some shale members seems to result from 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 may be stable on what otherwise would be abnormally steep slopes.



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Two general conclusions can be drawn. The first is that the Fort Randall Reservoir should have little effect on slope stability beyond the immediate shoreline unless wave erosion is on an unexpectedly large scale, or unless there is an unforeseen major rise in the ground-water table within the Pierre Shale. It would take a major change from the present environment to restore, or even approach, the conditions of glacial times. The second conclusion is that slope stability is directly related to the percentage of montmorillonite in the clay-size fraction of the Pierre Shale, although other moderating factors may also influence slope stability.



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## Ground-water investigations

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

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## Purpose

The ultimate goals were 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. The first creates permanent, or semipermanent, ground-water changes. The second includes periodic or cyclic processes that either occur rather ~~more or less~~ regularly or have occasional random occurrences.

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 in order to redevelop a state of equilibrium. This adjustment is a long range process controlled by the amount of water available and by the permeability of the ground. This report does not consider in detail the effects of the reservoir on the water table because the records available to date are insufficient.


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The most obvious of the cyclic factors are the seasonal climatic variations. Other factors are alternate periods of drouth 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 at least qualitatively on the basis of available ground-water records.

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## 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 by the author.

Porous tube piezometers were considered most suitable for several reasons. First, the materials are inexpensive. Second, several independent piezometers can be installed at different depths in a single drill hole providing ground-water data for several horizons at each site without materially increasing installation costs. Third, piezometric pressures will respond to relatively small changes in water volume. Small-diameter porous tube piezometers thus are quite sensitive to pressure changes in relatively impermeable materials. Finally, porous tube piezometers are nonmetallic and will 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 section and a standpipe made from three-eighths inch inside diameter plastic tubing (fig. 46). In the

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Figure 46.--Near here.

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drill hole the porous tube is packed in clean sand and seals of bentonite clay isolate it from water in horizons above or below the sand so that the water level in the standpipe reflects piezometric pressures at the horizon of the porous tube.





p. 228a follows

Figure 46.--Schematic diagram of piezometer installation.



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p. 229 follows

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


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## Piezometer locations

Twenty piezometers were installed at six sites along the reservoir shore and two sites downstream from the dam. General locations of the piezometer sites are shown on figure 2; detailed descriptions of the locations are given in appendix II. The two sites below the dam were selected as control installations for determination of regional ground water changes that may occur independently of the effects of the reservoir.

The piezometer sites are grouped generally about Pickstown and Chamberlain in order to make them easily accessible for periodic measurements. The piezometers in the vicinity of Pickstown include the ~~two~~ 2 control sites and ~~two~~ 2 sites in the lower part of the reservoir where ground water changes should be greatest. The piezometers in the vicinity of 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.



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Sites 1 and 2 at the town of Chamberlain were specifically intended to investigate seepage pressures in a terrace composed of fine surficial deposits quite similar to materials susceptible to landsliding 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 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 (site 4 and sites B to D), including one below the dam, are in Pierre Shale; the other four (site A and sites 1 to 3) are in alluvium. This distribution was selected in order 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. These depths and intervals were selected so that the piezometers would be far enough apart to ~~prevent~~ avoid ground-water migration from one horizon to another. An additional piezometer was installed at a depth of 12 feet at site B in order 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 1 to 3 piezometers were installed at each site.

## Results and interpretations

Water levels for each piezometer, together with the driller's log at each site, are shown graphically in figure 47. The reservoir

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Figure 47.-- Near here.

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levels at both Pickstown and Chamberlain are also plotted for comparison with ground-water levels. The following discussion of results and interpretations refers to the data in figure 47.

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



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Figure 47.--Piezometer readings and reservoir levels, with driller's log at each site.



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Relationship between reservoir level and ground-water level

As of November 1956 there had been little indication of a relationship between reservoir 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.


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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 reservoir. This could be expected because the piezometers are located near the reservoir in slow-draining alluvium, one of the more permeable materials along the reservoir. Piezometer 1-60 apparently shows a direct relationship between reservoir 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 have resulted from ground-water changes brought about by the rising reservoir level.



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During 1954 and 1955, the water level in 2-24 was high in the winter and dropped ~~fell~~ during the hot summer months. In 1956, however, there was a steady rise until in November it stood about 3 feet above the former normal level. This behavior suggests that the water table at site 2 began to respond in 1956 to the raised water level of the reservoir. If this assumption is correct there has been a lag of about 1½ years from the time the reservoir level started to rise to the time that the water table about 750 feet inland from the reservoir started to respond.

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Comparison between water levels in the reservoir and in piezometer 1-60 reveals an interesting relationship. The water level in the piezometer was above the reservoir level until about the end of September 1954. From that time at least until May 1956, conditions have been reversed so that the reservoir water level has been generally higher than the water level in the piezometer. The prereservoir hydraulic gradient obviously was toward the river, whereas from fall 1954 to at least spring 1956 water was moving from the reservoir into the terrace deposits. Judging from all the other piezometric data the normal hydraulic gradient along the Missouri River trench must be toward the reservoir. The reversed hydraulic gradient that has existed for at least 1½ years after the rise in the reservoir level is, therefore, a local anomalous condition that should exist only until ground-water conditions become adjusted to the new reservoir level. If ground-water adjustments in slow-draining alluvium are incomplete at a distance of less than 150 feet from the reservoir after 1½ years, it seems safe to assume that the overall readjustment along the reservoir may take at least 10 years.


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## Relationship between cyclic processes and ground-water levels

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

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


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## Behavior of ground water

This section includes a brief discussion of ground-water behavior at each piezometer installation followed by a general summary of ground-water behavior in alluvium and in Pierre Shale.



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## Site A

All of 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 located 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 located approximately at the ~~surface~~ top of the water table. There rarely is much more than 2 feet of water in the system and frequently it is dry during part of the summer and early fall.

The water level in A-30 generally stands 8 to 10 feet above the base of the porous tube, or about 7 to 9 feet below the water table surface. Fluctuations in the water table are crudely reflected at 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.



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Piezometric pressures in piezometer A-81 apparently are smaller than in A-30 despite the greater depth. The water level in A-81 has rarely been 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 for a short period from December 1954 to February 1955. There was a sudden rise of the water level in piezometer A-14 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 attributed either to unknown hydrologic factors or simply to clogging of the piezometer tube.



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Site B

All of the piezometers at site B are located in shale ~~is~~ 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.

Unexpected conditions exist in the water levels of the piezometers at site B. Piezometer B-12 has been dry almost half ~~of~~ the time, but 14 out of 24 measurements have shown up to 1.9 feet of water. Piezometer B-30, 18 feet below B-12, has been dry at all measurements. Piezometer B-59, 29 feet deeper than B-30, has consistently contained about 10 feet of water, and the minor water-level fluctuations roughly correlate with the fluctuations in B-12. The water level in B-59, moreover, was essentially 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 ~~depth~~ height of about 34 feet above the piezometer.



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The behavior of the piezometers at site B is anomalous for normal ground-water conditions in material of uniform permeability, but is logical 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 could become saturated even though fresh shale below was 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, records nothing.

Although piezometer B-59 is in fresh unweathered shale, the comparatively rapid development of equilibrium and the sizable monthly water-level fluctuations suggest that the water is moving through fractures rather than the shale itself. The presence of water in B-12 and B-59 and the lack 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 would be reflected by piezometers intersecting that system.


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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 required for piezometer B-92 to reach equilibrium in comparison with 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 substantiate 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, personal communications).



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



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Such behavior of the water levels at site C might be expected if ground-water movement in the shale matches the movement postulated for 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.


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Site D

All three piezometers at site D are ________ in unweathered Pierre shale. Water in the shallowest piezometer, D-32, has an average depth of about 10 feet, which suggests that the water table lies about 20 feet below the ground surface. Although the water level in the piezometer has been relatively constant there has been a slow unexplained rise totaling about 2 feet in the 3-year period 1954-1956. Small fluctuations, generally less than a foot in magnitude, are superimposed on the gradual upward trend of the water level.

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



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Behavior of piezometer D-92 implies that it is subject to slightly different conditions than the shallower piezometers. After the piezometers were installed the water level in D-92 lagged behind D-32 and D-62 and did not reach approximate equilibrium 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 difference in water levels contrasts with a difference of only 0.5 feet between water levels in piezometers D-32 and D-62.



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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 reflects the relative ease with which water moves along the fractures. Slow development of equilibrium in D-92 indicates less permeab~~le~~ility than at the shallower horizons. The greater difference of water levels between D-92 and D-62 than between D-62 and D-32 add to the impression that D-92 has either encountered a small relatively closed fracture or lies entirely in unfractured shale.



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Sites 1 and 2

The piezometers at sites 1 and 2 were installed to measure water pressures in the Chamberlain terrace (see p. 235). 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 between 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 edge. 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 recorded at 33 feet below the surface of the terrace.


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Site 3

The single piezometer installed at site 3 measures water pressures in a 6-foot thick bed of fast-draining alluvium that is underlain by the Niobrara formation and overlain by about 39 feet of slow-draining alluvium.

Measurements of one piezometer can give no data on either 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.



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Site 4

The drill hole at site 4 penetrated 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 pressures in the shale. The deepest piezometer, 4-93, measures pressures in both the Pierre shale and the Niobrara Formation near their contact.

This independence in the water levels in these piezometers is in direct contrast to the other shale piezometer installations, each of which showed an interdependence between the water pressures at the different levels. The water in 4-32 has had an average depth of 16 feet, greatest amount of the three, and has shown a maximum variation in level of 1.5 feet. Piezometer 4-57, in contrast, had a maximum depth of 1.8 feet and has been dry periodically despite the apparently permanent ground-water body above it. The deepest piezometer has had remarkably stable depth of 2 feet. With the exception of 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 relationship at all with the conditions at the shallower piezometers. There is neither correlation between water-level fluctuations nor a rise in water level toward a water table defined by one of the shallower piezometers.


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## Summary

The following brief generalities of ground-water behavior at the piezometer installations are valuable only as working hypotheses, inasmuch as they are based on data from only 8 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 layer with less permeability 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 of the ground-water movement in Pierre Shale apparently is along fractures. The water can move either vertically or horizontally. The rate of movement depends on the degree of development of fractures in that 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 ground-water 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 this circumstance is quite similar to that of water in stratified alluvium.


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Piezometer measurements were continued until December 1959 by the Ground Water Branch, U.S. Geological Survey, Huron, S. Dak. These have not been plotted on figure 47 nor have they been correlated with reservoir levels and precipitation surpluses. In general, these later measurements can add little to the previous discussion except the following minor corrections and additions:

Site A. The difference in water level between A-14 and A-30 through the latter part of 1958 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 1958-1959.

Site C. Minor fluctuations in C-55 and C-88 were apparently unrelated. C-29 became plugged in early 1959 and the shelter destroyed and other tubes plugged later in the year.

Site D. Water levels in D-32 and D-62 became practically coincident in the Spring of 1957 and remained so 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 lies 3 - 4 feet lower.

Site 1 and 2. Piezometer 1-29 remained dry or had less than 2 feet of water during 1957-1959, similar to conditions in 1954-1956. The water level in 1-60 rose to elevation 1351.9 in July 1958, and dropped through the rest of the year to 1344.9 in December. The

Landslide movements

Methods of investigations

An estimate of the time, rate, methods, and magnitude of movements was attempted for five individual landslide groups (see fig. 2). A network of control points was established in each area. These points were relocated periodically and compared with previous readings to determine the amount of movement.

Control points on the Highway 16, Landing Creek, and Paulson no. 2 slides and on the Cable School earthflows were located by transit traverse. At least two points on each traverse were located on presumably stable ground as reference points. The main traverses were closed in order to check surveying errors, but short open traverses to isolated points were extended from the closed traverses.

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



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The control points were 1- by 2-inch stakes 2 to 4 feet long driven into the ground until about only the upper 6 inches of the stake was exposed. At many points, however, more than 6 inches had to be exposed so that 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 had to be either replaced or abandoned.

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



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## Descriptions of landslides

Except for the Landing Creek slump-earthflow the geologic maps represent the geology before movement occurred. Landslide material and other shallow colluvial deposits as well as bedrock geology were mapped at the Landing Creek slump-earthflow to demonstrate the masking effects of the landslide materials. Mapping also showed some of the larger stratigraphically displaced blocks of Pierre shale that probably are remanents of old landslide blocks. Landslide and colluvial debris were not shown at the other three slides so that the causes of the landslides could be more clearly separated from their effects.



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Highway 16 slump

The Highway 16 slump (figs. 10 and 48) is in NE¼NE¼ sec. 22,

Figure 48.--Near here.

T. 104 N., R. 71 W., on U.S. Highway 16 about 1 mile southeast of the center of Chamberlain. The landslide occurred on the nose of a Pierre Shale ridge at the contact between the De Grey and Verendrye Members and is one of the few pure slump movements in the vicinity of the reservoir. Instead of disintegrating into an earthflow the base of the landslide has pushed upward and forward over the stable ground as a single unit forming a pressure ridge along the toe.

The location of the landslide is not one where exceptional quantities of ground water would be expected, but during 1952 and 1953 lush vegetation in and around 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 has contained standing water every spring.

The slump (fig. 48) consists of 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 backward into the open fracture behind them to form the two smallest blocks. At about the same time, or shortly after, the laterally unsupported edge of the stable ground directly upslope from the fracture also collapsed into the opening.


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(p. 258a follows)

Figure 48.--Highway 16 slump. Topography planetabled in 1953 by C.F. Erskine and D. Kroenlein. Geology mapped in 1956 by C.F. Erskine.

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p. 259 follows

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 were 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 of the 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 stable from 1954 to 1956. The error in horizontal closure of the closed transit traverse was 0.74 ~~feet~~ foot in 1953 and was never greater than 0.19 ~~feet~~ foot in the following years. The minor difference in locations for point 6 between 1953 and 1954 implies that most of the error in the 1953 measurements must have been 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 greater than 0.2 ~~feet~~ foot.



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The vectors plotted for each point show that there has been some relatively minor movement between 1953 and 1956. Control point 3 at the top of the lowest slump block had about 0.4 feet of movement, 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 have been essentially north and northeast downslope and toward the pressure ridge. Exceptions to this rule, point 2, and point 1 below the slide, showed southeastward movements of about 0.3 feet between the 1954 and 1955 measurements. Since no adequate reasons could be found to explain a localized change in ground movement, it has been assumed that these two stakes probably had been disturbed by cattle. The magnitude of movement at points 2 (with the exception of the (presumably disturbed) 1954-55 lateral displacement), 3, and 4 was greater between 1953 and 1954 than between the later measurements. The points beyond the limits of the landslide, with the exception of number 1, had neither individual nor cumulative movements as great as 0.2 feet, which is within the limits of surveying error.



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The apparent movements of points 5, $Z_2$, $Z_5$, and $Z_6$ are the only ones outside the landslide boundaries that may have some significance. Although small and within the limit of error in surveying, 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 unslope from the landslide, (3) plastic deformation of the unsupported material upslope from the landslide. At present, however, the available information is too limited to assign a specific cause to these movements.

Prior to 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 small distance and the high cohesion of the shale enabled the toe of the landslide to remain more or less intact instead of disintegrating into an earthflow.

The decrease from an initial major movement in 1952 through minor movements between 1953 and 1954 to very minor movements after 1954 implies that the landslide was becoming stabilized under the existing climatic conditions. A change, however, to climatic conditions more conducive to landslides might reactivate movement.


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## Cable School earthflows

The Cable School earthflows /figs. 49 and 15 (two larger flows)/

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Figure 49.--Near here.

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in NW¼NW¼NW¼ sec. 30, T. 103 N., R. 71 W., Brule County (fig. 42) 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 in maximum dimensions, the smaller one is about 100 by 110 feet. The greatest depth of disturbed material is less than 10 feet.

The two earthflows are located 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 stratum of gravel (fast-draining alluvium) that is in turn capped by 15 to 20 feet of till. Several feet of colluvium blankets all the earlier deposits. Although much of the colluvium normally moved by slope wash and creep, apparently former landslides also transported large quantities.



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p. 262a follows

Figure 49.--Cable School earthflows. Topography planetabled in 1952-1953 by C. F. Erskine, D. Kroenlein, and W. Smyth. Geology mapped in 1956 by C. F. Erskine.



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Both earthflows are quite similar in appearance. Below the prominent scarps at the head of each slide there is a zone of slump blocks. This zone makes up about half of the larger flow but less than one-fourth of the smaller one. The remainder of each slide has moved as a true flow. The toes have a 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 (fig. 49) on the larger earthflow when it was first examined in July 1952. The water must have been derived from local precipitation since 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 water.



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When the earthflows were first examined in July 1952, the still partially saturated materials, the freshly uprooted vegetation, and the active seeps indicated that ~~they~~ the flows must have occurred in the 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~~s~~ (fig. 49) toward the smaller one; the crack showed no change after 1953.

Eight control points were established on and in the vicinity of 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 feet; movements, therefore, are assumed accurate to at least 0.2 feet. All movements of the control points from 1953 to 1956 were randomly oriented and neither single nor aggregate movements at any point totaled more than 0.2 feet. These measurements indicated that there was essentially 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 essentially all of their movement occurred at one time, or at least during a short interval of time. They now appear to be stabilized and give no indication of future movement.


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Cable School slump-earthflow

The Cable School slump-earthflow in SE¼NE¼NE¼ sec. 25, T. 103 N., R. 72 W., Brule County (figs. 11, 50) was one of the most active

Figure 50.--Near here.

landslides during the investigations along the Fort Randall Reservoir. The slide is about 400 by 700 feet in maximum dimensions and is divided nearly equally between slump and earthflow movements.

The slump-earthflow is located on the southwest wall of the Cable School area directly across a small valley from the two Cable School earthflows (see p. ~~303-305~~ 262-264). The geologic setting, as a result, is similar: Sharon Springs and Gregory members of the Pierre shale are overlain by gravel (fast-draining alluvium) that in turn is capped by till. Colluvium forms a thin blanket over most of the surface and obscures much of the underlying geology.

A zone of springs exists at and just below the gravel-Gregory member contact. The ground-water source for the springs must be extensive because the flow was continual, although never large, 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.

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p. 265a follows

Figure 50.--Cable School slump-earthflow. Topography planetabled in 1953 by C. F. Erskine and D. Kroenlein. Geology mapped in 1956 by C. F. Erskine.


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The upper portion of the Cable School slump-earthflow is a broad but relatively short area, about 400 by 250 feet, composed of numerous slump blocks. Differential settlement of the slump blocks has produced a horst-and-graben structure both in the major blocks and, on a smaller scale, in the subsidiary units. A steep scarp 20 to 30 feet high forms the upper limit of the slump. The lower limit is approximately at the ground moraine-gravel contact (fig. 50) 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 portion of the landslide narrows from a width of almost 300 feet at the gravel-moraine 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. 11 pt. 4), behaves more as a debris slide than an earthflow because there is not enough moisture to permit flowage. Instead the springs sap the base of the slump blocks and the material slides down slope in semiplastic masses. This debris slide material is then saturated by the springs and moves as an extremely viscous liquid.



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The slide probably started as a predominantly slump-type movement with failure occurring in the saturated Pierre shale immediately below the gravel. Aerial photographs flown in June 1949 show the landslide as a small slump-earthflow less than half its size in figure 50. By July 1952 it had grown to the approximate size shown in figure 50. The lack of vegetation on the earthflow and freshness of the cracks and scarps showed that much of the movement had probably occurred that spring. Although no other ~~more~~ 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, the ninth point was just above the upper scarp, and the tenth ~~10th~~ on the ridge top above the landslide (figs. 11 and 50). Measurements were made approximately monthly during the fall of 1953 and the summers of 1954 and 1955.


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Control point 10, on stable ground, was the farthest from the triangulation base line, 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~~s~~. The loci of measurements of point 10 all fall within a circle slightly less than 0.5 feet in diameter; the greatest error in any measurement, therefore, is presumed to be less than 0.5 feet. The movements of point 9 that are greater than 0.5 feet are believed to represent the effects of creep on the steep slope immediately above the landslide.

Control points on the slump and flow portions 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 even 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 has 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 5 times as far as point 3.



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



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Another interesting observation derived from the control point measurements is the changes in rates of movement. Points 3, 4, and 6 have all shown a decrease 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. During the second year, movement at these points accelerated so that they moved about 3 times as far during the summer of 1955 as during the summer of 1954. Since the rate of movement at the toe of the earthflow reflects changes that occur farther up the slide, the acceleration at the toe probably resulted from addition of materials to the earthflow from the slump portion 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 forward movement of the individual slump blocks.



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## Landing Creek slump-earthflow

The Landing Creek slump-earthflow (figs. 17 and 51) is the largest

Figure 51.--Near here.

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. A number of smaller slumps and earthflows are superimposed along the toe of the main slump.

The landslide is located 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 pre-reservoir Missouri River level, and moved forward onto the terrace. There is no direct connection, consequently, between river erosion and landslide activation.



p. 271a follows

Figure 51.--Landing Creek slump-earthflow. Topography planetabled in 1953 by D. Kroenlein and C. F. Erskine. Geology mapped in 1956 by C. F. Erskine.


(p. 272 follows)

The ridge consists of the lower five members of the Pierre shale underlain by the Niobrara Formation. In pre-reservoir time the Niobrara Formation formed bluffs about 35 to 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 elevation 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 elevations ranging from 1,375 to 1,420 feet. Above the Crow Creek, and comprising essentially 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-Verendrye landslide material, but only the largest are indicated on the geologic map (fig. 51).

An understanding of the Landing Creek slide can be obtained 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 stabilized sooner. Although considerably modified by erosion, the southeast third is a jumbled series of slump blocks below a 30- to 40-foot 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 to 200 feet wide. The southeast margin of the graben is bounded by a second scarp up to 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 comprise about one-half its total length. Earthflows predominate; slump-type movement is dominant only in the northwest part of a subsidiary slide located at the grid location of 1,100 to 1,400 feet north and 1,600 to 2,000 feet east (fig. 51). Two prominent earthflows extend down small valleys to the pre-reservoir Missouri River level. The largest head in the north corner of the main slump and is about 750 feet long. It has a dumbbell-like 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 gulley.

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 explanation is that weathering processes over a long period of time had weakened the material so much that it became unstable on the existing slopes. Ground-water and minor gulley erosion probably then triggered movement in the weakened mass.

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Direct erosion by the Missouri River had no effect because the terrace remnants acted as a buffer to protect the upper slopes. The terrace surface, however, was dissected by many small gulleys which formed furrows in the general surface. As the slump moved onto the terrace, some material collapsed into these furrows and destroyed the unity of the original slump. Where the gulleys were broad and shallow the slump broke into fairly large blocks; where the gulleys 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 (fig. 51). 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 pre-slide 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 and thus appreciably increased the amount of ground water available.



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Little is known about the movement of the landslide before the control points were established in August 1953. A Corps of Engineers' map of the area _/, which was compiled photogrammetrically from aerial photographs taken in 1945, shows that the landslide was already quite extensive with at least one subsidiary earthflow that reached the river. taken in 1949 aerial photographs show the landslide as virtually the size that it was from 1952 to 1956. The largest earthflow has a larger lobe than on the earlier Corps of Engineers' map and the smaller earthflow southeast of it 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 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.

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_/ Missouri River Survey, Gavins Point near Yankton, S. Dak. to Stanton, N. Dak.; Omaha District, Corps of Engineers, U. S. Army, sheet 37, 1947.

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

The measurements indicate that there has been appreciable movement of all the points on the landslide during the three-year period, and also that the range in amount of movement has been very great. Point 4 had the least movement, about 2 feet, over the entire three-year period. Measurements are available for only two years at points 3 and J, but during that time their movements were ~~on~~ of the same order of magnitude as point 4. The other extreme is point 16, which had 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 stable and were used as controls for orienting the annual traverses. The location of point 15, below the slide, has been within the error of measurement during the three-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 control points showed some interesting contrasts between the movement of the earthflows and the slump. All of 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 exhibited by 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, located near the earthflow's edge, was at the side of the main "current" of the earthflow. Its movement, consequently, was much smaller than for 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 an angle of about $45^{\circ}$ to the other points on 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 a given set of 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 from 1953 to 1954 than did point 17 a short distance below it. Point 14 moved essentially 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 more rapidly than the slump.

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.

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## Paulson slump

The Paulson slump (figs. 52 and 53) is an old landslide that has

Figure 52.--Near here.

Figure 53.--Near here.

been reactivated by the development of two subsidiary landslides on its lower slopes. It is about 500 feet wide by 500 feet long 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 5 feet thick in this locality and is, in turn, 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 essentially outline the top and sides of the original slide. Other fractures were becoming common on the slump itself, especially between the scarps of the subsidiary slides.

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Figure 52.--Paulson slump. Topography planetabled in 1956 by C. F. Erskine and J. A. Sharps. Geology mapped in 1956 by C. F. Erskine and J. A. Sharps.



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Figure 53.--The Paulson slump. A reactivated slump in Pierre shale. The two active landslides are in the toe of the main slump, and the slump includes the area between the two active slides and the rather flat surface just above them. Photographed August 19, 1956.

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The subsidiary landslides are slump-earthflows although the earthflow portions were not very well developed on either slide. The slide on the southwest side of the old slump was about 225 feet long and 175 feet wide. Although the one on the northeast side was about 275 feet long and 200 feet wide, stream erosion had cut away its northeast corner until its total area was not much greater than that of the southwest 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.

in the slide

The causes of original movement cannot be determined today. The low scarp remnants at the head of the block and the relatively undisturbed shale exposed in the newer scarps indicate that the movement was small and that the landslide was predominantly a slump. Following this movement the slump became stable. Erosion subsequently modified it until the only remaining evidences of the landslide were the small scarp at the head and the flattened slope on the surface of the rotated slump block.

Aerial photographs show that by 1949 the subsidiary slides were almost as well developed as when they were first examined in 1952. At that time the subsidiary slides were active but the main slump was still stable. Renewed movement of the main Paulson slump started between 1952 and 1953, and continued to increase as long as the slide was under observation.

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Stream erosion probably was the most important cause of the subsidiary landslide activity. The stream bed, although normally dry, can contain a swiftly flowing stream after heavy rains or during wet seasons. Shortly after a cloudburst in the area the author observed a boulder more than 1 foot in maximum dimension that had been lodged against a tree near the stream channel. A stream that could transport boulders of this size certainly could perform major erosion in the Pierre shale. Although ground water may have made some contribution to the renewed instability of the old slump, there was never evidence of appreciable amounts of ground water in the landslide during ~~the~~ this study.

The northeast subsidiary slide was definitely activated by stream erosion. The southwest slide was probably also due to erosion although it is farther from the stream channel. 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.

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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 on figure 52 because they were only 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 feet and consequently all movement vectors are assumed accurate to at least 0.14 feet.

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From 1952 to 1956 the general trend 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 no more than 0.5 feet. Movement of all points on the slump between August 1953 ~~to~~ and October 1954 indicated that the entire slump was now active. The magnitudes of movement ranged from about 0.3 feet at points B-2, B-3, and B-4 to about 0.8 feet 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 feet, which is within the possible limits of error, and point $B_1$-5 moved about 0.5 feet, 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 of the points on the northeast half of the slide were destroyed, either by movement or by cattle, but on the southwest half the movements ranged from about 0.6 feet 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 potentiality was developing in the Paulson area between 1952 and 1956. Renewed activity of the old slump started 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 implied that the mass was losing its unity and that 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.

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Correlation between surplus precipitation available for ground-water storage, landslide activity, and ground water

The final section attempts to correlate surplus precipitation available for ground-water storage, landslide activity, and ground water. Emphasis throughout the report continually has been on the relationship between ground water and landslide activity, but little has been said about the possibility of correlations between (1) percentage of precipitation available for storage as ground water, (2) the ground-water conditions, and (3) slope stability. Such a correlation can be valuable as a basis for prediction and interpretation of slope stability conditions.


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Available precipitation for ground water

The amount of moisture that is potentially available to replenish and increase ground-water supplies is not easily determined. Precipitation can be recorded at a given area, 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 to 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.

Since runoff represents such a small part of the total precipitation, only the relationship between evapotranspiration and precipitation was used to determine approximately the abundance or scarcity of water 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 ~~yet~~ because there is no direct way to satisfactorily measure the moisture that plants release to the atmosphere.



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Thornthwaite (1948) has worked with basic climatic data in an attempt 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," which is the amount of evapotranspiration that will occur under given climatic conditions if there is an unlimited source of water. Thornthwaite 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 invaluable because it provides a comparative method of estimating the rate at which moisture can return to the atmosphere from the ground.

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## Relationship 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 relationship. Variations along the reservoir should be fairly small, however, and Pickstown is assumed typical of the entire area.

The potential evapotranspiration values computed by Thornthwaite's method and the precipitation record from the beginning of climatic records at Pickstown in 1948 through 1956 are graphed at the bottom of figure 54. Periods and amounts of precipitation surplus and

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Figure 54.--Near here.

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deficiency are shown by patterns; total annual values and surpluses or deficiencies are given below the graphs. Averaged annual curves for the entire period of record are at the left side.


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Figure 54.--Correlation between precipitation and potential evapotranspiration.

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The averaged curves 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 essentially nothing during the winter months to more than 6 inches during July. The precipitation curve is bracketed by the evapotranspiration curve; evapotranspiration is less than precipitation during the winter and greater during the summer. Precipitation in excess of the needs for evapotranspiration can be expected during the winter rather than during the summer.

The annual precipitation and potential evapotranspiration curves have a general resemblance to the 1948-1956 curves. Precipitation has a range of annual values from 12.5 to 29.9 inches; the range for potential evapotranspiration, on the other hand, is from 24.1 to 30.3 inches. Similarly, the annual evapotranspiration curves vary less from their average than the precipitation curves. Occasional irregularities point up disparities between individual months, but the variations from the 1948-1956 curve for any given 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, the records for individual months often reverse this pattern.


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The annual precipitation-potential evapotranspiration relationships are similar, despite fluctuations from the average. Every winter and spring the precipitation is greater than the amount used for evapotranspiration.

The precipitation deficiency of summer, , almost always exceeds the precipitation surplus of winter. Only one year, 1951, had a precipitation surplus. During the year there was a surplus of 5.7 inches, and just one month, July, had an appreciable deficiency. Ground-water intake probably was continuous, therefore, for most of that year.

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Twelve months' cumulative precipitation surplus

Ground water can trigger movement, or prepare a potential landslide mass for the trigger action. The triggering effect is rapid and should occur during or very soon after a period of excess precipitation. It is most important, therefore, during individual periods of excess precipitation. Ground water as a general cause requires an influx of ground water over an extended period of excess precipitation whereby it involves a slow buildup of stresses and other factors that affect the stability of a given mass.

A 12 months' cumulative precipitation surplus curve (fig. 54 top) was constructed from monthly precipitation and potential evapotranspiration data in order to point up 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 up to and including the month for which the point is plotted. The 12-month period was chosen because (1) it seemed to be a reasonable length of time to permit ground water to permeate the shale, and (2) cumulative values for 12 months automatically cancel out the effects of normal seasonal variations in precipitation surplus.



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The curve shows that a cumulative precipitation surplus for any 12-month period is rare. Over the 8 years for which records are available only 1 year had a surplus. From June 1951 to May 1952 the cumulative precipitation surplus was about 13 inches higher than the average values for the rest of the time from December 1948 to December 1956. The monthly surplus during that year was about 5 inches, in contrast with an average monthly deficiency of almost 8 inches during the rest of the period. June 1951 to May 1952 was, therefore, the only period in which enough precipitation was available to appreciably affect ground-water supplies.

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Correlation between available precipitation and slope stability

The limited data imply a definite relationship between available precipitation and slope stability. Although this correlation cannot yet be accepted as fact, it is at least a valuable working hypothesis. The following portion of this report first presents this hypothesis theoretically and then offers evidence that substantiates it.



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## Theoretical hypothesis

The theoretical hypothesis is based on the information in figure 54 and the assumption that most landslides are the result of a general cause of instability supplemented by a trigger action that initiates movement. There may actually be several causes of instability, but in this section only the apparent effects of ground water will be considered in detail.

Temporary increases in the available ground water may raise the water table and may also cause partial saturation of the material above the water table. If such an influx continues for a period of many months, the resulting changes in ground-water conditions may be sufficient to trigger landslides on some slopes. On most slopes, however, it will act as a general cause of instability. Following this line of reasoning, abnormally high peaks in the 12 months' cumulative precipitation curve (for example, from June 1951 to May 1952, fig. 54) may be expected to coincide with an appreciable increase in the amount of ground water, and a corresponding decrease in general slope stability.



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Periodically both general ground-water conditions, as evidenced by the 12 months' cumulative precipitation surplus curve, and temporary ground-water conditions, as shown by precipitation-potential evapotranspiration relationships, should be favorable to landslide activity and exceptional amounts of activity should occur on both new and old landslides. Only one such period has occurred during the time that climatic records have been kept at Pickstown. In the spring of 1952 the 12 months' cumulative precipitation surplus curve was at a peak and the precipitation and potential evapotranspiration curves showed a precipitation surplus.



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## Substantiating evidence

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

The slides described in the preceding section on landslide movements (p. 255) either originated or were active in 1956. The Highway 16 slump and the two Cable School earthflows all originated in 1952 and only minor subsequent movements had 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 is no way to compare the relative amount of movement that year with later movements. The Paulson slump alone has shown increasing activity after 1952. It also is the only one of the landslides in which the role of ground water apparently has been secondary to another cause--in this case, stream erosion at the toe.

Following the drop in the 12 months' cumulative precipitation surplus curve starting in May 1952 (fig. 54), there has been a gradual increase in overall slope stability. The ~~behavior of~~ control-points ~~movements~~ on all the landslides except the Paulson slump have shown either negligible movements or ~~else~~ a general decrease in rate of movement.

At Pierre, S. Dak., about 60 miles northwest of Chamberlain, Crandell noted a concentration of landslide activity in the spring of 1952. He states (~~in press~~ 1958, p. ~~336~~ 77): "Fewer than 10 large new landslides were observed in the Pierre area during the period 1948-51 ~~1951~~; 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 great amount of ground water.

Precipitation, potential evapotranspiration, and 12 months' cumulative precipitation curves were prepared for the Pierre area in order 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 months' 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 at Pierre between observed conditions and theory seems to further validate the hypothesis that precipitation surpluses can be correlated with periods of excessive landslide activity.

## Correlation between available precipitation and ground-water levels

A comparison was made between the available precipitation and ground-water levels in the piezometers in order to check the presumed relationship between precipitation surplus and ground-water conditions. An essentially qualitative analysis was made for those piezometers in the vicinity of 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.

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The results of the analysis were plotted graphically with percentage values as ordinates and number of months' cumulative precipitation surplus as abscissae (fig. 55). Graphs of individual

Figure 55.--Near here.

piezometers are divided into three groups on the basis of depth: (a) 12-30 feet, (b) 55-59 feet, and (c) 81-88 feet. Percentage values on the curves represent the degree of correlation between 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.

The curves show a relatively short-term relationship between precipitation surplus and ground-water levels in the shallower piezometers (fig. 55, a, b), but effects are essentially random in the deepest ones (fig. 55, c). Percentage values of the shallowest piezometers are above 70 for the 2-, 3-, and 4-month periods of cumulative precipitation surplus in every case except piezometer A-30. For the 55- to 79-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 at depths up to at least 30 feet. These effects decrease with depth, becoming minor at about 60 feet and ineffective at appreciably greater depths.

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Figure 55.--Number of months cumulative precipitation surplus. Correlation between piezometer water levels and 1 to 12 months' cumulative precipitation surpluses.

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The correlation between near-surface ground-water conditions and the short-term cumulative precipitation surplus values coincides with the assumption that ground water triggers landslides. The annual winter and spring precipitation surpluses (fig. 54) give maximum values for 2 to 4 months' cumulative precipitation surpluses in the spring, the season when landslide activity is greatest.


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no ¶ The long-term cumulative precipitation surplus values (fig. 55), in particular the 12 months' surpluses, show no correlation with the piezometer water levels. Two factors may contribute to this apparent lack of correlation. The first is that the 12 months' cumulative precipitation values were considered as an indication of the 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. Only those piezometers in which water levels had developed 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 the conditions in massive shale. The second factor is the limited period for which comparisons were made since ground-water the data used extended only from February 1954 through November 1956. Although the correlation between landslide activity and 12 months' cumulative precipitation surplus was pronounced after only one period of excess precipitation, a relationship between ground-water levels and the cumulative precipitation surplus might not be apparent until at least two times of abnormally high rainfall are observed.



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## Conclusion

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

The best correlation seems to be between winter and spring precipitation surpluses and spring landslide activity triggered by ground water. The 2 to 4 months' cumulative precipitation surpluses reach maxima in the spring after several months of surplus precipitation during the winter and early spring. Near-surface (less than about 60 feet deep) ground-water levels roughly reflect the cumulative precipitation surplus values, and they, also, tend to be above average in the spring. Most landslide activity, moreover, occurs in the spring. It seems reasonable, therefore, to postulate that after 2 to 4 months of surplus precipitation the surplus water that infiltrates the ground, together with the pre-existing ground water, can create conditions sufficiently unstable to promote landsliding.

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

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followed by 308a

## Appendix I

### Shale sample locations

Site 1. SE¼ SW¼ sec. 13, T. 95 N., R. 66 W., Gregory County.

Site 2. SW¼ SE¼ sec. 13, T. 95 N., R. 66 W., Gregory County.

Site 3. SW¼ SW¼ sec. 21, T. 104 N., R. 72 W., Lyman County.

Site 4. SW¼ SE¼ sec. 22, T. 104 N., R. 72 W., Lyman County.

Site 5. NW¼ SE¼ sec. 22, T. 104 N., R. 72 W., Lyman County.

Site 6. SE¼ SE¼ sec. 22, T. 104 N., R. 72 W., Lyman County.

Site 7. SW¼ SE¼ sec. 14, T. 104 N., R. 72 W., Lyman County.

Appendix II

Piezometer locations

Site A. NE¼ NE¼ SE¼ sec. 20, T. 95 N., R. 65 W., Gregory County, South Dakota; elevation, 1,360 feet.

Site B. SE¼ NE¼ NW¼ sec. 28, T. 97 N., R. 68 W., Gregory County, South Dakota; elevation, 1,444 feet.

Site C. SE¼ SE¼ SE¼ sec. 28, T. 96 N., R. 66 W., Charles Mix County, South Dakota; elevation, 1,463 feet.

Site D. SW¼ SW¼ SW¼ sec. 31, T. 95 N., R. 64 W., Charles Mix County, South Dakota; elevation, 1,498 feet.

Site 1. About 600 feet southwest of water plant intake building, Chamberlain, South Dakota; elevation, 1,381 feet.

Site 2. About 450 feet northeast of intersection of 11th Avenue and Courtland Street, Chamberlain, South Dakota; elevation, 1,390 feet.

Site 3. SE¼ SW¼ SW¼ sec. 11, T. 105 N., R. 71 W., Brule Co., South Dakota; elevation, 1,414 feet.

Site 4. SE¼ NE¼ NW¼ sec. 18, T. 103 N., R. 71 W., Brule County, South Dakota; elevation, 1,450 feet.

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