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GEOLOGICAL AND ENGINEERING ASPECTS OF UPPER CRETACEOUS SHALES IN WESTERN CANADA

(Report, 5 plates and 26 figures)

J. S. Scott and E. W. Brooker

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ABSTRACT

Overconsolidated shale is herein defined as a sedimentary deposit composed primarily of silt-and clay-sized particles dominated by members of the montmorillonite group of clay minerals and the deposit has been subjected to consolidation loads in excess of those provided by the present overburden. From an engineering standpoint these materials constitute one of the major problem soil types. Predictions of the engineering performance of overconsolidated clay shale either on a short or long term basis cannot as yet be made with reasonable confidence in spite of proven statical theory and modern laboratory testing techniques.

Overconsolidated clay shales of Upper Cretaceous age occur extensively in Western Canada where they cause stability problems for foundations and slopes. In order to assess the geological factors affecting the engineering behaviour of overconsolidated clay shales the Bearpaw Formation was selected as the primary unit for study. Slope failures involving the Bearpaw Formation or its stratigraphic equivalents were examined at various areas in Manitoba, Saskatchewan, and Alberta, from which it is concluded that the prime geological factors affecting the engineering behaviour of overconsolidated shales are: 1) depositional environment; 2) lithology and stratigraphy; 3) stress history; 4) structure; 5) climate; 6) geomorphology; and 7) groundwater.

The geometry of the failure surface of the retrogressive slides that are common in the overconsolidated clay shales of Western Canada is best described by a number of plane surfaces rather than by a circular arc. Therefore the wedge method of stability analysis appears to offer a promising approach to slope stability prediction for these materials.

Peak (ϕ') and residual (ϕ'_R) strengths of overconsolidated shales obtained from drained direct shear tests are usefully applied in an upper and lower limit approach to the problem of stability analysis.

The important unknown in stability analyses for prediction purposes is the value of the Residual Strength Index:

$$R = \frac{\phi' - \phi'_A}{\phi' - \phi'_R}$$

The value of ϕ'_A, the field angle of shearing resistance, is indeterminate but can be assessed through the analysis of well instrumented failed slopes.

GEOLOGICAL AND ENGINEERING ASPECTS OF UPPER CRETACEOUS SHALES IN WESTERN CANADA

INTRODUCTION

The analysis and prediction of earth mass stability has been of interest for many years. As early as 1776 Coulomb expressed in mathematical form his concept of the shearing resistance of soils. At this time the means of assessing the parameters involved did not exist, nor was the nature of the stresses involved understood. In 1846 Collin's study of the geometry of slope failures combined with an uncomplicated strength test established a rational method for assessing stability. This latter work was not immediately appreciated, and was given little subsequent attention. Increasing complexities of construction at the turn of the century resulted in the development of the circular arc method of analysis and the use of an undrained shear strength by Fellenius et al., in Sweden (Bjerrum, 1960).

Terzaghi (1936) defined three major clay soil mass types, namely: soft or normally consolidated; stiff intact; and stiff fissured. The last two usually occur as preconsolidated masses. Accumulating experience has indicated that the methods developed in Sweden for calculating the stability of normally consolidated clay masses are not applicable to overconsolidated soils. The shortcomings of the usual 'undrained' strength analysis is demonstrated by Figure 1 in Peck (1960). This diagram shows the Factor of Safety calculated by the undrained or total stress method, for slopes which have in fact failed, as a function of the liquidity index.

$$I_L = \frac{w - w_p}{w_l - w_p}$$

where w = natural moisture content
w_p = plastic limit
w_l = liquid limit

In normally consolidated soils the liquidity index is near and usually greater than 0.5, whereas for preconsolidated soils the liquidity index approaches zero. Highly overconsolidated soils often display liquidity indices that are negative. Thus, the diagram demonstrates that the reliability of the undrained method rapidly diminishes as the soil becomes highly overconsolidated.

Because the clays that have a low or negative value of I_L are usually fractured or fissured as well, the determination of soil strength is often not reliable. Hence, the uncertainties in assessing stability of clay shale slopes presents an ominous problem.

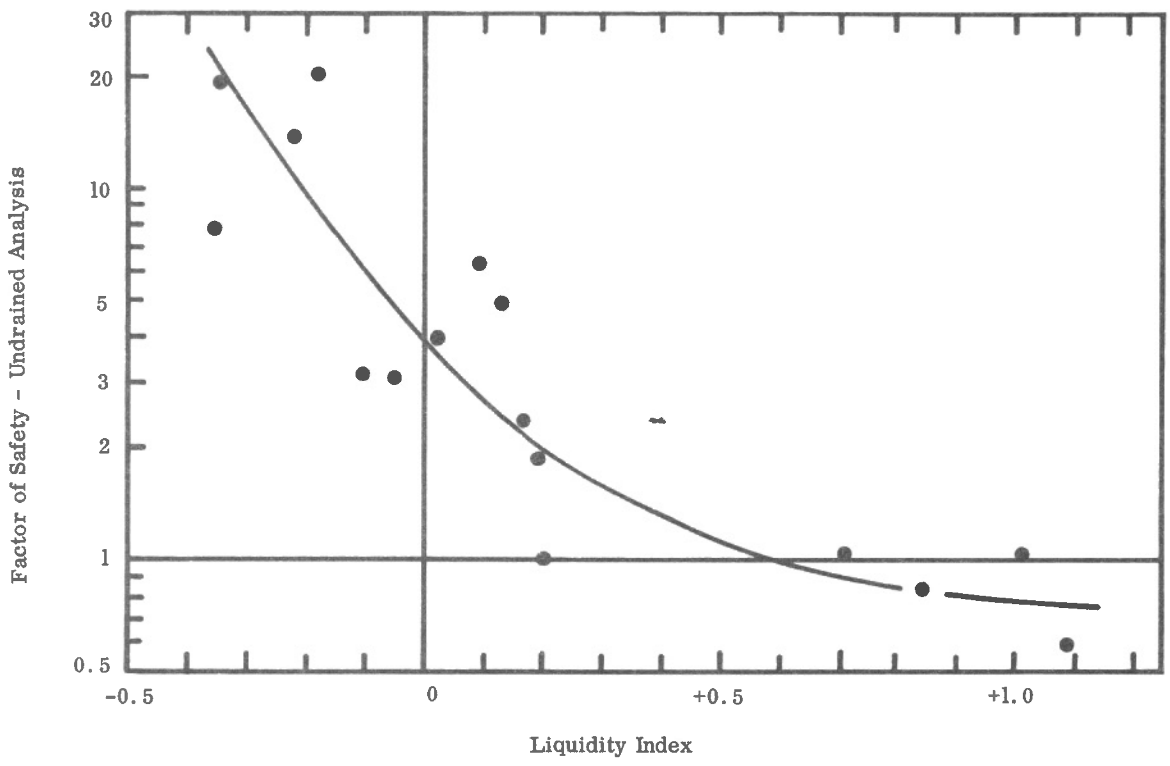


Figure 1. Factor of safety as a function of liquidity index for undrained analyses (Peck, 1960).

Since 1936 the development of soil mechanics has been rapid. The nature and fundamental aspects of shearing resistance of soils were developed under Terzaghi's guidance by Rendulic (1936) and Hvorslev (1960). Subsequently, sophisticated means of defining what appeared to be the necessary strength parameters applicable in refined analysis of soil masses were developed by Skempton and Bishop (1954), and Bishop and Henkel (1957). Accumulating laboratory and field evidence substantiated the concepts of short term and long term stability for many soil types. The initiation of field devices to document performance, govern construction activity, and thus relate actual to predicted performance, stimulated a confidence in the methods of analysis and the results of tests for physical constants of the soil.

Yet, through this rapid and fascinating advancement two dominant types of problem soil masses emerged. These are the sensitive quick marine clays, and the overconsolidated stiff fissured as well as the stiff intact clays. The problems associated with these clays have been the cause of considerable research. Rosenquist (1960) and Bjerrum (1954) have shown the influence of physico-chemical environment on the behaviour of the Norwegian Quick Clays. Although their explanation has not been directly applicable to the Leda Clays of Canada, Penner (1963, 1965) has, again on the basis of physico-chemical research, been able to define fundamental reasons for the observed performance of Leda Clay.

The fundamental actions involved in the performance of highly overconsolidated stiff fissured and stiff intact clays have, however, remained largely unresolved. In spite of the proven validity of theories based on statics combined with physical soil properties as obtained by modern laboratory testing techniques we have not yet been able to predict with reasonable confidence the performance, on either a short term or long term basis, of highly overconsolidated clays or, more particularly, clay shales (Peterson,1960; Hardy, 1957). The records all suggest that a revised approach to the diagnosis and treatment of clay shales may be in order. The authors believe that this is best achieved by the combined resources of the geologist and the engineer. This report is the product of such a combination of resources: geologist, J. S. Scott of the Engineering Geology Section, Geological Survey of Canada, and engineer E. W. Brooker, Associate Professor of Civil Engineering, University of Alberta.

Definition of 'Overconsolidated Clay Shales'

From a geological standpoint shales are the most abundant of sedimentary rocks. Estimates of the relative proportion of shale to other common sediments range from 70 to 82 per cent (Pettijohn, 1949). Because of their fine-grained texture and large variations in mineralogical composition and physical properties, however, they are difficult to study and, as yet, no generally acceptable scheme of classification for these sediments has been formulated. In the absence of a definitive classification the use of the term 'overconsolidated clay shale' in this paper requires some definition, as various interpretations may be applied to the several components of the term.

In essence the term 'overconsolidated clay shales' as used herein refers to a sedimentary deposit composed primarily of silt- and clay-sized particles dominated by members of the montmorillonite group of clay minerals, and these deposits have been subjected to consolidation loads in excess of those provided by the present overburden. The interparticle bonds of these materials are such that disaggregation of the sediment can be effected by immersion in water.

Selection of Field Study Areas

Large areas of Western Canada are underlain by formations of Upper Cretaceous age. Several of these formations comprise overconsolidated clay shales. Engineering problems resulting from the presence of these materials are, therefore, neither stratigraphically nor geographically restricted but are of widespread occurrence throughout the area.

In order to assess the geological factors affecting the engineering behaviour of overconsolidated clay shales in general, however, it was decided to select for study a stratigraphic unit having widespread occurrence and with which engineering problems had been encountered. The selection of such a stratigraphic unit would reduce the number of variables to be considered in the study and could provide the basic data required for the assessment of the engineering behaviour of overconsolidated clay shales. For these reasons the Bearpaw Formation was selected as the primary unit for study.

Although the Bearpaw Formation per se is not recognized in eastern Saskatchewan and Manitoba, stratigraphic equivalents of the formation are present. These stratigraphic equivalents, the Marine Shale Series in eastern Saskatchewan and the upper part of the Riding Mountain Formation in Manitoba also consist mainly of overconsolidated clay shales. Strata of the Riding Mountain Formation are better exposed than those of the Marine Shale Series, thus for a comparison of the engineering behaviour of these formations with the Bearpaw Formation outcrops of the Riding Mountain Formation only were examined.

Outcrops of the Bearpaw Formation and its stratigraphic equivalents occur mainly in river valleys where the overlying sediments have been removed by stream erosion. Landslides involving the overconsolidated shales are also prevalent along the valley walls, thus suitable sites for field investigation were readily determined from examination of aerial photographs in conjunction with published geological maps.

A preliminary field examination was made at various localities in Manitoba, Saskatchewan, and Alberta, at which both stable and failed slopes in the Bearpaw Formation or its stratigraphic equivalents were present (Fig. 2). Samples of the formation, from which regional variations in lithology and index properties could be determined, were obtained from most of the localities. The preliminary field examination also provided a basis for establishing the significant geological parameters affecting slope stability and for the selection of specific areas in which more detailed studies should be made.

Sites along South Saskatchewan River in central Saskatchewan, along the St. Mary River in southern Alberta and along Red Deer River in central Alberta were selected for more detailed examination as a result of the preliminary field work. These sites were examined jointly by the authors in order to derive the benefits of mutual observation relating to the mechanism of slope failures in these areas.

Physiography and Climate of the Study Area

The part of Western Canada underlain by the Bearpaw Formation and its stratigraphic equivalents is physiographically situated within the Saskatchewan Plains and Alberta High Plains Regions of the Interior Plains Province (Bostock, 1964).

Throughout the Saskatchewan Plains, which extend west from the Manitoba escarpment to the Missouri Coteau, surface elevations range from 1,500 to 2,150 feet. To the west of the Missouri Coteau much of the Alberta Plains Region is about 2,500 feet in elevation. Scattered ranges of hills or uplands rising from several hundred to over 1,000 feet above the surrounding terrain, occur in both of the Plains Regions. Many of these uplands are underlain by strata younger than the Bearpaw Formation.

The major drainage system of the southern Plains Region is that of the South Saskatchewan River, which flows east and northeast across the region before reaching its outlet in Lake Winnipeg at an elevation of about 715 feet a.s.l. Tributaries such as the St. Mary, Oldman, and Bow Rivers rise in the foothills and join the main stream of the South Saskatchewan River in southern Alberta. The Red Deer River, which also rises in the foothills, has its confluence with the South Saskatchewan just east of the Alberta-Saskatchewan boundary.

Throughout most of the Plains Region the South Saskatchewan River has an average gradient of 1.9 feet per mile and flows generally in a broad valley entrenched from 200 to 300 feet below the upland surface. North of Medicine Hat, however, the river flows through a narrow, steep-walled gorge that is from 450 to 500 feet deep.

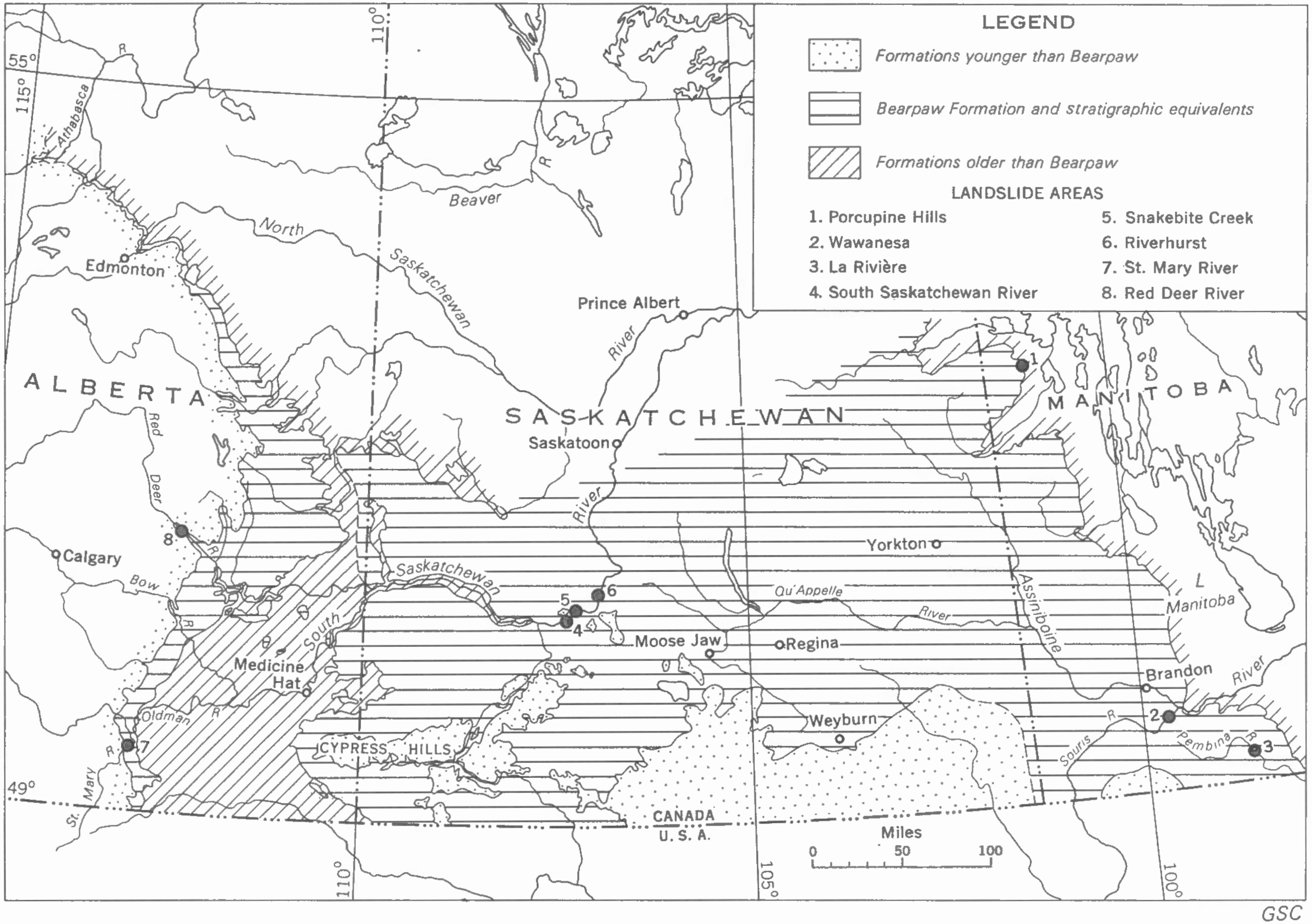


Figure 2. Generalized geology and locations of landslide areas.

Discharge of the South Saskatchewan River is characterized by two periods of maximum flow. The first occurs generally in mid-April, occasioned by run-off of local snowmelt throughout the drainage basin, and the second occurs late in June, as the result of meltwater flow derived from the tributary headwater regions of the foothills and mountains. Lateral erosion of the banks along the South Saskatchewan River is particularly active during periods of maximum flow.

The eastern part of the Plains Region is drained by the Assiniboine River and its main tributaries, Qu'Appelle and Souris Rivers. These rivers are, on the average, 200 feet below the level of the plain and flow as an underfit stream in a former glacial lake spillway channel, which has a broad level bottom and precipitous sides.

The total mean annual precipitation on the southern Plains Region ranges from 12 to 18 inches. In most areas, however, wide variations in precipitation occur from year to year and the differences in extreme annual amounts commonly exceed the mean annual total.

The area of low precipitation is centred about the junction of the Red Deer and South Saskatchewan Rivers. Fairly uniform increases in mean annual precipitation occur to the east and west of this area. On a seasonal basis, the highest amounts of precipitation occur during the period from June to August.

Infiltration of precipitation is dependent upon the complex interrelationship of such factors as soil porosity, permeability, and moisture content in conjunction with air temperature and the chemistry and viscosity of water. The amount and rate of infiltration of precipitation as it affects overconsolidated shales throughout the Plains Region is essentially unknown; however, the amount is probably a very small percentage of the total precipitation. As a consequence pore pressure increases in overconsolidated shales resulting from precipitation infiltration may well be of small magnitude from a mechanical standpoint. Addition of even small amounts of water to overconsolidated shales by infiltration, however, may lead to the generation of high swelling and/or osmotic pressures with consequent decreases in shear resistance of the soil mass. The seasonal distribution of precipitation is such that, for the valley of the South Saskatchewan River, a combination of excessive infiltration and active toe erosion could occur simultaneously.

Normally warm summers and long, very cold winters are also a climatic characteristic of the Plains Region. Mean monthly temperatures for January range from 10°F in the southwest to -5°F across the northern part of the area. In July mean monthly temperatures of about 65°F prevail over most of the area. Wide departures from the mean temperatures occur during all months of the year, but are particularly prevalent during the winter months. The range in extreme temperatures recorded throughout the area is from -50°F to 110°F.

From November to March mean temperatures are below 32°F. Frost penetration occurs to a depth of approximately 8 to 10 feet, with maximum penetration being reached either in late March or early April.

The extent of frost penetration into fractures in clay shale slopes, caused either by desiccation or stress relief, is unknown. It seems likely, however, that frost action in fractures is a factor contributing to the instability of slopes in overconsolidated clay shales.

GEOLOGY OF THE BEARPAW FORMATION

Distribution and Thickness

The name Bearpaw Formation was assigned by Hatcher and Stanton (1903) to a sequence of marine shales that occur in the Bearpaw Mountains of Montana. Through subsequent work by the same authors the formation was traced into Canada. The base of the formation is placed at the top of the uppermost coal-bearing and carbonaceous shale beds in the underlying Belly River Formation. The upper contact of the formation is transitional and passes vertically or laterally into sandstone beds of the overlying Edmonton Formation or its stratigraphic equivalent.

The Bearpaw is present throughout central and southern Alberta except where it has been removed by erosion from the axial portion of a bed-rock structural high known as the Sweetgrass arch. East of the Sweetgrass arch the formation occurs under the plains of central and southern Saskatchewan to a line that extends irregularly from Saskatoon to the southeast corner of Saskatchewan.

As with most sedimentary strata the thickness of the Bearpaw Formation varies from place to place. Approximately 50 miles northwest of Edmonton the formation is practically non-existent, having graded into the overlying beds of the Edmonton Formation, but in the Pembina oil field, west of Edmonton its thickness has increased to about 100 feet (Williams and Burk, 1964). Along the Red Deer River east of Drumheller, Alberta, the thickness of the formation is probably between 400 and 500 feet according to Allan and Sanderson (1945). In southern Alberta Russell (1932) measured 731 feet of the formation along St. Mary River south of Lethbridge. On the basis of bore-hole data obtained from the northwestern part of the east block of the Cypress Hills in southern Saskatchewan, Loranger and Gleddie (1953) reported the presence of 1,050 feet of the formation. Along the South Saskatchewan River valley, east of 109°W Long., a thickness of about 850 to 900 feet has been estimated for the Bearpaw Formation (Evans, 1961).

Age and Correlation

The Upper Cretaceous age of the formation has been established by a number of palaeontologists through examination of the faunal content of the sediments. Fragments of sanidine and biotite from a bentonite bed 65 feet above the base of the Bearpaw Formation near Lethbridge, Alberta, were used by Folinsbee et al. (1961) to obtain a date of 75 ± 4 million years for this material by the potassium-argon method. It is apparent from the correlation chart (Fig. 3), however, that the age range of the Bearpaw Formation is not

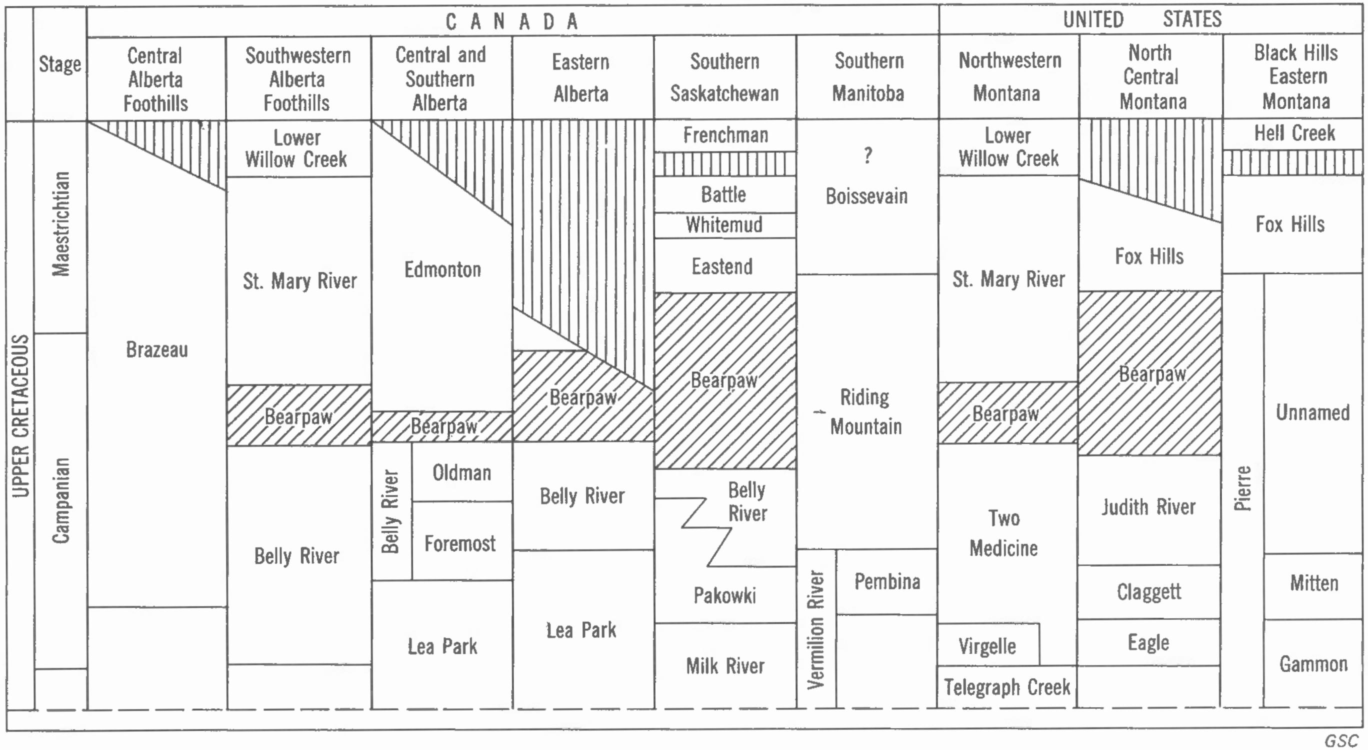


Figure 3. Correlation chart of some Upper Cretaceous Formations, Canadian plains and foothills and adjacent United States (after Williams and Burke, 1964).

everywhere the same. These age differences are the result of fluctuations with time in the extent of the marine environment in which most of the sediments of the formation were deposited.

Throughout eastern Saskatchewan the upper part of the Marine Shales Series, as named by Fraser et al. (1935), is the stratigraphic equivalent of the Bearpaw Formation. Farther east, in Manitoba, Bearpaw strata are probably correlative with the upper part of the Riding Mountain Formation as described by Wickenden (1945). Beds in the upper part of the Riding Mountain Formation are light grey, hard, siliceous shales of marine origin, which had been originally named the Odanah Series by Tyrrell (1890).

From the reference locality in Montana, shales of the Bearpaw Formation extend eastward into North and South Dakota and north-central Wyoming, where either the formation itself is recognized or it is represented by a stratigraphic equivalent within the Pierre Shale.

Depositional Environment

In the Plains Region Upper Cretaceous sediments were deposited in a broad shallow sea, which was flanked on the west by the highlands of the ancestral Cordillera and on the east by the Canadian Shield. At its maximum the sea extended from Mexico to the Arctic. Sediments of the Bearpaw Formation, however, were deposited during an epoch of fluctuating but slowly waning level of the sea. Withdrawal of the sea at the close of Bearpaw time is indicated by the transition of the marine sediments of the formation into the overlying non-marine deposits.

Most of the sediment that constitutes the Bearpaw Formation was derived from the western highlands. As a consequence the sandy phases of the formation are, in general, most prevalent close to the source area and the finer grained sediments, clay and silt, are more abundant at greater distances from the source area. These changes in sedimentary facies, however, are neither laterally nor vertically uniform as a result of the alternating transgressive and regressive phases of the sea produced by a continuously variable interrelationship between the rate of subsidence of the sedimentary basin and supply of sediment.

Volcanic activity, centred in the area that is now southwestern Montana, occurred at various times during the deposition of the Bearpaw sediments, resulting in the incorporation of layers of volcanic ash within the sedimentary sequence. Alteration of the volcanic ash in situ produced the layers of bentonite that are prevalent throughout the formation.

Several lines of evidence, such as bedding characteristics, form and structure of fossils contained in the sediments, and the presence of concretions, have led to the interpretation (Reeside, 1957) that Bearpaw sediments probably accumulated at a slow rate in relatively quiet waters. Radiogenic argon dates obtained by Folinsbee et al. (1961) have provided further evidence for a slow rate of sediment accumulation. Dates obtained from an apparently continuous sedimentary sequence in the Alberta and Peace River basins that spans the time of Bearpaw deposition indicate that sediment accumulated at the rate of 1 foot in 7,000 years.

Texture and Mineralogy

On the basis of textural analysis (Table I), sediments of the Bearpaw Formation and its stratigraphic equivalents may be classified into three main groups (Fig. 4). The clay shales form the dominant textural group, having an approximate textural composition of 5 per cent sand, 55 per cent silt, and 40 per cent clay, with substantial variations from these percentages being shown by any specific sample. The fine- to medium-grained sands that form the second textural group also contain up to 50 per cent of silt and clay sizes. Bentonites form the third textural group, which consists (as might be expected) mainly of clay- and silt-sized particles with minor amounts of sand.

Subdivision of the Bearpaw Formation into lithologic zones, based primarily on a separation between sand and clay shale members of the formation, has been made in several areas by various investigators. Although these subdivisions are useful for correlation within localized areas they have not proved satisfactory for correlation on a regional basis. Variations in lithology, however, affect the engineering behaviour of the formation and this aspect of the geology will be discussed further in the description of specific areas.

Clay-mineral composition is one of the important factors that affects the properties of soil materials, as has been demonstrated by Grim (1948) and others. In order to determine if any significant regional variations occur in the clay mineralogy of the Bearpaw Formation or its stratigraphic equivalents, samples were obtained for analyses by X-ray diffraction techniques. Results of the clay-mineral analyses, as presented in Figure 5 show that montmorillonite exceeds, with few exceptions, 55 per cent of the clay minerals present, followed by illite and chlorite. Most of the bentonite layers analyzed contain more than 95 per cent montmorillonite.

Identification of the exchangeable cation within the crystal lattice of the montmorillonite was not obtained from the X-ray diffraction analysis. Geochemical analyses of Bearpaw sediments from a section along St. Mary River (Fig. 18) show a marked increase in the presence of Na_2O in the bentonite layer at the base of the section, from which it may be inferred that sodium is present as an exchangeable cation in this material. Further analytical work would be required, however, to establish the precise chemical composition of the clay minerals.

TABLE I

Location and analysis of samples from Bearpaw Formation and stratigraphic equivalents

SAMPLE NO.	LOCATION						FORMATION	LITHOLOGY	MECHANICAL ANALYSIS(1)			CLAY MINERAL ANALYSIS(2)			PLASTICITY				ACTIVITY
	PROV.	N. LAT.	W. LONG.	SEC.	TP.	RGE.			% SAND	% SILT	% CLAY	% M	% I	% C	W	Wp	Wl	Ip	
SK-63/7	Man.	49°.04'.05"	98°.28'.10"	NE 28	1	8W1PM	Riding Mtn. (Odanah Beds)	Hard silicic shale	8	55	37	78	22	0		X[3]	X	X	
SK-63/8	Man.	49°.03'.45"	98°.28'.10"	SE 28	1	8W1PM	Riding Mtn. (Odanah Beds)	Hard silicic shale	11	54	35	87	13	0		X	X	X	
SK-63/13	Man.	49°.14'.30"	98°.90'.30"	SE 30	3	9W1PM	Riding Mtn. (Odanah Beds)	Hard silicic shale	0	55	45	83	17	0		X	X	X	
SK-63/15	Man.	49°.35'.15"	99°.40'.30"	NC 23	7	17W1PM	Riding Mtn. (Odanah Beds)	Hard silicic shale	3	49	48	83	17	0		X	X	X	
SK-63/17	Man.	49°.34'.20"	99°.42'.40"	NE 16	7	17W1PM	Riding Mtn.	Bentonite	0	10	90	95	5	0	70.2	71	180.4	109.4	1.21
SK-63/19	Man.	50°.44'.58"	101°.17'.15"	SW 27	20	29W1PM	Riding Mtn.	Arenaceous shale	32	51	17								
SK-63/20	Man.	50°.31'.30"	101°.17'.10"	SW 15	19	29W1PM	Riding Mtn.	Grey shale	8	44	48	73	19	8	30.1	27.0	80.0	53.0	1.10
SK-63/21	Man.	50°.31'.30"	101°.17'.06"	SW 15	19	29W1PM	Riding Mtn.	Silty shale	2	53	45	56	34	10	26.5	31.4	70.4	39.0	0.87
SK-63/24	Man.	50°.44'.10"	101°.17'.10"	S½ 3	19	29W1PM	Riding Mtn.	Arenaceous shale	9	57	34	75	25	0	27.7	35.9	80.3	44.4	1.31
SK-63/25	Man.	50°.44'.10"	101°.17'.00"	SE 3	19	29W1PM	Riding Mtn.	Silty shale	17	51	32	80	13	7	22.3	27.6	69.1	41.5	1.30
SK-63/39	Sask.	50°.40'.15"	107°.30'.45"	NW 6	20	11W3M	Bearpaw	Arenaceous shale	20	29	51	75	19	6	12.2	25.2	66.1	40.9	0.80
SK-63/39A	Sask.	50°.40'.15"	107°.30'.45"	NW 6	20	11W3M	Bearpaw	Grey silt	3	59	38				16.6	18.6	40.2	21.6	
SK-63/41	Sask.	50°.47'.30"	107°.15'.40"	EC 24	21	10W3M	Bearpaw	Silty shale	0	75	25	44	39	17	33.8	33.6	79.2	45.6	1.83
SK-63/42	Sask.	50°.39'.15"	107°.34'.35"	SE 33	19	12W3M	Bearpaw	Silty shale	7	54	39	73	20	7	35.5	34.4	87.2	52.8	1.35
SK-63/43A	Sask.	50°.38'.05"	107°.36'.30"	NE 27	19	12W3M	Bearpaw	Silty shale	18	57	25	70	28	2					
SK-63/43B	Sask.	50°.38'.07"	107°.36'.15"	NE 27	19	12W3M	Bearpaw	Grey Line sand	50	35	15								
SK-63/45	Sask.	50°.41'.15"	107°.19'.45"	NC 9	20	10W3M	Bearpaw	Silty shale	3	57	40	64	26	10	32.1	36.7	88.6	51.9	1.30
SK-63/96	Sask.	50°.40'.00"	107°.19'.45"	C 4	20	10W3M	Bearpaw	Silty shale	3	49	48	69	24	7	25.7	30.2	84.0	53.8	1.12
SK-63/46A	Sask.	50°.40'.00"	107°.19'.45"	C 4	20	10W3M	Bearpaw	Clayey Sand	72	16	12								
SK-63/47	Sask.	50°.40'.00"	107°.51'.00"	SE 3	20	14W3M	Bearpaw	Arenaceous shale	52	34	14								
SK-63/47A	Sask.	50°.40'.00"	107°.51'.00"	SE 3	20	14W3M	Bearpaw	Silty shale	14	52	34	83	17	0					
SK-63/49	Sask.	50°.41'.40"	107°.55'.00"	SE 18	20	14W3M	Bearpaw	Silty shale	3	60	37	57	29	14	17.8	22.0	49.1	27.1	0.73
SK-63/55A	Alta.	49°.33'.30"	112°.50'.40"	SW 18	7	21W4M	Bearpaw	Silty shale	0	57	43	51	32	17					
SK-63/55B	Alta.	49°.33'.30"	112°.50'.40"	SW 18	7	21W4M	Bearpaw	Silty shale	1	55	44	96	4	0	28.3	42.5	82.0	39.5	0.90
SK-63/55C	Alta.	99°.33'.30"	112°.50'.40"	SW 18	7	21W4M	Bearpaw	Silty shale	10	69	21	65	26	9					
SK-63/56A	Alta.	99°.34'.55"	112°.51'.05"	NE 24	7	22W4M	Bearpaw	Bentonite	1	20	79	100	0	0	54.3	54.7	139.0	84.3	1.07
SK-63/56B	Alta.	99°.34'.55"	112°.51'.05"	NE 24	7	22W4M	Bearpaw	Silty shale	1	63	36	54	38	8					
SK-63/57	Alta.	49°.29'.08"	112°.57'.06"	SW 19	6	22W4M	Bearpaw	Silty shale	1	68	31	96	4	0	8.16	28.6	70.2	41.6	1.34
SK-63/58A	Alta.	51°.21'.20"	112°.30'.30"	NW 32	27	18W4M	Bearpaw	Silty shale	4	58	38	100	0	0					
SK-63/58B	Alta.	51°.21'.20"	112°.30'.30"	NW 32	27	18W4M	Bearpaw	Silty shale	2	49	49	100	0	0					
SK-63/59	Alta.	51°.18'.40"	112°.24'.45"	NW 13	27	18W4M	Bearpaw	Silty shale	1	45	54	96	4	0	3.02	28.9	89.1	60.2	1.11
SK-63/61A	Alta.	51°.16'.05"	112°.19'.08"	NW 33	26	17W4M	Bearpaw	Silty shale	1	51	48	77	16	7	1.14	24.2	76.1	51.9	1.08
SK-63/61B	Alta.	51°.16'.05"	112°.19'.08"	NW 33	26	17W4M	Bearpaw	Silty shale	1	48	51	97	3	0					
SK-63/61C	Alta.	51°.16'.05"	112°.19'.08"	NW 33	26	17W4M	Bearpaw	Bentonitic shale	0.50	17.5	82	100	0	0		60.4	291.0	230.6	2.82
SK-63/61D	Alta.	51°.16'.05"	112°.19'.08"	NW 33	26	17W4M	Bearpaw	Bentonite	10	52	38	100	0	0	4.63	38.0	141.0	103.0	2.71
SK-63/61E	Alta.	51°.16'.05"	112°.19'.08"	NW 33	26	17W4M	Bearpaw	Silty shale	2	53	45	91	9	0					
SK-63/63A	Alta.	51°.07'.30"	112°.04'.15"	SW 17	25	15W4M	Bearpaw	Silty shale	3	53	44	96	4	0	3.46	22.7	99.3	76.6	1.74
SK-63/63B	Alta.	51°.07'.30"	112°.04'.15"	SW 17	25	15W4M	Bearpaw	Silty shale	4	36	60	71	18	11					
SK-63/67A	Alta.	51°.07'.08"	112°.06'.10"	NE 12	25	16W4M	Bearpaw	Silty shale	3	52	45	88	9	3					
SK-63/67B	Alta.	51°.07'.08"	112°.06'.10"	NE 12	25	16W4M	Bearpaw	Sandstone	57	21	22								
SK-63/67C	Alta.	51°.07'.08"	112°.06'.00"	NE 12	25	16W4M	Bearpaw	Carbon silt	8	62	30								
SK-63/67D	Alta.	51°.07'.08"	112°.06'.10"	NE 12	25	16W4M	Bearpaw	Silty shale	3	30	67	93	7	0		40.0	125.7	85.7	1.28
SK-63/68A	Alta.	51°.06'.08"	112°.06'.35"	C 1	25	16W4M	Bearpaw	Silty shale	2	65	33	72	19	9					
SK-63/68B	Alta.	51°.06'.08"	112°.06'.35"	C 1	25	16W4M	Bearpaw	Silty shale	4	60	36	56	30	14					

(1) Mechanical Analysis Sand 0.060 - 2.0 mm; Silt 0.002 - 0.060 mm; Clay <0.002 mm

(2) Clay Mineral Analysis M - Montmorillonite; I - Illite; C - Chlorite

(3) Plasticity X denotes samples that did not disaggregate upon agitation in water.

(4) Activity = $\frac{Ip}{\% < 2\mu}$

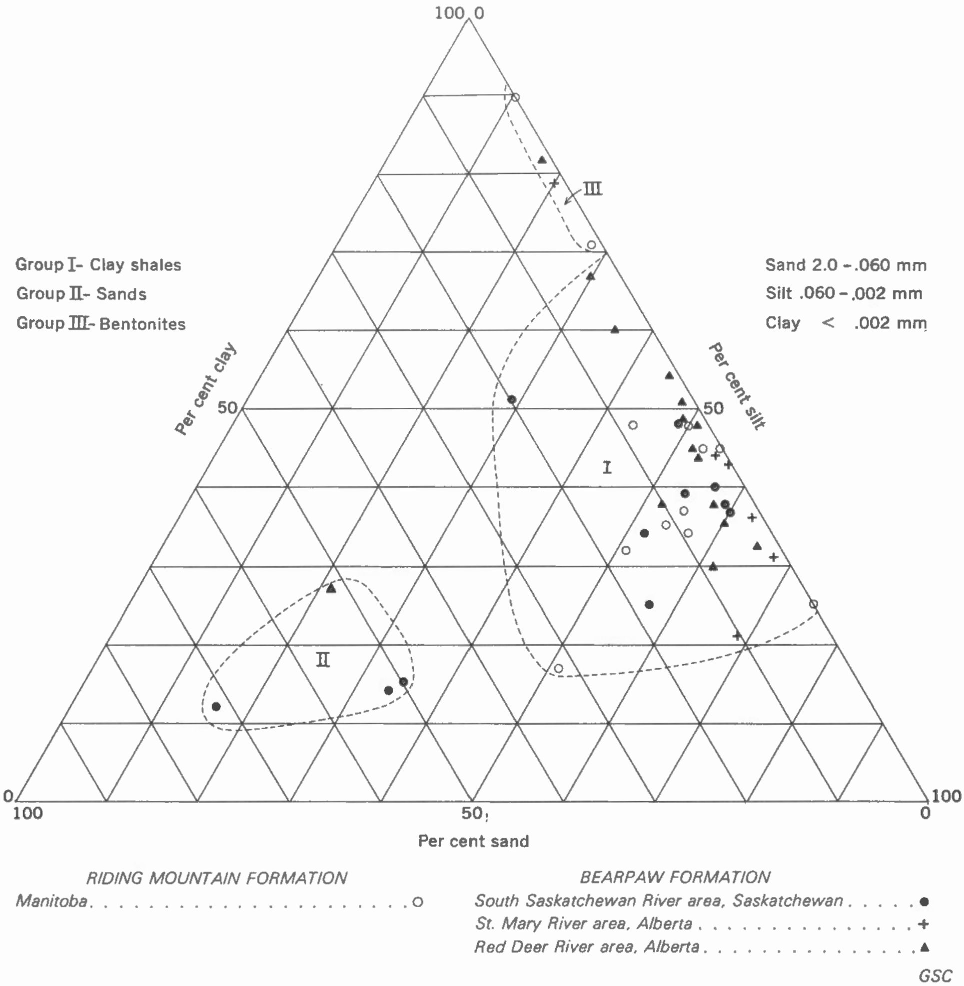


Figure 4. Textural analyses of Bearpaw sediments.

Although variations in the relative proportions of the three principal clay minerals (montmorillonite, illite, chlorite) occur throughout the Bearpaw Formation (Fig. 5) no discernible preferred pattern of regional distribution could be detected. Weaver (1958) has shown that clay minerals are predominantly detrital in origin and reflect primarily the character of the source area and that these minerals are not strongly modified by the depositional environment. It would appear, therefore, that the source areas of the Bearpaw sediments provided materials of relatively uniform mineralogical composition.

Although variations in grain-size distribution of the clay shales do occur the variations are not distributed in such a way that significant changes in the environment of deposition of the sediments can be detected.

From the mineralogical and textural evidence it is concluded that the clay shales were deposited in a relatively homogeneous geological environment.

Plasticity Characteristics

Plasticity characteristics of soils have been used for classification purposes and to distinguish soils that are different in origin and composition. The results of Atterberg limit determinations for a number of clay shale samples from the Bearpaw Formation and its stratigraphic equivalents are shown in Table I. Plasticity determinations were made on unweathered samples of the clay shale obtained from natural exposures. It is apparent from the field moisture contents (w), which are generally below the plastic limit (w_p), that the samples were obtained from a near surface zone of desiccation, as the water content of the shale is generally above the plastic limit for samples obtained from bore-holes as shown by Peterson (1958). Disaggregation of samples of hard, siliceous shale of the Riding Mountain Formation could not be effected by agitation in water, thus this material is essentially rock and could not be used for plasticity determinations.

The relationship between liquid limit (w_l) and plasticity index (I_p) are plotted in the form of a plasticity chart (Fig. 6), which classified the clay shales as mainly inorganic clays of medium to high plasticity. The chart was devised by Casagrande (1948) on the basis of analysis of plasticity characteristics of many soil samples of varying geological origin and mineralogical composition. It was found by Casagrande (1948) that soils having a common geological origin plotted along a line parallel with the empirically derived 'A' line. The plot of plasticity determinations for Bearpaw sediments is approximately parallel with the 'A' line, thereby indicating their common geological origin. The large range in liquid limit is due largely to the presence of montmorillonite in the sediments. Although no definite relationship between montmorillonite content and liquid limit can be demonstrated, Figure 7 shows a distinct trend toward an increase in liquid limit with an increase in the montmorillonite content of the sample.

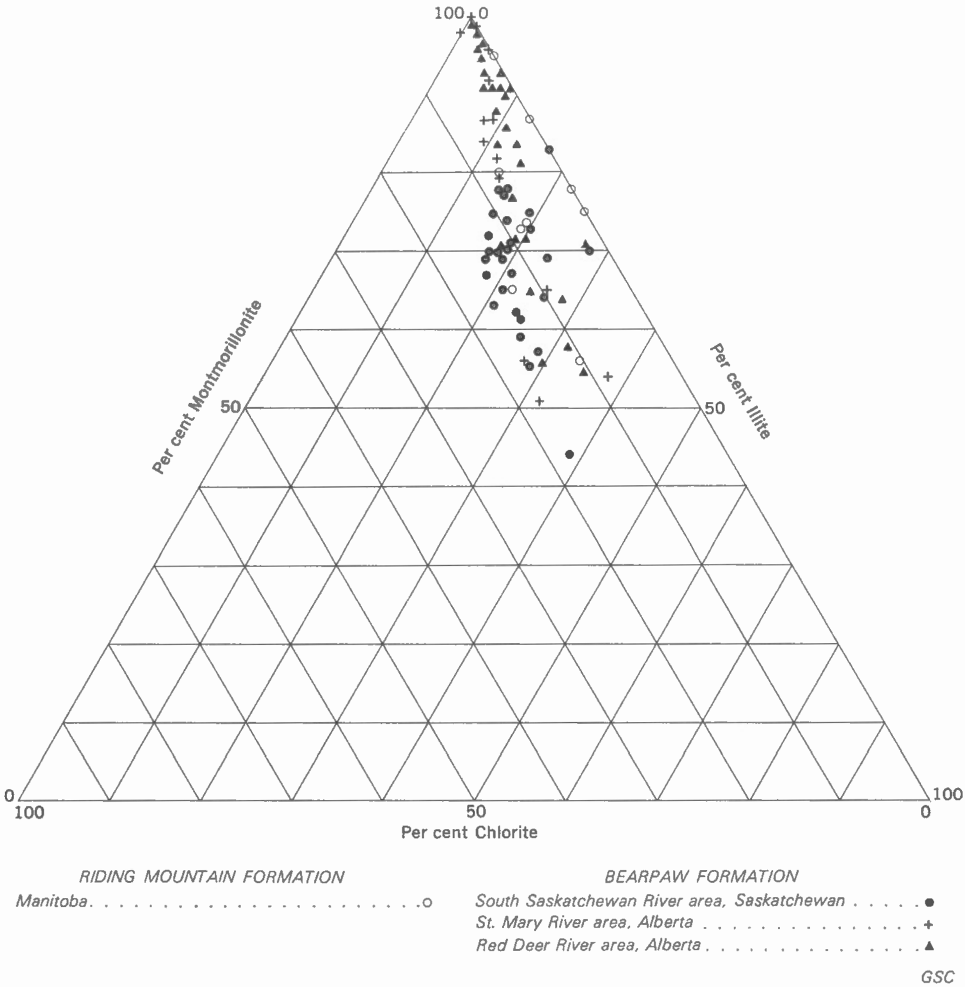


Figure 5. Clay mineral composition of Bearpaw sediments.

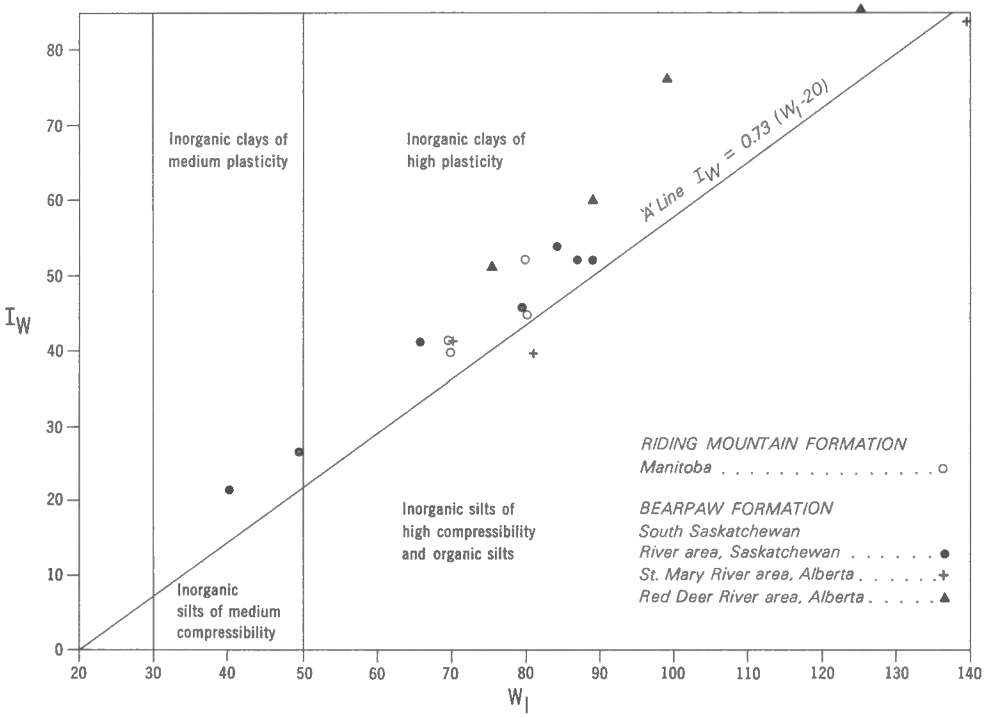


Figure 6. Plasticity chart.

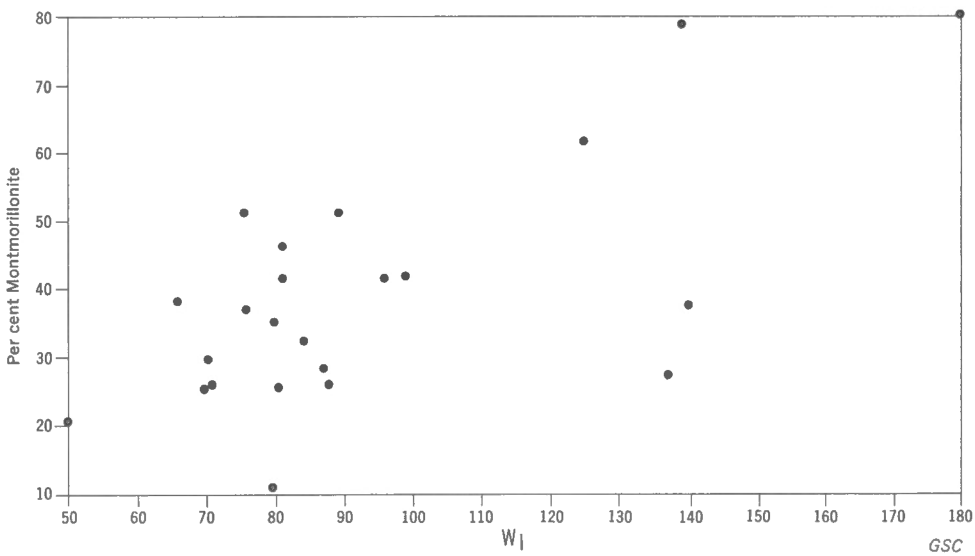


Figure 7. Relationship between liquid limit and montmorillonite content.

An index property known as 'Activity' in which the plasticity index (I_p) of a soil is divided by the per cent of clay fraction in the soil was proposed by Skempton (1953). In general the type of clay mineral present in the soil determines the Activity value. Skempton (op. cit.) found that if the plasticity index is plotted against the clay fraction for a given soil the points lie about a straight line, which extrapolates back to the origin. Activity values were determined from samples of the Bearpaw Formation (Table I). A plot of plasticity index vs. clay-mineral content for these samples, however, shows a wide scatter of points through which a number of straight lines that pass through the origin could be drawn. From the standpoint of Activity therefore, the Bearpaw sediments do not behave as uniform soil but reflect variations both in the clay-mineral composition of the samples and probably in the degree of crystallinity of the component minerals.

The presence of montmorillonite in relatively large quantities, whether disseminated through the mass of the sediment or concentrated, as in bentonite layers, is probably the most significant geological factor affecting the engineering behaviour of the formation. Minerals of the montmorillonite group, as is well known, have a high affinity for water. In the presence of water these minerals typically show swelling or expansion accompanied by a decrease in shear resistance of the soil. The extent to which swelling with consequent loss in shear strength will occur is controlled by such factors as the composition of the clay and its abundance relative to other soil constituents, the nature of exchangeable cations, the presence or absence of cementing materials, and the chemistry of the pore fluid. The relationship of these controlling factors to the soil mass as a whole is governed by the overall fabric of the soil mass, which includes such physical features as structure and texture on both microscopic and macroscopic scales. As a consequence of the complex interrelationships among these various factors no simple correlation between clay-mineral composition and engineering behaviour of Bearpaw sediments can be expected.

Stress History

In addition to the influence of mineralogy, the physical properties of a sedimentary deposit are determined by the kind and magnitude of forces to which the deposit has been subjected throughout its geological history. Transformation of a clayey sediment from a highly hydrated ooze at the time of deposition to an argillaceous rock at a later time in its geological history is effected primarily by the vertical forces exerted by overlying sediments. The extent of the transformation, whether to soft clay, a highly fissile shale, or some intermediate stage is a function of the magnitude of the overburden load and the duration of its application. Changes that occur during the transformation are a reduction in pore volume occasioned by expulsion of pore water, which, in some sediments, is accompanied by the deposition of cementing material. Throughout the consolidation part of a sediment's geological history it tends toward a condition of equilibrium with its stress environment, and in the process it attains a stress condition that reflects the confining pressure. If confining pressures are reduced, as for example through erosion, the sediment then tends toward a new condition of equilibrium in a state of reduced stress. During this phase, which has been variously termed unloading, relaxation, or rebound, void spaces tend to increase and moisture is free to migrate into a system that is now in a state of reduced potential energy. Other physical changes that can occur during unloading are the formation of joints and fissures and the rehydration of expanding-lattice clay minerals, such as the montmorillonites, if present.

The post-Bearpaw sedimentary record is sufficiently complete, thus the kinds of forces and their relative magnitude to which the Bearpaw sediments have been subjected can be interpreted.

According to Webb (1964) the tectonic and depositional environment prevailing during the Late Cretaceous Epoch continued without important change until late in the Paleocene Epoch. Throughout this time interval sediments derived from the highlands to the west of the present Rocky Mountains continued to be deposited in a marginal basin occupying the site of the present Rocky Mountains and Foothills and to the east across the plains. In the marginal basin the sediments attained a thickness of at least 5,000 feet, but a much thinner section of strata was deposited across the plains region. During the succeeding Eocene Epoch these and older strata were subjected to the orogenic deformation that formed the Rocky Mountains and uplifted the Plains Region.

Geological activity across the Plains Region throughout the remainder of the Tertiary was primarily that of massive erosion. The extent of denudation is indicated by the preservation on isolated uplands, such as the Cypress Hills, of but a minor portion of the vast volume of debris that moved eastward across the Plains during the Tertiary. Some indication of the amount of material removed can be obtained from the local relief shown by these isolated uplands, which are remnants of a more extensive older erosion surface. The Cypress Hills rise about 1,800 feet above the plains; to the north and east other erosional remnants capped by Tertiary sediments exhibit from 200 to 600 feet of local relief.

Prior to the onset of glaciation in the Pleistocene Epoch the plains area was deeply dissected by an integrated drainage network and most of the prominent topographic features of the present landscape were in existence at the time (Barton et al., 1964).

During the Pleistocene Epoch the continental Laurentide ice-sheets, which developed in the area immediately west of Hudson Bay, advanced to the south and west across the Plains Region. The continental ice masses coalesced in the foothills area with alpine glaciers that advanced to the east from the Cordilleran Region. Although four major glacial advances separated by interglacial stages have been recognized in the midwestern United States, deposits older than the last glacial advance (Wisconsin) have not been recognized with certainty in the Plains Region. Thicknesses of the glacial deposits that cover most of the Plains Region range from a few feet to hundreds of feet, but it is the conclusion of various workers (Barton et al., 1964) that neither glacial erosion nor deposition altered the pre-existing landforms to any great extent and that the continental glacier was not an exceptionally potent agent of erosion.

On the basis of the elevations of the unglaciated parts of the upland surfaces in southern Saskatchewan and southern Alberta, Stalker (1965) has calculated the probable maximum thickness of Laurentide glaciers that covered this area. In the Lethbridge area the thickness of ice during the strongest glaciation was 2,200 feet. An estimate of ice thickness of 4,000 feet over the Red Deer Valley east of Drumheller, Alberta and 3,700 feet over the South Saskatchewan River valley in central Saskatchewan is obtained by extrapolation of Stalker's data.

Apart from erosion and deposition of material one of the principal physical effects of glaciation was crustal depression under the influence of the ice load. Upon deglaciation rebound of the crust occurred and a measure of the extent of this rebound has been obtained from the extent of deformation of glacial lake beaches. Johnston (1946) has shown that the Campbell beach of glacial Lake Agassiz, which parallels the Manitoba escarpment, has a differential increase in elevation of about 200 feet over a distance of approximately 560 miles from its southern extent in Minnesota to its location on the east flank of the Pasqui Hills in Saskatchewan. The amount of post-glacial uplift in the western part of the Plains Region is unknown, but it may well be less than that for the Lake Agassiz area because of lesser ice thicknesses.

It is therefore apparent that sediments of the Bearpaw Formation have been subjected to loading by deposition of younger sediments and to unloading by erosion during the period of uplift, followed by a further cycle of loading and unloading as a result of glaciation. The relative extent of consolidation of the sediments attributable to a particular loading phase in the history of sediments cannot be assessed with any certainty. From the geological evidence, however, it is probable that in central and southern Alberta the load of sediments overlying the Bearpaw Formation that were eroded prior to glaciation exceeded the load imposed by glacial ice. In central Saskatchewan, in comparison, where the overlying sediments were thinner and the glacial ice relatively thicker than in Alberta the two main loading cycles may have been of nearly equal magnitude.

The rate of loading and unloading during the two cycles is vastly different. Accumulation of Bearpaw and younger sediments occurred over a period of about 25 million years and the period of uplift and erosion from the end of the Paleocene to the beginning of the Pleistocene was of some 49 million years duration. The advent of glaciation in North America occurred about 1 million years ago and if the Plains Region was affected only by the later stages of glaciation the glacial loading and unloading cycle occurred within perhaps the last 100,000 years.

Peterson (1958) has reported the presence of lateral stresses within the Bearpaw shale that are 150 per cent of the vertical stresses imposed by the existing overburden. These high lateral stresses are a residuum of higher overburden stresses imposed upon the shale throughout its geological history. Whether the residual stresses are the result of loading from Paleocene sedimentation, glacial loading during the Pleistocene or a combination of both is not definitely known.

Sheet-jointing, rock bursts, and similar stress relief phenomena have been observed in near-surface excavations in Palaeozoic igneous rocks in Vermont (White, 1946), and large scale exfoliation attributable to stress relief has been observed by Bradley (1963) in Mesozoic and older sandstones of the Colorado Plateau. These observations indicate that the magnitude of residual stress is a function of the magnitude of the pre-existing confining pressure and that residual stresses in geological materials, regardless of geological age, are relieved only to shallow depths in a zone that tends to parallel the surface of the ground. Stress relief effects are most pronounced where relatively rapid rates of overburden removal have occurred, for example, by man-made excavations or rapid downcutting by streams.

Stress relief in the Bearpaw Formation is accompanied by increases in moisture content rather than by the brittle fracture phenomena displayed by more competent rocks. Along the South Saskatchewan River valley increases in moisture content are most pronounced within the upper 50 feet of the shale as shown by Peterson (1958). The maximum thickness of the zone of increased moisture content occurs in the valley slopes and the minimum thickness on the upland and below the alluvial fill in the valley.

Below the near-surface zone of stress relief in the Bearpaw Formation residual stresses resulting from the loading history of the sediments remain. Excavations below the zone of stress relief permit the residual stresses to be relieved with consequent increases in moisture content and decreases in shear strength.

The probable stress history of Bearpaw sediments in terms of effective overburden stress in relation to water content is shown in Figure 8.

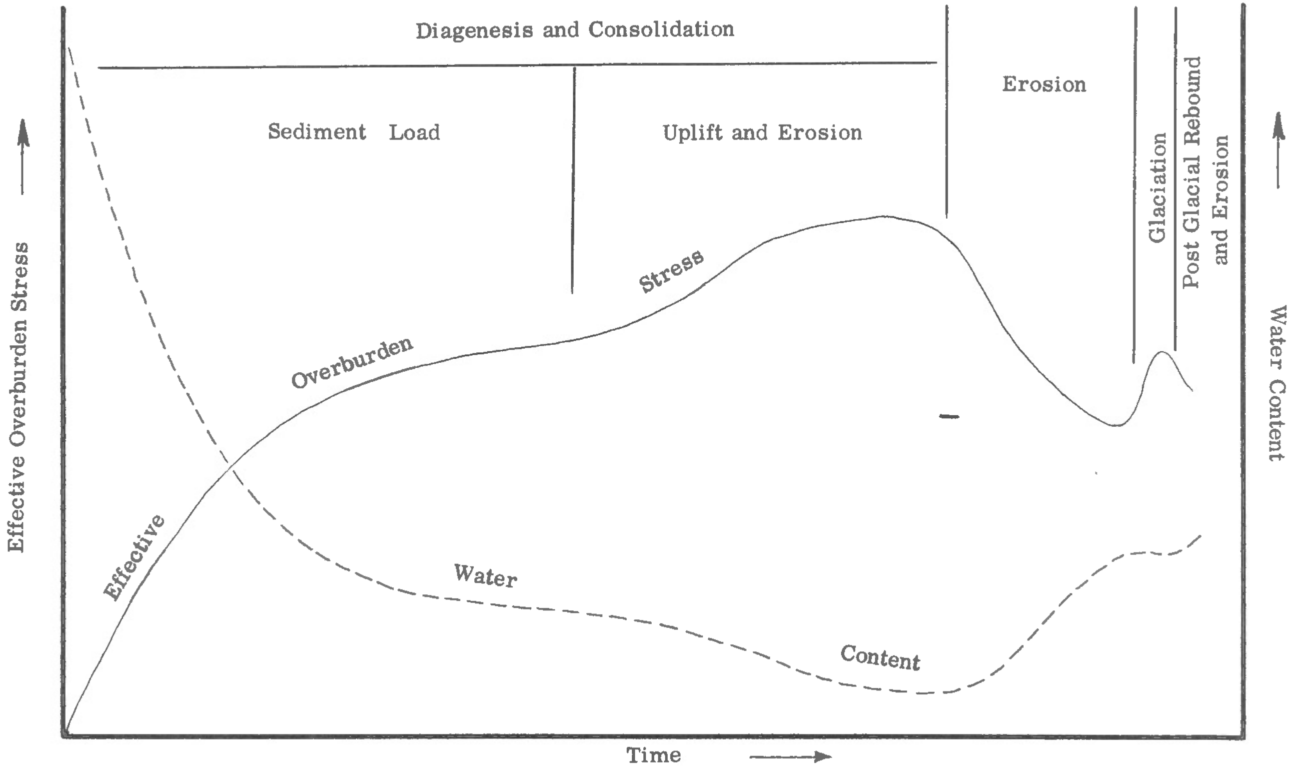


Figure 8. Probable stress history of Bearpaw sediments.

DESCRIPTION OF LANDSLIDE AREAS

Porcupine Hills Area, Manitoba

Along the eastern flank of the Porcupine Hills, a part of the Manitoba Escarpment, landslides have occurred over an area that extends continuously for 30 miles from tp. 40, rge. 26 to tp. 44, rge. 28. Throughout most of its length the landslide area is from 2 to 3 miles wide. Although the overconsolidated clay shale strata involved in this area of slope failures are, for the most part, older than the Bearpaw Formation, the area is included in the present study because of the unique morphology of the slide and the geological environment in which the slope failures have occurred.

Dense tree cover and lack of both access and reliable outcrops in the slide area prevented a detailed field examination being made in this area during the preliminary study. Most of the information concerning the area has been obtained, therefore, from interpretation of aerial photographs and from published geological reports and maps.

One of the most striking features of the slide is the parallelism and extent of the slump-block ridges (Fig. 9). The most prominent ridges, which occur adjacent to the head of the slide, show marked backward rotation and some of the ridges extend continuously for several miles. Regardless of the length of the ridges they all trend parallel with the original scarp face and show practically none of the coalesced arcuate trend of slump block ridges that is commonly developed, as for example, along the South Saskatchewan River. This linear trend is very likely due to the absence of major drainage channels across this part of the escarpment, which would tend to orient failures in a direction parallel with the tributary valley walls.

Graben structures are prevalent in the head region of the slide. This fact accompanied by the absence of a bulge at the toe of the slide indicates that the principal surface of rupture is one of low curvature.

According to Wickenden (1945, Map 637A) the slide area is underlain by Upper Cretaceous strata of the Riding Mountain, Vermilion River, and Favel Formations and by the Ashville Formation, which contains beds of both Upper and Lower Cretaceous age. The bedrock formations are covered by a substantial thickness of glacial drift. All of the bedrock strata consist primarily of grey and black shale. Bentonite layers occur in all of the formations, but are most abundant in the upper beds of the Vermilion River Formation and the lower beds of the Riding Mountain Formation. Although erosion has removed much of the upper part of the Riding Mountain Formation in this area some of the hard silicic shale characteristic of upper Riding Mountain strata is probably present in the 'razor-back' ridges that occur in the head region of the slide area.

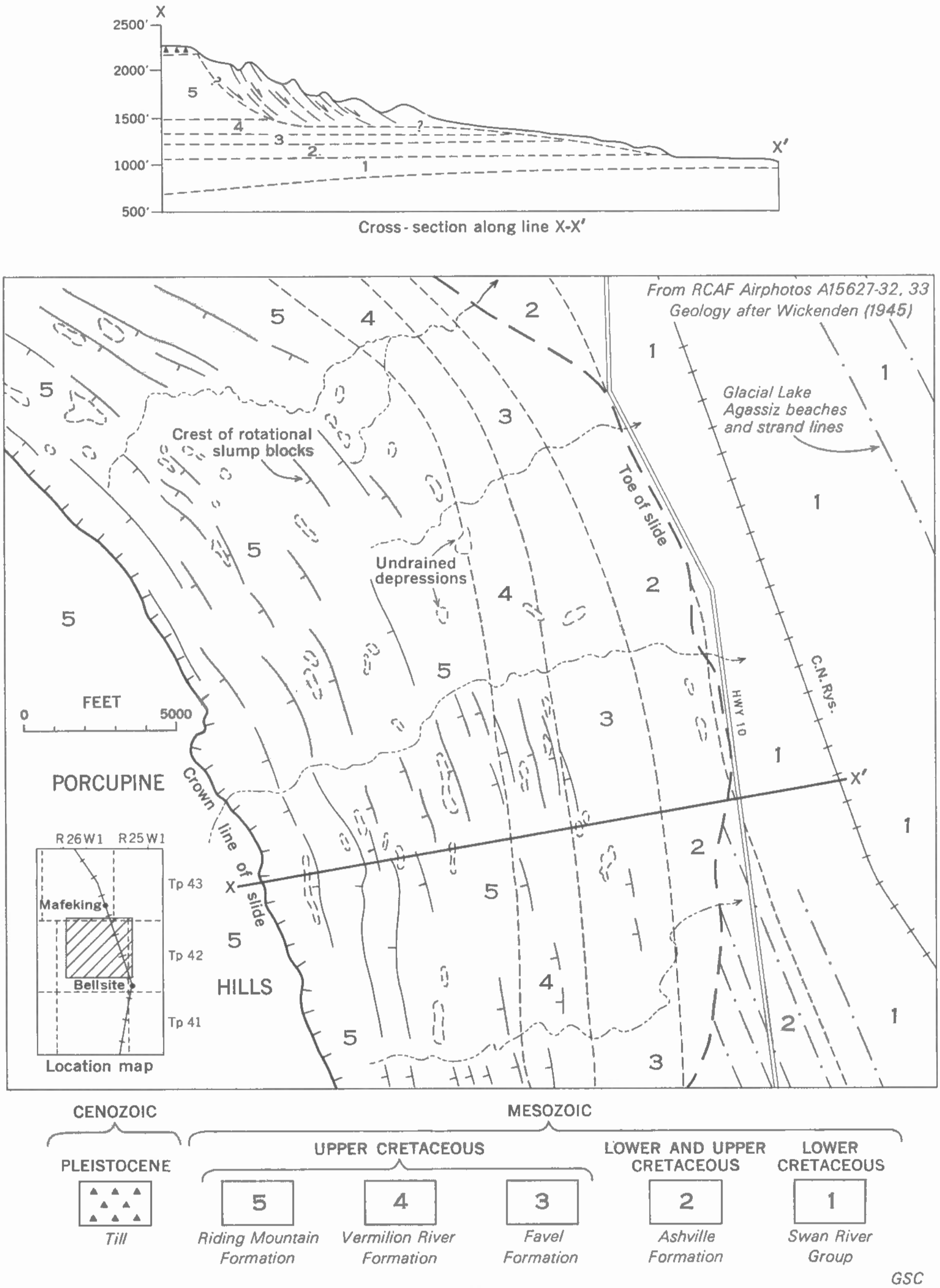


Figure 9. Part of landslide area, Porcupine Hills.

Throughout the landslide area the strata dip west at about 5 to 10 feet to the mile (Wickenden, 1945). The structure and stratigraphy of a typical part of the landslide area are shown in Figure 9.

Wickenden (1945, p. 55) reported that beds down to the Ashville Formation have been disturbed by the landslide, but did not state in what part of the slide the disturbed beds of the Ashville Formation were observed. From the location of the outcrop area of the Ashville it is assumed that these strata have been disturbed mainly in the toe area of the slide.

The most probable location of the main part of the failure surface is within the highly bentonitic strata of the upper part of the Vermilion River Formation and the lower part of the Riding Mountain Formation as shown in the cross-section in Figure 9.

Truncation and disturbance of glacial Lake Agassiz beaches by the slide (Fig. 9) indicate that the slide occurred at a time following the high level stages of Lake Agassiz. It is possible that drainage of Lake Agassiz may have contributed to the instability of the original slope by producing a rapid draw-down condition at the toe of the slope.

The presence of groundwater discharge in the form of springs and seepage zones was observed in the toe area of minor slope failures that had occurred along the valley walls of creeks that drain the northern part of the slide area. In view of the relatively few creeks that drain the slide area it is probable that groundwater discharge on a larger scale along part of the escarpment prior to the failure was in the form of a series of contact springs located at or near the top of the highly bentonitic layers within the stratigraphic sequence. Decreases in shear strength of the bentonites in the zone of ground-water discharge could have initiated small failures, which retrogressed to form the broad belt of failed slopes now present.

Post-glacial uplift, evidenced in the slide area by the deformation of glacial Lake Agassiz beaches (Johnston, 1946), may also have been a factor in contributing to slope instability by increasing the groundwater discharge gradient.

Wawanesa and La Rivière, Manitoba

In southern Manitoba minor slope failures involving strata of the Riding Mountain Formation occur at various places along the valley walls of Souris and Pembina Rivers. Typical examples of these failures were examined along Souris River, approximately 3 miles southwest of Wawanesa (Fig. 10) and 1 mile northeast of La Rivière on the wall of a small valley tributary to the Pembina River (Fig. 11).

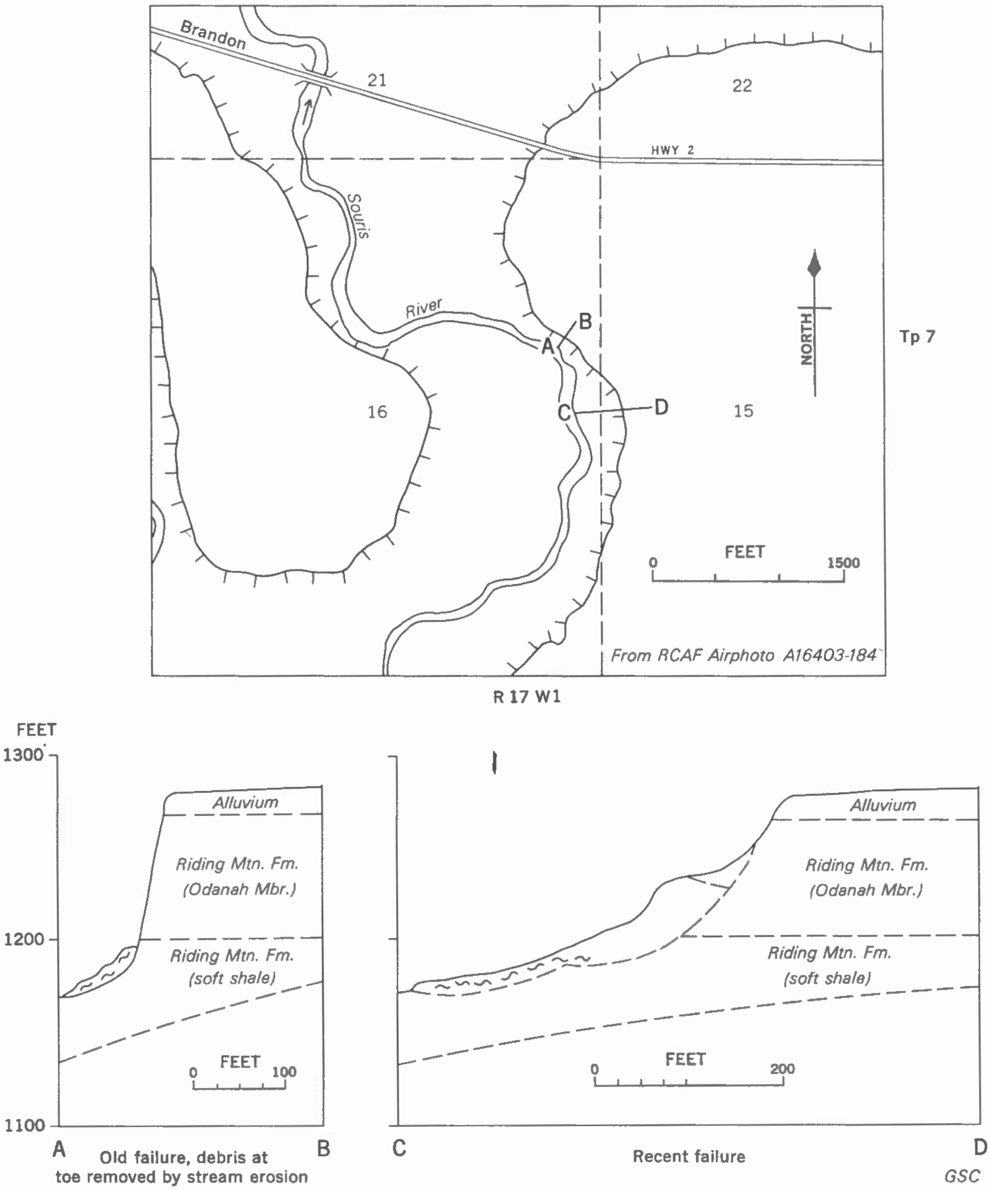


Figure 10. Location of landslide along Souris River southwest of Wawanesa.

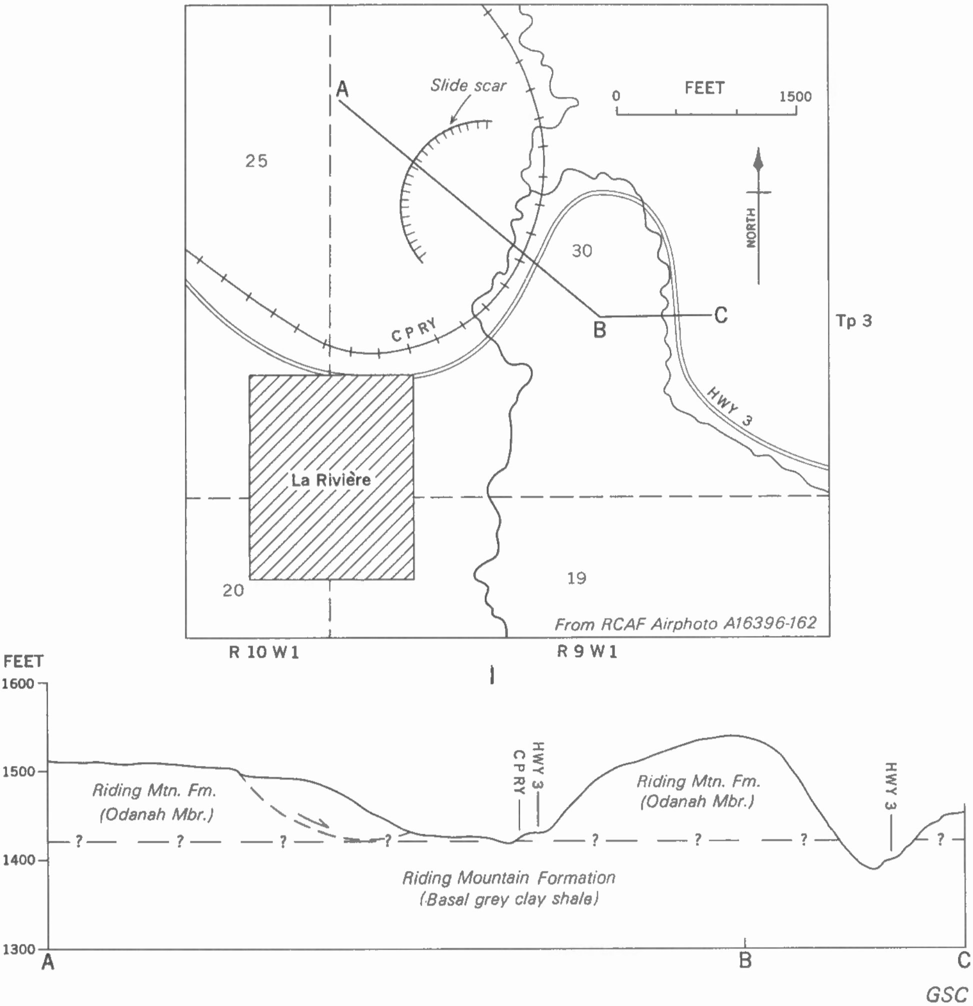


Figure 11. Location of landslide at La Rivière.

Plate I. Failure surface in sediments of Riding Mountain Formation along Souris River. (GSC 118522)

The valley of Souris River in the Wawanesa area, is characterized by an asymmetrical cross-section resulting from the development of incised meanders. Slope failures are confined, therefore, to the undercut slopes which rise 100 feet or more above the stream channel at angles ranging from 30 to 60 degrees.

Most of the failures are relatively small involving from a few thousand to 50 thousand cubic yards of material. The failure surfaces are typically spoon-shaped (Plate I), indicative of a rotational slump type of failure, but break-up of the slide material through movement has produced some of the characteristics of a debris slide.

At the site of the slide area the stratigraphic section consists of two shale members of the Upper Cretaceous Riding Mountain Formation. The lower member, which consists of dark brown, soft, moist shale, is exposed for 30 feet above river level and is overlain by approximately 65 feet of pale grey, hard, highly fractured silicic shale. The boundary between the shale members is marked by an 18-inch thick layer of bentonite. Throughout the area the river banks are capped by from 10 to 15 feet of stratified fine sand which according to Elson (1960) is of glacial lake origin. The regional dip of the bedrock strata is to the southeast at from 5 to 10 feet per mile.

The upper surface of the bentonite layer is the locus of a zone of groundwater discharge, which along a 100-foot length of the discharge face amounts to a flow of about 1 gallon per minute. Although the discharge quantity is small the presence of groundwater discharge and the bentonite layer are probably the most significant geological factors contributing to the instability of the slopes.

Retrogressive development of some of the slides is occurring as evidenced by the presence of tension fractures in the crown area of the failure scars and by active lateral stream erosion at the toe of the failed slopes.

The valleys of both Pembina River and its tributary, which contains a branch line of the Canadian Pacific Railway at La Rivière, were eroded primarily by glacial meltwater during an early phase in the deglaciation of southern Manitoba. A small slide of the rotational slump type is present in the north wall of the tributary valley about one half mile northeast of La Rivière (Fig. 11). It is probable that this slide occurred during the erosion of the valley by glacial meltwater. Although debris at the toe of the slide has been removed by stream erosion the slide mass appears to be stable and is completely overgrown by vegetation. Bedrock is not exposed in the slide area, but good exposures of the dark grey, highly fractured, hard silicic shale of the Riding Mountain Formation in essentially horizontal beds are found in road cuts in adjacent valley walls.

Several bands of bentonite, each about one inch thick, are interstratified with the hard shale. The bentonite layers are moist where exposed in the road cuts and appear to be the locus of groundwater seepage. Pocket penetrometer readings of 0.30 T.S.F. were obtained for the bentonite layers as opposed to penetrometer readings of 4.5 T.S.F. or greater for the hard shale.

Although the exact stratigraphy involved in the slide at La Rivière was not directly observed it may be inferred from the local stratigraphy that the presence of either a bentonite layer or a soft shale zone was a geological factor contributing to the slope failure.

Landslide scars may be observed elsewhere along the valley walls of Pembina River and some of its tributaries. These slides also probably occurred during the active erosion of the valley by glacial meltwater. The present underfit character of Pembina River and some of its tributaries is such that active stream erosion is confined to the meander belt. The erosion of the valley walls occurs therefore in few places, as the meander belt is generally located within the central part of the valley.

Both the Wawanesa and La Rivière areas are underlain by similar rock types, but the prevalence of active landslides along Souris River and the generally stable slopes along the Pembina Valley are reflections of the differences in stream regimen between the two areas.

South Saskatchewan River Area

Throughout most of central Saskatchewan the valley of the South Saskatchewan River is underlain by overconsolidated shales of the Bearpaw Formation, which are exposed by river erosion. Extensive landslides have occurred on both sides of the valley and, in fact, this form of mass wasting is primarily responsible for the widening of the valley. As might be expected landslides are most prevalent on the undercut slopes of the river and along the most deeply entrenched reaches of the river. For example, where the river is entrenched 200 feet below the upland surface landslides have created a valley width of approximately 6,000 feet, but where the river valley is 600 feet deep landslides have extended the valley width to approximately 24,000 feet.

The effectiveness of landslides in valley widening may be further illustrated by comparing valley development in an area underlain by overconsolidated shales with that in an area underlain primarily by sandstones with interbedded overconsolidated shales. At the Rapid Narrows of South Saskatchewan River, 20 miles northeast of Medicine Hat, Alberta, the river has eroded a steep-walled gorge through shales of the Bearpaw Formation and sandstone and shale of the underlying Belly River and older formations. There the valley is 450 feet deep and 3,400 feet wide. Farther east in central Saskatchewan, where only the Bearpaw Formation has been eroded, a valley 450 feet deep would have a width of approximately 13,000 feet due to landslides.

In addition to the extensive areas of slumping that occur along the main valley of the South Saskatchewan River slumping has also occurred along many of the stream valleys that are tributary to the river. The extent of slumping along the tributaries is proportional to the size of the tributary and, in general, decreases upstream from the confluence of the tributary with the South Saskatchewan River.

During the field study several areas of slumping along tributary streams were examined, as these provided examples of geological and other factors relating to the mechanism of landslides that were not so apparent in the main valley slide areas. Three of these are discussed below.

Area 'A'

One of the most extensive areas of landslides along the South Saskatchewan River occurs within tp. 20, rge. 10, W. 3rd mer., herein designated as Area 'A' (Fig. 12). The upland in this area is a part of the Missouri Coteau through which the river has eroded a valley 600 feet deep. The valley width in Area 'A' is approximately 24,000 feet, but extensive slumping along tributary valleys, particularly on the south side of the river, has contributed extensively to the widening of the South Saskatchewan River valley in this

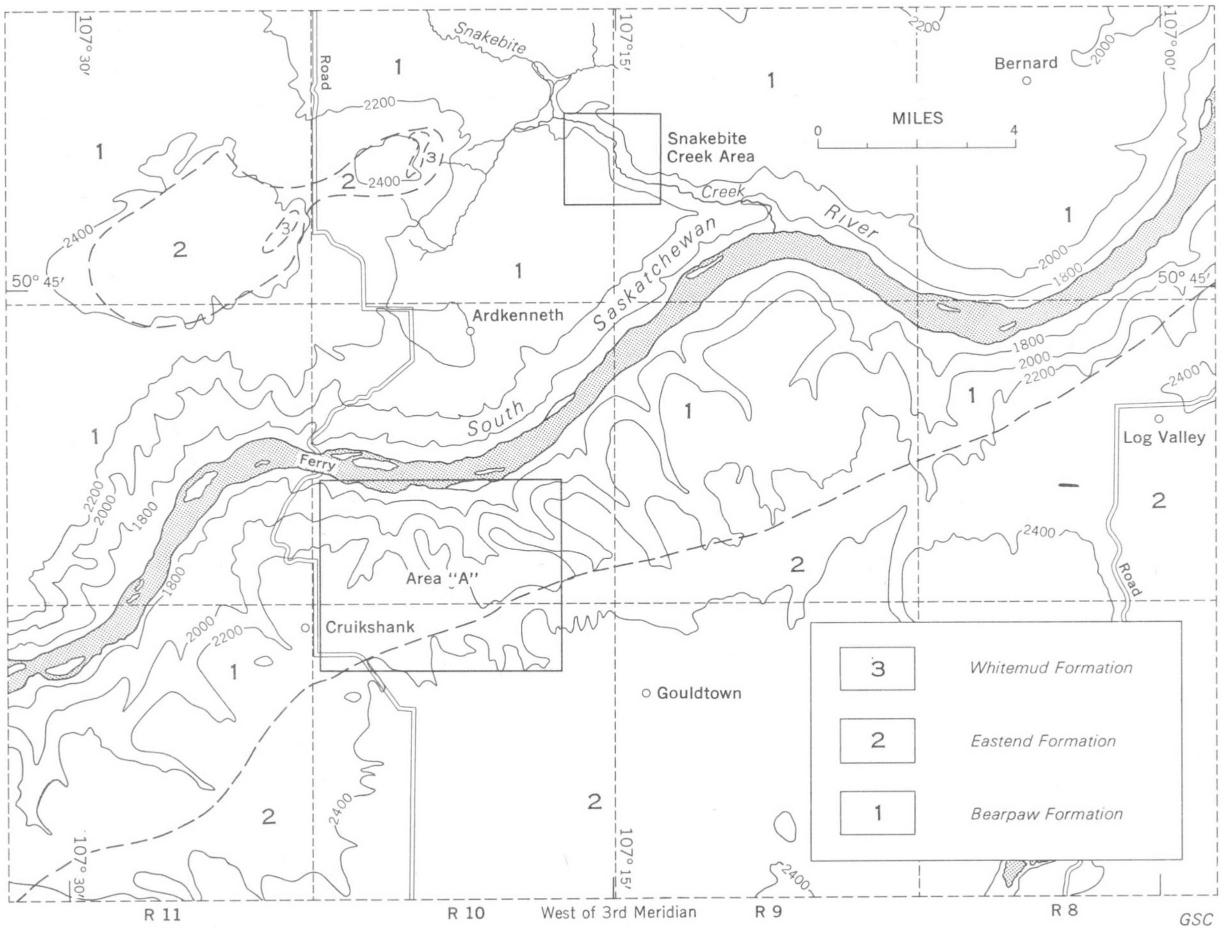


Figure 12. Bedrock geology, South Saskatchewan River area.

area. Although landslides are more extensively developed in Area 'A' than elsewhere along the South Saskatchewan River the processes of mass movements are typical of those found elsewhere along the river valley.

Two main zones of mass movement can be recognized within Area 'A', viz. a zone of rotational slumping in the head region of the valley walls and a zone of soil creep on the lower valley slopes, which are composed of bare exposures of weathered clay shales of the Bearpaw Formation (Fig. 13).

The slump-block ridges are aligned parallel with the main valley or with tributary drainage valleys (Fig. 14). Graben structures are common in the zone of rotational slumping indicating lateral spreading along a zone of failure having a large radius of curvature. Retrogression of slides along the tributary valleys has caused many of the smaller slide areas to coalesce and thus expedite widening of the South Saskatchewan River valley.

Strata of the Bearpaw Formation that occur within Area 'A' and elsewhere along the South Saskatchewan River valley have been subdivided into three lithological members by Evans (1961) (Fig. 13). The lowest unit exposed is the Snakebite member composed of dark grey shales containing numerous bentonite seams, nodule horizons, and layers of calcite. Soil creep in the weathered zone of these shales is particularly active on the lower slopes of the valley walls. The weathered zone consists of a 1 to 2 inch surface of desiccated, nodular, exfoliated shale underlain by 8 to 10 inches of soft, flaky shale fragments with an abundance of disseminated selenite crystals.

The lower shales are overlain by approximately 40 feet of light brown, fine-grained, moderately compact sandstone of the Cruickshank member. The pale weathering character of the sandstone and its relative resistance to erosion make this unit easily recognizable about the mid-height of the valley slope throughout Area 'A'.

Dark grey, silty shales of the Vermilion member overlie the Cruickshank sand member and are the principal strata involved in the zone of rotational slumping. Bentonite seams also occur in the Vermilion member but are less abundant than in the lower shales.

Throughout Area 'A' and the adjacent parts of the South Saskatchewan River valley strata of the Bearpaw Formation dip east at 8 to 10 feet per mile.

Fraser et al. (1935) have indicated the presence of the Eastend Formation in the southeast part of Area 'A' (Fig. 12). Where exposed the Eastend Formation consists of fine sand and silt in transitional contact with the underlying Bearpaw Formation. No exposures of the Eastend Formation, however, were observed in Area 'A', as the Bearpaw Formation in this area appeared to be directly overlain by glacial till.

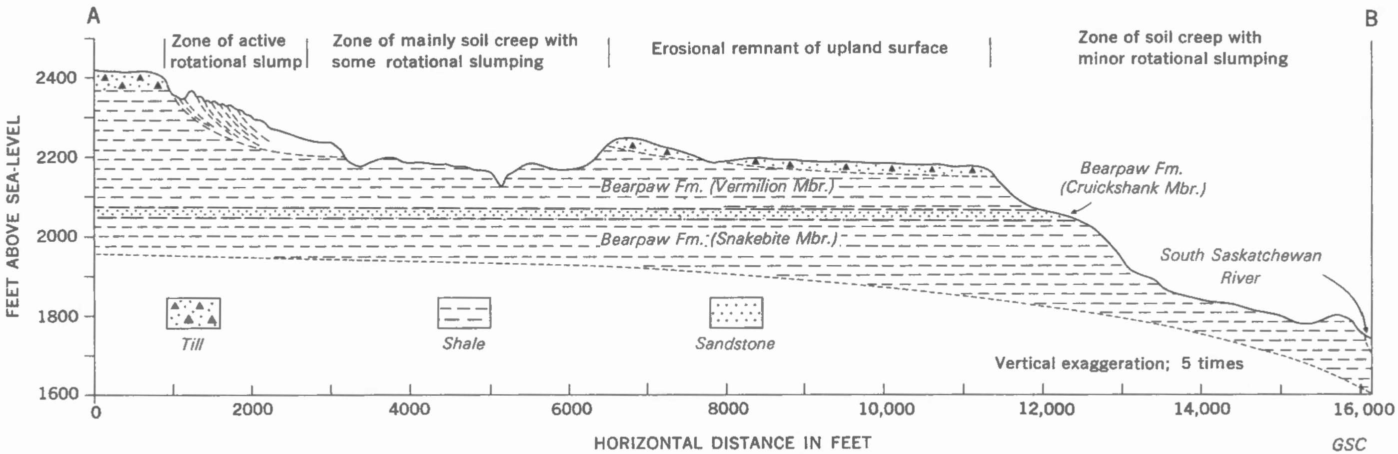


Figure 13. Section along line A-B, Area "A". (See Figure 14 for location)

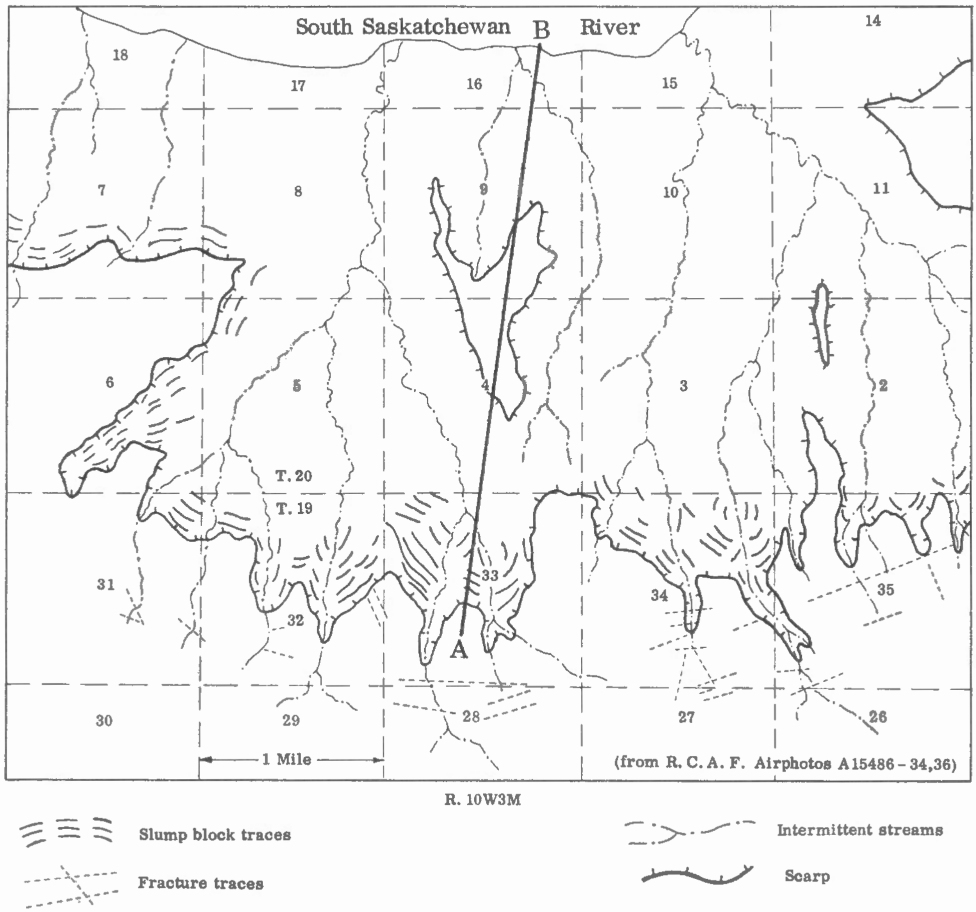


Figure 14. Slope failures along south bank of South Saskatchewan River, area "A".

North of the South Saskatchewan River in tp. 21, rges. 10-11, W. 3rd mer. sands and silts of the Eastend Formation and kaolinitic sands and clays of the overlying Whitemud Formation are preserved beneath the glacial till on a local topographic high (Fig. 12). These strata are erosional remnants of a thicker section of younger sediments that covered the Bearpaw Formation. The exact thickness of the post-Bearpaw section in the South Saskatchewan River valley is unknown, but on the basis of the thickness of the post-Bearpaw section in southern Saskatchewan (Fraser et al., 1935) it probably did not exceed 1,000 feet.

The presence of these younger strata indicates that this area has undergone less preglacial erosion than areas farther east where several hundred feet of Bearpaw strata have been removed.

The presence of bentonite layers and the high percentage of montmorillonite in shales of the Bearpaw Formation are significant factors contributing to the instability of slopes underlain by the formation. Although the coincidence of slope failure surfaces and bentonite layers was not directly observed in the field it is probable that such a relationship existed at least during the early phases of landslide development.

Studies of the Pleistocene geology of the South Saskatchewan River valley and adjacent areas by Christiansen (1959), Scott (1962), and others have shown that erosion of the river valley through most of central Saskatchewan began approximately 10,000 to 12,000 years ago following the deglaciation of southern Saskatchewan. During the early stages in the development of the South Saskatchewan River the steepest gradients and consequently the most active erosion would occur along the face of the Missouri Coteau. It is probable, therefore, that the rate of erosion of the river valley through Area 'A' exceeded the rate of residual stress relief in the freshly exposed shales of the Bearpaw Formation, which contributed to the instability of the valley walls.

Most of the stream courses that drain Area 'A' were dry during July, 1963 when the area was examined and there was no sign of springs or seepages that would indicate active groundwater discharge in the slide area. Many of the slopes show evidence of rill erosion and the effects of sheet wash from which it is concluded that the tributary stream channels are the result of rapid run-off during spring break-up and during infrequent heavy rains.

Although the tributary stream pattern is essentially dendritic, which is to be expected in an area of flat-lying sediments with relatively uniform lithology, the upper reaches of the tributary streams and their secondary tributaries show a marked tendency toward an angular drainage pattern that appears to be controlled by incipient bedrock fractures.

The fracture traces shown on Figure 14 are not visible on the ground, but are clearly shown as linear trends on aerial photographs. The origin of the fractures is uncertain, but their pronounced trend parallel with the main valley scarp suggests they are vertically oriented and may be tension fractures associated with stress relief during development of the valley. Regardless of origin the fractures constitute planes of weakness within the bedrock and are, therefore, a factor to be considered in a stability analysis of this area.

Snakebite Creek

Slumping has developed extensively along the lower reaches of Snakebite Creek, a small perennial stream tributary to the South Saskatchewan River in tp. 21, rges. 9-10, W. 3rd mer. (Fig. 15). For a distance of 5 miles upstream along the creek from its confluence with the South Saskatchewan River slumping has occurred on both valley walls, but is more extensively developed on the west side of the valley. The landslides are of the rotational slump type, thus the valley walls comprise a series of linear arcuate ridges separated by undrained depressions. On both sides of the valley the depressions are the locus of greatest vegetation density; however, the west side of the valley contains shrubs and small trees which are generally absent on the prairie grass-covered slopes on the east side of the valley.

Beyond a distance of 5 miles slumping has occurred only on the west side of the valley producing an asymmetrical valley profile as shown in the cross-section A-B (Fig. 15). It may be noted that the landslides in the area represented by the cross-section have occurred on a convex bend in the creek, thus the erosion is probably not a significant factor contributing to the instability of these slopes.

Dark blue-grey, soft clay shale is exposed in the slide area beneath a cover of approximately 80 feet of till overlain by a thin veneer of glacial lake sediments. On the basis of Evans' (1961) stratigraphic studies the clay shale strata occur near the base of the Vermilion member of the Bearpaw Formation. Bentonite seams near the base of the Vermilion member were observed by Evans (1961) farther downstream in Snakebite Creek. Although these bentonite layers were not observed in the slide area as shown in Figure 15 they are probably present. The attitude of undisturbed bedrock strata in Snakebite Creek is essentially horizontal but they probably conform to the regional easterly dip of 8 to 10 feet per mile.

Removal of the drift cover and relief of residual stresses in the underlying Bearpaw Formation by stream erosion is the prime factor contributing to the instability of slopes along Snakebite Creek. The more extensive development of landslides on the west side of the valley, however, suggests that forces contributing to slope instability are not of equal magnitude on both sides of the valley. Although the annual flow characteristics of Snakebite Creek are unknown there is no field evidence to indicate more extensive erosion along the toe of the west wall of the valley. The presence of perennial stream flow and the greater density of vegetation on the west side of the valley indicate groundwater seepage, which probably originates

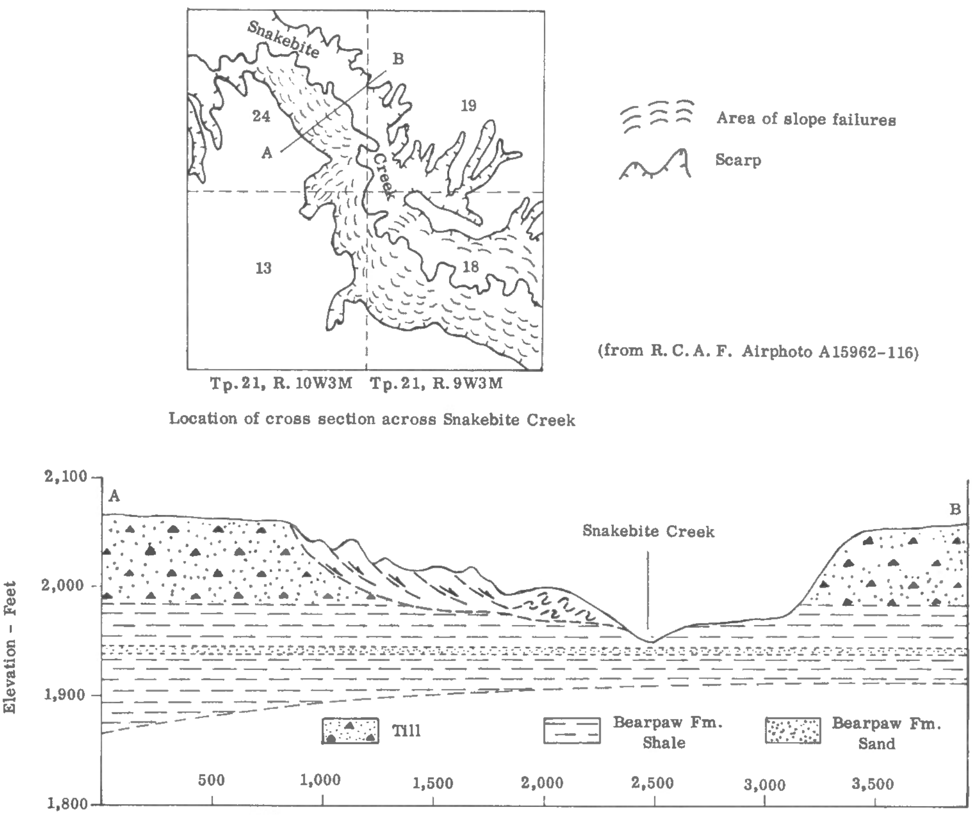


Figure 15. Asymmetrical valley profile caused by slumping along Snakebite Creek.

primarily from the west side of the valley. A local topographic high located 4 miles west of Snakebite Creek provides approximately 400 feet of local relief whereas to the east of the creek the maximum local relief is about 200 feet. If the local groundwater flow system is related to the topography then it is reasonable to infer that the groundwater gradient and seepage pressures on the west side of Snakebite Creek will be greater than on the east side of the creek.

The more extensive development of landslides along the west side of Snakebite Creek also suggests that the north and northeast exposure of the slope may be a contributing factor to slope instability. The climatic conditions of the area are such that greater retention of moisture from snowmelt and infrequent precipitation and consequently higher seepage pressures could be expected on slopes receiving the least amount of insolation.

It has been suggested by Beaty (1956) that slope exposure in an area east of Berkeley, California, plays an important role in determining on which slopes landslides are most likely to occur. Throughout most of the South Saskatchewan River valley, however, there is no apparent relationship between slope exposure and the extent of landslides because of the dominance of other factors, but in local areas such as Snakebite Creek slope exposure may be significant, particularly during the initial phases of landslide development.

Riverhurst, Saskatchewan

The reservoir of the Gardiner Dam will occupy a part of the river valley in which extensive landslides have occurred. In order to examine the effect of filling the reservoir on the stability of failed slopes an area of slope failures along a tributary to the South Saskatchewan River (sec. 36, tp. 22, rge. 8, W. 3rd mer., sec. 30, tp. 22, rge. 7, W. 3rd mer.) 5 miles west of Riverhurst, Saskatchewan was selected for detailed study. In this area 30 feet of alluvial sand and gravel and 80 feet of till overlie the overconsolidated clay shales that are a part of the Snakebite member of the Bearpaw Formation.

Eighteen well-point piezometers were installed at varying depths in the shale along a section across the valley to determine the configuration of the groundwater flow pattern in the landslide area and to monitor changes in the groundwater regimen as reservoir filling progresses (Scott, 1965). A more detailed description of the piezometer installations and the results obtained from them will be incorporated in a forthcoming report on the land-slide study in the Riverhurst area.

Because of the low permeability of the Bearpaw Formation the water level readings obtained from the well-point piezometers are influenced by a lengthy time lag in equilization of hydrostatic pressure in the piezometer with

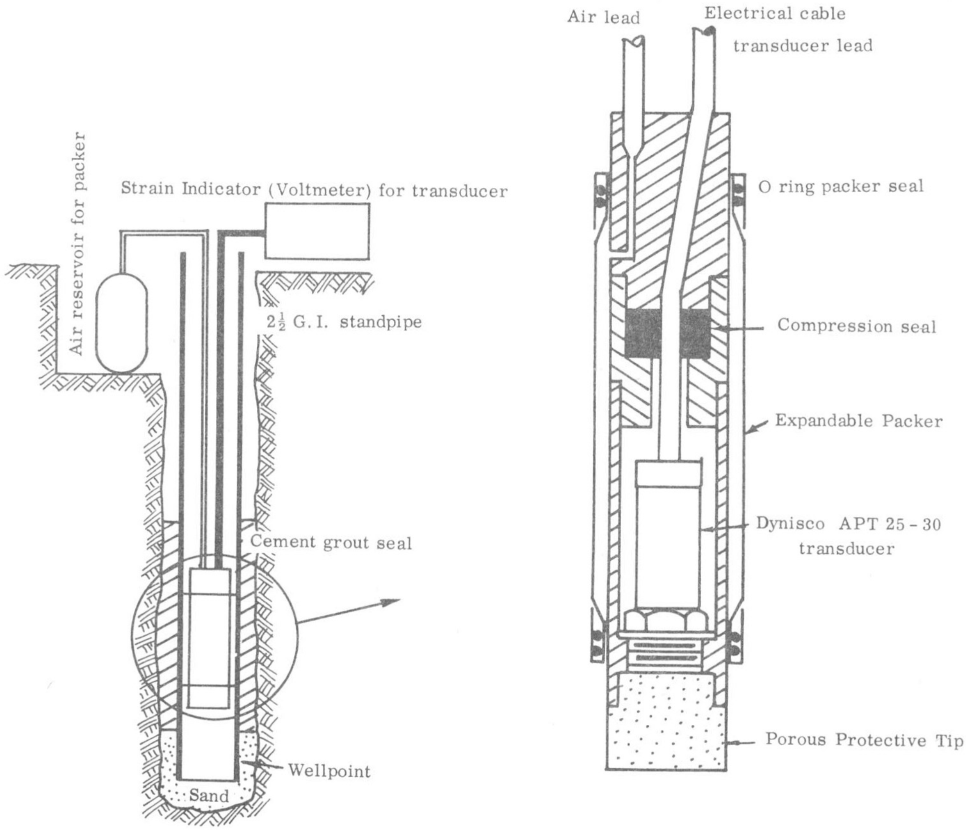


Figure 16. Schematic diagram of G. S. C. – U. of A. transducer piezometer.

that in the formation. Consequently the piezometers are not sufficiently responsive to reflect either the magnitude or temporal aspects of reservoir level fluctuations that could contribute to reactivation of the landslides.

An attempt to overcome the time lag in response of the piezometers has been made by the Geological Survey of Canada in cooperation with the Department of Civil Engineering, University of Alberta, through the development of a closed system diaphragm-type piezometer using a pressure transducer sealed by a pneumatic packer in the well-point standpipe. In laboratory tests reported by Brooker and Lindberg (1965) the response of the transducer piezometer in a soil having a permeability of 2.5×10^{-7} con./sec. was so rapid that readings could not be accumulated and the time for 100 per cent response under laboratory conditions was less than 10 seconds in the soil used. A preliminary field trial of the transducer piezometer in one of the well-point installations at Riverhurst (Scott, 1966) indicated an increased responsitivity of the transducer piezometer over the open standpipe by a factor of at least 1,000. A typical transducer piezometer installation is schematically illustrated in Figure 16. The results of further field trials of the transducer piezometer are contained in a report by Brooker, Scott and Ali (in press).

St. Mary River, Alberta

In southern Alberta landslides have also occurred in various areas along stream valleys where the Bearpaw Formation has been exposed by erosion. Typical occurrences of landslides are present along the reach of St. Mary River within tps. 6 and 7, rge. 22, W. 4th mer. This part of the river also contains a complete section of the Bearpaw Formation, which had been studied in detail by Russell (1932) and Byrne and Farvolden (1959), thus the area provided an excellent location for the landslide study (Fig. 17).

The valley of St. Mary River throughout the study area is characterized by incised meanders, which create a cross valley profile of pronounced asymmetry. Undercut slopes on the outside of the meander curves rise from 100 to 200 feet above the valley floor with inclinations having a horizontal to vertical ratio of 1.5-1.2:1. Active vertical and lateral erosion by the river is indicated by the presence of recently formed meander cut-offs.

Slope failures that occur along this section of St. Mary River are confined primarily to the steep undercut slopes and comprise two main types. The larger failures are of the retrogressive rotational slump type, which show the characteristic backward rotation of the slump blocks in the head region of the slide. The upper part of the failure surface of these slides is probably a circular arc, but in the toe region the failure surface appears to be tangent to a horizontal plane. The smaller failures are of the conical wedge type in which the movement has been vertically downward along a steeply inclined failure surface of low curvature. The lower part of the

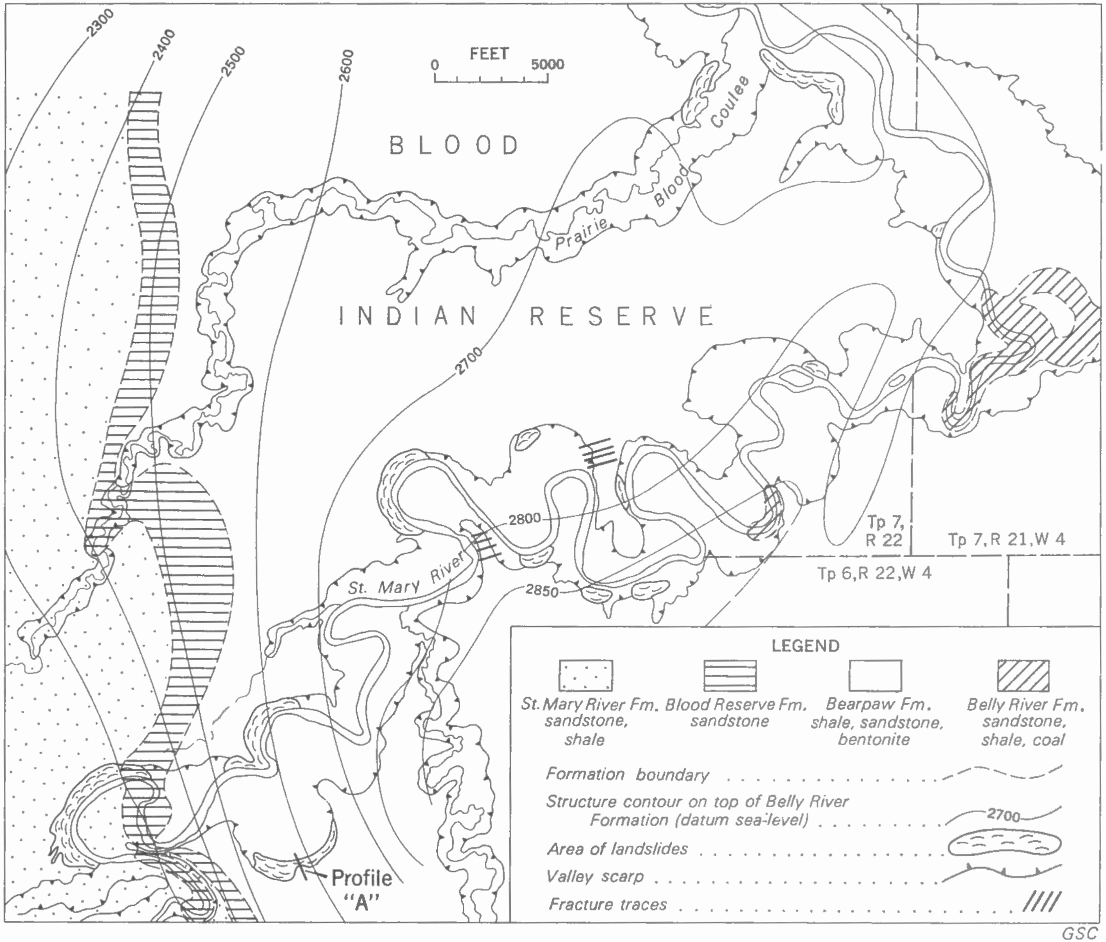


Figure 17. Occurrence of landslides in Bearpaw Formation along St. Mary River.

failure surface of the smaller slides may also be tangent to a horizontal plane. The cross-section, Profile 'A', shown in Figure 18 represents a stable part of a steep undercut slope between two small slope failures. The position and shape of the probable failure surface as shown in Figure 18 is inferred from the exposed failure surfaces of the adjoining failed slopes.

In contrast with the South Saskatchewan River area, slope failures are not continuous along the reach of St. Mary River where continuous exposures of the Bearpaw Formation occur. Stable slopes 200 feet in height at an inclination of 1.2:1 were observed in some places along St. Mary River and in other places steeper but lower slopes underlain by shales of the Bearpaw Formation are present (Plate II).

The typical sediment of the Bearpaw Formation, as exposed in the St. Mary River valley, is dark grey and dark brown, friable, blocky or fissile shale commonly containing an abundance of crystalline selenite. Numerous seams of light yellow, grey, and olive-green bentonite ranging in thickness from 0.1 foot to 2.0 feet are interstratified with the shale. The bentonites consist primarily of the clay mineral montmorillonite as determined by X-ray diffraction analysis, and from geochemical analyses it is probable that sodium is the principal adsorbed cation (Fig. 18).

The top of the formation is marked by approximately 90 feet of massive, cliff-forming, fine-grained sandstone locally designated as the Blood Reserve member. Throughout the remaining 730 feet of the formation, composed primarily of shales with bentonite seams, silty and sandy shales and minor sandstone layers occur, but these constitute only about 5 per cent of the section.

Pleistocene sediments composed of till, alluvium, and lake deposits mantle the bedrock formations in the St. Mary River area. The aggregate thickness of these sediments ranges from 25 to 200 feet depending upon the configuration of the preglacial topography (Stalker, 1963). In places along the valley the surficial deposits are exposed in almost vertical slopes that have been carved by rill erosion into castle-like forms (Plate II).

Bedrock strata in the St. Mary River area are exposed along the east limit of the Alberta syncline, thus the general structure of the area is monoclinal with a prevailing westward dip of several degrees. In the vicinity of the landslide study area along St. Mary River, however, the bedrock structure contours (Fig. 17), as determined by Russell (1932), show the presence of a minor anticlinal structure that trends northeast. Hence the direction of dip of the Bearpaw strata in the study area varies from northwest to west, but the magnitude of the dip remains at less than 5 degrees.

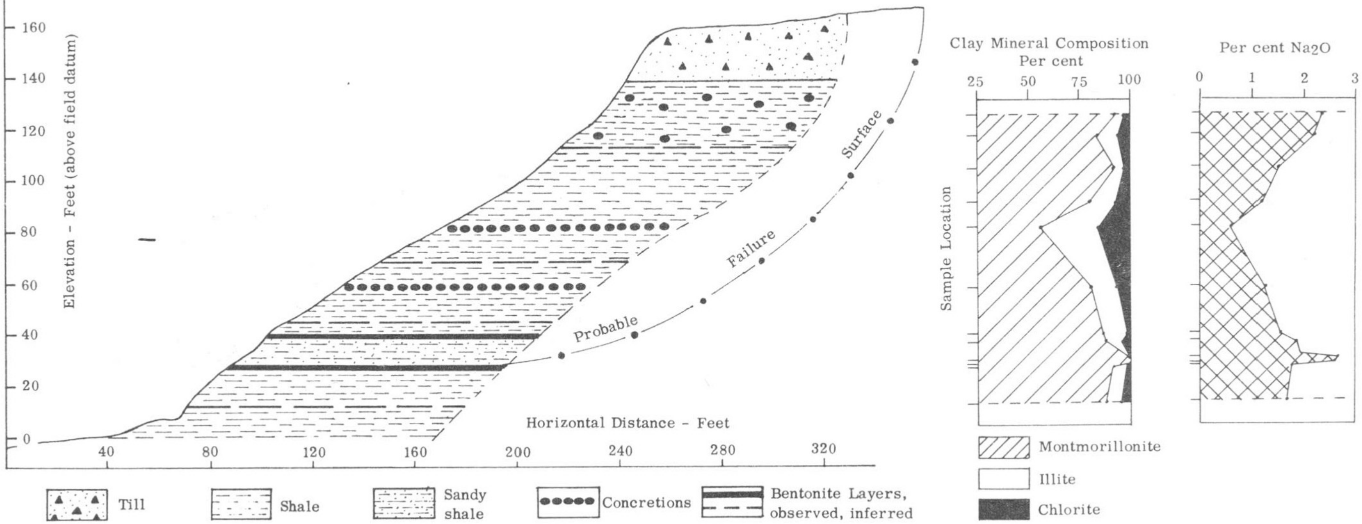


Figure 18. Profile "A", St. Mary River, NW Sec. 19, Tp. 6, R. 22 W. 4th.

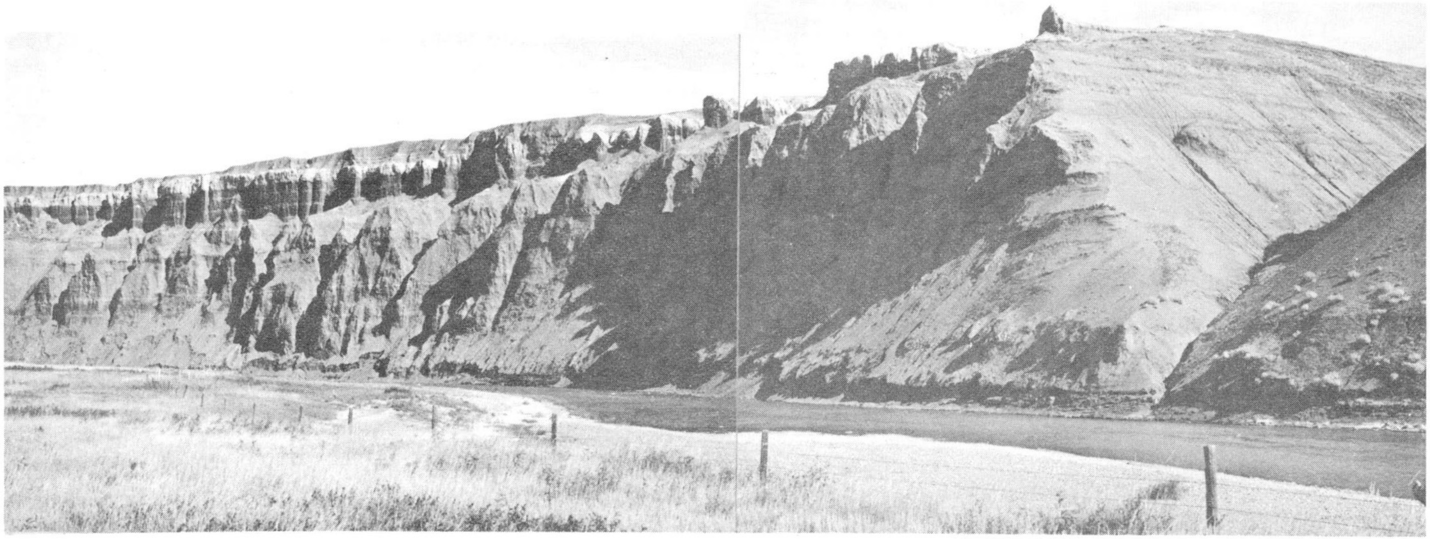


Plate II. Exposure of Bearpaw Formation along St. Mary River.
(GSC 118600, 118601)

Several linear features, accentuated by differential erosion, having a trend parallel with the axis of the minor anticline, were observed on aerial photographs and on the ground. These features, as shown on Figure 17, are interpreted as fracture traces in the bedrock associated with the formation of the anticlinal structure.

The distribution of landslides throughout the exposed section of the Bearpaw Formation along St. Mary River is such that there is no apparent preferential relationship between the incidence of slides and any particular part of the stratigraphic section. Bentonite seams, however, were observed in the toe area of most of the slides (Fig. 18), thus the presence of bentonite at the base of a steep undercut slope is one of the prime geological details contributing to slope instability. Many of the bentonite seams can be recognized on the weathered shale surface by the presence of a thin band of sulphate efflorescence or a row of sparse vegetation. Upon removal of the weathered material from the bentonite a zone of soft, moist, plastic shale is commonly found overlying the bentonite whereas below the bentonite seam the shale is hard, highly fractured, and at a much lower water content. These conditions suggest that groundwater, although minor in quantity, is actively discharging from the slopes and can be an effective agent in reducing the shear strength of the bentonites and immediately overlying shales. The presence of bentonite seams retards the downward movement of groundwater, thus the flow pattern within the slopes is complex and probably contains elements of both saturated and unsaturated flow.

In both the St. Mary River and South Saskatchewan River areas the texture, clay mineralogy, and general lithological composition of the Bearpaw Formation are similar, although bentonite layers are more abundant in the St. Mary River area. Both areas exist under essentially the same climatic conditions, thus the groundwater regimen in the two areas is similar. In spite of these similarities there is a marked difference in slope behaviour between the two areas.

Along the South Saskatchewan River slope failures are continuous over many miles and generally occur wherever erosion of the overlying till has exposed shale of the Bearpaw Formation, regardless of the depth of erosion. Along St. Mary River, on the other hand, slope failures are discontinuous, and apparently stable slopes of the Bearpaw Formation with inclinations of 1.2:1 exist at heights of up to 200 feet. A comparison of the two areas suggests that differences in slope behaviour can be attributed to significant differences in the geological history of the two areas.

On the west side of Blood Indian Reserve, 16 miles west of the St. Mary River section, approximately 2,300 feet of younger Upper Cretaceous and Tertiary strata overlie the Bearpaw Formation. It is probable that in early Tertiary time a comparable thickness of younger strata covered the Bearpaw Formation in the St. Mary River area.

Although Bearpaw strata in the St. Mary River area have been subjected to greater preglacial overconsolidation loads than the strata in the South Saskatchewan River valley the loading was not sufficient to produce permanent cohesive bonding in the shale, which would prevent its disaggregation in the presence of water.

Uplift of the Sweetgrass arch during the early Tertiary produced westward dips of 2 to 5 degrees in the Bearpaw Formation and other strata of the St. Mary River area that occur in the western flank of the arch. The uplift also initiated a lengthy period of preglacial erosion, during which time the Bearpaw and younger strata were removed from the central part of the arch. As a result of the inclination of Bearpaw strata the entire formation in the St. Mary River area is exposed in a band 4 miles or less in width as compared with the extensive areal distribution of the formation over Saskatchewan occasioned by the very low dip of the formation in that area (Fig. 2).

Erosion of the sedimentary strata across Western Canada prior to the advent of glaciation undoubtedly produced a near-surface zone in which relief of stresses due to geological loading occurred. The depth and configuration of the zone of stress relief is unknown but it probably closely reflected the topographic surface with depths varying from a minimum under topographic highs to maximum in valley areas. In an area of uniform topography and dipping strata such as the St. Mary River area the zone of stress would extend over the entire formation. In contrast, in an area of similar topography but essentially horizontal strata such as the South Saskatchewan River valley only the upper beds would occur within the zone of stress relief and high residual stresses could exist in older beds of the same formation.

According to Stalker (1963) the Blood Indian Reserve area, which lies immediately west of the St. Mary River, was subjected to several stages of glaciation during the Pleistocene Epoch. Both Cordilleran and Laurentide glaciers covered the area, but the Cordilleran glaciers were near the limits of their advance in this area; consequently deposits of the Laurentide glaciers are more prominent. The probable maximum thickness of Laurentide glaciers that covered the St. Mary River area is approximately 2,200 feet as given by Stalker (1965) for the maximum ice thickness at Lethbridge, Alberta, 10 miles north of the St. Mary River section. The maximum ice thickness is almost equivalent to the thickness of post-Bearpaw sediments. However, the unit weight of glacier ice is about one half the unit weight of consolidated sediments, thus the load imposed by the glacier on Bearpaw strata would not exceed the preconsolidation load imposed by the sediments. Conversely the effects of glacial unloading would be subordinate to the effects of preglacial erosion.

Erosion of the St. Mary River valley was initiated through the drainage of preglacial lakes that formed during the recessional phases of the last glacier that covered the area. The meandering course of St. Mary River and the presence of well developed terraces on the slip-off slopes within the valley suggest a low initial gradient for the river and that a gradual lowering of base-level occurred throughout the valley development. Under these conditions rate of relief of any residual stresses in the Bearpaw Formation could be compatible with the rate of downcutting and consequently stress concentrations that could contribute to slope instability were not extensively developed. In addition the base level of erosion of the valley may be within the zone of stress relief, thus higher residual stresses, which could be present at greater depths, have not been encountered by erosion as appears to have been the case in the South Saskatchewan River valley.

Although the quantitative differences between the stress history of the St. Mary River and South Saskatchewan River areas cannot be specified the differences in geological history between these areas are of sufficient magnitude to suggest that variations in slope behaviour of the Bearpaw Formation are related to the stress history of the area.

Red Deer River, Alberta

Strata of the Bearpaw Formation are well exposed along the valley of Red Deer River from the mouth of Willow Creek,tp. 28, rge. 18, W. 4th mer. southeast to beyond Finnegan,tp. 25, rge. 15, W. 4th mer. (Fig. 19). Landslides have occurred at various places throughout this reach of the valley, but the development of slope failures, in spite of the extensive exposures of the Bearpaw Formation, is appreciably less along the main valley than along some of the tributary valleys. In comparison with the South Saskatchewan and St. Mary River valleys the Red Deer Valley is relatively lacking in landslide development, thus this area provided an excellent location for a comparative study of the geological factors affecting slope stability.

Throughout the study area the valley of Red Deer River follows a relatively straight southeasterly course and is entrenched from 300 to 400 feet below the upland surface. The width of the valley at the upland surface varies from 4,000 to 7,500 feet, but the width of the floor of the valley in the study area is essentially constant at about 2,000 feet. As a consequence of the variations in width of the top of the valley and the constant width of the valley floor the inclinations of the valley slopes range from 7:1 to 1.5:1.

The steeper slopes of the valley have been subjected to rapid run-off and erosion with the consequent development of badland topography. Terraces are also well developed along the valley, but in many places the terraces have been deeply dissected by stream erosion and are thus either obscured or obliterated.

The stream channel of Red Deer River is approximately 400 feet wide and is entrenched within the alluvial sediments that mantle the valley floor.

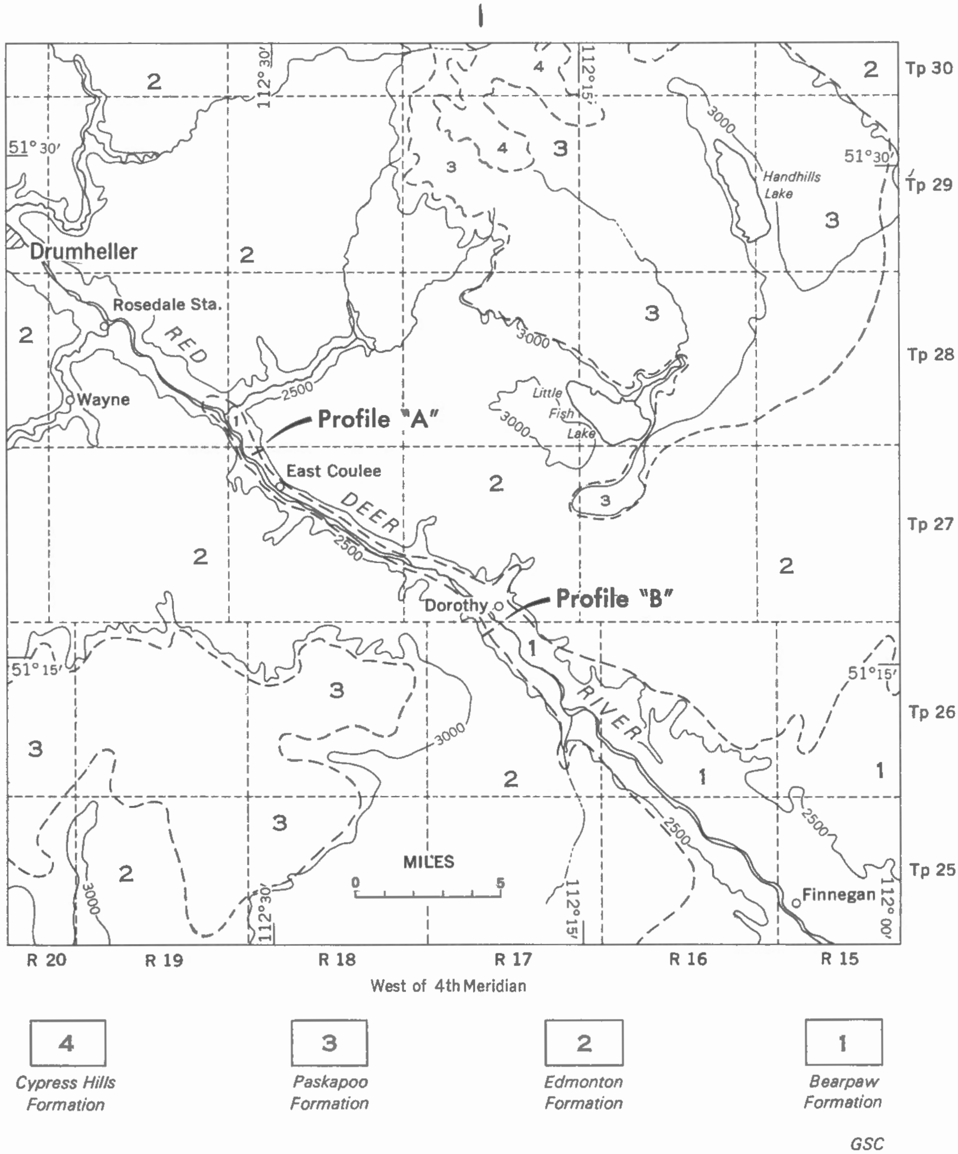


Figure 19. Bedrock geology of part of Red Deer area.

Plate III. Graben structure in landslide area on south side of Red Deer River at Dorothy, Alberta. (GSC 118541)

The course of the channel is characterized by gentle open meanders that sweep across the entire width of the valley floor, thus the undercut slopes of the meanders commonly impinge against the valley walls.

Between the mouth of Willow Creek (sec. 7, tp. 28, rge. 18, W. 4th mer.) and the village of Dorothy (sec. 3, tp. 27, rge. 17, W. 4th mer.) only one area of landslides involving Bearpaw strata was observed along the main valley of Red Deer River. The failure area extends for approximately 1 mile along the north wall of the valley in sec. 23, tp. 27, rge. 18, W. 4th mer. Rotational slumping has occurred in the head region of the slide, but in the lower part of the slide the geometry of the failure surface is obscured by slope wash deposits.

Several of the large tributary valleys to the south of the main valley between Willow Creek and Dorothy contain areas of landslides; however, these slides have occurred entirely in strata overlying the Bearpaw Formation.

Southeast along Red Deer Valley from Dorothy to Finnegan (sec. 18, tp. 25, rge. 15, W. 4th mer.) landslides in the Bearpaw Formation are more prevalent, but are not extensively developed. Some of the slides display well developed graben structures in the head region of the slide (Plate III), indicative of lateral spreading along an essentially horizontal failure surface. Other slides are of the retrogressive slump-type with pronounced backward rotation of the failure blocks in the head region. In the toe region of these slides, however, the failure surface appears to be tangent to a horizontal plane.

The most extensive development of retrogressive rotational slump failures has occurred in the tributary valleys to Red Deer River. One of the best examples of slope failures along tributary valleys occurs in Circus Coulee on the north side of Red Deer Valley at Dorothy (Plate IV) where the valley has been abruptly widened from about 1,300 feet to 3,500 feet by retrogressive rotational slumping.

The Bearpaw Formation in the Red Deer Valley has a thickness of about 550 feet according to Allan and Sanderson (1945) and contains lithologies similar to those found in the South Saskatchewan River and St. Mary River sections. The shale and sandstone components of the Bearpaw Formation have not been designated as separate members in the Red Deer Valley section. Williams and Dyer (1930) have suggested, however, that an intermixture of sandstone and shale near the middle of the Bearpaw section on Red Deer River may represent a southward extension of the Bulwark sandstone that is recognized as a distinctive member of the Bearpaw Formation in areas to the north.

Dark grey and dark brown fissile clay shale containing abundant crystalline gypsum is characteristic of most of the formation. Sandy shales and weakly cemented fine-grained sandstone layers are interbedded with the shales and are most abundant in the upper 100 feet of the formation where they form a transition into the overlying sandstone beds of the Edmonton Formation (Fig. 20).

Layers of bentonite, consisting primarily of montmorillonite, as determined by X-ray diffraction analyses, are present at various levels throughout the shale sequence. The bentonite layers are generally only a few inches thick, but the section at Dorothy exposes a layer of impure bentonite approximately 20 feet thick (Fig. 20). The weathered surface of the impure bentonite is characterized by an intricate pattern of polygonal desiccation cracks and a pale grey colour, which provides a sharp colour contrast with the dark shales of the surrounding area.

Nodular calcareous concretions in thin layers are also present at various levels throughout the Bearpaw Formation in the Red Deer River valley.

Strata of the Bearpaw Formation dip west into the Alberta syncline at an inclination of approximately 7 feet per mile as shown on the cross-sections presented by Allan and Sanderson (1945).

Approximately 1,500 feet of younger sediments of Upper Cretaceous and Tertiary age overlie the Bearpaw Formation in the Hand Hills to the north of Red Deer Valley (Fig. 19). The younger rocks consist of 1,050 feet of

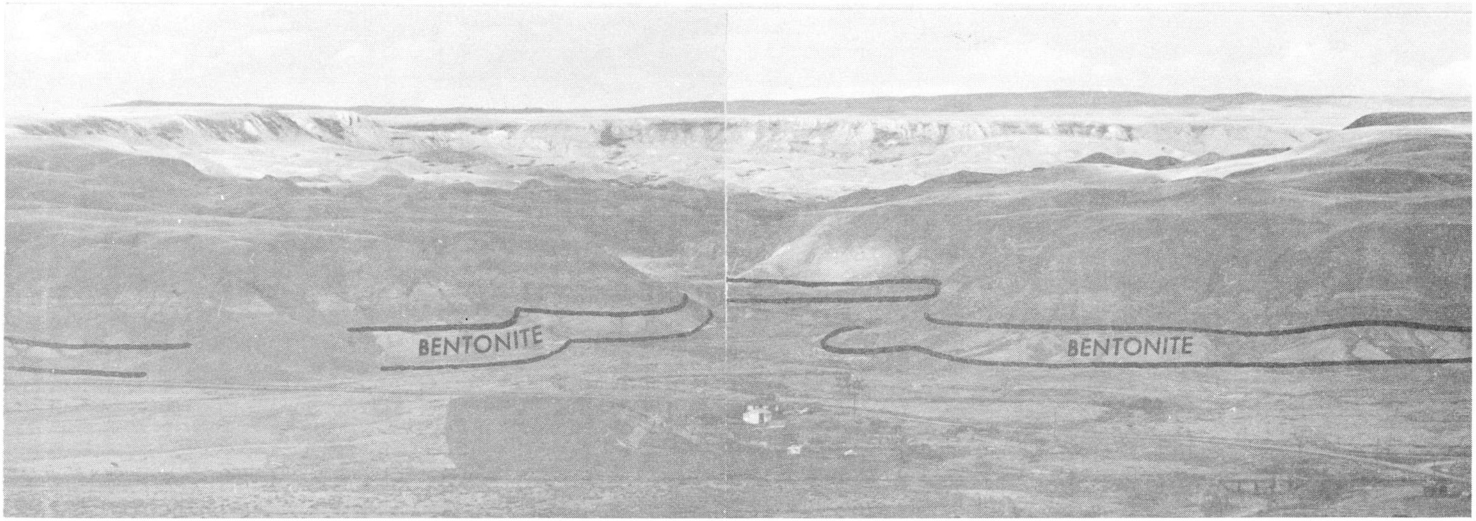


Plate IV. Slump area in tributary valley, north of Dorothy, Alberta.
(GSC 118607, 118608)

sandstone, shale, coal, and bentonite beds of the Edmonton Formation, which is unconformably overlain by approximately 360 feet of clayey sandstone and clay shales of the Early Tertiary Paskapoo Formation. These formations are capped by about 75 feet of loosely cemented conglomerate, sand, and marl of the Cypress Hills Formation, which is probably Oligocene in age according to Allan and Sanderson (1945).

The presence of most of these younger strata also in the Wintering Hills to the south of Red Deer Valley (Fig. 19) indicates that prior to the period of extensive preglacial erosion the Bearpaw strata in the Red Deer Valley were overconsolidated by at least 1,500 feet of younger sediments. The exact thickness of sedimentary rock strata removed from the Red Deer Valley prior to glaciation is unknown, but the bedrock topography of the area suggests a thickness in the order of 1,000 feet.

The probable maximum thickness of glacier ice that covered the area during the Pleistocene Epoch is estimated, on the basis of data reported by Stalker (1965), at 4,000 feet. On the basis of the unit weight of ice being approximately one half that of sedimentary rock it is therefore apparent that glacial loading of Bearpaw strata exceeded the load imposed by superincumbent sediments.

Although erosion of the valley of Red Deer River in the area east of Drumheller, Alberta was initiated during an interglacial stage (Stalker, 1961) the main downcutting of the valley occurred during post-glacial time. Changes in the base level of erosion occurred during the development of the valley, as evidenced by the presence of well defined terrace levels (Plate V). Some of the upper terraces, which can be correlated with terraces along Crawling Valley in tps. 25, 26, rge. 17, W. 4th mer., were developed by ice-controlled drainage of glacial lake Drumheller, which occupied the lowland west of Hand Hills (Craig, 1956).

Tributaries to Red Deer River in the area between Willow Creek and Dorothy, however, have developed by headward erosion following development of the main valley. The tributaries, therefore, are characterized by steep valley walls and gradients and the absence of terraces.

No evidence of groundwater seepage was observed in the sections of the Bearpaw Formation examined in the Red Deer Valley. Because of the low permeability of the shales and the relatively low amounts of precipitation that occur over the area only minor amounts of groundwater flow through the shales could be expected. In parts of the Oyen map-area, approximately 25 miles to the east of Fennigan, the Bulwark sandstone member of the Bearpaw Formation yields water to wells at the rate of 5 to 10 gallons per minute (Kunkle, 1962). Recharge to the sandstone aquifer occurs through overlying shales of the Bearpaw Formation and the rate of groundwater movement through the shales as calculated by Kunkle (1962, p. 11) is 0.035 g.p.d./ft.^{2}.

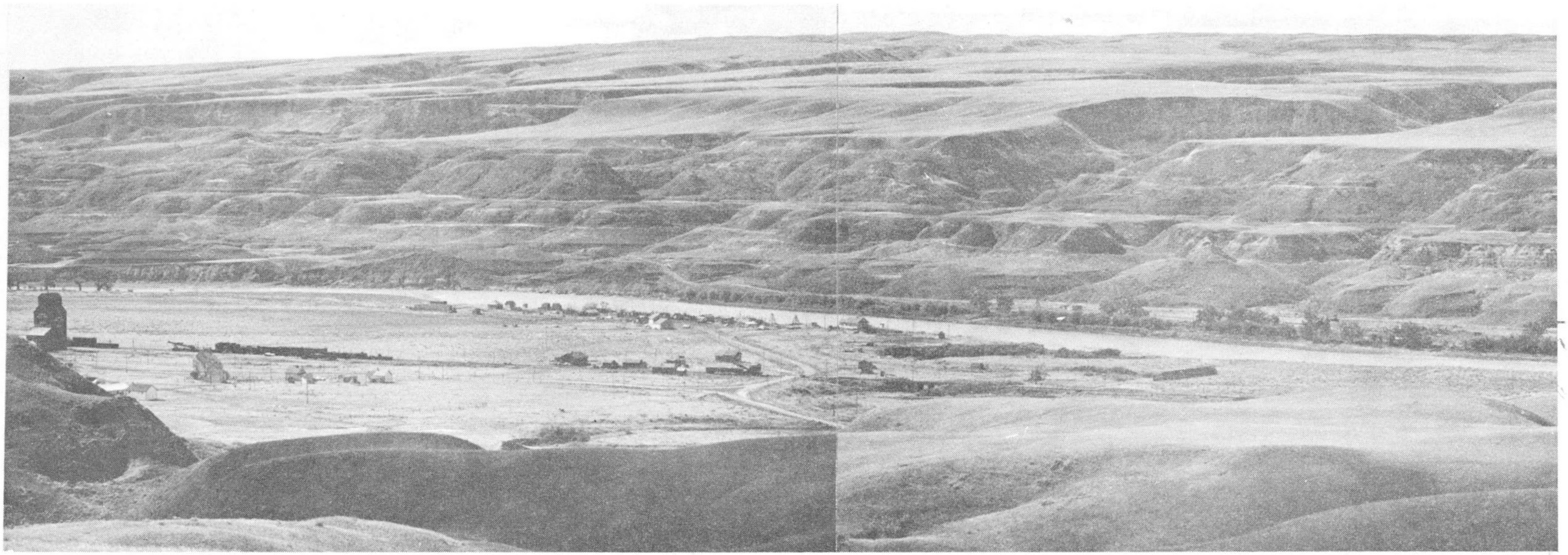


Plate V. Terraces along Red Deer River at Dorothy, Alberta.
(GSC 118605, 118606)

The lithological characteristics of the Bearpaw Formation in the Red Deer Valley and the hydrogeological regimen of this area are essentially similar to those conditions that prevail in the St. Mary River and South Saskatchewan River areas. The three areas, however, show marked contrasts in the engineering behaviour of Bearpaw slopes as a result of variations in the type and magnitude of geological factors that have prevailed throughout the geological history of each of the areas.

The magnitude of the residual stresses that are now present in the Bearpaw Formation in the various areas and the configuration of the near-surface zone of stress relief are unknown. However, the geological history of sedimentation, erosion, and glaciation, which is common to all of the areas, suggests that each of the areas has been subjected to geological loads that are of the same order of magnitude but not necessarily of the same origin. In the St. Mary River area the sediment load exceeded the glacial load whereas in the Red Deer River and South Saskatchewan River areas glacial loading appears to be of greater magnitude. If the stress history was the only significant factor then it might be expected that the engineering behaviour of Bearpaw slopes in the South Saskatchewan River and Red Deer River areas would be similar. Tributary valleys along Red Deer River, e.g. Circus Coulee, display similar landslide occurrences to those found along the South Saskatchewan River, but the relatively stable slopes of the main valley of Red Deer River contrast sharply with the extensive areas of slope failures along the South Saskatchewan River.

The most apparent geological factor contributing to the stability of slopes along the main valley of Red Deer River is the extensive development of terraces. The terracing by river erosion is analogous to the multiple benching of cut slopes, which increases slope stability by reducing the shearing stresses that would exist in an unbenched slope of the same height. The greater prevalence of slope failures in the unterraced tributary valley walls indicates the presence of greater shear stresses in these slopes than exists in slopes along the main valley.

Although the rate of downcutting of the Red Deer Valley is unknown the terraces indicate that downcutting occurred in several stages rather than as a continuous process. Under such conditions it would have been possible for residual stresses in the valley walls to be relieved at a rate compatible with the erosion, thus reducing residual stress concentrations that have contributed to failures in the South Saskatchewan River valley and in the tributary valleys of Red Deer River.

Although the extent of the near surface zone of stress relief in the Red Deer Valley is unknown it is possible that as a result of the combination of westward dip of Bearpaw strata and eastward flow of Red Deer River the entire formation is exposed within the zone of stress relief. Higher residual stress levels may exist within the formation, but at depths below the present base level of erosion of the river.

Where landslides have occurred in the Red Deer Valley area the geometry of the failures suggests that failure surfaces are, in part, tangent to a horizontal plane. The most probable locus of failure surfaces is, therefore, along bentonite layers as has been found in other areas examined in the landslide study. In the Circus Coulee failure north of Dorothy the 20-foot layer of impure bentonite occurs at a level that is coincident with the lower part of the failure zone (Plate IV).

The differences in behaviour of slopes along Red Deer Valley as compared with both the South Saskatchewan River and St. Mary River areas provides further indication of the significance of the geological history of an area in assessing the stability of slopes in overconsolidated shales.

INFLUENCE OF GEOLOGICAL FACTORS ON ENGINEERING BEHAVIOUR

The detailed geological data already described has been resolved into seven factors as indicated in Table II. This table suggests the physical action of each factor, along with the influence it has on the engineering behaviour of overconsolidated clay shale slopes.

The action of geological processes is to cause variations either in shear stresses (Fig. 21a) or in shear strength (Fig. 21b) (Terzaghi, 1949). In this connection the action of seasonal variations in piezometric level may be examined. Because seasonal variations are a normal part of the life of a slope it is not believed that this factor alone can cause a failure. However, when the influence of seasonal variations is superimposed upon the other active processes it may finally trigger a slope failure. The precise nature of the combined effects of the seven geological factors involved in a failure is indeed complex. Although the factor of safety when failure occurs must be unity, it is doubtful that we can consider a constant value of either shear strength or stress, while the other varies. It is more probable that the combined action of all the factors results in decreasing shear resistance and progressive decrease in the factor of safety (Fig. 21c).

The complexity of the combined action involved in a slope failure may be illustrated by the following example. A typical landslide area along the St. Mary River in southern Alberta is shown in Figure 17. The cross-section of this area is shown in Figure 18.

The primary factor identified with this failure was toe erosion. Yet outcrops of bentonite seams were apparent at several elevations. At each outcrop deposits of salts leached from the natural ground by percolating water were observed. As a consequence the following actions may be described, all of which are in some degree responsible for the final failure.

1. Increased shear stresses due to removal of toe load by erosion.

2. As a result of toe erosion and valley deepening there is a net reduction in confining pressures. This induces negative pore-water pressures, approximating in magnitude the stress reduction, which dissipate with time resulting in increases of moisture content and consequent decreases in shear resistance. This constitutes, in fact, rebound or swelling.

3. Due to the lowered confining stress and time-diminishing negative pore-water pressures the coefficient of earth pressure at rest (K_o) increases. The variation of K_o with stress history as defined by the over-consolidation ratio is illustrated in Figure 22 (Brooker and Ireland, 1964). The maximum value of K_o is K_p, the coefficient of passive earth pressure, which represents a limiting case of plastic equilibrium according to the Mohr-Coulomb failure criterion:

$$\sigma'_1 = \sigma'_3 \tan^2 (45 + \frac{\phi'}{2}) + 2c \tan (45 + \frac{\phi'}{2})$$

where

σ'_1 = major principal effective stress

σ'_3 = minor principal effective stress

c = effective cohesion

ϕ' = effective angle of shearing resistance

The development of these stresses has a definite, but rather indeterminate influence on stability. The indeterminate nature of this factor results because the complete stress history of the Upper Cretaceous soils is not clearly defined because the preconsolidation load is only approximately known. Therefore, the value of the overconsolidation ratio (OCR) can only be roughly estimated. Nevertheless, the value of OCR will be high and as a result the value of K_o will also be large. If the ground remains intact the influence of this lateral force should be included in a stability analysis. This can be conveniently done by the methods described by Bishop (1955), Bishop and Morgenstern (1960). The result of this stress history may manifest itself as the development of joints and slickensides (Skempton, 1961; Terzaghi, 1961). In this event new paths for percolating groundwater may be formed. Therefore the processes described below may be accelerated.

4. During the period when the above factors are active an increase of moisture content occurs. At the same time, along pervious strata percolation of groundwater is occurring. Both of these actions cause migration of fresh, relatively salt-free water into the soil and, as a consequence, the leaching of natural salts from the soil. Although the mechanism of changes in the chemistry of the pore water is not clearly understood the leaching action results in diminished salt concentration of the free pore water. Van't Hoff's equation (Yong and Warkentin, 1965)

$$P = RT (C_c - C_o)$$

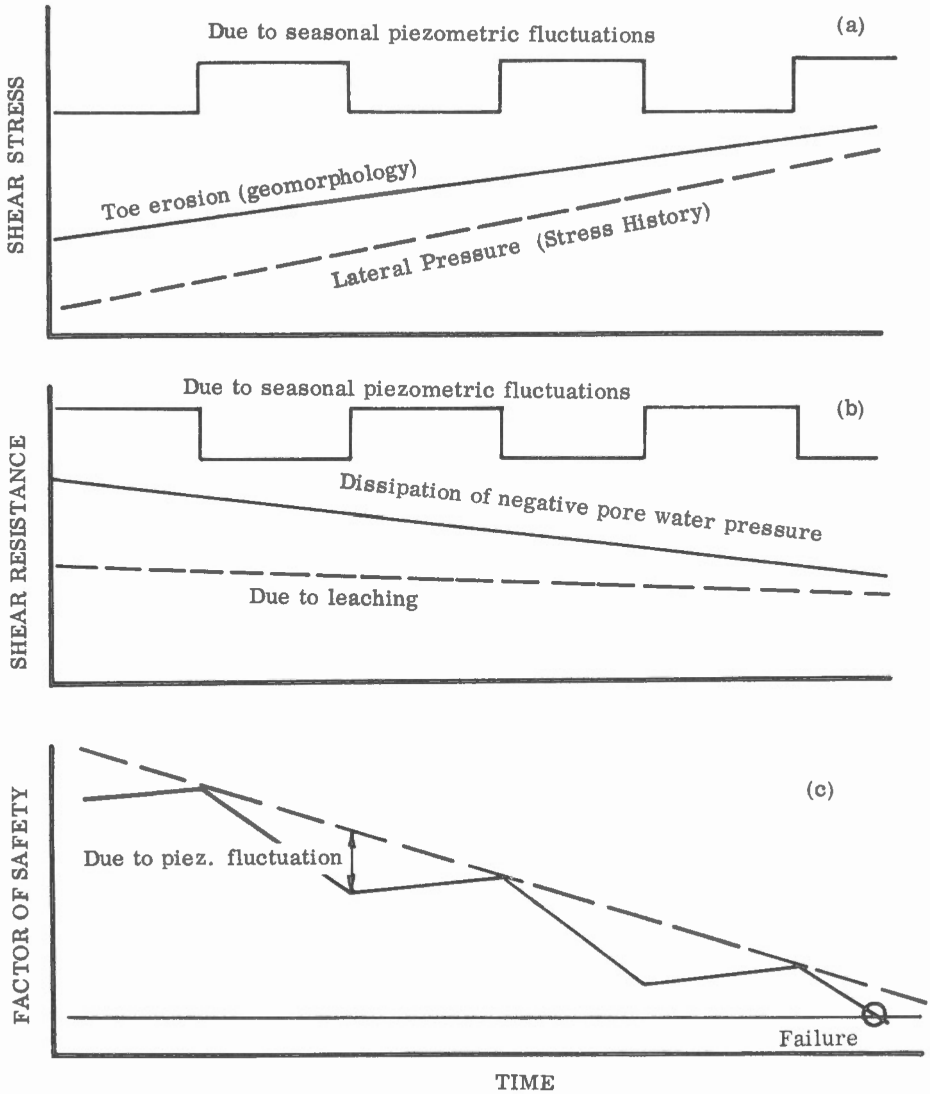
describes the osmotic pressure (P) resulting from a difference in salt concentration between diffused double layers of the soil particles C_c and the salt concentration of the pore-water fluid C_o. R is the general gas constant and T is the absolute temperature. This equation suggests that as leaching of natural salts occurs, causing C_o to decrease, the repulsive forces between clay particles increases and the volume must increase. The volume increase will be accompanied by a decrease in shearing resistance. The physical manifestation of this action may be an entirely new soil water system with different values of the parameters c' and ϕ' in Coulomb's equation of strength

$$S = c' + \sigma' \tan \phi'.$$

TABLE II

Geological Factors Controlling Slope Stability and their Engineering Consequence

GEOLOGICAL FACTORS		INFLUENCE	ENGINEERING CONSEQUENCE
1. Depositional Environment	(a) Marine deposition.	High sodium conc. in pore fluid.	High osmotic swelling potential.
	(b) Fine-grained clastic sediments and volcanic ash.	Clay sizes dominant, montmorillonite abundant.	High plasticity soils.
	(c) Slow rate of sedimentation.	Sediments generally of uniform texture and structure.	Low permeability.
	(d) Relatively shallow depth of burial.	Lithification incomplete, interparticle bonds weak.	Low shear strength, high rehydration swelling potential.
2. Lithology and Stratigraphy	(a) Bentonite layers interstratified with clay shales.	Retards downward movement of groundwater; leaching along top of bentonite layers.	Zones of high plasticity, high swelling pressure, low shear strength.
	(b) Fine-grained sandstone layers widely spaced.	Drainage layers widely spaced.	Layers of relatively high shear strength related to distribution of sandstone.
3. Stress History	(a) Loading by younger sediments.	Consolidation.	Increase in shear strength with loading.
	(b) Diastrophism and preglacial erosion.	Removal from marine environment, leaching, dilatancy, sediments in condition of overconsolidation.	Residual stresses relieved only in near surface zone.
	(c) Glacial erosion and loading.	In some areas consolidation loads may exceed those imposed by sediments.	May cause increase in O.C.R.
	(d) Glacial unloading.	Initiation of residual stress relief; sediments in condition of over-consolidation.	Residual stress concentrations.
	(e) Postglacial rebound.	Terrain uplift – erosion base level lowering, increase in groundwater gradients.	Stream erosion accelerated, downcutting exceeds rate of residual stress relief.
	(f) Valley erosion.	Exposure of zone of high residual stress. Fracture development parallel with the valley.	High lateral stresses in valley walls.
4. Structure	(a) Attitude of bedding.	Controls outcrop width of formation.	Older beds of formation may occur below near-surface zone of stress relief.
	(b) Fracture development.	Vertical planes of weakness.	Loss in mass shear strength.
5. Climate	(a) Precipitation.	Affects rate of groundwater movement and leaching.	Seasonal pore pressure fluctuations leaching may decrease shear strength.
	(b) Temperature.	Alternate wetting and drying produces fractures. Freeze/thaw action in fractures.	Decrease in shear strength in near-surface fracture zone.
6. Geomorphology	(a) Position of base level.	Controls depth of valley and in part, rate of erosion.	Failures occur only where base level of erosion is below critical height of slope. Critical height, however, is not constant.
	(b) Stream channel configuration.	Asymmetrical cross valley profile, steep undercut slopes, toe erosion.	Decrease in shearing resistance of slope.
	(c) Terrace development.	Reduction in effective slope height.	Decrease in shear stresses.
	(d) Rate of erosion.	Rate of downcutting may exceed rate of residual stress relief in valley walls.	Increase in shear stresses.
	(e) Slope exposure.	Exposure to insolation may depress groundwater flow regimen.	Increase in effective stress due to decrease in neutral stresses.
7. Groundwater	(a) Quantity.	Seasonal variations in flow may increase leaching and pore pressures.	Decrease in shear resistance.
	(b) Quality.	Differences between groundwater chemistry and pore-water chemistry can create osmotic swelling pressures.	Decrease in shear resistance.



INFLUENCE OF GEOLOGIC FACTORS ON
(a) Shearing Stress in a Soil Mass
(b) Shearing Resistance of the Mass
(c) Factor of Safety against Failure

Figure 21. Variation in shear strength, shear stress and factor of safety with time.

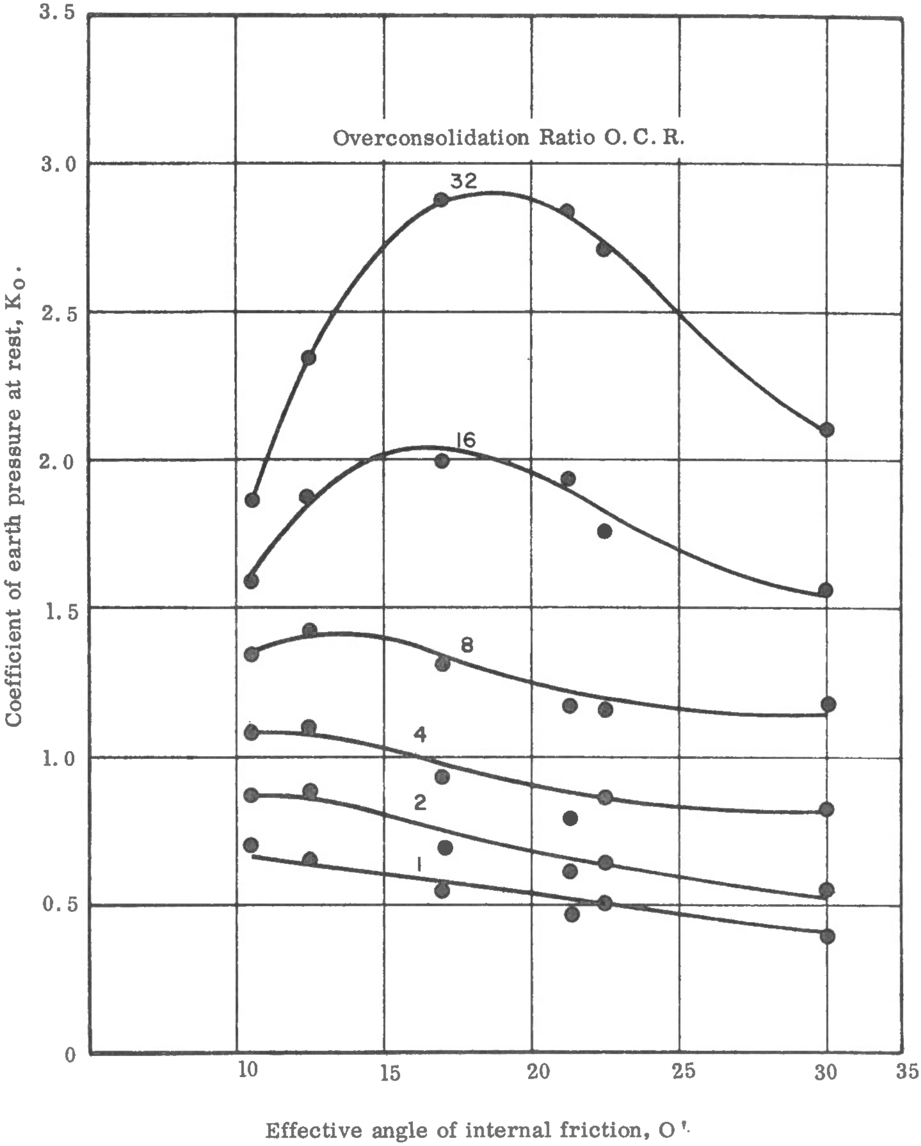


Figure 22. Relationship between coefficient of earth pressure, over-consolidation ratio and angle of shearing resistance.

The initiation of a slide may be caused by a number of processes. Toe erosion, leading to oversteepening and increased shear stresses has been well established by the present study. Continued downcutting by river action may cause a slope to exceed the critical height consistent with the slope angle. The field slope chart approach to regional stability problems (Lane, 1961) has demonstrated the well known fact that for a given slope angle there is a critical maximum height for slope. A typical relationship between slope height and slope cotangent developed by Lane is illustrated in Figure 23. This particular chart was developed for an area in South Dakota. It was found during this present study that no such chart could be developed for the Bearpaw-Shale over the Alberta-Saskatchewan area. However, if restricted to a more finite area it might prove to be a functional approach.

The Prediction of Slope Stability

The foregoing discussion indicates that prediction of stability of natural slopes is most difficult. However, the needs of modern projects demand a simple rational approach. This approach is provided by combining the knowledge of soil behaviour derived from soil mechanics with geologic information, which defines not only the complexity of a given region but also the significance of the factors involved. When these aspects are considered relative to cases of known behaviour then a reasonable interpretation and predication may be made for similar circumstances in other regions (Peck, 1962). Clearly then, the prediction of stability is an art which can only be learned through application.

The following paragraphs summarize concepts which have been developed from observations and laboratory evidence. In some instances the evidence was not obtained from this study, but the findings are believed to be applicable to it. The concept of residual strength was developed by Skempton (1964) and appears to provide the most rational means of assessing the shear resistance of clay shales. The method of stability analysis proposed herein has proven rational in a number of cases in Western Canada, but not within the study areas. These concepts will be applied and the results reported as accumulating evidence permits. There appears to be substantial evidence to support the introduction of these concepts into practice at this time, concerning analysis of slopes (R. Peterson and D. J. Henkel, personal communication, 1966).

Geometry of Failure Surface

Wherever definite landslide topography occurred in the study area the impression was gained that the scalloped topography is a result of a series of retrogressive slides. In the Alberta area this retrogressive action of the slides has often been observed and several cases of this nature have been reported in this study.

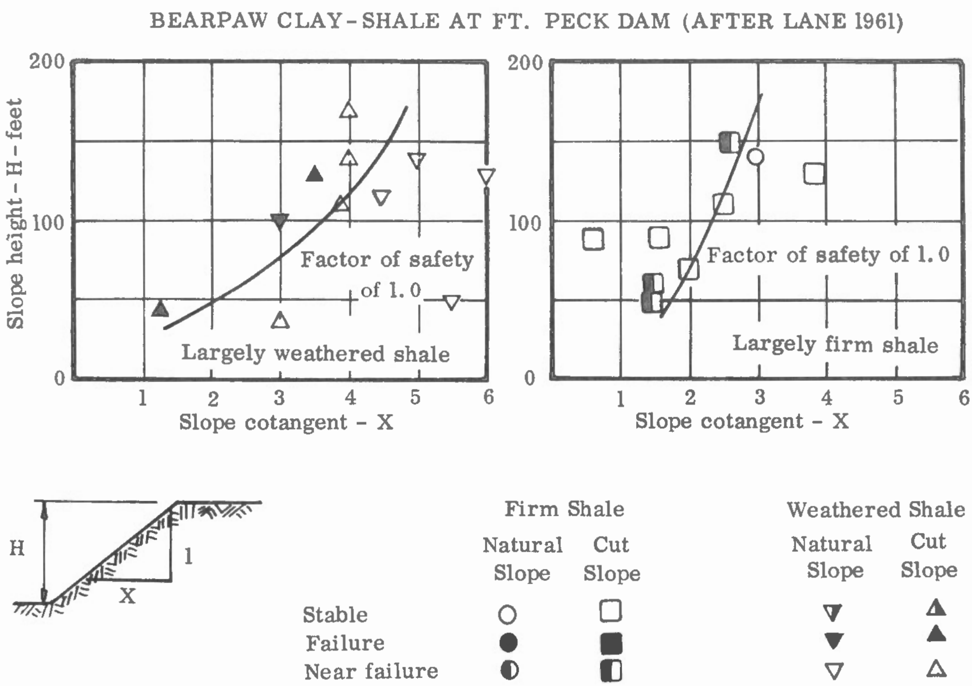


Figure 23. Relationship between slope height and slope cotangent for Bearpaw shale (Lane 1961)

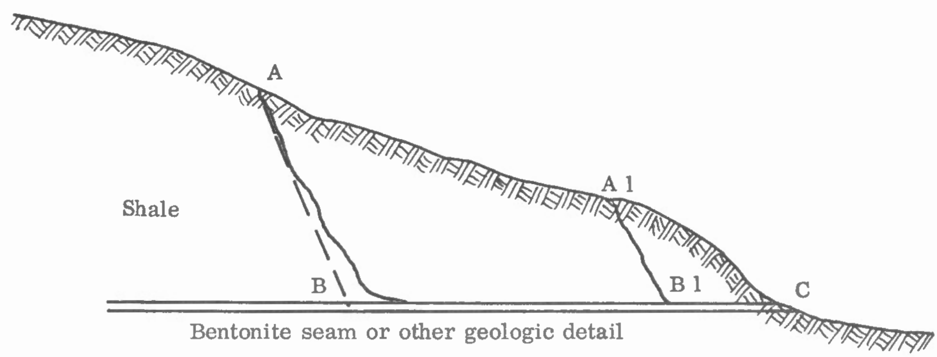


Figure 24. Approximate geometry of failure surface in Cretaceous formations in western Canada.

In most cases where subsurface information is available, and the geometry of the slide is reliably inferred, the surface of failure is composed of two portions (Fig. 24). The portion AB is a flat concave-up arc, connecting with the portion BC. The surface BC is nearly horizontal and is often seated in a relatively weaker material such as bentonite or montmorillonitic clay. The slide may commence on the surface A1 B1 and retrogress to some other surface AB. Retrogression ceases when the overall slope angle becomes sufficiently small. The occurrence of landslides that are best approximated by this type of failure surface appears to be much more frequent than slides that approximate a circular arc. This clearly establishes the dominant influence of geologic detail.

In order to facilitate the analysis of slides of this nature the failure surface may be approximated by the surface A B C (Fig. 24). The knowledge of the geometry is of importance and must, therefore, be carefully determined. In those cases where movements are occurring, or suspected, the installation of slope movement indicators is recommended. The slope movement indicator is a precise device for measuring the slope of a special borehole casing. Integration of slope measurements taken at specified intervals yields slope movements. These procedures are now well developed and the value of this technique proven (Wilson, 1962).

Method of Analysis

The stability analysis, per se, is a tool of statics that is an aid to the prediction of stability. Many forms of stability analysis exist, each one suitable under a certain set of circumstances. The circular arc analysis, in terms of effective stresses developed by Bishop (1955) is very useful in some cases. This method is a refined method of slices utilizing a procedure of successive approximations. The method considers the moment equilibrium of a soil mass, with each slice placed in vertical and horizontal equilibrium. It is most suited to the stability analysis of earth dams because of the geometric restriction to a circular arc failure surface.

Most natural slope failures in stratified shales are not of a circular arc geometry. Instead, the failures may often have their seat in thin seams of bentonitic material or other geologic discontinuity. Moreover, it is seldom that a huge mass of earth slips at once. Rather, the landslide may be retrogressive in nature. Due to this fact it seems more reasonable to use a method of analysis that is based primarily upon the equilibrium of horizontal forces without restrictions to a circular failure surface. The wedge, sliding block, and modified Swedish analyses satisfy this requirement to various extents. What appears to be a very useful method of analysis for the purposes of valley slope stability in a heterogeneous cross-section is the wedge method of stability analysis. This method considers the statical equilibrium of adjacent soil masses. Based on the observations reported the following method of analysis is recommended.

The probable surface of failure typical of landslides is shown in Figure 24 by the plane A B C. The wedge with all the forces acting on it is shown in Figure 25a. The neutral force vectors are obtained from estimated or measured piezometric levels. For convenience the surface of failure is assumed to be composed of two distinct parts, A D and D C. The corresponding soil masses are termed the uphill or active wedge and downhill or passive wedge, respectively (Fig. 25a). Vector diagrams for each wedge are shown in Figure 25b and c. The value of the lateral forces E_1 and E_2 on the wedges is unknown, but the direction of these forces on the wedges may be assumed. For an assumed factor of safety the forces E_1 and E_2 may be found by solution of the appropriate vector diagrams. The correct value of safety factor will be that for which the values of E_1 and E_2 are equal. This is a compatibility requirement for the plane E D. By plotting the values of E_1 and E_2 against the factor of safety as in Figure 25d, the correct value of FS may be determined for the assumed failure surface.

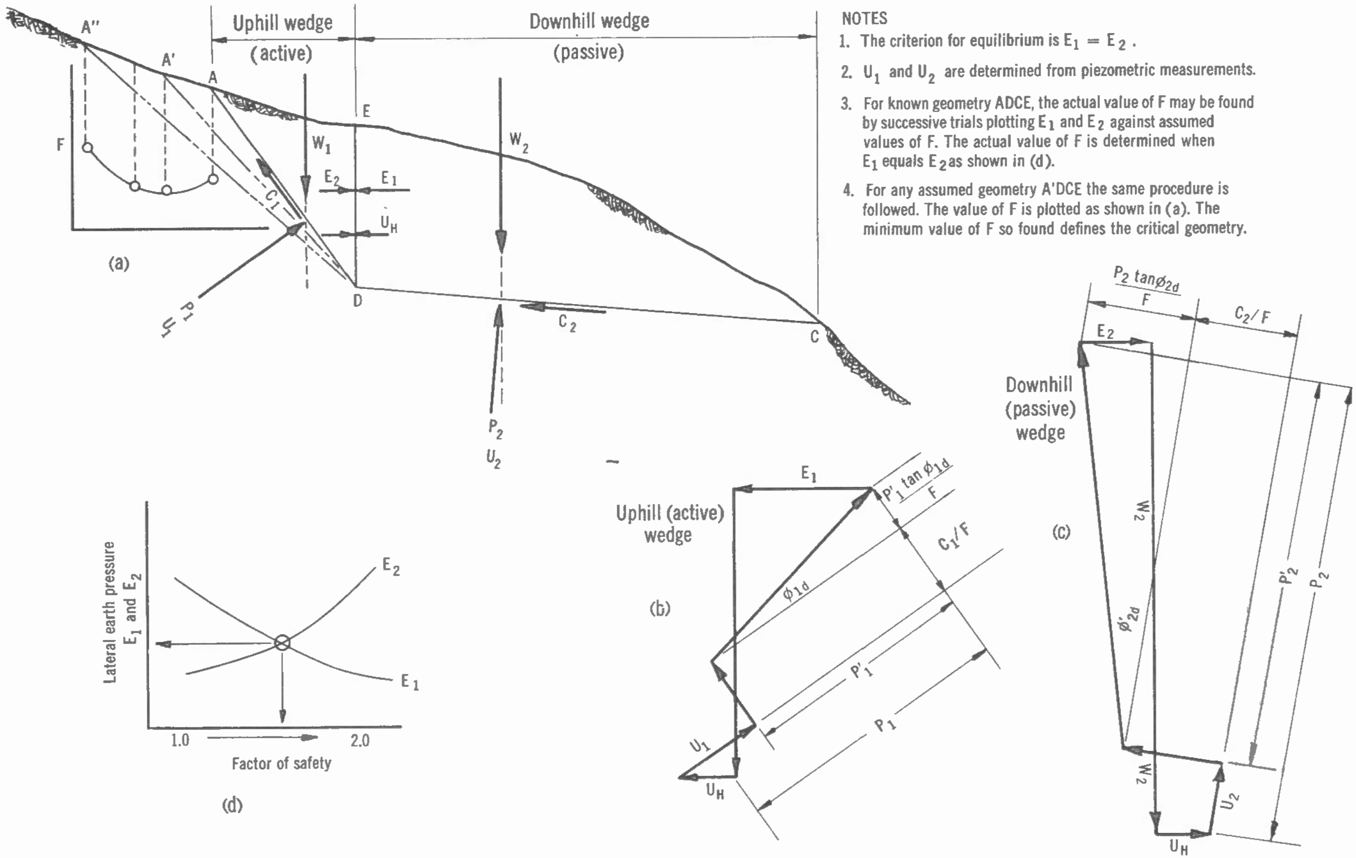
In the case of stability assessments the probable geometry of the segment D C may be more readily approximated than the segment A D. Thus it may be advisable to examine various attitudes of plane A D. The procedure suggested above is followed for each assumed surface A' D, A'' D, etc., and the value of FS found plotted beneath the point A in question. The most probable failure surface will correspond to the minimum factor of safety ($\overline{FS}$, $\overline{A}$, Fig. 25a). To date two cases, applying this method, have been assessed with a reasonable degree of agreement with observed behaviour.

The requirements for application of this method are; knowledge not only of the failure surface and corresponding pore-water pressures, but also of the correct strength parameters; the strength parameters in question are the effective cohesion C' and effective angle of shearing resistance ϕ'.

The Determination of Strength Parameters

The values of c' and ϕ' for use in analysis are usually determined in the laboratory by triaxial and, more recently, direct shear tests. These tests have appealing scientific aspects applied to clay shales. However, experience over the past 20 years has suggested that the value of triaxial test results is very limited (Peterson, 1960; Skempton, 1964). The real purpose of laboratory testing is to produce results that are applicable to field conditions. That is, laboratory testing must be intimately related to field phenomena.

Recent laboratory triaxial test results at the University of Alberta have shown that the results may not be descriptive of the soil mass behaviour. In the field the clay shales of Western Canada are fractured, jointed, and slickensided. Thus the soil mass behaviour is controlled by those discontinuities, because both the shearing resistance of the soil mass and the soil mass



NOTES

1. The criterion for equilibrium is $E_1 = E_2$.
2. U_1 and U_2 are determined from piezometric measurements.
3. For known geometry ADCE, the actual value of F may be found by successive trials plotting E_1 and E_2 against assumed values of F. The actual value of F is determined when E_1 equals E_2 as shown in (d).
4. For any assumed geometry A'DCE the same procedure is followed. The value of F is plotted as shown in (a). The minimum value of F so found defines the critical geometry.

Figure 25. Wedge method of stability analysis.

permeability will be a function of such geologic detail. Testing an intact specimen in the laboratory may, therefore, have little relationship to field conditions. In fact it has been found that it is necessary to test the soil over extended periods of time in order to achieve proper measurements of pore-water pressures. Inasmuch as the soil strength is greatly influenced by rate of testing, it seems desirable to utilize slow testing rates.

A test that shows greater possibilities is the drained direct shear test. While this test has been used since the time of Coulomb in 1776, it has recently gained in popularity with the introduction of the residual strength concept by Skempton (1964). The translation of adjacent planes under a normal stress relates to field conditions to a better degree than the mechanism of failure in the triaxial test. Tests on intact specimens display the characteristics shown in Figure 26. At first the intact soil is very rigid and carries substantial load without large strains. At some point the peak shear resistance is obtained whereupon continued strain results in a diminished and ultimately residual shear resistance. The shearing resistance is plotted against normal stress on the plane in a Mohr diagram. The slope of the rupture line through a series of peak strength points is the peak angle of shearing resistance ϕ'. The intercept of this rupture line on the shear axis is called the cohesion, c'. The rupture line drawn through residual strength points yields the residual angle of shearing resistance, ϕ'_R and the intercept of this line with the shear axis gives a corresponding value for the residual cohesion c'_R. Both the residual strength parameters are significantly less than the peak strength parameters.

The residual shear strength parameters describe the resistance to movement or surfaces that have been translated. In a real soil mass only a portion of a potential sliding surface will pass along such planes. A certain portion of the plane will pass through intact chunks of soil. As a result the actual resistance to shear along a potential sliding surface must lie between the peak and residual strengths. The ratio

$$R = \frac{\phi' - \phi'_A}{\phi' - \phi'_R}$$

where ϕ'_A = Actual effective angle of shearing resistance applicable in the field

ϕ'_R = Residual effective angle of shearing resistance

will be termed the residual strength index. The important unknown in stability analysis for prediction purposes is the value of R. Although this value is indeterminate the concept that the values ϕ' and $\phi'r$ exist provides an upper and a lower limit approach to stability prediction. Some insight into the value of R on a regional basis may be obtained through the analysis of failed slopes. However, to be of significance failure cases must be well instrumented so that

SHEAR CHARACTERISTICS OF OVER-CONSOLIDATED CLAY

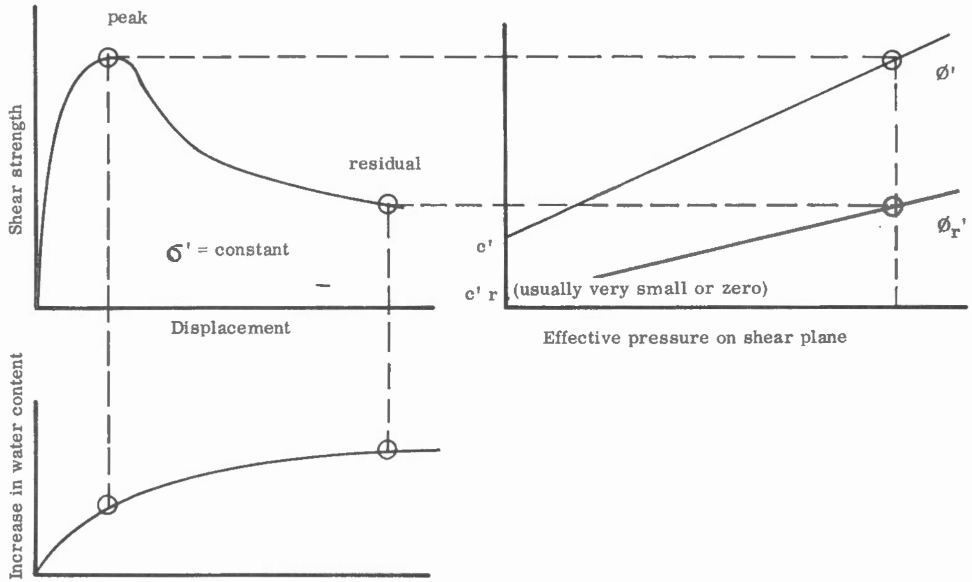


Figure 26. Illustration of direct shear test results.

the only unknown is the value of ϕ'_A. This necessitates the accurate measurement of pore-water pressures in the field. Hvorslev (1954) has set out criteria for the rational design of piezometers. A unit that meets the time lag requirements in clay shales was developed jointly by the Geological Survey of Canada and the University of Alberta (Brooker and Lindberg, 1965). The unit involves the use of a high sensitivity pressure transducer, placed in a probe with a surrounding expandable packer. This unit is illustrated in Figure 16.

The use of strength parameters, determined from laboratory direct shear tests, combined with observations of actual slides having complete piezometric information in a rational method of analyses should therefore yield useful results. Most of the analyses reported to date are deficient in at least one of these aspects.

CONCLUSIONS

From the foregoing it may be concluded that:

1. The stability of slopes in overconsolidated shales is governed by the complex interrelationship of geological and environmental factors that have prevailed throughout the history of both the slope and the slope-forming material. These factors and their influence on the stability of slopes are tabulated in Table II.

2. Sources of uncertainty in analyzing the performance of clay shales are created by geological details, which may remain undisclosed during normal subsurface investigations.

3. The geometry of the surface of failure in the clay shale deposits of Western Canada is best described by a number of plane surfaces. In order to assess slope stability it is important to have knowledge of this geometry. This can be accomplished by observations from slope movement indicator stations.

4. The wedge method of stability analyses appears to offer a promising approach to slope stability prediction in the Upper Cretaceous clays of Western Canada.

5. In order to obtain useful results from a stability analysis complete field measurements are needed. These include not only slope movement indicator stations, but also a sufficient number of compatible piezometer points to establish the phreatic surface and pattern of groundwater movement.

6. Strength parameters obtained from laboratory shear tests provide a measure of peak and residual strengths. These parameters are usefully applied in an upper and lower limit approach to the problem. The values are most realistically determined by direct shear tests. The Residual Strength Index, R, is an important unknown, which can be assessed on the basis of field observations.

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