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Paleohydraulics of pleistocene drainage development of the Souris, Des Lacs, and Moose Mountain spillways, Saskatchewan and North Dakota

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PALEOHYDRAULICS OF PLEISTOCENE DRAINAGE DEVELOPMENT

OF THE SOURIS, DES LACS, AND MOOSE MOUNTAIN

SPILLWAYS, SASKATCHEWAN AND NORTH DAKOTA

by

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This thesis submitted by Mark L. Lord in partial fulfillment of the requirements for the degree of Master of Science from the University of North Dakota is hereby approved by the Faculty Advisory Committee under whom the work has been done.

(Chairman)

This thesis meets the standards for appearance and conforms to the style and format requirements of the Graduate School of the University of North Dakota, and is hereby approved.

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Signature Mark L. Lord

Date August 1, 1984

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ABSTRACT

Recent recognition of the rapid draining of numerous glacial lakes, including some in the Northern Plains, has revealed a need for further research concerning this process. Geomorphic interpretation of the Souris, Des Lacs, and Moose Mountain Valleys, and the gravel deposits in them, has resulted in the recognition of five phases in the development of the drainageways. Textural analyses and paleohydraulic methods were applied to the sediments associated with each phase to distinguish and characterize the discharges. Four of the five phases of development involved short-lived, high velocity (>4 m/s) discharges resulting from the rapid draining of glacial lakes; the other phase (2) involved deposition by glacial meltwater.

Phase 1 discharges (3×10^4 m^3/s) initiated the development of the Des Lacs Valley; the source probably was a supraglacial lake in the vicinity of Bowbells, North Dakota. Deposits of generally unstructured sandy gravel occur relatively high on the valley walls and are confined to the lower Des Lacs and Souris spillways.

Phase 2 discharges (2.1×10^3 m^3/s) deposited outwash sediment, consisting of cross-bedded gravelly sand, in the Moose Mountain and lower Souris Valley. Glacial meltwater from the Moose Mountains, which commenced when the ice sheet divided around the Moose Mountains, was the source of these flows.

Phase 3 and 4 discharges (1.9×10^4 m^3/s) probably each resulted from the rapid draining of Glacial Lake Arcola. Phase 3 was an

erosional stage that incised the Moose Mountain and lower Souris Valley and induced landslides along the valley walls. Deposition of unstructured pebble gravel throughout much of the lower Souris spillway occurred during phase 4. The deposits commonly are inset into the valley walls, indicating that they were later truncated by erosion.

The upper Souris Valley was developed and the lower Souris and Des Lacs spillways were enlarged during the cataclysmic discharges of phase 5 (2.0×10^5 m^3/s) from Glacial Lake Regina. Huge bars of unstructured pebbly cobble gravel were deposited during this event. Several geomorphic features associated with this phase are similar to features described in the Channeled Scabland. Phase 5 concluded the development of the spillways.

A glacial chronology proposed by Clayton and Moran (1982) indicates that the drainage described in this study occurred between 11,700 BP and 11,300 BP. Although active ice is generally assumed to have been northeast of the spillways during the time of drainage development, the presence of stagnant ice is indicated by streamlined collapse topography within a fluvially scoured zone, several inconsistencies in paleohydraulic calculations, and the abundance of landslides in the lower Souris Valley.

The information developed from this study, and the relatively recent recognition of other floods resulting from the rapid draining of glacial lakes, indicates that this form of 'instantaneous' drainage development may have been common during the Pleistocene Epoch, especially in areas of ice stagnation.

INTRODUCTION

General Statement

Drainage development during the Wisconsinan has been traditionally thought to be a gradual, uniform process similar to the present-day conditions, although somewhat accelerated due to the glacial environment. More recent research, especially during the past 10 years, has revealed that many large drainageways developed very rapidly (geologically, instantaneously). For example, the Souris spillway of southern Saskatchewan and north-central North Dakota may have developed catastrophically as Glacial Lake Regina drained (Kehew, 1982; Kehew and Clayton, 1983). In order to understand better Wisconsinan drainage development in the Northern Plains region, detailed research of the spillways and the sediments within them are needed. This study will attempt to bridge some of these deficiencies.

Purpose

The purpose of this study is to investigate the gravel deposits within the Souris, Des Lacs, and Moose Mountain Valleys, in order to develop a better understanding of fluvial erosional and depositional processes occurring during deglaciation. The specific objectives of this study are:

1) To distinguish and characterize the different drainage events which led to the development of the spillways.
2) To determine the paleohydraulics of the discharges responsible for the gravel deposits in the spillways studied.

Area of Study

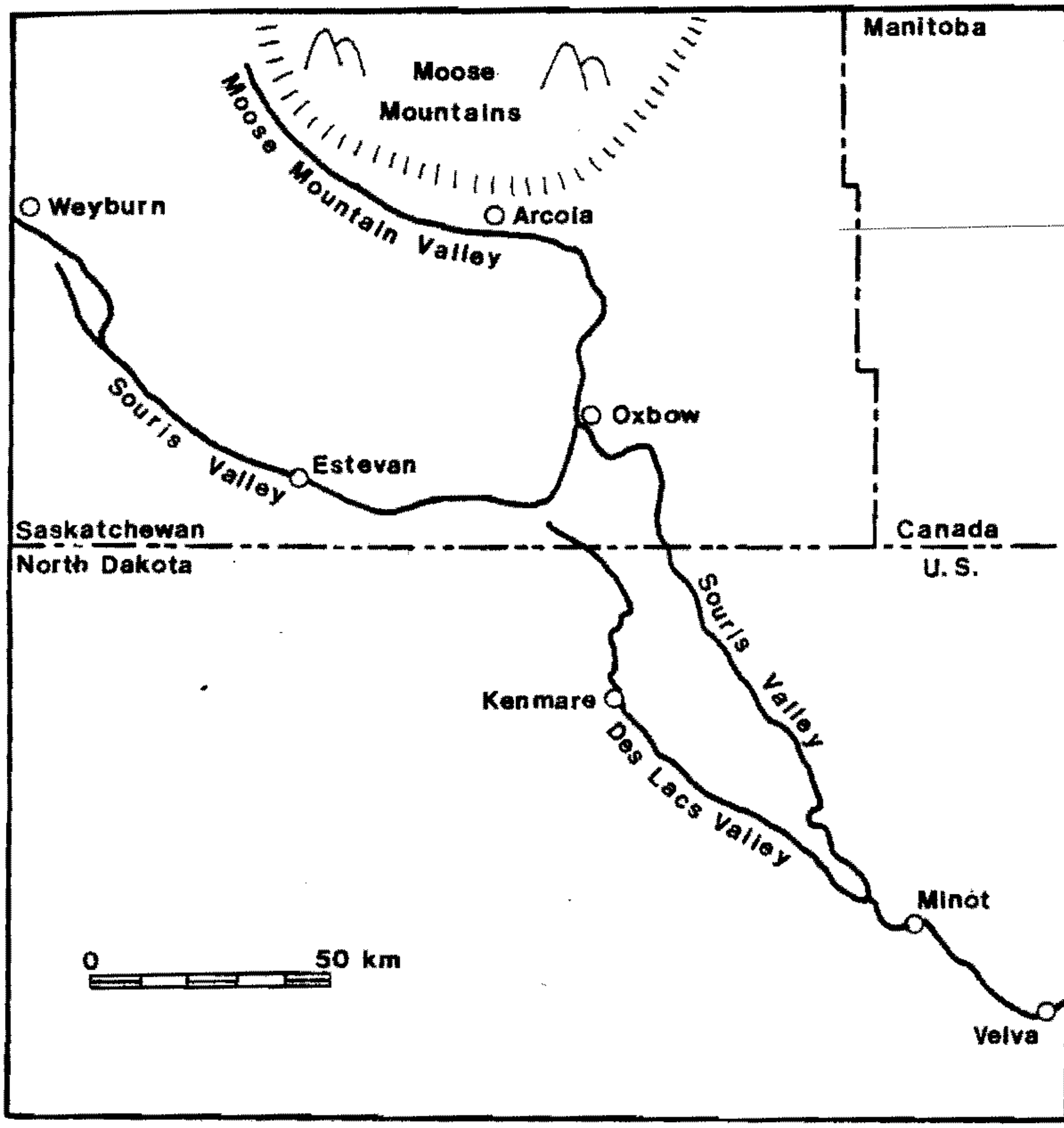
The area of study is the Souris, Des Lacs, and Moose Mountain Valleys of southeastern Saskatchewan and north-central North Dakota (Figure 1). Study was limited to the Souris Valley from Weyburn, Saskatchewan downstream to Velva, North Dakota, the Moose Mountain Valley from Arcola, Saskatchewan to its confluence with the Souris Valley at Oxbow, Saskatchewan, and the entire Des Lacs Valley.

Previous Work

Glacial Chronology

Glacial chronologies of the Northern Plains region recently have been proposed by Clayton and Moran (1982), Clayton and others (1980), and Christiansen (1979). In the following discussion, the chronology proposed by Clayton and Moran will be used for two reasons. First, it is the most recent chronology proposed for the Northern Plains region; consequently, it includes the most recent data. Second, the dating of the margins is based only on radiocarbon dates of wood; other materials commonly dated are more susceptible to contamination by old carbon (Clayton and others, 1980; Clayton and Moran, 1982).

Figure 1. Location of valleys of study in southern Saskatchewan and north-central North Dakota.



Manitoba
Moose
Mountains
Moose Mountain Valley
Weyburn
Arcola
Souris Valley
Oxbow
Estevan
Saskatchewan
North Dakota
Canada
U. S.
Souris Valley
Kenmare
Des Lacs Valley
Minot
0
50 km
Velva

Clayton and Moran (1982) recognized three major glacial events in the midcontinent region prior to late Wisconsinan time. The nonglacial middle Wisconsinan interval was followed by late Wisconsinan advances that reached a maximum extent in central Iowa about 14,000 BP. The Laurentide ice sheet began its last retreat from north-central North Dakota and southern Saskatchewan, the area of study, about 11,700 BP. Clayton and Moran (1982) depict this retreat as phases K and L (Figure 2). During this time, the Souris Lobe probably advanced to the Martin ice margin (about 100 km east-southeast of Minot) at least once and readvanced to the Minot ice margin (about 30 km east-southeast of Minot). Initial development of drainageways in the area of study probably began at this time.

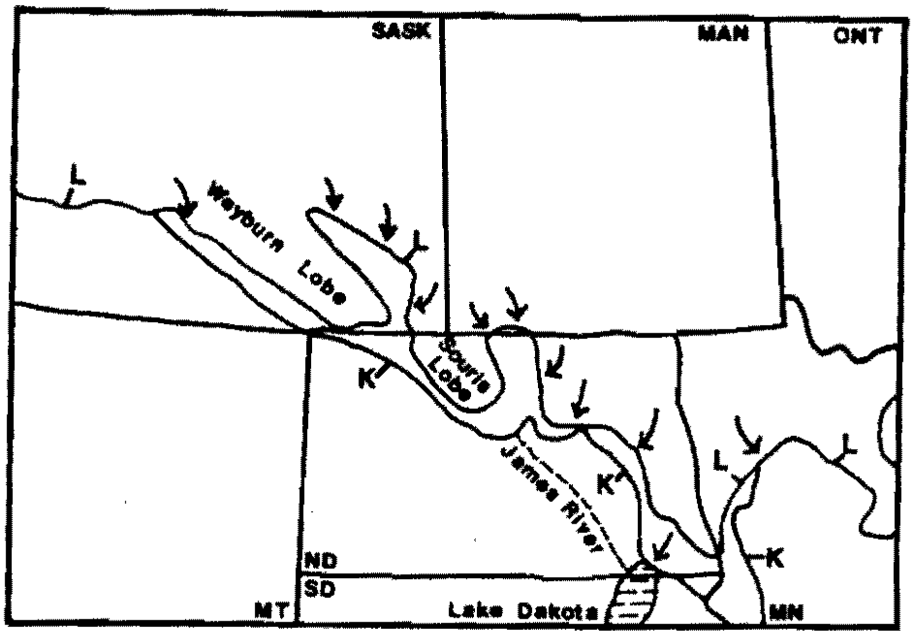
The recession of the glacier was interrupted by two readvances, phases M and N of Clayton and Moran (Figure 3). Ice marginal lakes were common during these phases; pertinent to this study were the presence of Lake Regina, Lake Souris, and Lake Agassiz. By the end of these phases, the major drainage paths in existence today were probably established. The glacier had wasted north of the midcontinent region by about 9500 BP (Clayton and Moran, 1982).

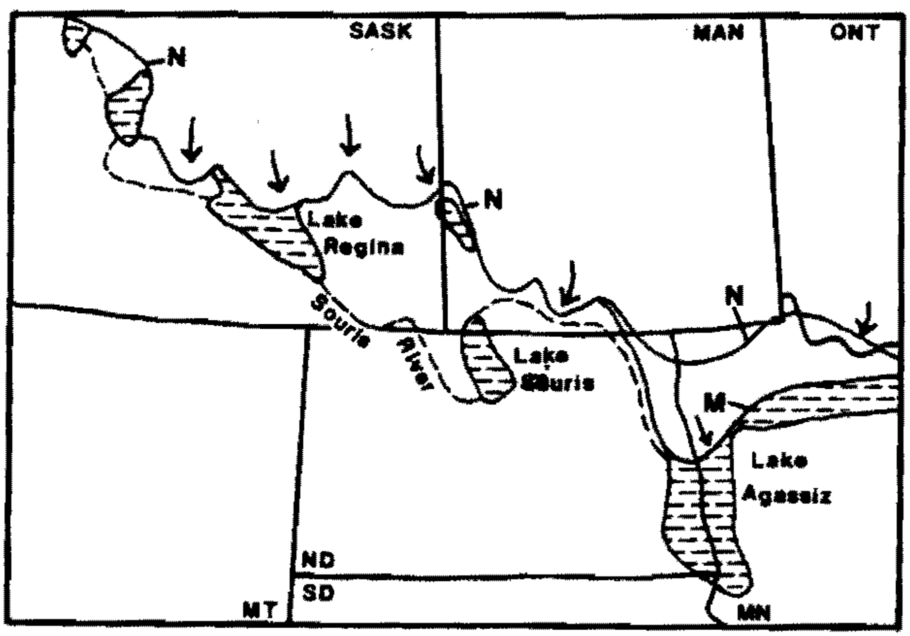
Surficial Geology of Study Area

Low-relief hummocky collapsed glacial sediment predominates over much of the study area (Clayton, 1980). However, high relief collapse topography exists in the Moose Mountains of Saskatchewan (Christiansen, 1956). Most of the surficial material is till of late Wisconsinan age assigned to the Pleistocene Coleharbor Group (Bluemle, 1971).

Figure 2. Ice-margin positions of phases K and L (about 11,700 BP) (Clayton and Moran, 1982).

Figure 3. Ice-margin positions of phases M and N (about 11,300 BP) (Clayton and Moran, 1982).


SASK
MAN
ONT
L
Weyburn Lobe
Souris Lobe
K
James River
K
L
L
K
ND
SD
MT
Lake Dakota
MN


SASK
MAN
ONT
N
N
Lake Regina
Souris River
Lake Souris
N
M
Lake Agassiz
ND
SD
MT
MN

In distinct contrast with the till plains are the Souris, Moose Mountain, and Des Lacs Valleys. Sections of these valleys are up to 2 km wide and up to 60 m deep. Recent literature refers to these valleys as spillways, outlet channels of glacial lakes (Kehew, 1982; Kehew and Clayton, 1983). Spillways are quite distinct from glacial meltwater valleys. Spillways are erosional systems characterized by deeply incised valleys with relatively low width to depth ratios; outwash valleys, on the other hand, have relatively high width to depth ratios and are depositional systems. Spillways in the Great Plains primarily developed as river channels, not river valleys (Clayton and others, 1980).

Terraces are common throughout much of the Souris, Moose Mountain, and Des Lacs Valleys. Such terraces are underlain by coarse gravel, and like the till, are assigned to the Coleharbor Group (Kehew, 1983). Kehew and Clayton (1983) recognized that these are not typical river terraces representing remnants of former flood plains. Landslides are common on the valley walls, especially in the Souris. Where the gravel terraces and landslides occur together, the terraces are nearer to the center of the valley.

The Des Lacs and Souris Valleys are partially filled with Holocene alluvium of the Oahe Formation (Kehew, 1983; Clayton and Moran, 1979). Limited exposures of Fort Union Group rocks, of Paleocene age, exist along the valley walls. Exposures are most common in the Minot, North Dakota and Estevan, Saskatchewan regions (Lemke, 1960; Kehew, 1982).

The Souris, Moose Mountain, and Des Lacs Valleys have not received much attention in geologic studies of the area. Lemke (1960), however, discussed the origin of the Souris and Des Lacs in some detail. He suggested that both valleys originated before the last glacial advance. This belief is supported by observations of till draped over bedrock, within the valley, in the vicinity of Kenmare and Verendrye.

The late Wisconsinan ice covered the entire area, including the valleys. Lemke (1960) proposed that as the ice front retreated, meltwater deposited sand and gravel upon ice remnants lying in the valleys. Post-depositional melting of the ice left the sand and gravel deposits high on the sides of the valleys. Supporting evidence for a kame terrace origin of the deposits is that they do not appear to be paired. Lemke (1960) stated that a problem with this interpretation is the lack of collapse structures within all of the terrace deposits except one near Minot.

Studies up until the 1980's continued to classify the Souris and Des Lacs Valleys as meltwater valleys, although none included detailed investigations of the valleys or terraces (Dury, 1964; Bluemle, 1972; Pettyjohn and Hutchinson, 1971; Armstrong, 1971). Dury (1964) estimated that a Souris meltwater valley had a discharge of 3.6×10^4 m^3/s based on meander wavelength. Clayton and others (1980) suggested that the valleys were former river channels. Evidence cited for this belief is the presence of discrete cutbanks, consistent bank heights, uniform widths, and regular meander shapes of the valleys. They did not propose an origin for the valleys, except to indicate that they were once occupied by bankfull flow.

The first detailed studies of the Souris and Moose Mountain Valleys of Saskatchewan were done by Christiansen (1956, 1961). He recognized that the Souris and Moose Mountain Valleys were spillways, with sources at Glacial Lake Regina and Glacial Lake Arcola, respectively. Christiansen (1956) noted that the Souris spillway includes the deep valley and the eroded till plains which flank the valley. The eroded till plains reach up to 13 km in width and are commonly mantled with lag boulders. The Moose Mountain spillway is not as deeply entrenched as the Souris spillway and is not flanked by eroded till plains.

Most recently, a catastrophic flood hypothesis has been proposed for the origin of the Souris spillway in a series of papers by Kehew (1979, 1982, 1983) and Kehew and Clayton (1983). It was suggested that a rapid draining of Glacial Lake Regina induced a catastrophic flood which created the upstream section of the Souris Valley and then flowed into the pre-existing Souris and Des Lacs Valleys.

Topographic evidence for this cataclysmic event includes a fluvially scoured surface adjacent to the Souris Valley, channel divide crossings, anastomosing channels, channel bifurcations, and fluvially streamlined hills. Sedimentologic evidence includes huge gravel bars (the kame terraces proposed by Lemke, 1960) up to 2 km in length and 0.6 km in width, and fluvially transported boulders up to 3 m in diameter. Kehew (1982) and Kehew and Clayton (1983) estimated the duration of the flood have been about 9 days at a peak discharge of 1×10^5 m^3/s; discharge estimates were made using the Manning equation. It was also proposed that the flood from Glacial Lake Regina was triggered by the flooding of a proglacial lake farther upstream in a domino fashion.

Studies of Large-discharge Fluvial Events

Paleohydrology is the study of past occurrences, distributions, and movements of continental waters. Paleohydraulics, a subdiscipline of paleohydrology, applies the techniques of engineering mechanics to determine past water flows (Baker, 1983). Paleohydraulic methods were used in this study to determine paleovelocities and paleodischarges of the flows responsible for the deposition of the coarse gravels in the Souris, Moose Mountain, and Des Lacs Valleys.

An abundance of paleohydraulic methods is currently in use. Numerous characteristics of river systems have been studied and related to paleovelocities and paleodischarges. Among these are: silt-clay ratios of channel banks, bed roughness (resistance to flow), climatic records, hydraulic radius, channel widths and depths, energy gradient (usually assumed to be equal to the slope of the water surface), meander wavelengths, drainage patterns, sedimentary structures, and maximum particle sizes (Ethridge and Schumm, 1978; Maizels, 1983; Schumm, 1965; Baker, 1973; Dury, 1964; Bradley and Mears, 1980).

Ideally, combinations of the above methods should be used in paleohydraulic calculations. The multi-event origin of most of the valleys within the study area precludes the use of most of these methods. Only individual gravel deposits can be assured to be the result of a single event. For this reason, only paleovelocity formulas based on maximum particle size were used (sedimentary structures are rare in the deposits). Three groups of paleovelocity techniques have been developed. They are: 1) theoretically-derived formulas for natural streams, 2) empirically-derived formulas for natural streams, and 3) empirically-derived formulas for riprap.

Theoretically-derived formulas for gravel transport are the most abundant. Helley (1969) and Bradley and Mears (1980) derived formulas based on balancing fluid drag forces against the resisting forces of the sediment. A relationship between velocity and maximum particle size was determined in both studies. Baker and Nummedal (1978), Baker and Ritter (1975), and Baker (1973) have suggested that shear stress is a better indicator of competence than velocity. There is significant variability of the velocity profiles of streams; the use of shear stress normalizes some of the variability (Baker and Ritter, 1975). Baker (1973) and Baker and Nummedal (1978), in studies of the Channeled Scabland of Washington, showed good correlation between shear stress and maximum particle size. The theoretical basis of the equations was developed from DuBoys' equation for critical tractive force.

The most notable empirical work was done by Fahnestock (1963) on the braided White River of Washington. His method involved catching boulders in transport and determining the velocity at that point with a current meter. Most riprap studies arrived at formulas on a basis similar to that of Fahnestock (1963). The differences are that particle transport is observed in a flume instead of a natural stream and the particles are generally crushed rock (Mavis and Laushey, 1949; Torpen, 1956; Strand, 1973). From the data collected in these studies, a relationship between maximum particle size and velocity can be determined.

Compilations of existing data have also been conducted in order to relate either velocity or shear stress to maximum particle size. Most of these studies simply average out the more commonly used techniques

(Costa, 1983a; Bradley and Mears, 1980; Baker and Ritter, 1975; Novak, 1973). Common to all of the formulas generated by the above methods are certain assumptions and problems. One basic assumption is steady, uniform flow (Maizels, 1983). This presents special problems in the study of coarse gravel where turbulence is maximized. Macroturbulent lift forces can completely distort the velocity profile and can induce entrainment of particles at lower velocities than predicted (Baker and Ritter, 1975; Baker, 1973; Krumbein, 1942). A second assumption is that all sizes are available for transport and that all particles in a deposit were entrained in flow (Baker and Ritter, 1975). A survey of the local geology could adequately determine the size of particles available. Measuring the dimensions of the ten largest particles present in a deposit, versus one or five, reduces the probability of a particle being deposited by mass wasting, ice rafting, or other means which might bias the velocity determinations (Scott and Gravlee, 1968; Fahnestock, 1963; Bradley and Mears, 1980). Several paleohydraulic formulas are also based on the assumption that the particles are spherical and equal in density; this can also lead to minor errors in paleovelocity determinations (Baker and Ritter, 1975; Costa, 1983a).

METHODS OF STUDY

Preliminary Work

Interpretation of topographic maps and aerial photographs comprised the first phase of research for this study. Most maps were of the scale 1:24,000; most photographs were 1:50,000. The objectives of this phase were the following:

1) To determine the location, form, and areal extent of sand and gravel deposts within the Souris, Des Lacs, and Moose Mountain spillways.

2) To study the landforms directly related to the large discharges which formed the spillways.

3) To locate, correlate, and determine the elevations and slopes of terraces present within the spillways.

Field Work

Field work for this study was conducted during the summers of 1982 and 1983. The primary objective of the field work was to sample and collect data for textural and lithologic analysis which could be used to develop paleohydraulic interpretations of the water flows responsible for the deposits. The secondary objective was to study the morphologic relationships between the individual deposits and between the deposits and the spillways.

Terrace Correlations

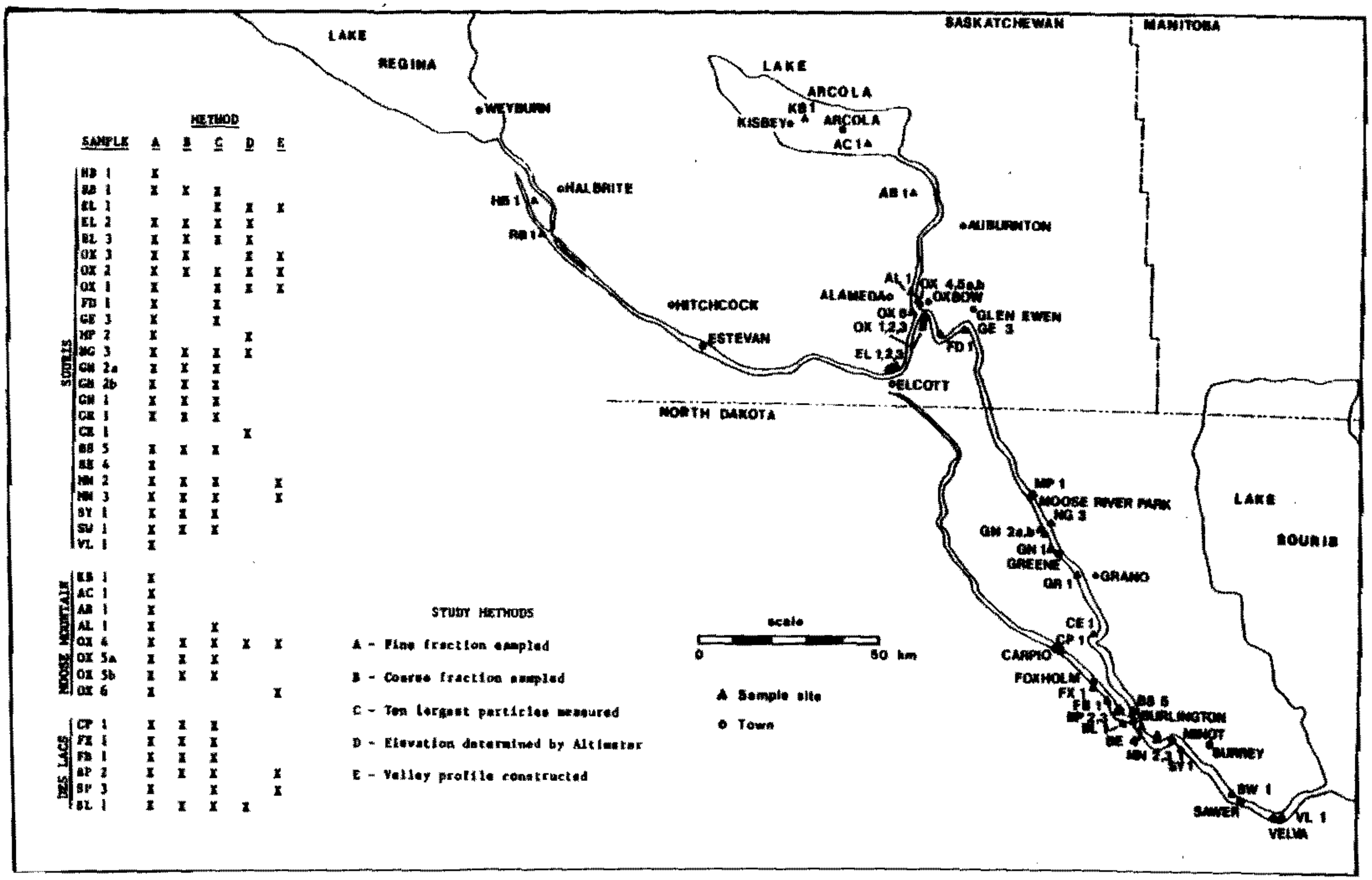
The sand and gravel terraces present in the Souris, Des Lacs, and Moose Mountain Valleys were correlated mainly on the basis of surface elevations determined with an altimeter, where elevation control in the field permitted, and from topographic maps. The morphologic relationships of the terraces to the valley walls also assisted in the correlation of terraces.

Sample Locations

Over thirty deposits, located by maps and airphotos, were sampled within the study area. Other areas were visited, but could not be sampled using the methods employed in this study. Figure 4 shows the locations of the samples and the study methods used at each site.

Ideally, sampling should have been done in the same relative location at each deposit. For example, if sampling was done near the surface in the planimetric center of each deposit, the number of controls could be maximized, thus permitting justifiable comparisons between results. However, in this study, numerous factors prohibited the use of this method of sampling. The primary factor was the fact that all good exposures of the deposits sampled were in gravel pits.

Figure 4. Map of sample locations and study methods used at each sample site.



	SAMPLE	A	B	C	D	E
SOURIS	HB 1	X				
	RB 1	X	X	X		
	EL 1			X	X	X
	EL 2	X	X	X	X	
	EL 3	X	X	X	X	
	OX 3	X	X		X	X
	OX 2	X	X	X	X	X
	OX 1	X		X	X	X
	FD 1	X		X		
	GE 3	X		X		
	MP 2	X			X	
	MG 3	X	X	X	X	
	GN 2a	X	X	X		
	GN 2b	X	X	X		
	GN 1	X	X	X		
	GR 1	X	X	X		
	CE 1				X	
	BB 5	X	X	X		
	BE 4	X				
	MN 2	X	X	X		X
	MN 3	X	X	X		X
	SY 1	X	X	X		
	SW 1	X	X	X		
	VL 1	X				
MOOSE MOUNTAIN	KB 1	X				
	AC 1	X				
	AB 1	X				
	AL 1	X		X		
	OX 4	X	X	X	X	X
	OX 5a	X	X	X		
	OX 5b	X	X	X		
	OX 6	X				X
DES LACS	CP 1	X	X	X		
	FX 1	X	X	X		
	FB 1	X	X	X		
	BP 2	X	X	X		X
	BP 3	X		X		X
	BL 1	X	X	X	X	

The relative position of the gravel pits within the deposits varied randomly from one site to the next.

In addition to the variation in sampling location, other factors also affected the specific location of sampling. Among these were: the freshness of the exposure, the amount of vegetative cover on the exposure, and the accessibility of the exposure. In some places, these factors narrowed possible sampling locations to several square metres. The possible effects of the variability in sampling will be discussed with the analysis of results.

Ten Largest Particles

At each sampling location, where possible, the long, intermediate, and short axes of the ten largest particles were measured. In some places, only two dimensions were measured because of partial burial. In addition to recording the dimensions of the largest particles, the lithology was also noted. The results of these determinations are shown in Appendix A.

Coarse-grained and Fine-grained Sediment Sampling

The main purpose for the sampling of the sediment was to obtain grain-size distributions. The expectations were that paleohydraulic data and distinguishing characteristics of the different discharge events could be determined. The coarse- and fine-grained fractions had to be sampled differently, because large machinery necessary to sample very coarse sediment properly was unavailable for this study.

The coarse-grained fraction (<-4 phi) was sampled by photographic methods. A 50 x 50 cm wooden frame was placed upon the exposure; a photograph was then taken of the frame and the particles in the frame (Figure 5). Channel samples of the fine-grained fraction (>-4 phi) were collected at the same locations as the photograph sites. Samples averaged about 2 kg.

Laboratory Work

Coarse-grained Sediment Size Analysis

Procedure

The coarse-grained sediment fraction required an unique method of analysis. The procedure used for analysis will be discussed first, followed by the theory behind the procedure. The procedure was as follows:

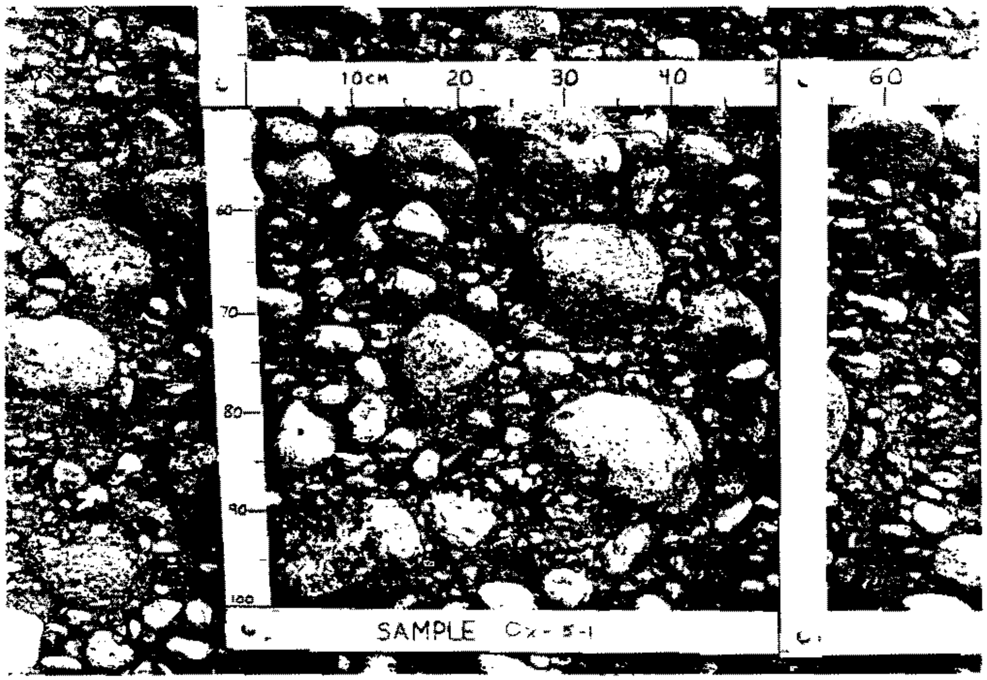
1) The photographs taken to sample the coarse material were all enlarged to 20 x 25 cm black and white prints.

2) The scales of the photographs were determined; most scales were approximately 1:3.

3) A transparency was placed over each photograph and the outline was traced of every particle with a short diameter coarser than -4 phi (usually about 5.5 mm on the photograph).

4) The apparent short diameter and area of each particle trace was measured using a Talos Electronic Digitizer. The areas

Figure 5. Wooden frame used to sample the coarse fraction of the sediment.


10CM
20
30
40
60
60
70
80
90
100
SAMPLE Cx-5-1

determined were tallied under the appropriate phi size class. Half phi intervals were used.

5) The total area of each size class was calculated, as well as the cumulative coarse-grained area.

6) The area percentages of each size class, the coarse-grained fraction, and the fine-grained fraction were then determined.

7) The weight percentages arrived at for the fine-grained fraction by seive analysis were then normalized to the fine-grained fraction area percentage determined from the photographs.

8) A complete cumulative size distribution curve, combining fine-grained weight percentages and coarse-grained area percentages, was then determined for each sample.

Theory of Procedure

The most commonly used method of size analysis, sieving, is not practical when particles are cobble size or larger. This prompted the developed procedure to be used.

In this study, and in the case of thin sections, size data are collected from a two-dimensional image. There is varied opinion regarding whether long or short particle axes should be measured when analyzing for size in thin sections. Friedman (1958) obtained good correlation with sieving, after using a simple correction factor, by measuring the long axes of particles. Packham (1955) measured the short axes and, with a complicated conversion, also achieved good correlation with sieving.

Friedman (1958), using the long axes, developed his relationship, between thin section and sieving, using sandstones with well-sorted, well-rounded sediment comprised predominantly of quartz. On the other hand, Packham's (1955) relationship is more accurate with poorly-sorted sediments where particles are spheres or biaxial ellipsoids having two equal shorter axes. The short axes were measured for this study because the sediment is much more analagous to that used in Packham's study. The result is that the true grain size distribution probably is coarser than calculated.

The procedure used in this study has two advantages over that of sediment size analysis using thin sections. Most important is that the actual area of each particle is determined, versus assigning areas of perfect spheres to measured diameters as done with thin sections. Also, with both methods, it is not known whether the true diameter of a grain is being measured or a diameter less than true. However, in exposures of sand and gravel, the coarser material tends to protrude beyond the fines because of differential erosion. Thus, in a photograph, the true diameter of the larger particles will be seen more frequently than in a thin section.

The principal disadvantage to the procedure used in this study is the lack of control. The validity, compared to sieving, is not known because no sieving of the coarse material was done. Some control was established at one site (MN 2); the exposure was extensive enough so that 5 photo samples could be taken. From the data derived from these samples, confidence limits were calculated for each coarse-grained size class using T statistics (Table 1). At the 90 percent confidence level, individual percentages expected range from $\pm$ 1.5 to $\pm$ 6.5; the average is $\pm$ 3.2. The question still remains as to whether the variability of the size analysis results is due to the method of sampling, the method of size analysis, or the intrinsic variability of the sediment.

No correction factors were used to convert the derived size data to sieve size data, as is done with thin sections, for the following reasons: there was no means to derive a correction factor, and, in addition, this method is significantly different from that of thin sections so that the generally small corrections developed for thin sections do not necessarily apply.

Some possible sources of error which could cause deviation from the true size distribution are:

1) The sample size may not have been large enough (sampling originally was done with a 100 x 100 cm square frame, but the size of the frame limited the number of exposures which could be sampled).

2) The variability of the sediment.

3) Not knowing from the photo whether the intermediate or short axes of particles were being measured.

4) Operator error in tracing particles, and measuring their diameters and areas.

5) The overall validity of the essentially untested procedure.

TABLE 1

Confidence Limits For Coarse-grained Size Classes
(Confidence limit is 90 percent)
(Sample size is 5)

<-6.5 phi	
	S = 6.80
	U = ± 6.48
-6.0 phi	
	S = 1.60
	U = ± 1.53
-5.5 phi	
	S = 2.04
	U = ± 1.95
-5.0 phi	
	S = 2.75
	U = ± 2.62
-4.5 phi	
	S = 1.63
	U = ± 1.55
-4.0 phi	
	S = 1.61
	U = ± 1.54

S = standard deviation
U = mean

Even though many possible sources of error exist, the data derived from the procedure used appear to be reasonable.

Another topic yet to be discussed is that of combining size data derived by two completely different methods (sieving and photo analysis). The basic assumption made is that the particle area is proportional to the particle weight. Given the characteristic lithologic composition of the sediment (mostly granitic with some local sedimentary rocks--not significantly different), this assumption should be valid.

Fine-grained Sediment Size Analysis

The fine-grained fraction (>-4 phi) sediment samples commonly were partially cemented together by calcium carbonate (caliche), and consequently had to be disaggregated before the size distribution could be determined. Numerous methods of disaggradation were attempted to avoid using an acid which, at least to some degree, would affect the carbonate content. The following sediment disaggregation techniques were tried without success: soaking the sample in water at room temperature, soaking the sample in boiling water, soaking the sample in boiling water immediately followed by soaking in very cold water, dry heating the sediment (to about 250 degrees Celsius) followed by soaking the sample in very cold water, soaking the sediment in an ultrasonic bath, and subjecting the sediment to an ultrasonic probe.

In lieu of the above listed methods, dilute hydrochloric acid was used to prepare the sediment for sieving. Each sediment sample was soaked in a 5.0 percent hydrochloric acid solution for about 10 minutes. Following the soaking, the sediment was drained of the acid solution, rinsed with water, and drained of water. The sediment sample was oven-dried at a temperature of about 80 degrees Celsius.

Several different size carbonate grains were weighed, put in a 5.0 percent hydrochloric acid solution for 10 minutes, and then weighed again to determine the percent weight loss to be expected for the carbonate grains. Based on 15 grains, an average weight loss of 8.3 percent was found. Given this small percent, it is doubtful that the acid treatment had a significant effect on the size distribution, and what little effect it did have, should have been similar for all of the samples.

The size distribution of the fine-grained size fraction (-4 phi to 4 phi) was determined using standard sieving methods (Folk, 1980). A half-phi sieve interval was used and the Ro-Tap machine was run for 12 minutes per sample. The original mass of the individual samples averaged about 1500 g.

After sieving was complete, the lithology of the very coarse sand fraction, based on 200 grains, was determined for each sample. Identifications were made using a low power binocular microscope. Four lithologic compositions were identified: crystalline rock, carbonate rock, shale, and siltstone and sandstone.

RESULTS

Geomorphology of Spillways and Deposits

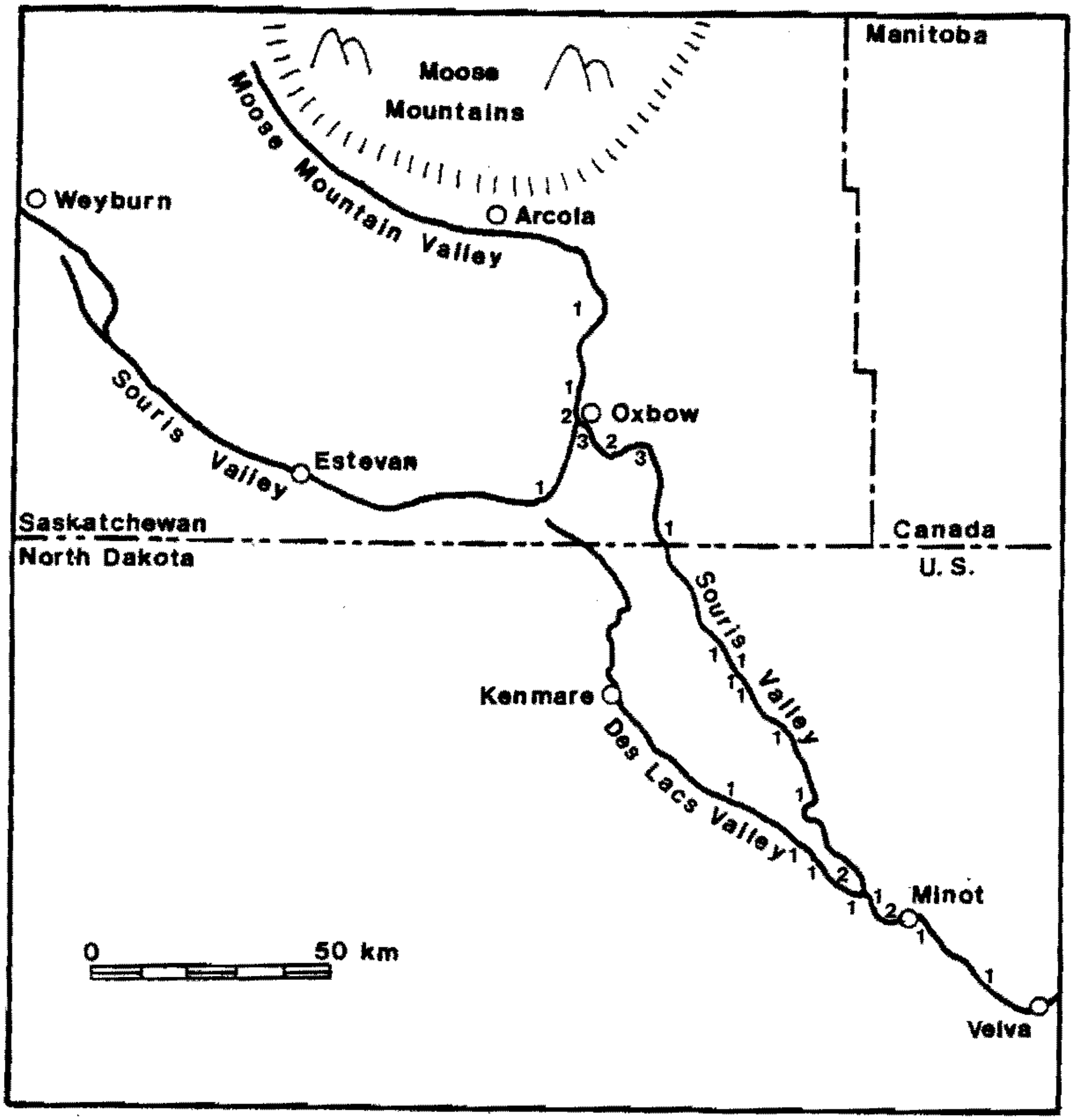
Depositional and erosional terraces exist at several places in the Souris, Des Lacs, and Moose Mountain spillways. The depositional terraces are remnants of former channel bottom sediment that remained after downcutting. The tread of the erosional terraces was formed primarily during downcutting by former fluvial action.

The terrace heights above the present valley floor vary from about 10 m to 60 m. The form of the terraces and their relationship to the surrounding topography also vary significantly. Most terraces occur singly, but some multiple terrace localities do exist (Figure 6). In the upper Souris Valley (upstream from the Souris-Moose Mountain confluence), only single terraces exist; in the Moose Mountain and Des Lacs Valleys, double terraces are present; and in the lower Souris Valley, triple terraces occur (Figure 7).

The correlation of the terraces indicates four separate sets of deposits (A, B, C, and D). Terrace profiles (Figure 8) also enable possible discharge source areas to be determined for each set of deposits. The oldest terraces (Figures 8, and 9), Set A Terraces, are located almost exclusively in the Des Lacs Valley. The profile of the terraces shows a constant slope of about 0.000406. These deposits typically occur high on the valley wall and are generally indistinct

Figure 6. Oblique air photo, view to south, of multiple terraces (arrows) (3) on the Souris Valley wall, 2.5 km south of Oxbow, Saskatchewan.

Figure 7. Location of single (1) and multiple (2,3) terrace sites in the Souris, Des Lacs, and Moose Mountain Valleys.

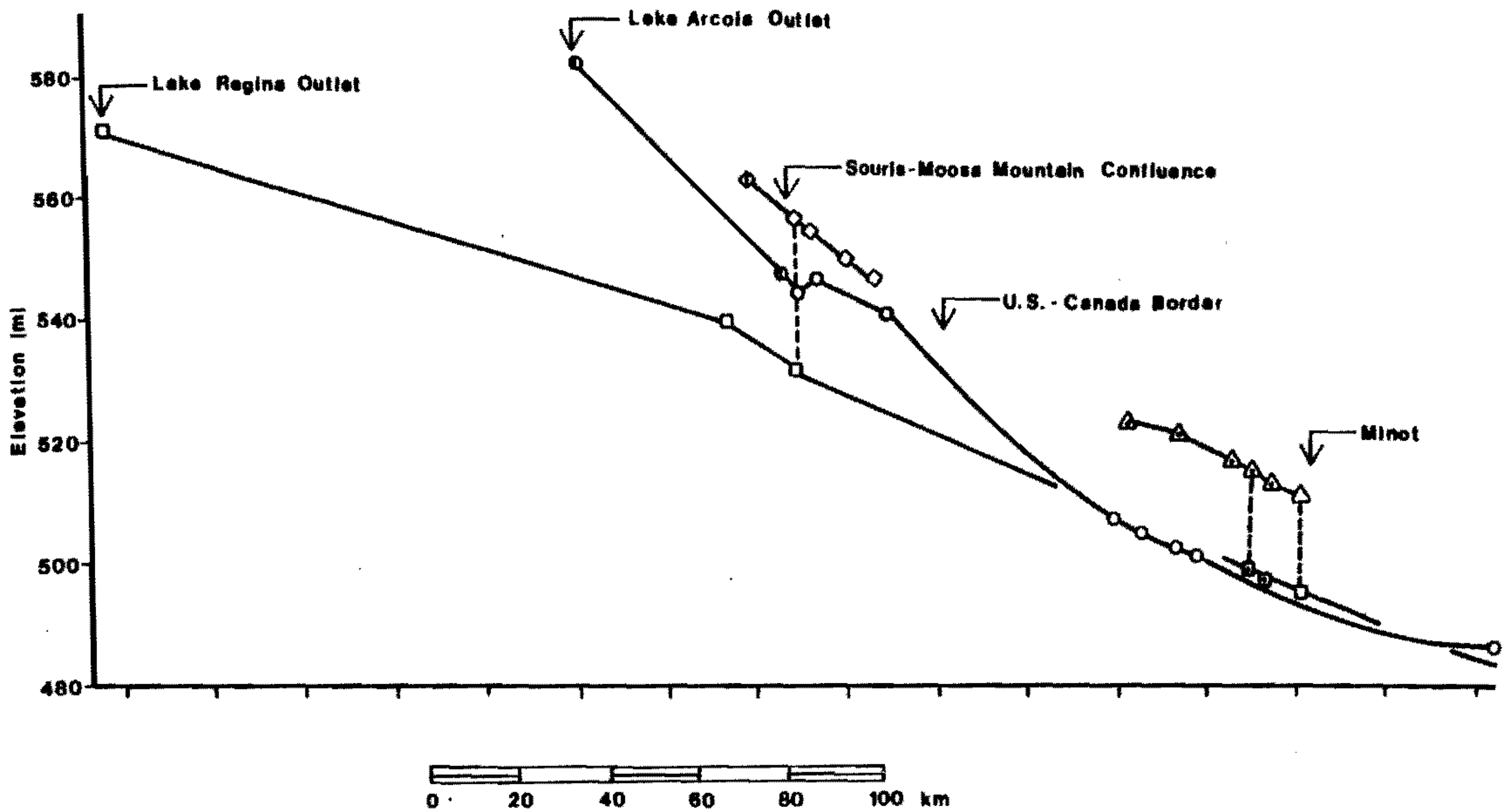


Moose
Mountains
Manitoba
Moose Mountain Valley
Weyburn
Arcola
Oxbow
Souris Valley
Estevan
Saskatchewan
North Dakota
Canada
U. S.
Souris Valley
Kenmare
Des Lacs Valley
Minot
Velva
0
50 km

Figure 8. Terrace profiles along the Souris, Des Lacs, and Moose Mountain Valleys.

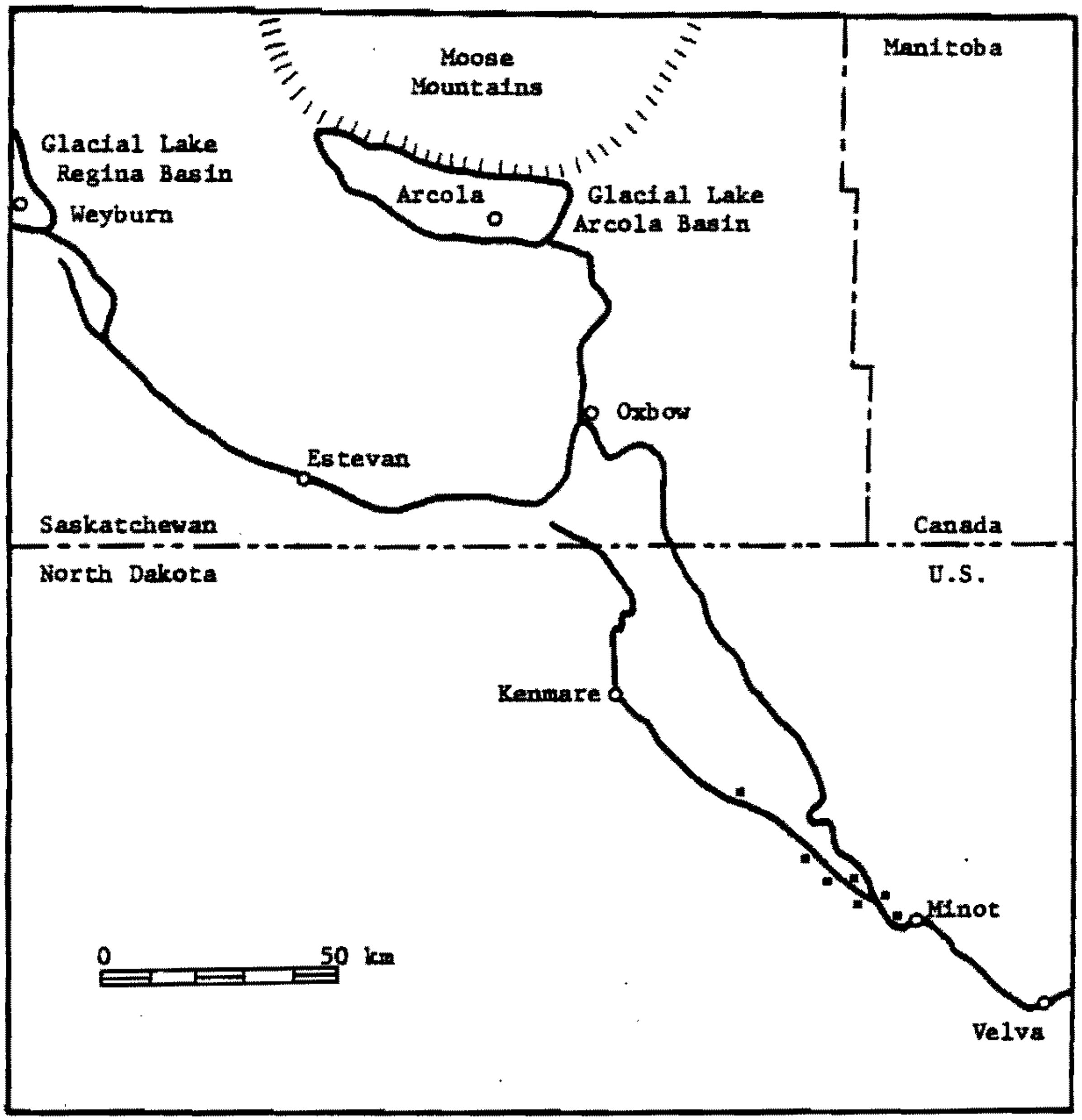
EXPLANATION

△ Set A Terrace deposit (see Figure 9)
◇ Set B Terrace deposit (see Figure 12)
○ Set C Terrace deposit (see Figure 15)
□ Set D Terrace deposit (see Figure 22)
¦ Multiple terrace location
□ Deposit occurs in Souris Valley
⊟ Deposit occurs in Des Lacs Valley
◫ Deposit occurs in Moose Mountain Valley



Lake Arcola Outlet
Lake Regina Outlet
Souris-Moose Mountain Confluence
U.S. - Canada Border
Minot
Elevation (m)
580
560
540
520
500
480
0
20
40
60
80
100 km

Figure 9. Location of Set A Terraces (darkened squares).



Moose Mountains
Manitoba
Glacial Lake Regina Basin
Weyburn
Arcola
Glacial Lake Arcola Basin
Oxbow
Estevan
Saskatchewan
Canada
North Dakota
U.S.
Kenmare
Minot
Velva
0
50 km

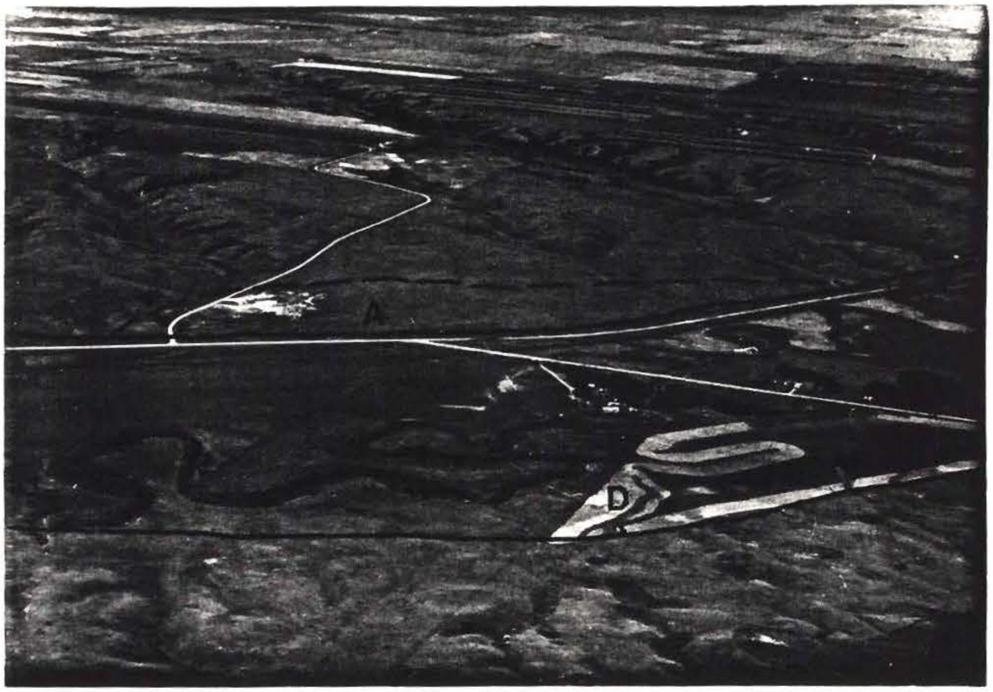
from the surrounding till plains (Figure 10). Only a slight vegetational and topographic change are present at the till contact at the margin of these terraces. These deposits, therefore, represent an early phase of the Des Lacs Valley development, probably the initial phase. The surface elevation of one of these deposits (BL 1, Figure 4) is approximately 514 m. A contact observed there between the sand and gravel and the underlying Paleocene bedrock is at an elevation of 494 m. This indicates a minimum incision of 20 m by this fluvial event. Most Set A Terrace deposits are homogeneous and consist of unstructured, unoriented sandy pebble gravel (Figure 11).

The source of the discharges responsible for these deposits was probably in the vicinity of Bowbells, North Dakota, where a fluvially scoured zone flanking the Des Lacs Valley terminates and collapse topography abuts the valley (Kehew, 1982). Because there is no observed evidence in this region of a glacial lake, proximal glacial meltwater or supraglacial lake discharges offer possible discharge sources for the Set A Terraces. This phase of development, when discharges initiated the development of the Des Lacs Valley and deposited the Set A Terraces, will be referred to as phase 1.

The second oldest terraces (Figures 8, and 12), Set B Terraces, are confined to southeastern Saskatchewan in the Moose Mountain and lower Souris (downstream from the Moose Mountain-Souris confluence) Valleys. The profile of these terraces also shows a uniform slope with a value of about 0.000558. Set B Terraces are the uppermost ones in the Moose Mountain and Souris Valleys, and like the Set A Terraces, commonly exhibit little topographic distinction between the sand and gravel

Figure 10. Oblique air photo, view to northwest, of a Set A Terrace (A) deposit (FX 1) in the Des Lacs Valley (D), about 4 km south of Foxholm, North Dakota.

Figure 11. Gravel exposure of a Set A Terrace deposit, BP 2; section 29, T.156N, R.84W. Shovel is about 60 cm long.


A
D

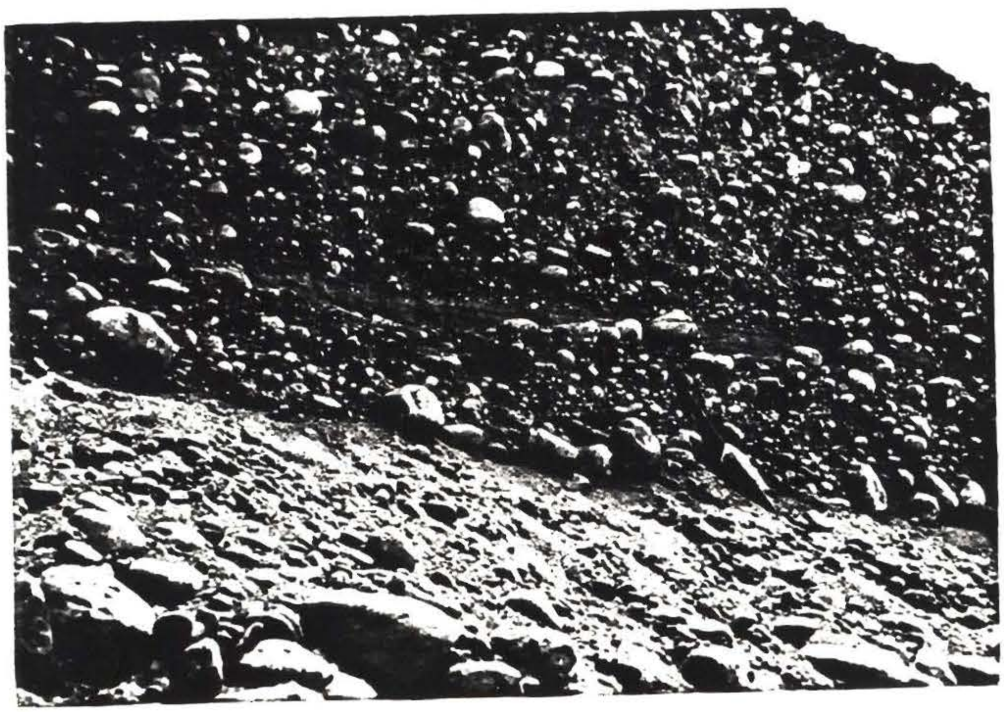
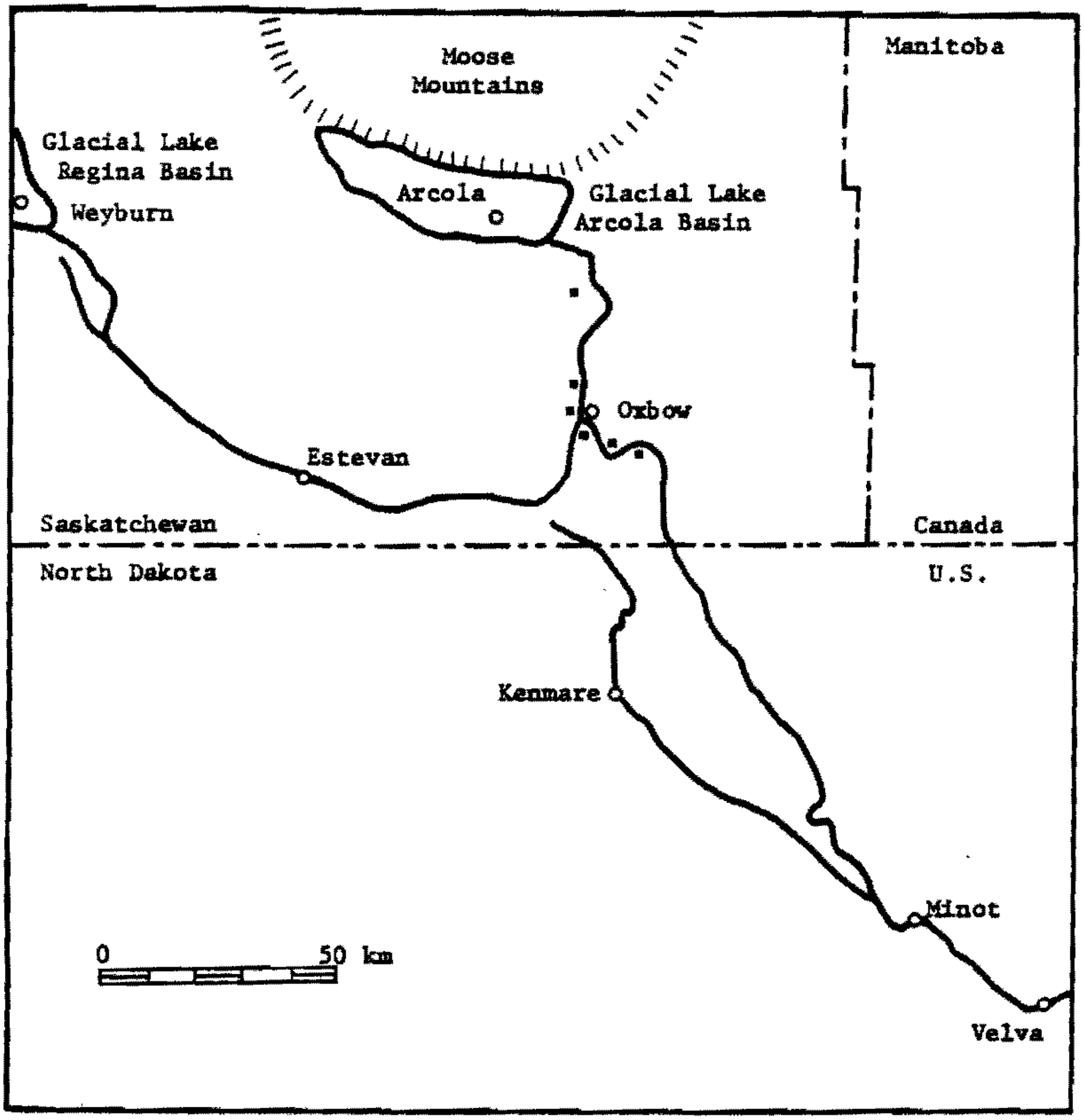


Figure 12. Location of Set B Terraces (darkened squares).



Moose
Mountains
Manitoba
Glacial Lake
Regina Basin
Weyburn
Arcola
Glacial Lake
Arcola Basin
Oxbow
Estevan
Saskatchewan
Canada
North Dakota
U.S.
Kenmare
Minot
0
50 km
Velva

deposits and the surrounding till plains (Figure 13). Essentially undisturbed remnants of the valley formed by this phase exist in the Moose Mountain Valley. These remnants suggest a wide, shallow valley, probably representing a former valley train. This stage, to be referred to as phase 2, probably initiated the first significant development of the Moose Mountain and Souris Valleys.

Discharges responsible for the Set B Terraces appear to have originated in the Moose Mountain area, as inferred by Christiansen's (1956) phase 2. The maximum thickness of these deposits is about 15 m, thus indicating a minimum incision of 15 m for this phase. Deposits typically consist of fairly well defined tabular cross-stratified to structureless gravelly sand (Figure 14). The texture and structure of these deposits, like the morphology of the valley, indicate a glacial meltwater origin.

The Set C Terraces (Figures 8, and 15) are abundant in the lower reaches of the Moose Mountain Valley and throughout the lower Souris Valley. The profile of these terraces has a concave form with a slope of 0.000704 in the Moose Mountain Valley, decreasing to 0.000168 in the Minot, North Dakota region. These terraces differ from other terraces in the spillways in that the majority (9 out of 13) have an arcuate form and are 'inset' into valley walls composed of till (Figure 16). In addition to the inset nature of the deposits, many of the deposits occur on the valley side of inactive landslides which occasionally form the valley walls in the lower Souris Valley (Figure 17). When lines connecting landslides are drawn on maps, they define a sinuous zone (Figure 18), presumably reflecting the existence of a sinuous channel at

Figure 13. Oblique air photo, view to northeast, of a Set B Terrace (B) deposit (AL 1) in the Moose Mountain Valley (M), 4 km west of Alameda, Saskatchewan.

Figure 14. Sand and gravel exposure of a Set B Terrace deposit, FD 1, showing tabular cross-bedding; section 31, T.2N, R.1W. Scale shown is in centimetres.


M
B

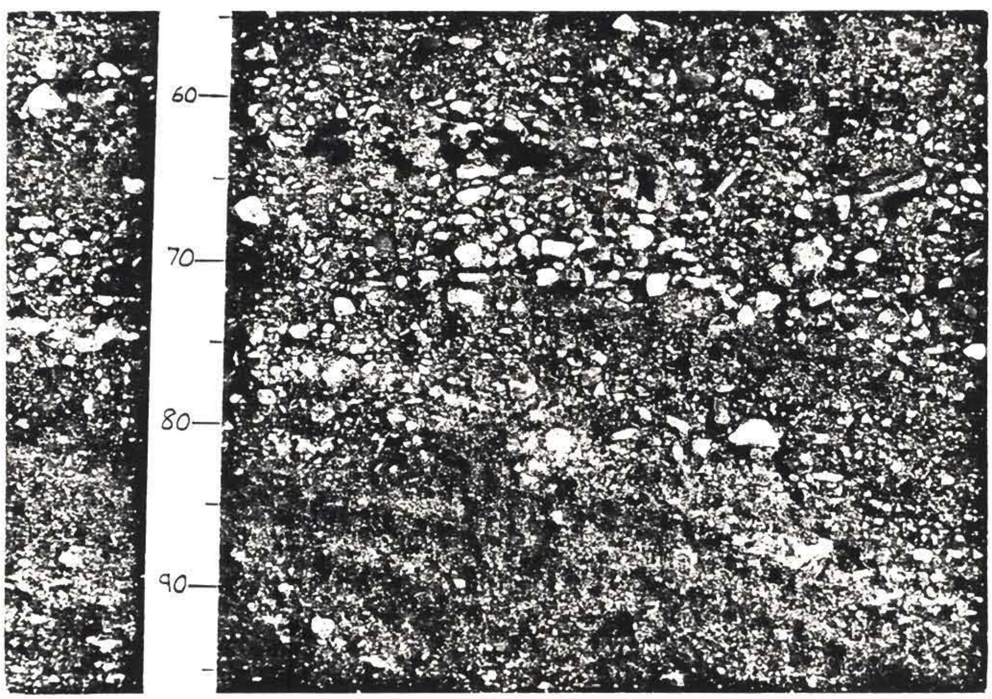
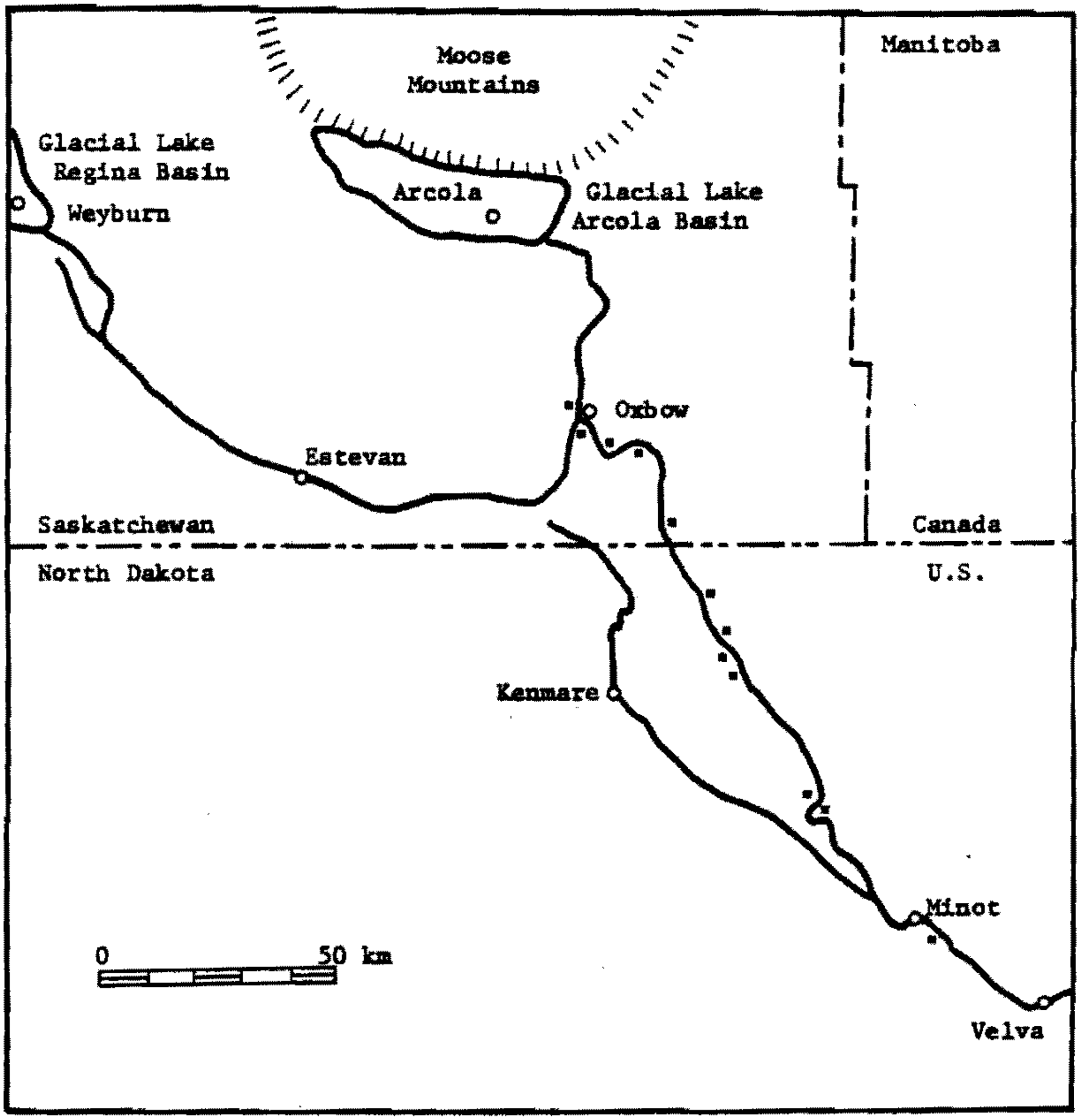

60
70
80
90

Figure 15. Location of Set C Terraces (darkened squares).



Moose
Mountains
Manitoba
Glacial Lake
Regina Basin
Weyburn
Arcola
Glacial Lake
Arcola Basin
Oxbow
Estevan
Saskatchewan
Canada
North Dakota
U.S.
Kenmare
Minot
0
50 km
Velva

Figure 16. Oblique air photo, view to northeast, of an inset Set C Terrace (C) deposit in the Souris Valley (S), just north of the United States-Canada border.

Figure 17. Vertical air photo of a Set C Terrace (C) deposit on the valley-center side of a landslide deposit (L) in the Souris Valley, 3.5 km north of Grano, North Dakota. Area shown is 6.7 km wide; north at top. Photo 340-5782 of U.S. Department of Agriculture.


C
S

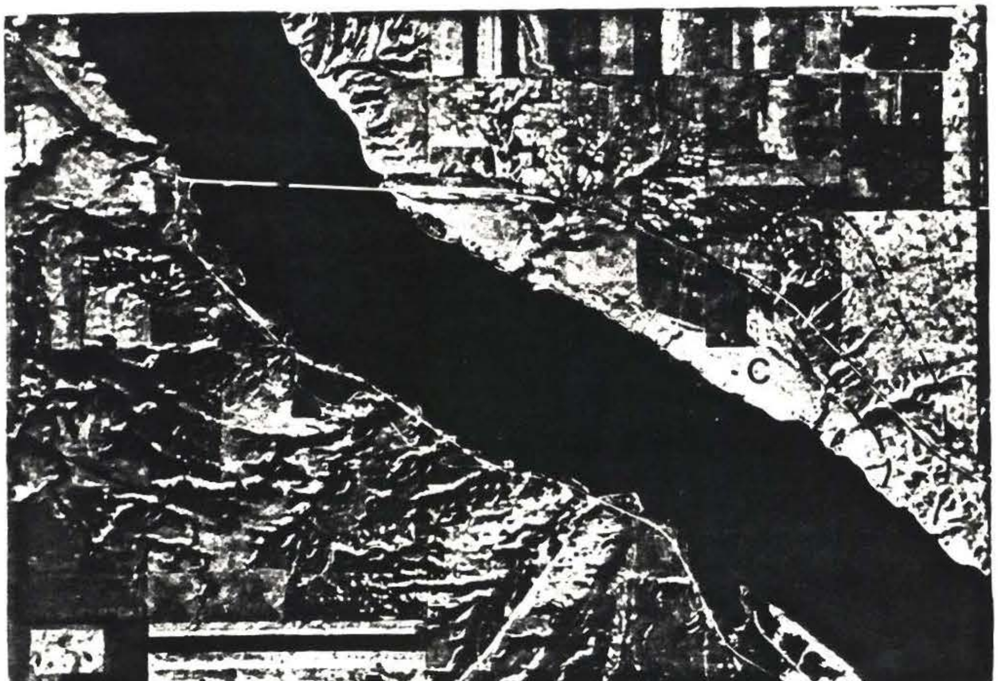
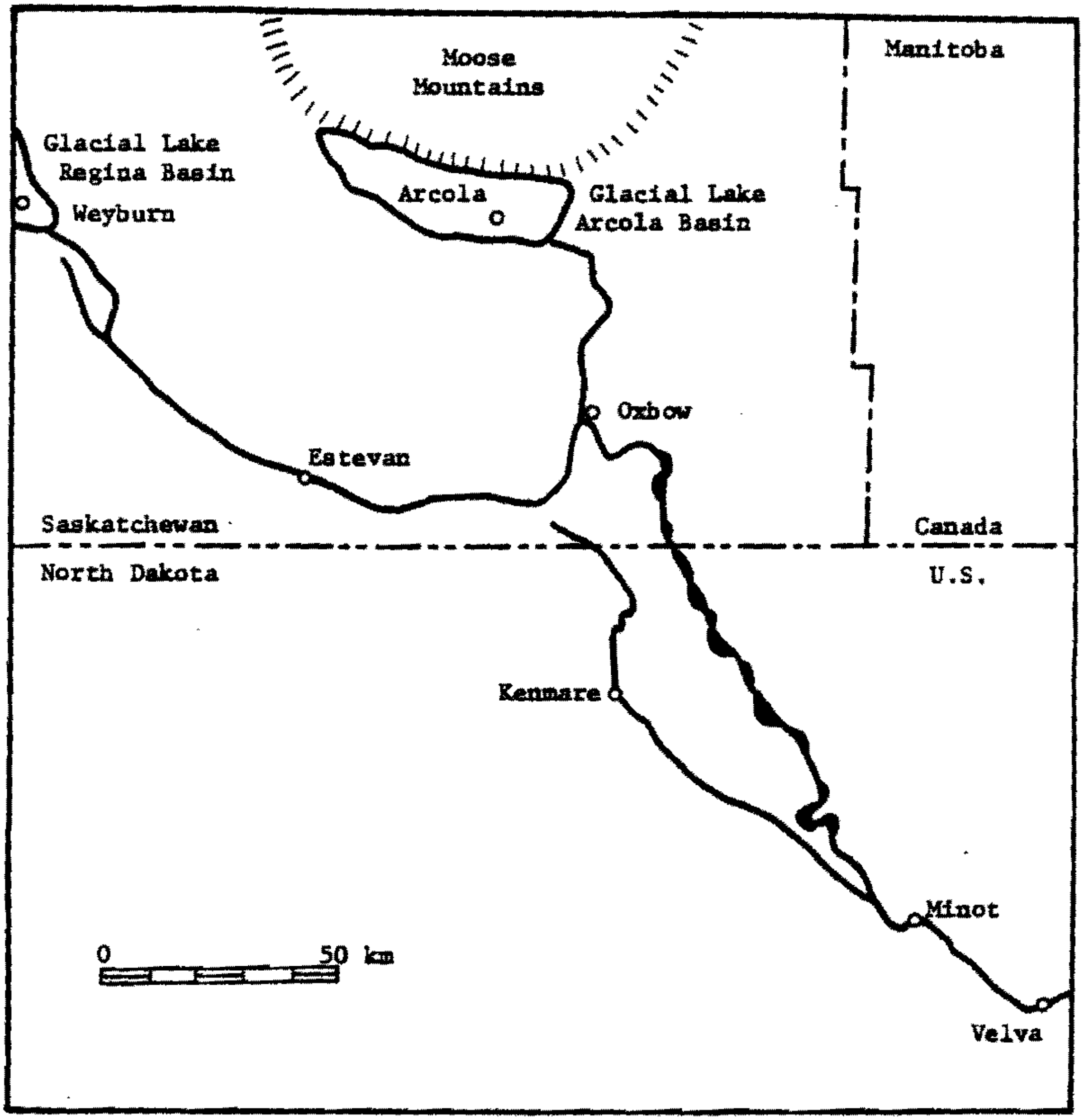

C

Figure 18. Location of landslide deposits in the lower Souris Valley. Landslides are indicated by darkened areas adjacent to the Souris Valley.


Moose
Mountains
Manitoba
Glacial Lake
Regina Basin
Weyburn
Arcola
Glacial Lake
Arcola Basin
Oxbow
Estevan
Saskatchewan
Canada
North Dakota
U.S.
Kenmare
Minot
Velva
0
50 km

the time of the slope failures. The relationship between the landslide deposits and the Set C Terraces indicates that the landslides preceded the deposition of the sand and gravel deposits. Therefore, a period of erosion, which incised the preexisting meltwater valley (Set B Terraces), preceded the deposition of the Set C Terraces. Other evidence for this conclusion is found in the Moose Mountain Valley in the vicinity of Alameda, Saskatchewan. In that area (Figure 19), the Moose Mountain Valley consists of a wide outwash valley with a well-incised channel on the eastern margin.

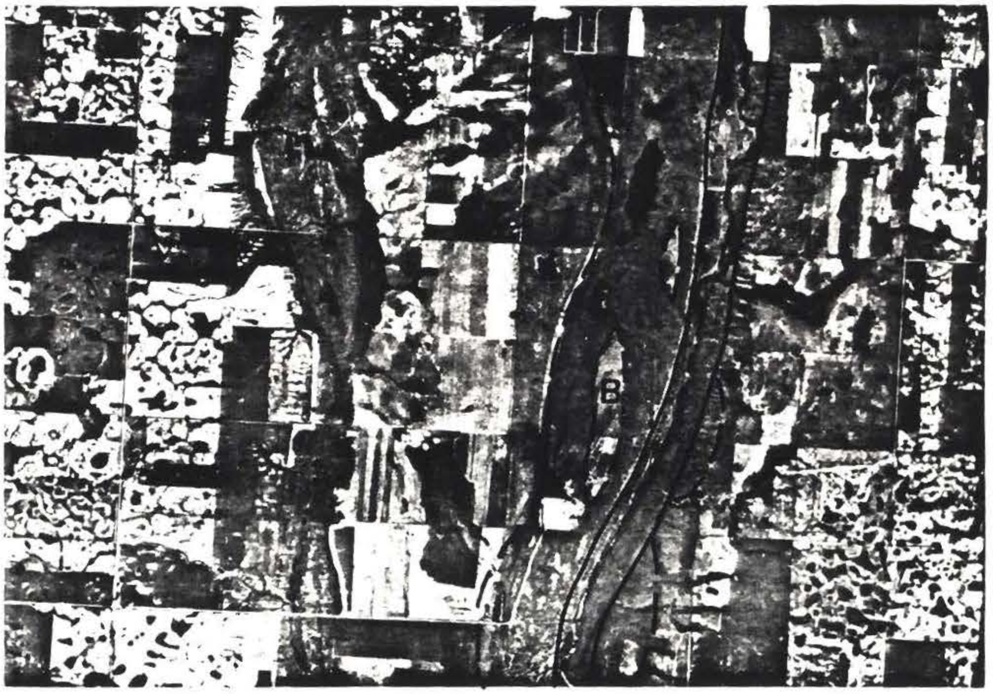
The deposit shown in Figure 19 is a Set B Terrace (AL 1), which was deposited by meltwater from the Moose Mountain area. The relationship between the wide shallow valley of the Set B Terrace and the well-incised channel within it, indicates a period of erosion following the deposition of the gravelly sand. The period of erosion which occurred between the deposition of the Set B and C Terraces will be referred to as phase 3; phase 4 represents the stage of development when discharges deposited the Set C Terraces.

The well-incised channel, resulting from the erosional phase 3, originates at the margin of Glacial Lake Arcola (located about 6.5 km south-southeast of Carlyle, Saskatchewan), indicating that this channel served as a spillway of Lake Arcola (Figure 20). The discharges that formed the Set C Terraces also appear to have commenced at the outlet of Glacial Lake Arcola as suggested by the lack of incision into the lake plain.

The Set C terrace deposits have a maximum thickness of about 15 m. They are very homogeneous accumulations of unstructured and unoriented pebble gravel (Figure 21).

Figure 19. Vertical air photo of a Set B Terrace (B) deposit and the wide, shallow valley (dashed lines) associated with it, and the well-incised channel (solid lines) within it that formed between the time of deposition of the Set B and C Terraces. Area shown is 7.3 km wide; north at top. Photo A21749-18, Saskatchewan Department of Energy, Mines and Resources.

Figure 20. Oblique air photo, view to northwest, of the southeast outlet, the Moose Mountain spillway (MS), of Lake Arcola (LA). The Moose Mountains (MM) are shown in the background. The outlet is located approximately 6.5 km south-southeast of Carlyle, Saskatchewan.


B

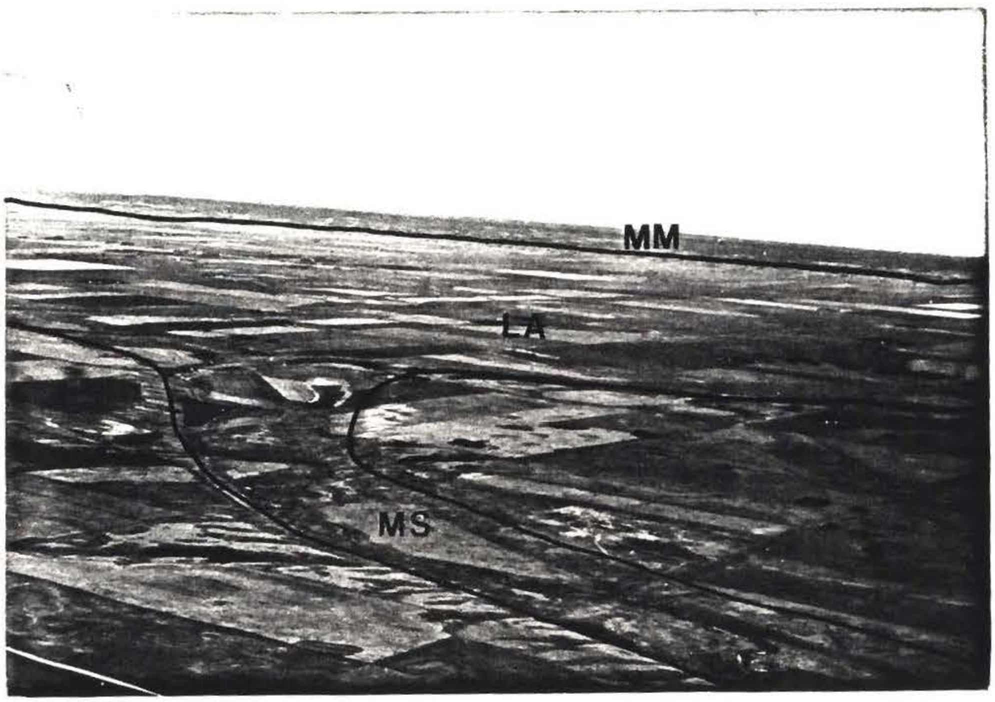

MM
LA
MS

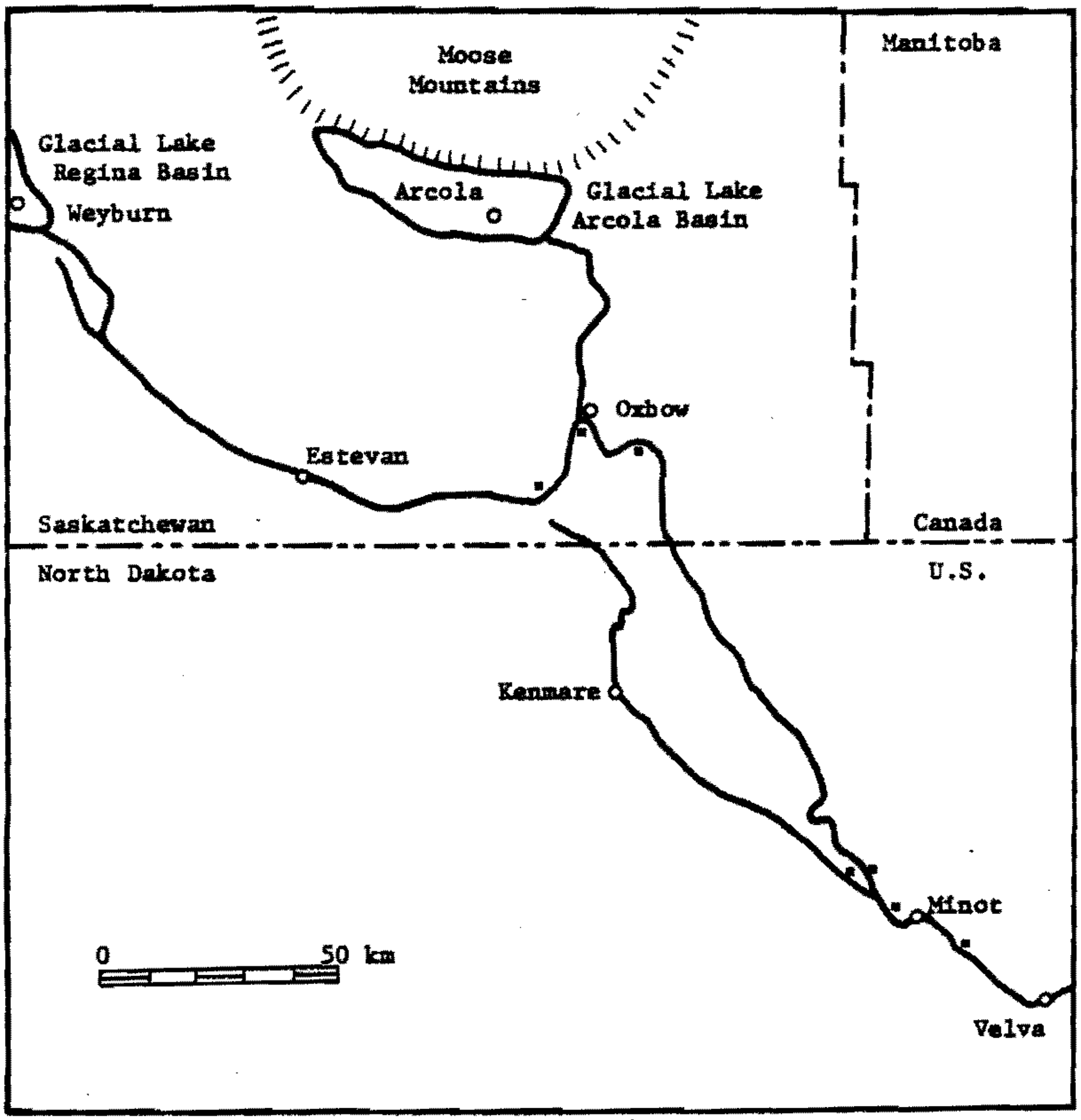
Figure 21. Gravel exposure of a Set C Terrace deposit, GN 1; section 19, T.160N, R.85W. Pick handle is about 1 m long.

The youngest terraces within the spillways, Set D Terraces (Figures 8, and 22) are located in the upper and lower Souris Valley as well as in the Des Lacs Valley. The profile of these terraces forms an irregular concave slope. The slope of the profile upvalley of Elcott, Saskatchewan is about 0.000209; from Elcott to Minot, North Dakota the slope is 0.000290. Possible explanations for this anomalous situation will be discussed later.

All the Set D Terraces are located at point bar positions within the valleys (Figure 23). These terraces, unlike most others present within the spillways, are probably not erosional remnants, but represent essentially unaltered depositional bars. Supporting evidence for this interpretation includes: an elevation lower than all other terraces, the point bar position of the deposits, and the non-inset nature of the deposits relative to the spillway walls.

One bar, in the Souris spillway near Elcott, Saskatchewan (Figure 23) is over 2 km in length, 0.6 km in width, and about 25 m in height above the present valley floor. Surface boulders on this bar are very large at the upstream end of the bar (up to 2.7 m in diameter), but markedly decrease in size toward the downstream end of the bar. Internally, the bar consists of homogeneous, very poorly-sorted unstructured and unoriented pebbly cobble gravel (Figure 24). The form and texture of the bar is similar to gravel bars in the Channeled Scabland of Washington (Baker 1973; Baker and Nummedal, 1978) and the Snake River Plain (Malde, 1968); both are interpreted to have been deposited by catastrophic floods. The Set D Terrace bars are most similar in form to pendant bars. Such bars occur just downstream of

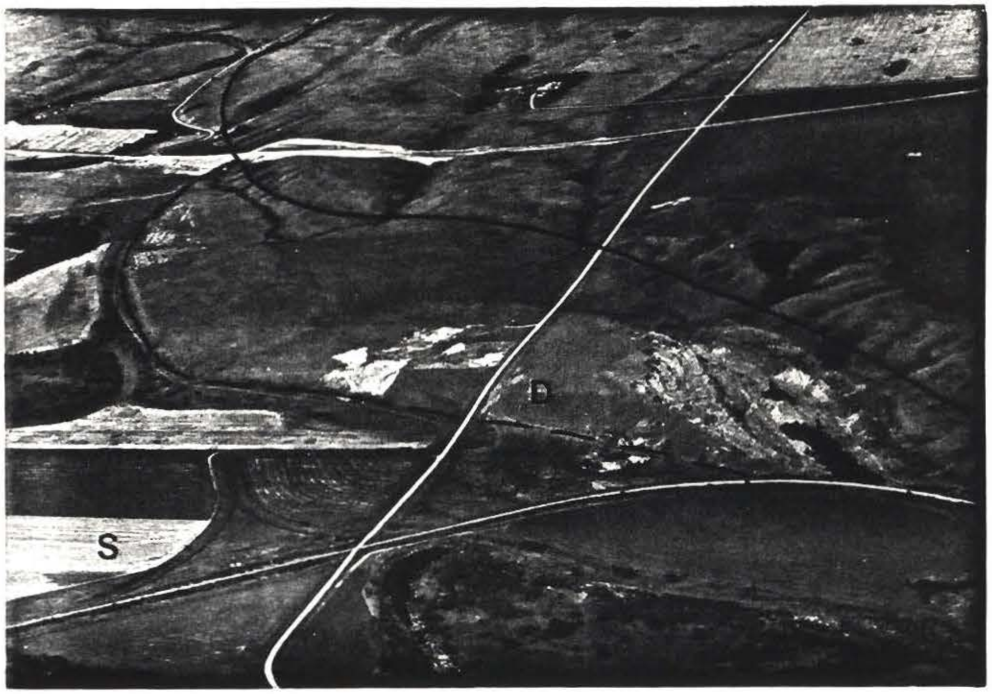
Figure 22. Location of Set D Terraces (darkened squares).

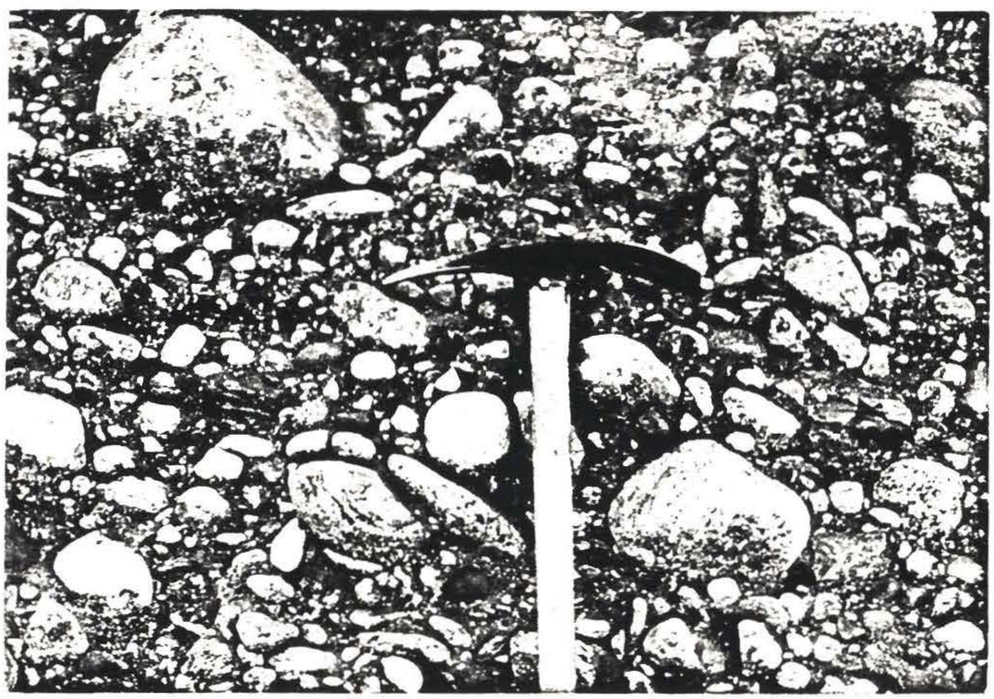


Moose
Mountains
Manitoba
Glacial Lake
Regina Basin
Weyburn
Arcola
Glacial Lake
Arcola Basin
Oxbow
Estevan
Saskatchewan
Canada
North Dakota
U.S.
Kenmare
Minot
0
50 km
Velva

Figure 23. Oblique air photo, view to west, of a Set D Terrace (D) deposit (EL 1) in the Souris Valley (S), 8 km north of Elcott, Saskatchewan.

Figure 24. Gravel exposure of a Set D Terrace deposit, EL 1; section 1, T.2N, R.3W. Pick is about 40 cm wide.


D
S



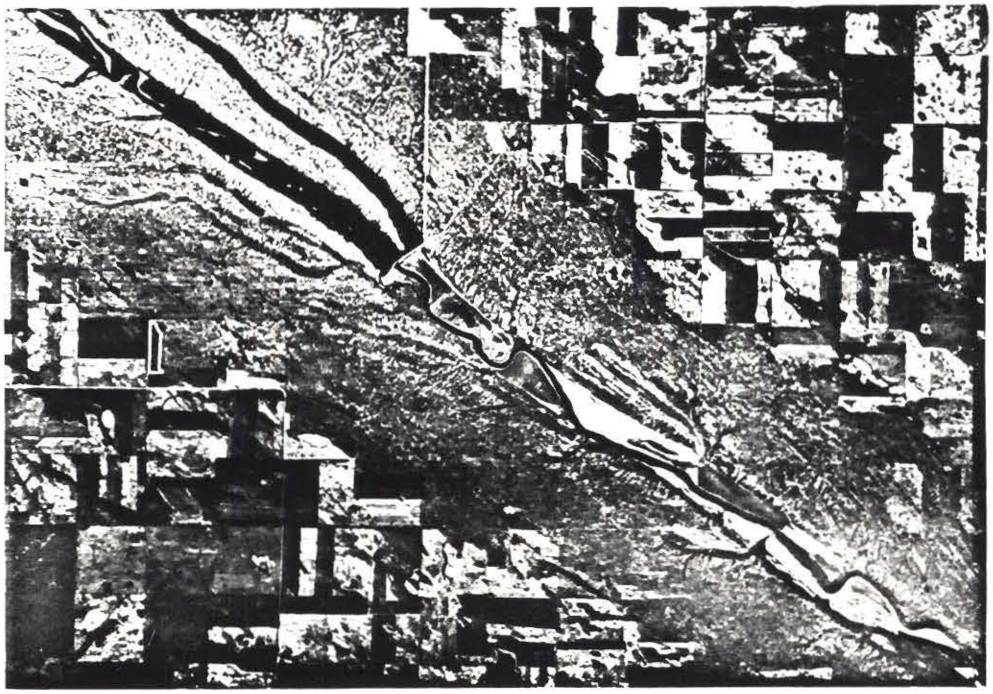
bedrock projections or at the inside of meanders, and commonly consist of huge foreset beds (Baker, 1973; Malde, 1968). However, the internal structure of the Set D Terrace bars is more appropriate to eddy bars. Eddy bars generally form in the mouths of tributaries or at other zones where there is expansion of flow, and consist of very poorly-sorted boulders, cobbles, sand, silt, and clay (Baker, 1973; Pardee, 1942).

The upper Souris Valley, which served as the southern outlet of Glacial Lake Regina, begins near Weyburn, Saskatchewan (Christiansen, 1956). Kehew (1979, 1982) and Kehew and Clayton (1983) proposed that discharges from a nonrepetitive catastrophic flood, starting from Glacial Lake Regina, formed the upper Souris spillway and enlarged the preexisting valleys of the lower Souris spillway and Des Lacs.

The Set D Terraces result from those discharges from Lake Regina, and apparently represent the last significant phase of development of the Souris and Des Lacs Valleys. Kehew (1982) and Kehew and Clayton (1983) noted a number of geomorphic features associated with the Souris and Des Lacs spillways that indicate a cataclysmic event. Among these are: a fluvially scoured zone with longitudinal grooves flanking the upper Souris and upper Des Lacs Valleys (Figure 25), lag-concentrated boulder armor fields within the scoured zones (Figure 26), a variety of streamlined hill types (Figure 27), anastomosing channels near Minot, North Dakota (Figure 28), major drainage divide crossings where the flood-water discharge exceeded the capacity of the channels (the Souris to the Des Lacs Valley and the Souris Valley to Glacial Lake Souris), and major channel bifurcations of the upper Souris spillway. Percussion marks were observed on cobbles and boulders at the Elcott bar (El 1)

Figure 25. Vertical air photo of fluvially scoured zone with longitudinal grooves flanking the Souris spillway, 14 km west of Macoun, Saskatchewan. Area shown is 5.4 km wide; north at top. Photo A21749-7, Saskatchewan Department of Energy, Mines and Resources.

Figure 26. Boulder armor field in the fluvially scoured zone flanking the Souris spillway, 5 km south of Hitchcock, Saskatchewan.



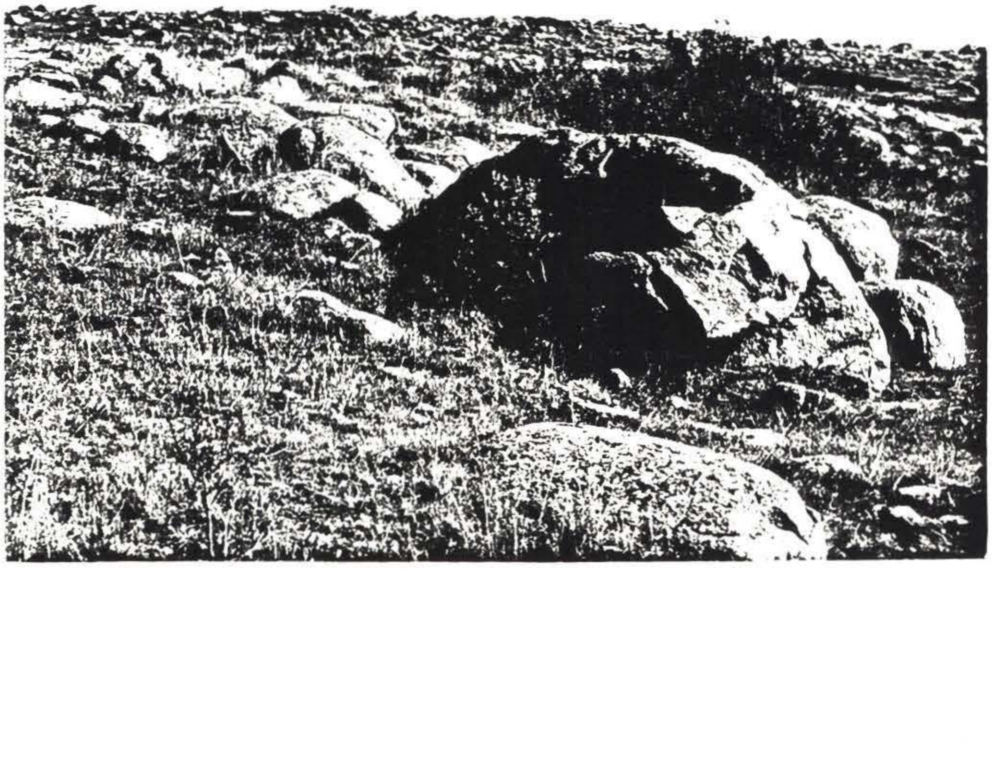


Figure 27. Vertical air photo of streamlined hill, composed of sand and gravel, in the Des Lacs spillway, about 8 km north of Burlington, North Dakota. Area shown is 2.7 km wide; north at top. Photo BAM-4M-148 of U.S. Department of Agriculture.

Figure 28. Topographic map of anastomosing channels 8 km west of Minot, North Dakota. Area shown is 14.9 km wide; contour interval is 1.5 m. Composite map made from portions of Deering SW, Deering SE, Surrey, and Sawyer NE, U.S. Geological Survey 7.5 minute topographic quadrangles.

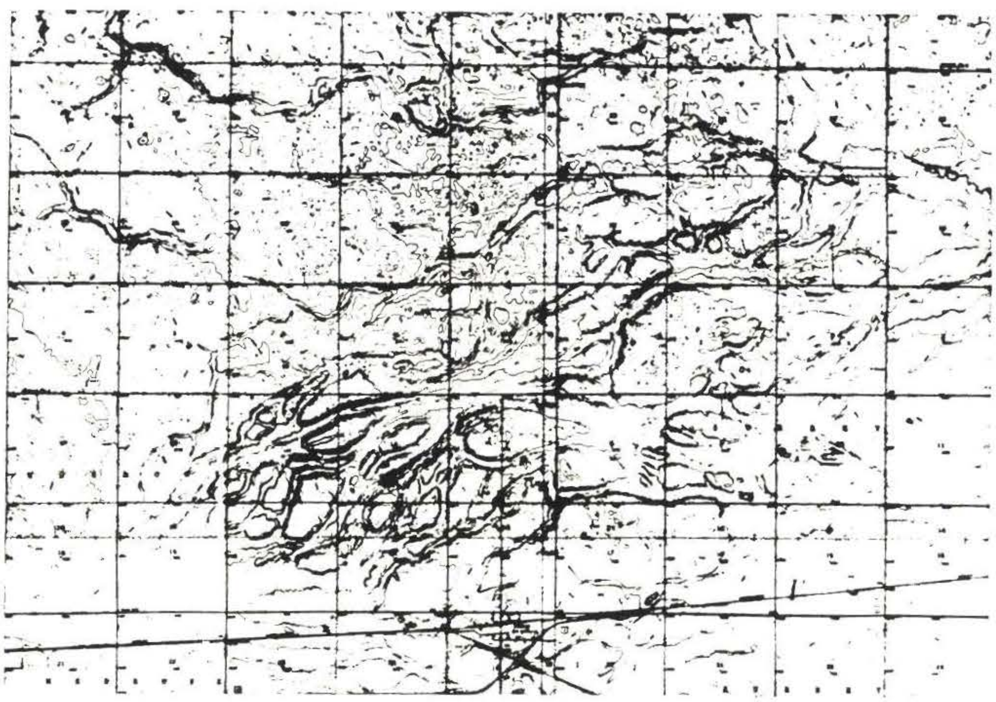


Figure 29. Percussion mark on a boulder at sample site EL 1, 8 km north of Elcott, Saskatchewan. Lens cap is 5 cm wide.

site during field work for this study (Figure 29). Percussion marks on boulders in the flood deposits of the Channeled Scabland have been attributed to the impact of large particles against one another during high velocity turbulent transport (Bretz, 1929).

The flood waters, resulting from the emptying of Lake Regina, not only formed the upper Souris, but also must have significantly enlarged the lower Souris and Des Lacs Valleys. The lower Souris Valley, before the flood, was an extension of the present Moose Mountain Valley, and consequently must have been about the same size. The Souris Valley, downstream of the Souris-Moose Mountain confluence, was incised about 40 m deeper than the Moose Mountain Valley by the flood. This stage of development, which includes the rapid draining of Glacial Lake Regina, is referred to as phase 5.

There are several indications that at least some stagnant ice existed throughout the scoured zone flanking the upper Souris spillway during phase 5. Kehew (1982) suggested that the longitudinal grooves in the scoured zone were formed by longitudinal roller vortices and that the more deeply incised channels were formed by the coalescing of these vortices. In several areas the longitudinal grooves are well developed and can be used as paleocurrent indicators. Because some areas show longitudinal grooves oriented toward the margin of the scoured zone, the water probably flowed onto ice (Figure 30). The very irregular boundary between the glacial collapse topography and the fluvially scoured zone is also best explained the former presence of ice there.

Several sites within the margins of the scoured zone appear not to have been scoured. Some of these non-scoured areas appear as though water flow was diverted around them (Figure 31). These features also indicate the presence of remnant ice. The land beneath the ice would be protected from flood erosion and would thus end up as a positive relief feature after the ice melted. The ice might have been more resistant to erosion because it assumed a streamlined equilibrium form rapidly, thus reducing the pressure drag. This would be analogous to the streamlined loess hills of the Scablands. The loess is hypothesized to have reached a streamlined equilibrium form rapidly and was preserved, while irregularities in the basalt surface generated macroturbulence and intensive erosion (Baker and Nummedal, 1978). It seems a likely possibility that the Lake Regina flood initially flowed on ice and subsequently cut down to the land surface as the flood progressed.

Textural Analysis

Particle size data were processed through a computer program written by Richard D. LeFever (verbal communication, 1984) to calculate various textural parameters. The determinations included: cumulative frequency, frequency, and statistical parameters including the mode, coarsest 1 percent, median, mean, sorting, and skewness. The latter three were determined using moment measures. The results of these calculations for each sample are shown in Appendix B. The individual sample results were averaged by drainage phases and are shown in Table 2.

Figure 30. Vertical air photo of longitudinal grooves indicating paleocurrent directions in the fluvially scoured zone flanking the Souris spillway, 8 km west of Estevan, Saskatchewan. Area shown is 9 km wide; north at top. Photo A21739-82 of Saskatchewan Department of Energy, Mines and Resources.

Figure 31. Vertical air photo of non-scoured, collapse topography (dashed lines) locality within the fluvially scoured zone flanking the Souris spillway, 14.5 km south of Halbrite, Saskatchewan. Area shown is 8 km wide; north at top. Photo A21654-4 of Saskatchewan Department of Energy, Mines and Resources.

TABLE 2

Average Phi Values of Textural Analysis by Phase

PHASE	MEDIAN	MEAN	SORTING	SKEWNESS	COARSEST 1 PERCENT	MODE
1	-4.00	-3.51	2.81	0.37	-7.05	-5.88
2	-1.32	-1.17	1.83	0.17	-4.23	-1.25
4	-4.83	-4.06	2.79	0.48	-7.15	-6.75
5	-4.90	-4.03	3.01	0.43	-7.46	-7.00

The textural values determined for the deposits of phases 1,4, and 5 are similar; those for phases 4 and 5 are nearly identical. Phase 2 deposits, however, are finer, better sorted, and less positively skewed than the other deposits (Figure 32).

The frequency and cumulative frequency curves (Figure 33) of each phase display the same relationship between the drainage phases as the textural parameters do; phase 2 is distinct from the others. Sediments of phases 1, 4, and 5 are almost exclusively bimodal and commonly show an inflection point at about -0.5 phi. Sediments of phase 2 are most commonly unimodal and have a relatively smooth cumulative frequency curve. Still another distinction between phase 2 deposits and the others is that phase 2 sediments consistently have a lower percentage of mud, reflecting better sorting.

The similarity of the exposures and the textural similarities of phases 1, 4, and 5 sediments suggest that the characteristics of the depositional flows were similar. There is strong geomorphic evidence that the flows of phase 5 were the result of the catastrophic flooding of Glacial Lake Regina (Kehew, 1982; Kehew and Clayton, 1983). By analogy, it is probable that the flows of phases 1 and 4 were also the result of floods. The fact that all the deposits associated with phases 1, 4, and 5 are homogeneous masses of unstructured and non-oriented gravels also indicates rapid aggradation of sediments in very turbulent waters (Baker, 1973). The bimodal distribution of these sediment sizes may imply a mixing of traction and suspended loads. The high turbulence generated as a result of the rapid flows and high bed roughness probably would have promoted such mixing.

Figure 32. Plot of sorting versus mean. The numbers plotted represent the value of the given phase textural analysis.

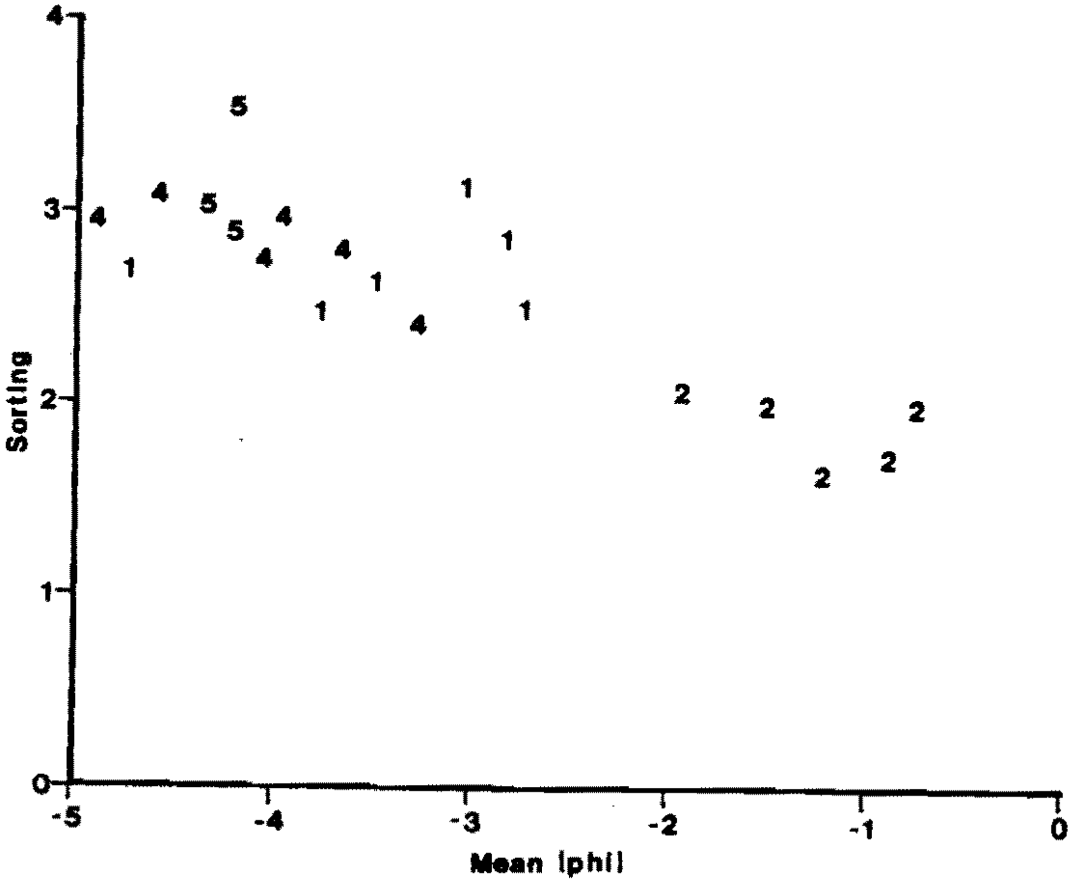
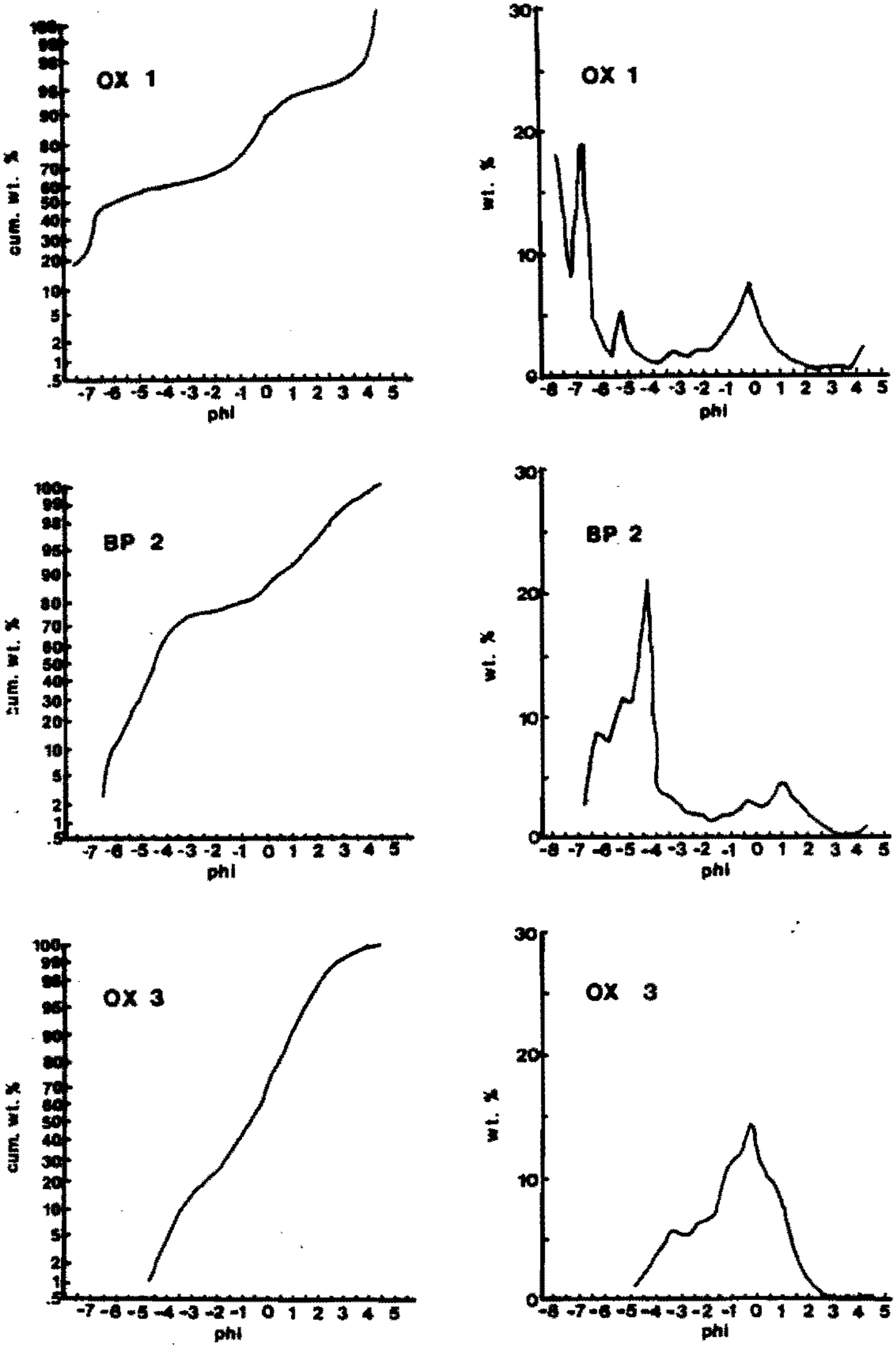

Sorting
Mean (phi)
4
3
2
1
0
-5
-4
-3
-2
-1
0

Figure 33. Representative frequency (right) and cumulative freqency (left) curves. Sample OX 1 is a phase 5 deposit, BP 2 is a phase 1 deposit, and OX 3 is a phase 2 deposit.


OX 1
cum. wt. %
phi
OX 1
wt. %
phi
BP 2
cum. wt. %
phi
BP 2
wt. %
phi
OX 3
cum. wt. %
phi
OX 3
wt. %
phi

The spatial variation of bedding, grain size, and sorting in the phase 2 deposits, together with the textural data, suggest a glacial meltwater origin for the sediments (Williams and Rust, 1969; Fahnestock, 1963).

Very Coarse Sand Lithology

Only minor differences exist in the very coarse sand lithologies associated with different drainage phases. Individual sample results are shown in Appendix C; phase averages are shown in Table 3. Phase 5 lithologies indicate a relatively stronger influence of the local bedrock than the other phases, possibly because the discharges were more erosive and cut through the Pleistocene drift. The absence of a clear distinction between the lithologies of each phase suggests that the influence of the local bedrock and local drift masked possible source area differences.

Paleohydraulic Characterization of the Discharges

Paleovelocity Estimating Techniques

Paleohydraulic methods have not yet established enough consistency so that one formula has been established as superior to others. For this reason, paleovelocity estimates were made by averaging the results of six different formulas, similar to the approaches used by Costa (1983a) and Bradley and Mears (1980). The six formulas were chosen on

TABLE 3

Coarse Sand Lithology Averages For Each Phase

(Percent: based on 200 grain counts per sample)

PHASE	CRYSTALLINE ROCK FRAGMENTS	CARBONATES	SHALE	OTHER SEDIMENTARY ROCK FRAGMENTS
1	60.7	26.3	8.1	4.9
2	62.5	26.7	5.9	4.9
4	61.8	26.4	6.4	5.4
5	59.0	23.2	8.7	9.1

the basis of their applicability to the study of coarse gravel and their degree of consistency with the small amount of empirical data available. Each of the formulas used determines velocity as a function of particle size. Unlike the methods employed by Costa (1983a) and Bradley and Mears (1980), formulas based on data compilations were avoided. Data compilations, because they often result in using different formulas based on the same data, may give undue weight to certain paleohydraulic approaches. Three of the formulas chosen have a theoretical basis; the other three used have an empirical basis--one derived for natural streams and the other two for riprap evaluation. Table 4 shows the velocity reconstruction methods used in this study.

Baker (1973), in a paleohydraulic study of the Channeled Scabland of eastern Washington, evaluated sediment transport using DuBoys' theoretically derived equation for critical tractive force (Tc):

$$Tc = Sf\ D\ S \quad ,$$

where Sf = specific weight of the transporting fluid, D = water depth, and S = energy slope. Baker calculated the critical tractive force in canyons that acted as constrictions in the flow path. By measuring the size of the transported boulders at these constrictions, a relationship between critical tractive force and particle size was determined:

$$Tc = 0.084\ d^{10/9} \quad ,$$

TABLE 4

Paleovelocity Reconstruction Methods

a. Baker (1973; plus Novak, 1973)

$$Vb = 0.367\ d^{0.37}$$

b. Helley (1969)*

$$Vb = 3.276 \left[\frac{(Ss - 1)\ dl\ (ds + di)^2\ MRl}{C'd\ ds\ dl\ (MRd) + 0.178\ di\ dl\ (MRl)}\right]^{0.5}$$

$$MRl = (di/4) \cos \theta + (3/16\ ds^2)^{0.333} \sin \theta$$

$$MRd = 0.1\ ds \cos \theta + ((3/16\ ds^2)^{0.333} \cos \theta - (di/4) \sin \theta)$$

c. Bradley and Mears (1980; modified by Costa, 1983)

$$Vb = (2\ (Ss - Sf)\ di\ g\ f\ /\ Sf\ (Cl + C'd))^{0.5}$$

d. Fahnestock (1963)

$$Vb = 0.337\ d^{0.385}$$

e. Mavis and Laushey (1949)

$$Vb = 0.199\ d^{0.444}$$

f. Strand (1973)

$$Vb = 0.187\ d^{0.5}$$

g. Average (this report)

(1) $Vb = 0.232\ d^{0.44}$ (2) $V = 0.287\ d^{0.44}$

*English units used; metric used in all other methods.

TABLE 4 (continued)

LIST OF SYMBOLS

C'd - adjusted drag coefficient; dependent on CFS (See Helley, 1969)

Cl - lift coefficient (0.178)

d - intermediate particle diameter (mm)

di - intermediate particle axis (mm)

dl - long particle axis (mm)

ds - short particle axis (mm)

f - coefficient of static friction (0.7)

g - gravitational acceleration (9.81 m/s^2)

MRd - drag turning moment

MRl - lift turning moment

Sf - specific weight of transporting fluid (1000 Kg/m^3; 9800 N/m^3)

Ss - specific weight of particle (2650 kg/m^3; 26,000 N/m^3)

V - mean channel velocity (m/s)

Vb - channel bed velocity (m/s)

θ - imbrication angle (degrees)

where d = intermediate particle diameter. Novak (1973), by compiling pre-existing data, arrived at a formula to convert critical shear stress to velocity (V):

$$V = (Tc \;/\; 1.7066)^{1/3} \quad .$$

The equation Baker (1973) derived was rewritten for this study using Novak's conversion to demonstrate a relationship between velocity and particle size (Table 4a).

The second theoretically based method used is one developed by Helley (1969). Bed velocities necessary to initiate motion were calculated by balancing the turning moments of fluid drag and lift against the resisting moment of the submerged particle weight (Table 4b). Helley's method is different from most others in that it considers most physical characteristics of the particle, the particle orientation, and the lift force. Helley (1969, p.4) used an adjusted drag coefficient which is dependent on the Corey shape factor ($CSF = ds/(di\ dl)^{1/2}$). The adjusted drag coefficient is corrected for the portion of the particle resting on the bed that is not exposed to the fluid drag force, by multiplying the drag coefficient by 0.75. The initial imbrication angle of the particles is not known for this study; an arbitrary value of 10 degrees was used. This value was also used by Costa (1983a) for a study where the initial imbrication angle was unknown.

The third theoretical approach used was developed by Bradley and Mears (1980) and simplified by Costa (1983a) (Table 4c). This relationship balances the drag, lift, and the gravitational and frictional forces to compute the bottom velocity. At incipient motion, the fluid drag force on the particle is equal to the net frictional force between the particle and the bed. The fluid drag force is dependent on the drag coefficient, the size of the particle, the density of the fluid, and the bottom velocity. The resisting frictional force is dependent on the densities of the fluid and particles, the particle volume, gravitational acceleration, and the coefficient of static friction. Bradley and Mears (1980) specifically developed this method for the study of coarse particles deposited by deep, low-gradient flows--a situation probably analogous to the late Wisconsinan flows that developed the drainageways in the area of study.

Fahnestock (1963), in a classic study of the White River of Washington, derived an empirical formula relating transporting bottom velocity to particle size (Table 4d). The White River is a meltwater stream from the Emmons Glacier on the northeast side of Mount Rainier. Particles in transport were caught either by a screen or by hand, and their dimensions were determined. Bed velocity measurements were made with a current meter just upstream from the point of capture. The relationship between particle size and velocity was then determined.

Mavis and Laushey (1949) and Strand (1973) determined sediment transport formulas on the basis of riprap studies, mostly in flumes. The derivation of the formula used by Mavis and Laushey (1949) is based on the assumption that the tractive force on a grain varies directly with the hydrodynamic pressure on the grain (Table 4e). Strand (1973), by using the abundant data of the United States Bureau of Reclamation on riprap stability, developed a formula to define the minimum diameter needed for riprap stability for any given bottom velocity (Table 4f).

At each sample location in this study where the average dimensions of the ten largest particles were determined, each of the six methods was applied to calculate an estimate of paleovelocity. An average of the methods for each of the locations was determined, from which a regression equation was calculated (Table 4g). This equation is within plus or minus 14 percent of the other six methods, a relatively small margin of error considering the diversity of the approaches used.

After determination of bed velocity, the values were converted to mean channel velocities to be consistent with most published data. There is varied opinion on what conversion factor should be used. Bradley and Mears (1980) and Strand (1973) chose to use a conversion factor of 1.43, a ratio used for modern-day streams and canals. Fahnstock (1963) noted that turbulence in flow can cause substantial mixing; in the White River, the bed velocity approximated the mean velocity. Conversely, Baker (1973) suggested that very deep flows may have large differences between mean and bed velocities. In parts of the Channeled Scabland, Baker estimated the mean velocity to have been twice the bed velocity. Costa (1983a) chose the formula,

$$V = 1.2\ Vb$$

as a compromise; the same has been done in this study. The calculated mean velocities and bed velocities, as determined from each of the six methods, for each of the sites sampled, are shown in Table 5.

When interpreting paleovelocity data, one should be aware of possible sources of error. The amount of turbulence in the flow can have a large effect on the entrainment of the particles. A local doubling of velocity will lead to a four-fold increase in drag force (Fahnestock, 1963). Entrainment of particles can also be induced by undercutting of a bank, impact with a particle during transport, and by removal of finer materials surrounding a larger particle. Each of these mechanisms would cause entrainment at velocities lower than predicted (Fahnestock, 1963; Bradley and Mears, 1980).

Paleovelocity Estimates for Phases

On the basis of the divisions of sample sites into geomorphic phases made earlier, the range and average paleovelocity of each phase was calculated. Shown in Figure 34 are the results of these calculations. Paleovelocity estimates of phase 3 are not depicted because none of the gravel deposits observed was assigned to this phase of development. Because the flows of phases 3 and 4 both initiated from Glacial Lake Arcola, and because phase 4 was depositional and phase 3 was erosional, it can be assumed that velocities achieved during phase 3 were greater than those of phase 4.

As expected, the phase 5 deposits, resulting from the flooding of Glacial Lake Regina, required the highest flow velocities. Velocities

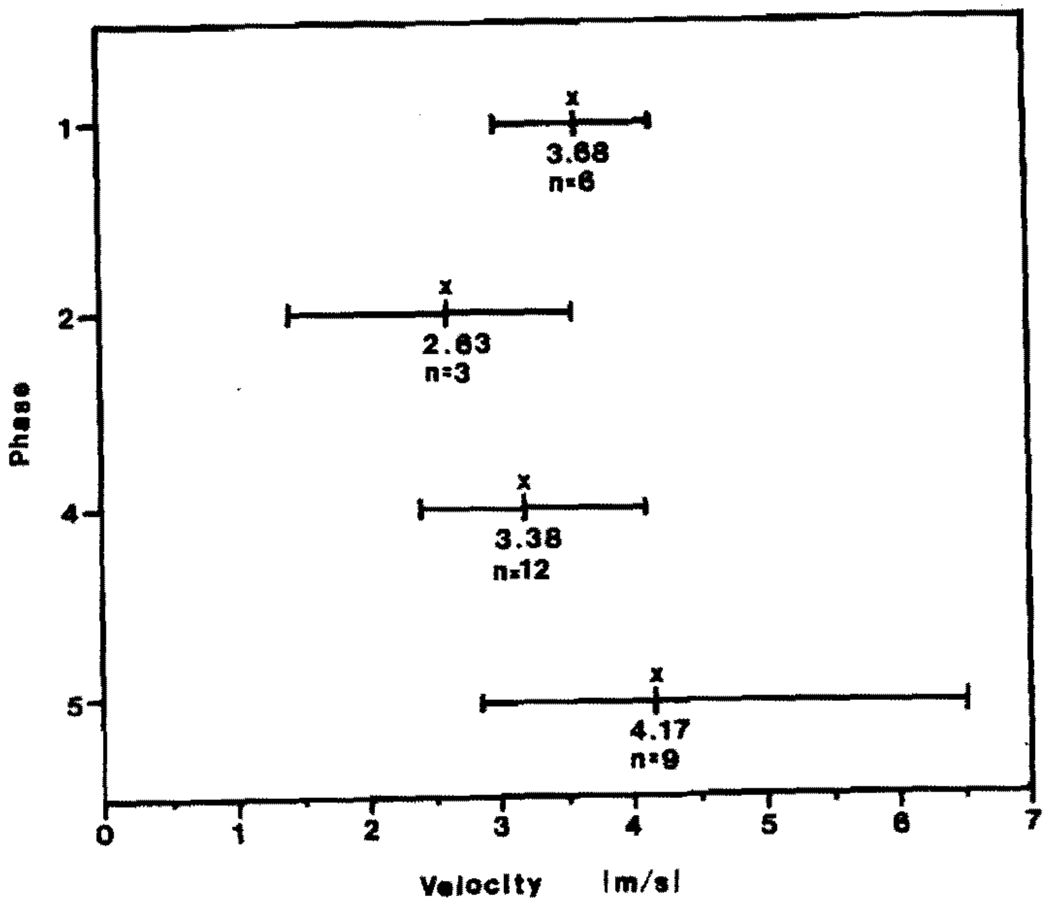
TABLE 5

Calculated Mean Velocites (m/s) For Each Site Sampled

SAMPLE	METHOD (see table 4) a	b	c	d	e	f	g(2)	PHASE
Souris (in Canada) samples								
RB 1*	3.19	2.47	2.99	3.20	2.67	3.47	3.60	5
EL 1	5.15	4.67	6.34	5.25	4.73	6.61	6.55	5
EL 2	3.56	2.70	3.43	3.58	3.04	4.02	4.07	5
EL 3	3.47	2.61	3.12	3.49	2.95	3.49	3.85	5
OX 1	3.51	2.90	3.86	3.53	2.99	3.94	4.15	5
OX 2	3.25	2.56	3.47	3.25	2.72	3.55	3.76	4
OX 3	1.47	0.78	1.29	1.43	1.06	1.23	1.45	2
FD 1	2.60	1.90	2.55	2.58	2.08	2.63	2.87	2
Moose Mountain samples								
GE 1	3.16	2.39	3.17	3.17	2.64	3.43	3.59	2
OX 4	3.18	2.40	3.08	3.18	2.65	3.45	3.59	4
OX 5a	3.33	2.56	3.27	3.34	2.80	3.67	3.80	4
OX 5b	3.00	2.24	3.09	3.00	2.48	3.19	3.40	4
OX 5c	2.92	2.16	2.89	2.92	2.40	3.08	3.28	4
Souris (U.S.) samples								
NG 3	2.90	2.14	2.86	2.89	2.38	3.04	3.24	4
NG 4	3.58	2.85	3.46	3.60	3.06	4.05	4.12	4
GN 2b	3.03	2.22	2.91	3.03	2.51	3.23	3.38	4
GN 2a	2.47	1.81	2.49	2.45	1.96	2.45	2.74	4
GN 1	3.15	2.35	3.07	3.15	2.63	3.41	3.55	4
GR 1	2.23	1.57	2.10	2.20	1.73	2.14	2.40	4
BS 5	3.38	2.51	3.51	3.39	2.86	3.75	3.85	5
MN 2	4.03	3.31	4.28	4.08	3.53	4.75	4.80	5
MN 3	3.20	2.39	3.13	3.20	2.67	3.48	3.61	1
SY 1	2.91	2.21	3.12	2.90	2.39	3.06	3.32	4
SW 1	3.31	2.52	3.22	3.32	2.78	3.64	3.76	5
Des Lacs samples								
CP 1	2.71	1.98	2.70	2.70	2.19	2.78	3.01	1
FX 1	3.49	2.90	4.04	3.51	2.97	3.91	4.16	1
FB 1	3.03	2.20	3.00	3.03	2.51	3.24	3.41	1
BP 2	3.38	2.73	3.61	3.38	2.84	3.72	3.94	1
BP 3	2.61	1.82	2.49	2.60	2.10	2.65	2.86	5
BL 1	3.45	2.66	3.44	3.47	2.93	3.85	3.96	1

* See figure 4 for sample locations.

Figure 34. Graph of paleovelocity estimates for each phase. The bars shown represent the range of estimates for each phase based on the average of the calculated values for each sample; the x represents the average.



Phase
1
2
4
5
x
3.68
n=6
x
2.63
n=3
x
3.38
n=12
x
4.17
n=9
0
1
2
3
4
5
6
7
Velocity |m/s|

exceeding 6 m/s are calculated to have occurred in the vicinity of Elcott, Saskatchewan during phase 5.

Velocities achieved during phase 1,3, and 4 were very high. These high velocities, along with the homogeneous nature of the deposits, imply steady and sustained rapid flows with rapid aggradation of sediment. The paleovelocity estimates for phase 2, although relatively low, are also high compared to most streams today. However, it should be kept in mind that there is significant variability in the texture of these deposits, implying that only occasional high discharges are represented by the calculated paleovelocity estimates.

Paleodischarge Estimates

As with paleovelocity, numerous formulas are used to estimate paleodischarge. However, unlike paleovelocity methods, all paleodischarge estimates require more information than just particle size. Parameters frequently required for these estimates are flow depth, energy slope, channel roughness, velocity, and hydraulic radius.

Four different formulas were applied in an attempt to arrive at paleodepth estimates (Table 6). These formulas were chosen on the same basis as the paleovelocity formulas--their applicability to the study of coarse gravels, and their degree of consistency with existing data. Once depth (D) is determined, accurate discharge estimates can be made.

Calculations for the Souris spillway at Elcott, Saskatchewan (sample locality EL 1) illustrate the problems involved in paleodepth estimates. This locality was influenced only by events during phase 5

TABLE 6

Paleodepth Reconstruction Methods

---

a. DuBoy:

$$D = Tc \ / \ Sf \ S$$

b. Costa (1983):

$$D = w \ / \ Sf \ S \ V$$

$$w = 0.009 \ d^{1.686}$$

c. Baker (1974):

$$D = 0.000102 \ d \ / \ S$$

d. Manning:

$$D = (n \ V \ / \ S^{1/2})^{3/2}$$

D - average water depth
n - Manning roughness coefficient
S - energy slope (m/m) (usually considered the channel bottom slope)
w = unit stream power (N/m/s)

and, therefore, should have the most straightforward results. The paleovelocity determined for the EL 1 site is 6.55 m/s; the slope is 0.000207. Paleodepth calculations for this site (Table 7) vary by more than an order of magnitude, and are all greater than suggested by field evidence. For these reasons, paleodischarges were not estimated through the use of existing formulas by incorporating the paleodepth estimates.

As an alternative method, paleodischarges were estimated by determining the cross-sectional areas of the spillways and multiplying them by the paleovelocity estimate at that point. At least some portion of a channel associated with each phase, except phase 1, has remained unaltered so that the form and dimensions of the channel could be reasonably estimated. To compensate for possible inaccuracies, minimum and maximum depths were estimated; this method resulted in a range of calculated discharges for each profile. Channel dimensions associated with phase 1 were estimated on the basis of the elevation of gravel-bedrock contacts and the lateral distribution of the phase 1 deposits.

Paleodischarges were calculated at five valley profile locations (Figure 35) chosen to represent each of the valley segments and be in close proximity to a gravel deposit. The profiles and paleohydraulic estimates at each location are shown in Figures 36, 37, 38, 39, and 40.

As with the paleovelocity estimates, range and average paleodischarge values were calculated for each phase (Figure 41). The relationship shown between the different phases is similar to that of velocities; the highest discharges occurred during phase 5 and the lowest during phase 2. The phase 2 discharge estimates represent maximum discharges only periodically achieved; the estimates for the

TABLE 7

Paleodepth Sample Calculations For Sample Locality EL 1

METHOD

a. $D = Tc / Sf\ S$:

$Tc = 479.6\ kg/m^2$
$S = 0.000207$
$Sf = 1000\ kg/m^3$

$D = 2316.9\ m$

b. $D = w / Sf\ S\ V$; $w = 0.009\ d^{1\ 686}$

$d = 1253\ mm$
$w = 1504.5$
$S = 0.000207$
$Sf = 9800\ N/m^3$
$V = 6.55\ m/s$

$D = 113.2\ m$

c. $D = 0.000102\ d / S$:

$d = 1253\ mm$
$S = 0.000207$

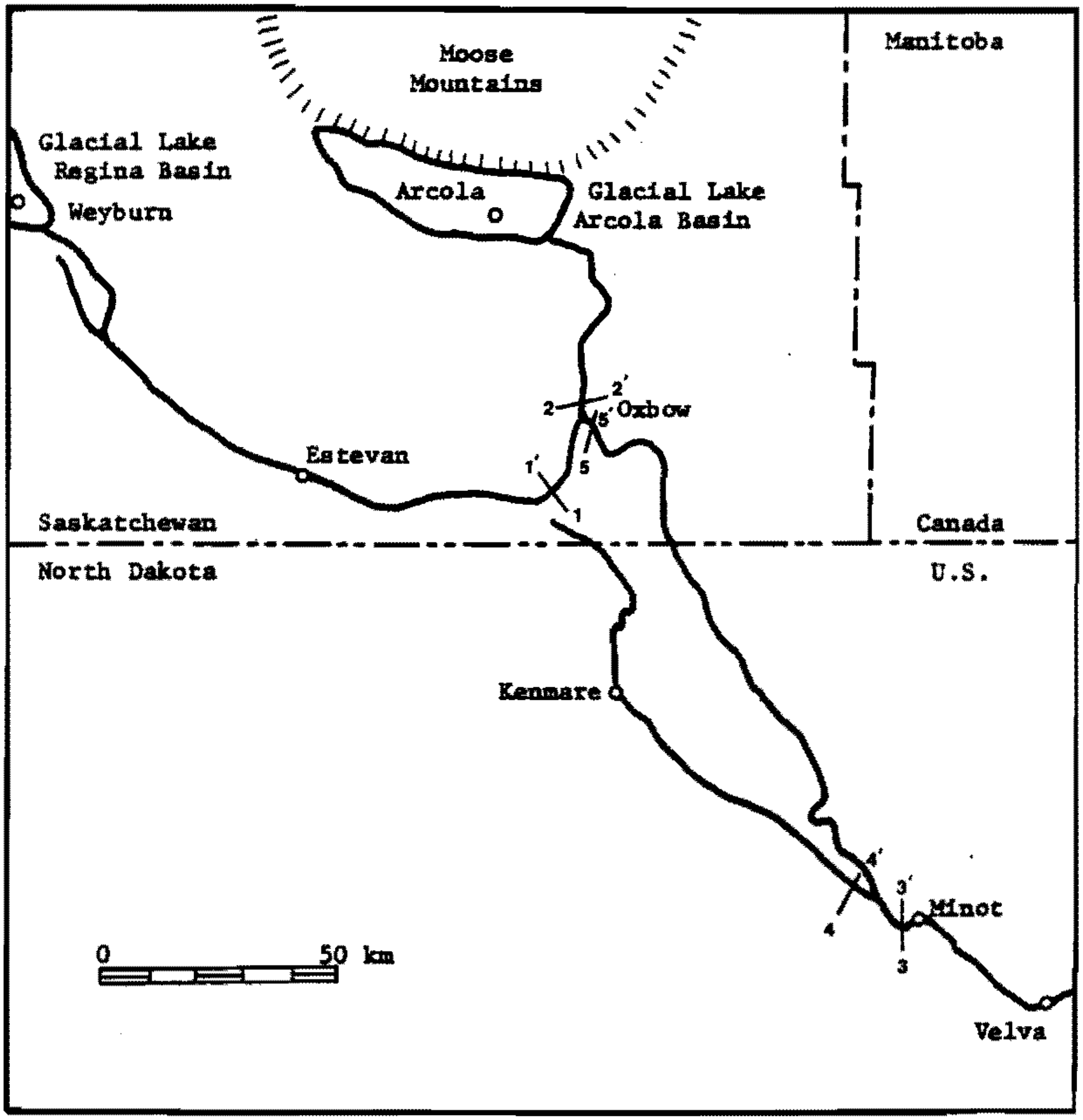
$D = 617.4\ m$

d. $D = (n\ V / S^{1/2})^{3/2}$:

$n = 0.04$
$V\ 6.55\ m/s$
$S = 0.000207$

$D = 77.7\ m$

Figure 35. Location of five valley profile sites chosen for paleodischarge estimates.



Moose Mountains
Manitoba
Glacial Lake Regina Basin
Weyburn
Arcola
Glacial Lake Arcola Basin
2
2'
Oxbow
5'
5
Estevan
1'
1
Saskatchewan
Canada
North Dakota
U.S.
Kenmare
4'
3'
Minot
4
3
0
50 km
Velva

Figure 36. Profile 1-1' of the Souris Valley at Elcott, Saskatchewan (see Figure 35 for location).

Figure 37. Profile 2-2' of the Moose Mountain Valley just north of Oxbow, Saskatchewan (see Figure 35 for location).

Explanation for Profiles

15 m
500 m
Scale

V=Velocity
D=Depth
Q=Discharge

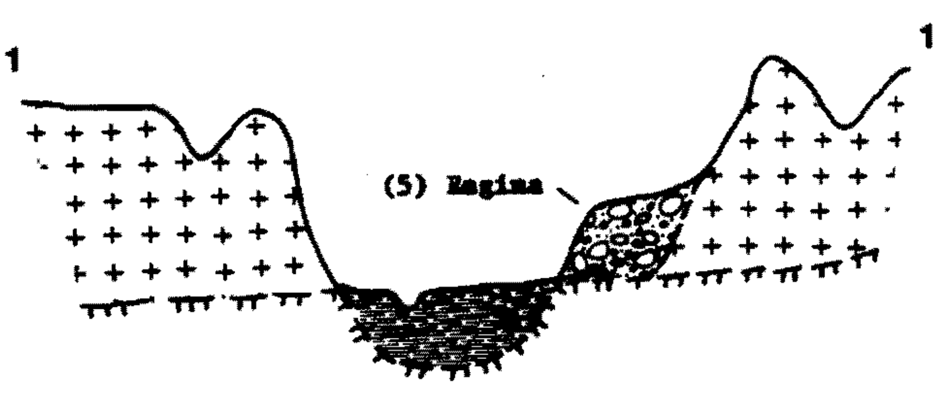
Sand and gravel

Bedrock

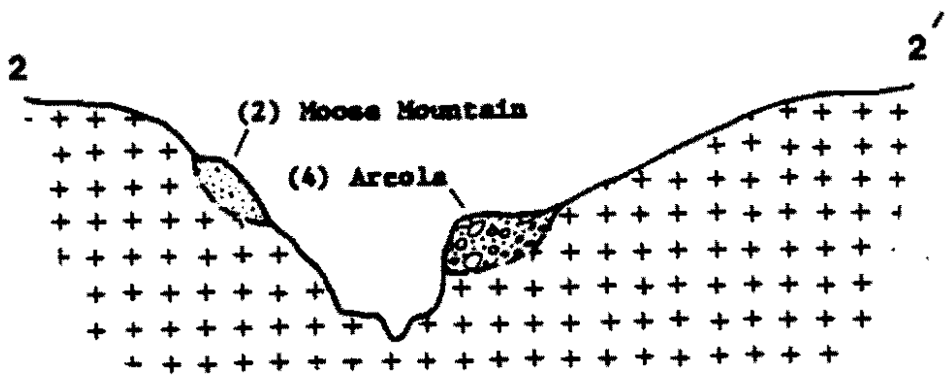
Till

Alluvium

The number in parentheses refers to the associated phase.



(5) $V=6.6$ m/s $D=30–65$ m $Q=1.2–4.8 \times 10^5$ m^3/s

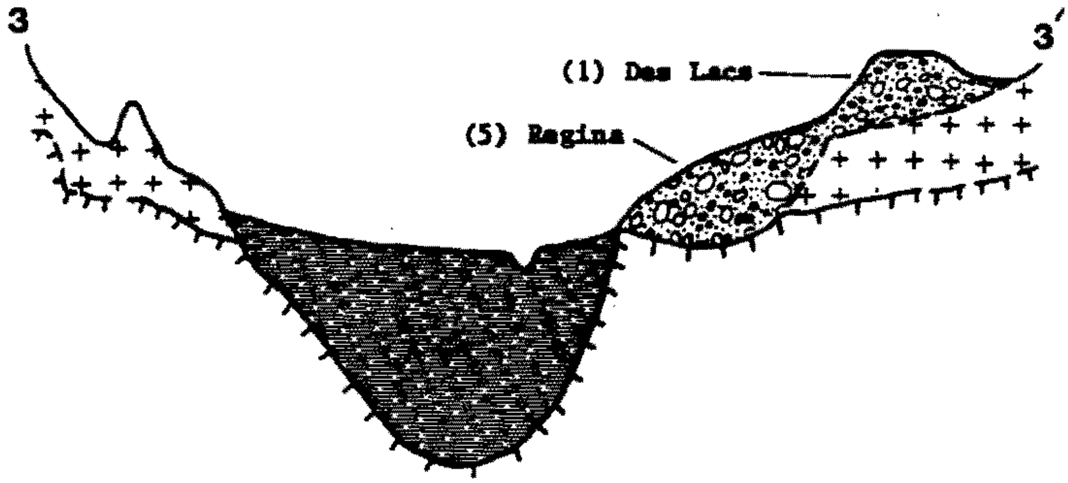


(2) $V=1.2$ m/s $D=1–7$ m $Q=5.6 \times 10^2–4.5 \times 10^3$ m^3/s

(4) $V=3.6$ m/s $D=5–15$ m $Q=1.0–3.0 \times 10^4$ m^3/s

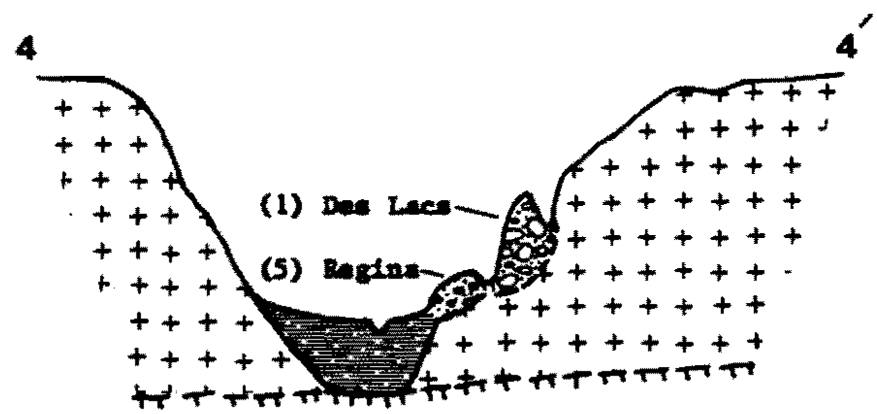
Figure 38. Profile 3-3' of the Souris Valley at Minot, North Dakota (see Figure 35 for location).

Figure 39. Profile 4-4' of the Des Lacs Valley just north of Burlington, North Dakota (see Figure 35 for location).



(1) V=3.6 m/s D=5-20 m $Q=1.6\text{-}7.4 \times 10^4\ m^3/s$

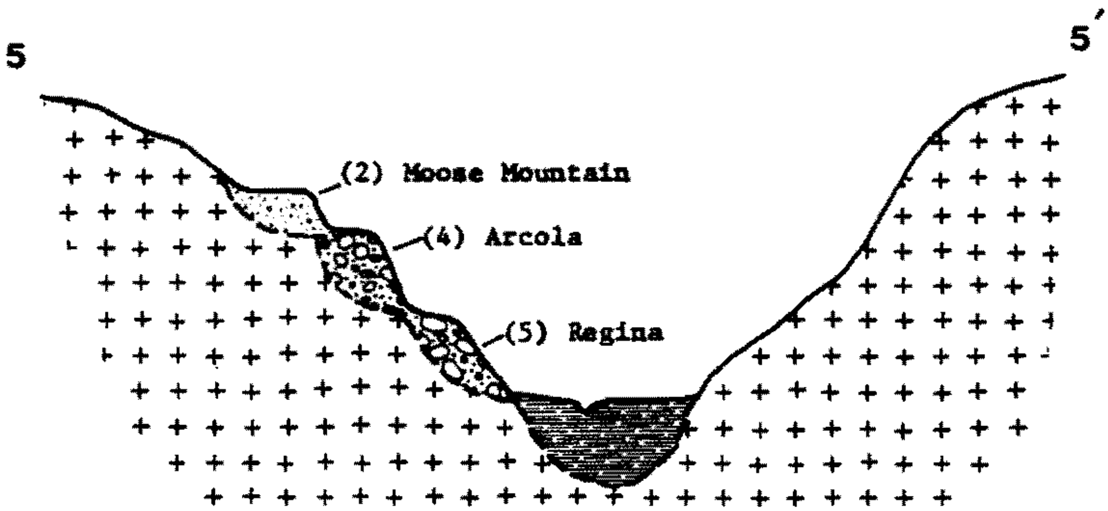
(5) V=4.8 m/s D=40-75 m $Q=1.8\text{-}5.1 \times 10^5\ m^3/s$



(1) V=3.9 m/s D=5-20 m $Q=1.5\text{-}7.2 \times 10^4\ m^3/s$

(5) V=2.9 m/s D=20-45 m $Q=4.3\text{-}9.5 \times 10^4\ m^3/s$

Figure 40. Profile 5-5' of the Souris Valley at Oxbow, Saskatchewan (see Figure 35 for location).

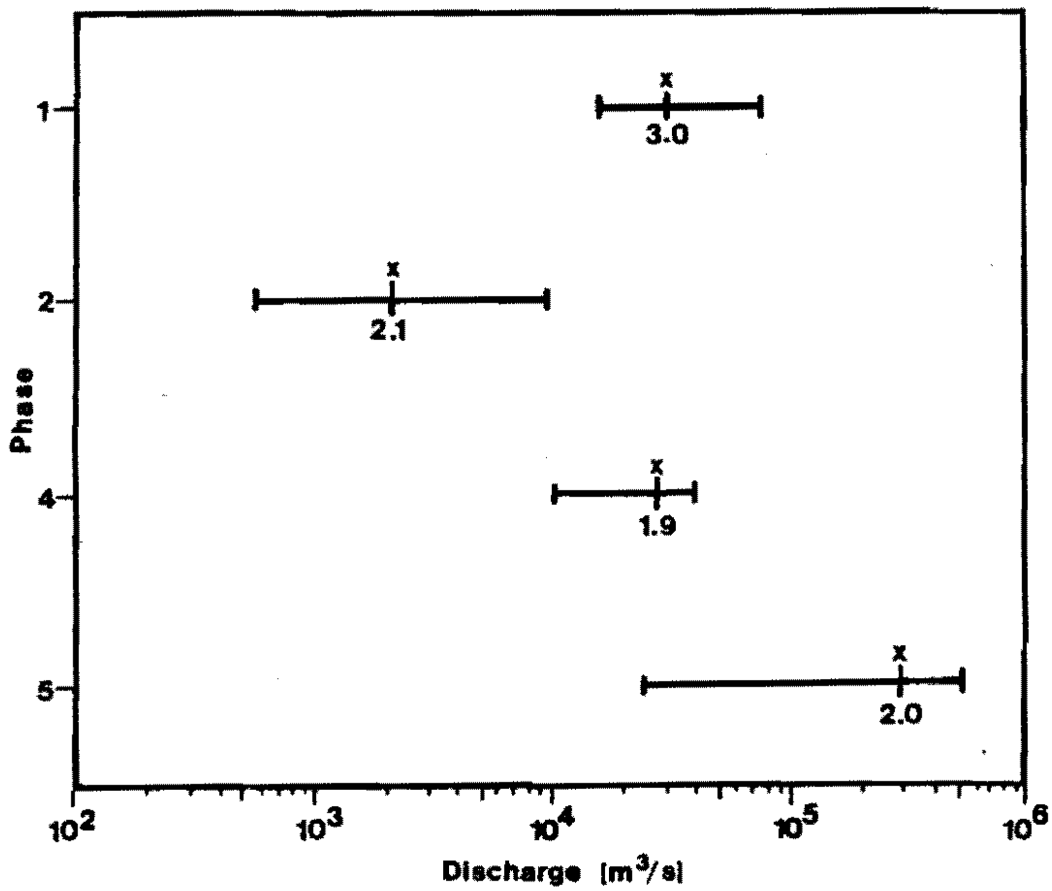


(2) V=2.6 m/s D=1-7 m $Q=1.6-9.3 \times 10^3\ m^3/s$

(4) V=3.7 m/s D=5-15 m $Q=1.3-3.9 \times 10^4\ m^3/s$

(5) V=4.2 m/s D=30-40 m $Q=1.0-1.6 \times 10^5\ m^3/s$

Figure 41. Paleodischarge estimates for each phase of drainage development. The bars shown represent the range of estimates; the x represents the average.



Phase
1
2
4
5
3.0
2.1
1.9
2.0
10^2
10^3
10^4
10^5
10^6
Discharge [m^3/s]

other phases probably represent average conditions. For reasons stated previously, phase 3 discharges probably were greater than those which occurred during phase 4.

Other Paleohydrologic Estimates

Paleohydrologic estimates other than velocity and discharge determinations are commonly made in studies of Pleistocene sediments. Such calculations have been made to determine the height of possible dams impounding glacial lakes, the water volume of lakes, and the duration of discharges required to drain lakes.

Costa (1983b) derived a formula relating the discharge, resulting from the failure of a dam, to the height of a dam:

$$H = 0.0969\ Qm^{0.541} ,$$

where H is the height of the dam and Qm is the maximum discharge. This relationship was based on data collected from 22 historic dam failures. The equation was applied to the study of phase 5, the catastrophic flood from Glacial Lake Regina. The average maximum discharge estimated to have been achieved during the flood is $2.0 \times 10^5\ m^3/s$; this yields a value of 71.5 m for the height of the dam. Given the presence of the nearly flat topography at the outlet to Lake Regina, any dam must have been glacial ice.

Calculations were also made to determine the height of dams necessary to cause in the discharges achieved during phases 1 and 4. These calculations were made on the assumption that these phases also involved a flood because of the similarity of the deposits of these phases to those of phase 5. Using Costa's equation, a dam height of 20.0 m is calculated for Lake Arcola (Phase 4) and 25.6 m for a dam of an unknown lake during phase 1. The dam height estimates for these three phases appear high. The major reason for this discrepancy probably is the scale problem of using a formula developed for artificial impoundments to predict the behavior of much larger Pleistocene proglacial lakes.

Clague and Mathews (1973) compiled data on lake volumes and discharges associated with jökulhlaups. A jökulhlaup is an Icelandic term for a glacier outburst which entails a sudden release of ice-impounded water. A formula was developed by Clague and Mathews (1973) relating maximum discharge to source lake volume using data from 10 jökulhlaups, including that from Glacial Lake Missoula:

$$V = 1607.5\ Qm^{1.49} \quad ,$$

where V is the lake volume. The application of this equation indicates that the volume of Lake Regina was 1.3 x 10^{11} m^3. This figure is 1.7 times larger than the value Kehew and Clayton (1983) estimated assuming an average depth of 8 m and an area of 9300 km^2.

Using the estimated discharges of phase 4, a volume of 3.8×10^9 m^3 is obtained for Lake Arcola; this implies an average depth of 6.5 m, assuming an area of 580 km^2. The application of Clague and Mathews' (1973) equation to the discharges of phase 1 implies a lake with a volume of 7.5×10^9 m^3. As stated previously, there is no known source lake for the phase 1 discharges.

Calculations of the duration of the discharge events were made on the basis of estimated discharges and lake volumes. The results of these and the other paleohydrologic calculations are summarized in table 8. Only a single value, based on the average paleodischarge estimates, has been calculated to allow for ease of comparison. All the durations show very rapid draining of the source lakes; this is typical during failure of ice dams during jökulhlaups (Stone, 1963; Marcus, 1960).

The paleohydrologic calcuations have not been applied to discharge estimates of phase 2 because the related sediments clearly exhibit characteristics typical of outwash deposits, in contrast to the sediment associated with phases 1,4, and 5.

There are large uncertainties associated with the paleohydrologic information obtained from the discharges of phases 1, 4, and 5. However, the relative magnitude associated with these estimates should be correct. In addition, these estimates provide some information on the sizes of lakes and ice dams not obtainable by geologic mapping.

TABLE 8

Summary of Paleohydrologic Results

PHASE	DISCHARGE (m^3/s)	(a) HEIGHT OF DAM (m)	(b) LAKE VOLUME (m^3)	AVERAGE LAKE DEPTH (m)	FLOOD DURATION (days)
1	3.0×10^4	25.6	7.9×10^9	?	3.0
4*	1.9×10^4	20.0	3.8×10^9	6.6	2.3
5**	2.0×10^5	71.5	1.3×10^{11}	13.8	7.5

* Glacial Lake Arcola area: 580 km^2

** Glacial Lake Regina area: 9400 km^2 (Kehew and Clayton, 1983)

DISCUSSION

Drainage Development of the Valleys of Study

Phase Descriptions

During the past 30 years, various methods of study have been used to interpret the origin of the Souris, Des Lacs, and Moose Mountain Valleys and the gravel deposits within them. Interpretations have varied from kame terraces (Lemke, 1960), to meltwater valleys (Dury, 1964), to outlet valleys of glacial lakes (Christiansen, 1956, 1961), and to catastrophic glacial lake flood channels (Kehew, 1979; Kehew, 1982; Kehew and Clayton, 1983). A problem in trying to decipher the history of drainage development is the failure of any previous studies to study the full length of all of the valleys and make full use of all geologic evidence.

Generally speaking, the more recent studies have been more thorough. This logical development has caused the suggested origins of the valleys to change from a relatively gradual process (formation of kame terraces), to a very rapid process (formation of cataclysmically-induced spillways). An objective of this study is to make full use of all aspects of geologic information in order to understand the complete developmental history of the valleys. Three general approaches were used to determine and characterize the development of the Souris, Des Lacs, and Moose Mountain Valleys: (1) geomorphic relationships, including terrace correlations, valley form, and morphologic relationships between the gravel deposits and spillways; (2) textural analysis of sediments and the very coarse sand lithologies; and (3) paleohydraulic interpretations based on the ten largest particles at each deposit. The geomorphic relations proved to be the most useful in distinguishing the different stages of development and provided the framework into which the other results could be integrated. On the basis of the geomorphic evidence, five phases of drainage development were recognized.

Phase 1: During phase 1, development of the Des Lacs Valley was initiated. It is possible that partial development of the Des Lacs Valley had occurred prior to the last glacial advance. Lemke (1960, p. 57) described a gravel pit, involving collapsed and/or sheared sand and gravel overlain by till, located just to the south of Donnybrook in the Des Lacs Valley. This and other evidence led Lemke to suggest that parts of the Des Lacs Valley were formed prior to the last glacial advance. The complex history of this valley makes it difficult to decipher the pre-phase 1 valley form and the amount of erosion caused by the phase 1 discharges.

The deposits assigned to this phase consist of sandy gravel (mean: -3.51 phi) in relatively high positions on the valley walls. The deposits are generally unstructured and homogeneous. Paleohydraulic estimates indicate velocities of 3.7 m/s and discharges of 3.0×10^4 m^3/s. The character of these deposits, like those of phases 4 and 5, indicates rapid aggradation of sediment with minimal velocity fluctuations of the depositional flows.

A glacial outwash origin for these deposits does not seem likely. Many studies indicate that variability in sediment size and sorting are basic to meltwater streams (Fahnestock, 1963; Miall, 1977). Williams and Rust (1969, p. 668) state that vertical and lateral textural changes are apparent over intervals of centimetres because of frequent channel abandonments and fluctuations of discharge.

The sedimentologic evidence suggests that the discharges responsible for the phase 1 deposits occurred over a short period of time and were not repetitive. The most likely sources of large volumes of water which could discharge nonrepetitively are glacial lakes. A lake volume of 7.9×10^9 m^3 has been calculated based on discharge estimates. Freers (1973) mapped a small amount of glacial lake sediment, about 35 km^2, in the Bowbells, North Dakota region, but the lake would not have been large enough to produce the phase 1 discharges. This still does not exclude the possibility of a substantial supraglacial lake.

Phase 2: The stratigraphically highest sand and gravel deposits of the Moose Mountain and lower Souris Valleys were laid down during phase 2. These deposits occur in a wide, shallow valley typical of an outwash valley. The valley head is about 15 km south of the Moose Mountains, but is abruptly truncated by Glacial Lake Arcola sediments. This meltwater valley apparently developed when the ice sheet split around the Moose Mountains forming the Moose Mountain Lobe to the east and the Weyburn Lobe to the west (Christiansen, 1956).

The deposits assigned to phase 2 consist of, on the average, gravelly sand (mean: -1.17 phi). The sediment is quite variable texturally, as is typical of outwash sediment. Paleohydraulic estimates indicate that maximum velocities of 2.5 m/s and maximum discharges of 2.1×10^3 m^3/s were achieved.

Phase 3: During phase 3, the Moose Mountain and Souris outwash valleys were deeply incised, inducing landslides along much of the Souris Valley. The resulting channel can be traced to the southeast corner of Glacial Lake Arcola, indicating that discharges from that lake were the source of the erosive flows. It seems likely that the velocities and discharges achieved during phase 3 exceeded those of phase 4. Phase 4 flows had the same source area as 3, but were depositional.

Phase 4: The phase 4 discharges, with a Glacial Lake Arcola source area, deposited pebble gravel (mean: -4.06 phi) throughout much of the lower section of the Moose Mountain Valley and the Souris Valley. These deposits commonly occur on the inside of the phase 3 landslide deposits. The homogeneous character of the coarse sediment, as with the phase 1 deposits, indicates rapid aggradation of sediment during a nonrepetitive discharge event. Thus, a rapid or catastrophic draining of Glacial Lake Arcola is suggested as the source of the phase 4 flows. Average velocities of 3.4 m/s and discharges of 1.9×10^4 m^3/s are estimated to have been achieved.

Phase 5: The catastrophic flood proposed by Kehew (1982) and Kehew and Clayton (1983) occurred during this phase. Kehew (1982) noted that there is only one level of terraces in the upper Souris Valley, indicating that the flood waters from Lake Regina developed this reach. The flood discharges then flowed into the existing Des Lacs and lower Souris spillways, widening and deepening them by a factor of about 3.

Very large bars of pebbly cobble gravel (mean: -4.03 phi), similar to those described in the Scablands, were deposited by phase 5 discharges. Average velocities of 4.27 m/s and discharges of 2.0×10^5 m^3/s were achieved. There are numerous geomorphic criteria to indicate that the draining of Lake Regina was catastrophic. But, as was the case for phases 1 and 4, there was no evidence observed to indicate repetitive discharges.

Since the draining of Glacial Lake Regina, Holocene sedimentation and valley side dissection have occurred, but the form of the valleys has remained essentially unaltered (Kehew, 1983).

Alternative Hypotheses for Phase Correlations

The deposits of phases 1,4, and 5 are remarkably similar in sediment grain size and texture; this makes it very difficult to distinguish these deposits in the field. The main basis for terrace correlations was elevation, and it is difficult to arrive at a representative elevation for each deposit. Correlating deposits was particularly difficult in the lower Souris Valley. The expected elevations, as determined from slopes of the phase 4 and 5 terraces are similar in this stretch. The inset nature of the gravel deposits, with the appearance of having been truncated since deposition, led to the interpretation that these deposits probably were deposited before the last erosive event, phase 5.

Alternatively, the inset gravel deposits could have been deposited by discharges from Glacial Lake Regina. The valley walls consisting of landslide deposits (phase 3) probably were relatively unconsolidated compared to the valley sides cut in till. This could have led to rapid erosion of landslide areas, thus allowing for expansion of flow, and causing deposition in the alcove areas of the valley walls. This situation would be analogous to eddy bars described by Baker (1973) which formed at the mouths of tributaries. A flaw in this logic is that not all inset gravel deposits appear to be associated with landslide deposits. However, it should be noted that not all the landslide deposits are readily obvious on air photos.

The distinction between phase 3 and phase 4 is also based on relatively meager evidence. Phase 3 is defined on the presence of landslide deposits between gravel deposits of phase 4 and the undisturbed valley wall, indicating that the landslides preceded gravel deposition. It is very possible, if not likely, that the flows which triggered the landslides also deposited the phase 4 sediment. Scott and Gravlee (1968, p. 17) reported many landslides initiated during high discharge flows from a flood surge on the Rubicon River, California. They also described the deposition of lateral berms or terraces somewhat analogous to the phase 4 deposits. The distinction between phase 3 and phase 4 will be maintained for the purposes of separating erosional and depositional processes in spite of the distinct possibility that the resulting features formed contemporaneously.

A third alternative to be considered is the possibility of a normal glacial meltwater phase or phases preceding the Des Lacs Valley discharges of phase 1. Kehew (1984, verbal communication) has observed typical outwash deposits, consisting of gravelly sand, in the vicinity of Verendrye, North Dakota and in the Spring Creek Valley near Velva (Figure 42). The deposits in the Spring Creek Valley are higher in elevation than those near Verendrye, and both appear to be stratigraphically higher than the deposits associated with phase 1. This relationship suggests that a period of glacial meltwater deposition preceded phase 1.

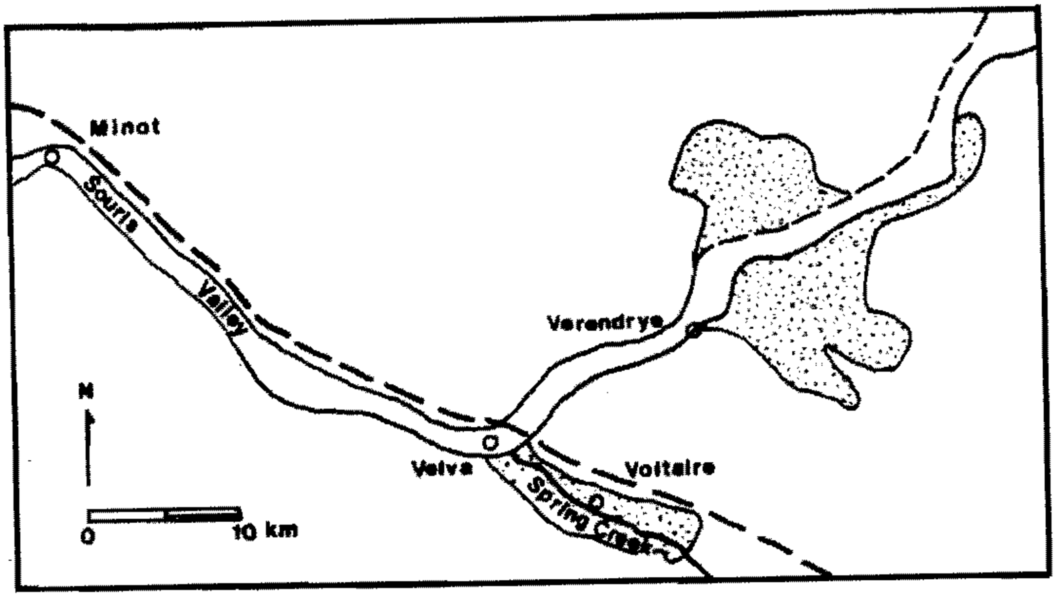
Kehew believes that the form of the Spring Creek Valley, since the deposition of the outwash, only has been altered by Holocene erosion. It seems likely that these deposits are the result of an ice-marginal stream. The deposits near Verendrye have been incised by subsequent discharges down the Souris Valley. It is possible that these deposits represent a later ice-marginal position which preceded phase 1. Discharges from phase 1 could have downcut through these deposits and flowed into Glacial Lake Souris I (Schnacke, 1982), thus accounting for the high stratigraphic position of the outwash deposits.

The uncertainty of the correlation of deposits with phases and exact sequence of drainage development does not alter the strong evidence indicating four sets of gravel deposits within the valleys. Nor does it alter the interpretation that three sets of the terraces were deposited by flood-like flows and one set by normal glacial meltwater.

Summary of Souris Valley Development

The Souris Valley, from its confluence with the Moose Mountain Valley downstream to its confluence with the Des Lacs, has a complex

Figure 42. Location of outwash deposits in Spring Creek Valley and near Verendrye. Dashed line represents a possible ice-margin position.



Minot
Souris
Valley
Verendrye
N
Velva
Voltaire
Spring Creek
0
10 km

history of development. A brief discussion of the development, with regard to the phases proposed, will follow (Figure 43).

The first event proposed, which initiated the development of the Souris Valley, occurred during phase 2 (discharges during phase 1 were confined to the Des Lacs Valley). The valley probably had a high width to depth ratio, as is common with many outwash valleys (Figure 43a).

Erosional discharges from Glacial Lake Arcola, during phase 3, next affected the valley. The valley was widened and deepened by rapid erosion, which also induced landslides to occur in the Souris (Figure 43b and 43c).

The depositional discharges of phase 4 from Glacial Lake Arcola deposited coarse gravels throughout the channel bottom (Figure 43d). The final significant development of the Souris Valley occurred during phase 5. Erosive discharges, from the rapid draining of Glacial Lake Regina, widened, incised, and straightened the Souris spillway markedly. The result of this erosion left the channel bottom deposits of phase 4 stranded and inset into the valley walls (Figure 43e). Since the end of Phase 5, gradual Holocene sedimentation has partially filled the Souris Valley, but the form of the valley has not changed significantly.

Presence of Stagnant or Dead Ice

The abundance of glacial stagnation features has been well documented in the Midcontinent region (Clayton and others, 1980). There are several indications that ice, probably stagnant, was present throughout much of the study area during drainage development.

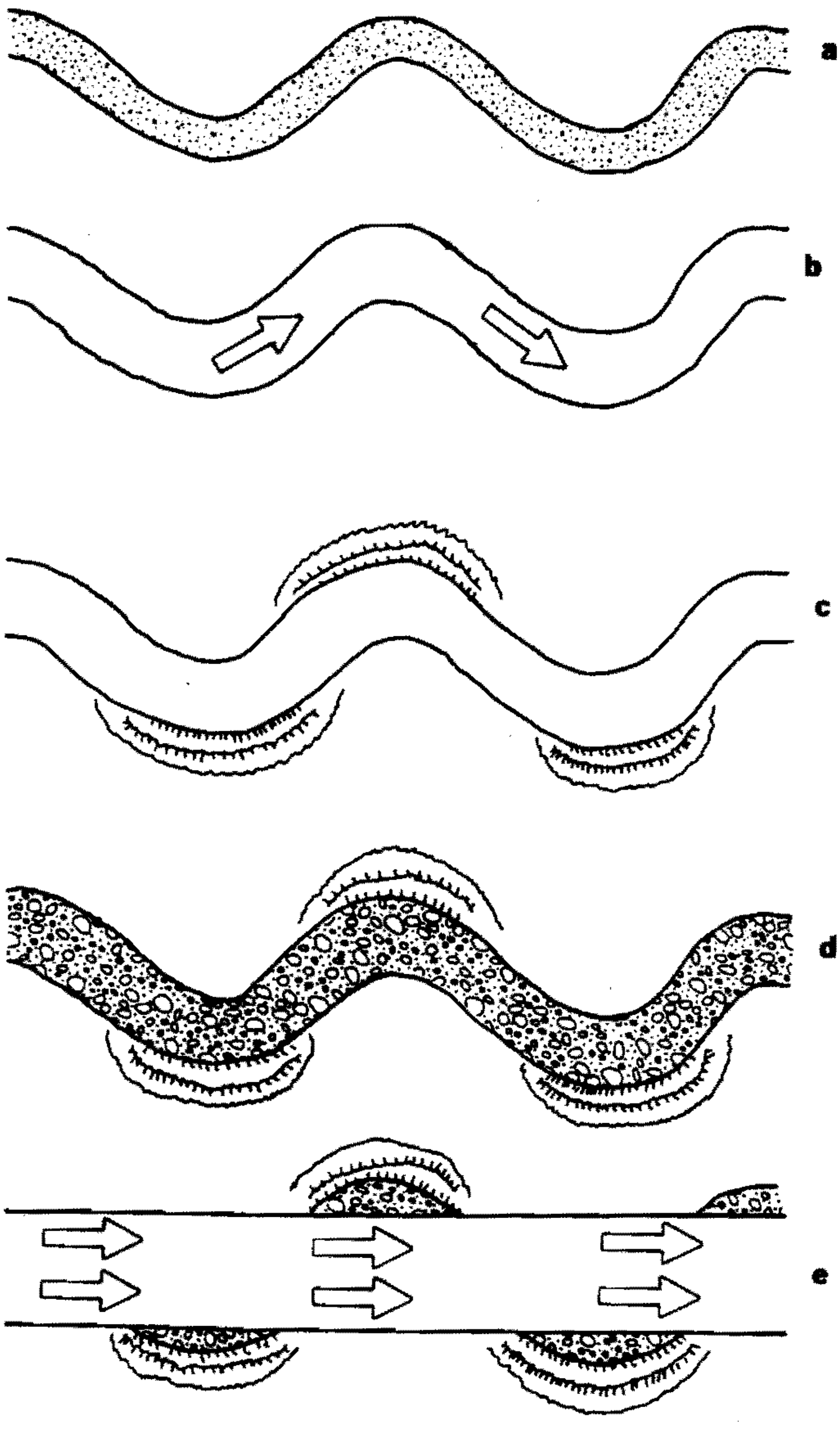
Figure 43. Stages of Souris Valley development.

a) Aggradational meltwater stream during phase 2 initially developed the Souris Valley.

b) and c) Rapid erosion by discharges from Glacial Lake Arcola, during phase 3, widened and incised the Souris Valley. Landslides occurred at cutbank positions as sediment was removed rapidly.

d) Discharges during phase 4, also from Glacial Lake Arcola, deposited coarse sediment throughout much of the valley.

e) The discharges from the rapid draining of Glacial Lake Regina, phase 5, further incised and widened the Souris Valley. The landslides and gravel deposits of phases 3 and 4 were left truncated by erosion.


a
b
c
d
e

Geomorphic evidence for stagnant ice includes: fluvially-formed longitudinal grooves oriented pointing out of the scoured zone onto collapse moraine, an irregular boundary between the fluvially scoured zone and collapse moraine, and the presence of small nonscoured areas within the scoured zone.

Paleohydraulic calculations also denote the presence of ice. Calculations of depths of flows, energy slopes, heights of lake dams, and lake volumes all indicate higher values than field evidence suggests. Even though the uncertainty in the specific values obtained by these calculations is high, the fact that all of the estimates are high is significant. The presence of some thickness of ice, at the time of drainage development, would make each of the determinations made much more reasonable than if no ice were present. For instance, an ice-walled lake would increase the potential volume of the lake, and flood discharges starting on ice would have a higher energy slope than the ground slope predicts. Rebound, since the retreat of the glacier, has not been figured into any calculations. However, the amount of rebound should be the greatest to the north, thus implying an even smaller ground slope during the late Wisconsinan than exists now.

It was noted previously that the slope associated with the Set D Terraces of phase 5 decreases upstream--the opposite of what is normal in stream systems. The presence of some thickness of stagnant ice at the time of drainage, especially in Saskatchewan, could have caused the surface slope to be sufficiently steep in the upstream reaches even though the ground slope was probably negligible. One last indicator of stagnant ice at the time of drainage development are the landslide deposits. While a certain thickness of ice cover is not necessary for landslide development, the increase in overburden pressure exerted by ice, in addition to the higher water table beneath the ice, would reduce the stability of the rapidly-eroded valleys.

Glacial Margins and Drainage Development

The objectives of this study did not include reconstructing glacial ice margins, although some evidence concerning the presence of ice has been presented. A series of paleogeographic maps, which includes the paleohydraulic estimates, was constructed to relate observations of this study with recent compilations of glacial chronologies (Clayton and Moran, 1982; Clayton and others, 1980). For the most part, the paleogeographic margins were chosen on the basis of a logical fit, similar to locating a piece of a puzzle; much more field work and airphoto interpretation will be required to locate the margins with assuredness.

Phase 1 is depicted with the active ice margin immediately to the east of the Des Lacs Valley (Figure 44 (legend) and Figure 45 (map)). This margin would fit somewhere between phases K and L of Clayton and Moran (1982). The extent of stagnant ice is uncertain, but probably covered the area southwest to the Missouri Coteau. It seems likely that the phase 1 discharges, after passing through the Minot region, continued southeast in the vicinity of Velva along the active ice margin. Three former southeasterly meltwater channels were recognized in the Velva region by Lemke (1960) and Clayton (1980).

Figure 44. Legend for paleogeographic maps.

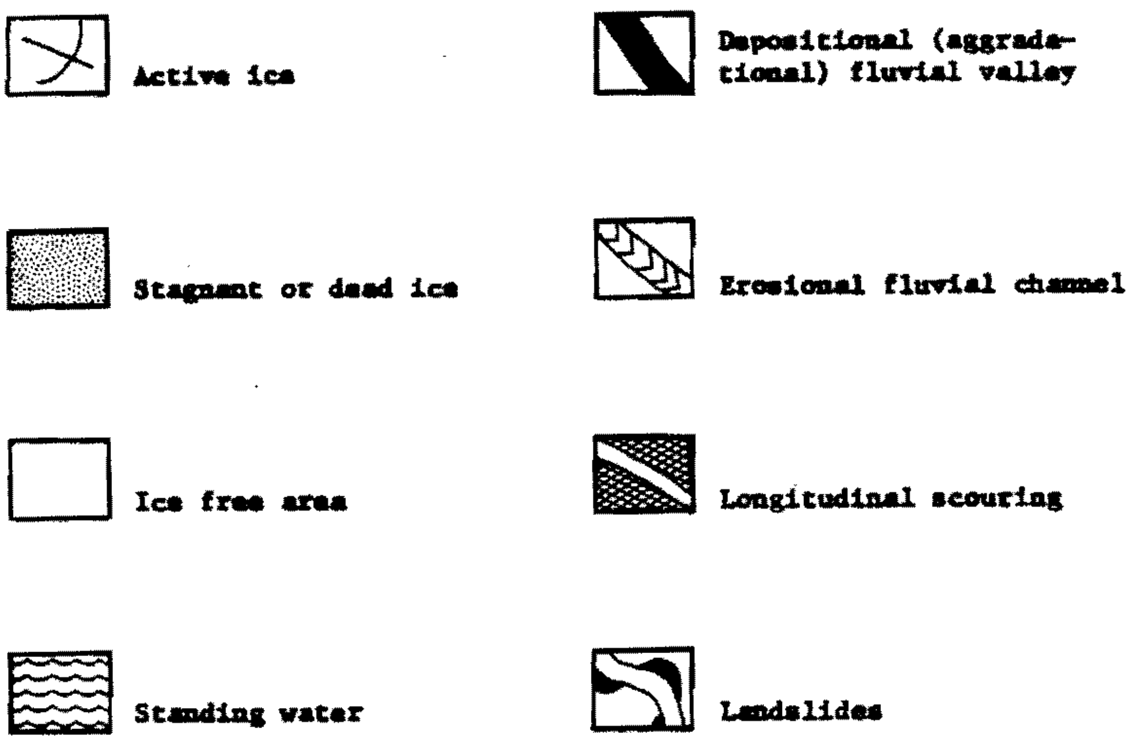
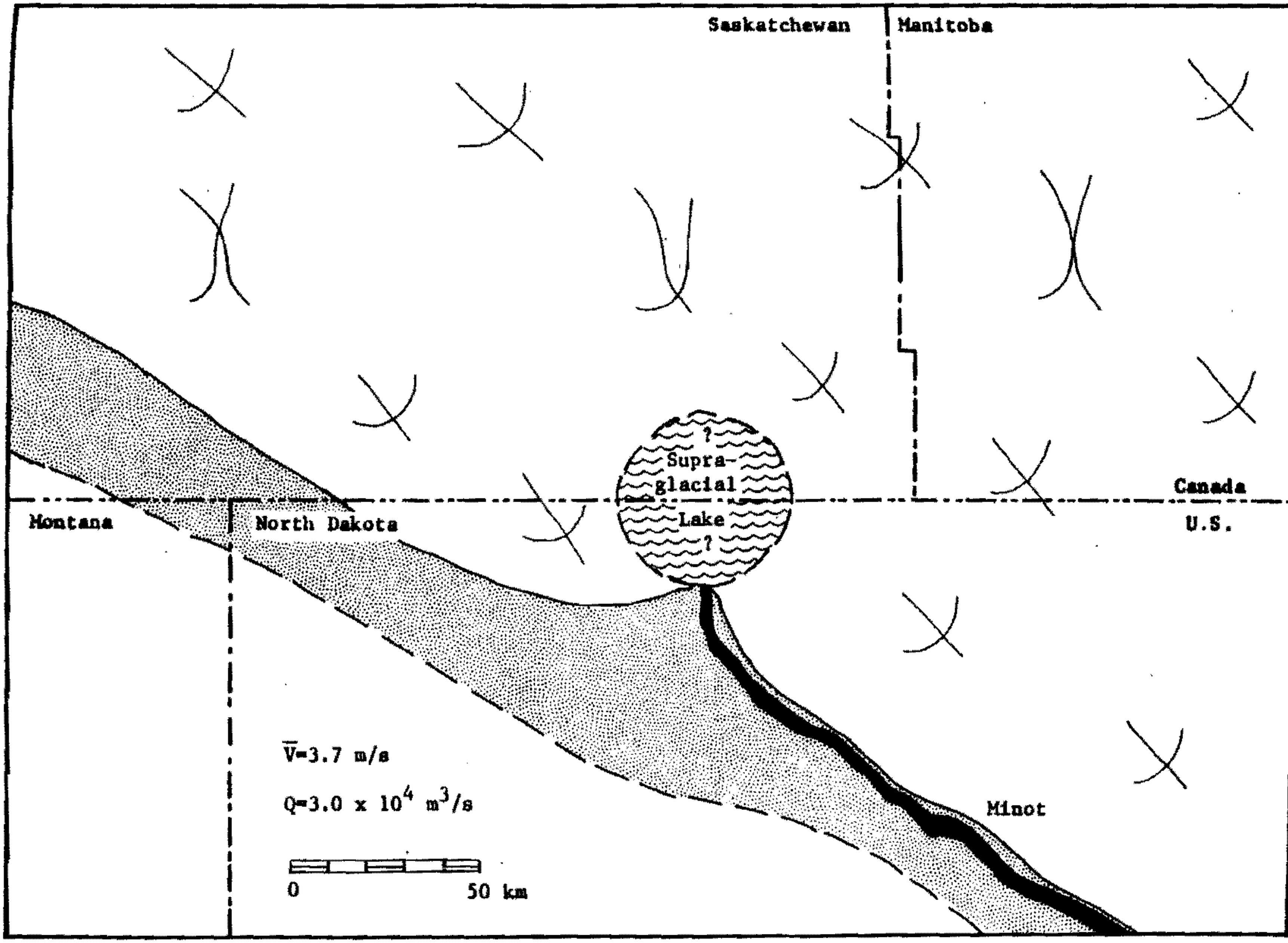

Active ice
Depositional (aggradational) fluvial valley
Stagnant or dead ice
Erosional fluvial channel
Ice free area
Longitudinal scouring
Standing water
Landslides

Figure 45. Paleogeographic map of phase 1.


Saskatchewan
Manitoba
?
Supra-
glacial
Lake
?
Canada
U.S.
Montana
North Dakota
$\overline{V}$=3.7 m/s
Q=3.0 x 10^4 m^3/s
0
50 km
Minot

Phase 2 development was initiated as the Weyburn Lobe formed by the splitting of the ice sheet around the Moose Mountains (Figure 46). Meltwater from the Moose Mountains commenced development of the Moose Mountain and Souris Valleys. It is assumed that the Des Lacs spillway was essentially dormant during phases 2, 3, and 4.

Glacial Lake Arcola developed as the ice continued to melt along the margins of the Moose Mountains (Figures 47 and 48). The draining of Lake Arcola could have been increased by the splitting of the ice sheet northwest of the lake, thus allowing a greater influx of meltwater. The discharges during phases 3 and 4 could have flowed into an early stage of Glacial Lake Souris, possibly supraglacial. The outlet at this time probably was the Girard Lake spillway (Kehew, verbal communication, 1984; Schnacke, 1982). Stagnant or dead ice was probably still widespread at this time; the position of the active ice margin is uncertain.

During phase 5, phases M and N of Clayton and Moran (1982), Glacial Lake Regina discharged catastrophically (Figure 49). The drainage paths shown are those described by Kehew and Clayton (1983). The margin of the stagnant ice probably represents a minimum southern extent. The active ice margin coincides fairly closely with that diagrammed by Clayon and Moran (1982) for phase N. Phase 5 concluded significant development within the study area as drainageways migrated northward with the ice.

Clayton and Moran (1982) dated phases K and L (phase 1) at 11,700 radiocarbon years before present and phases M and N (phase 5) at 11,300 radiocarbon years before present. Consequently, it can be concluded

Figure 46. Paleogeographic map of phase 2.

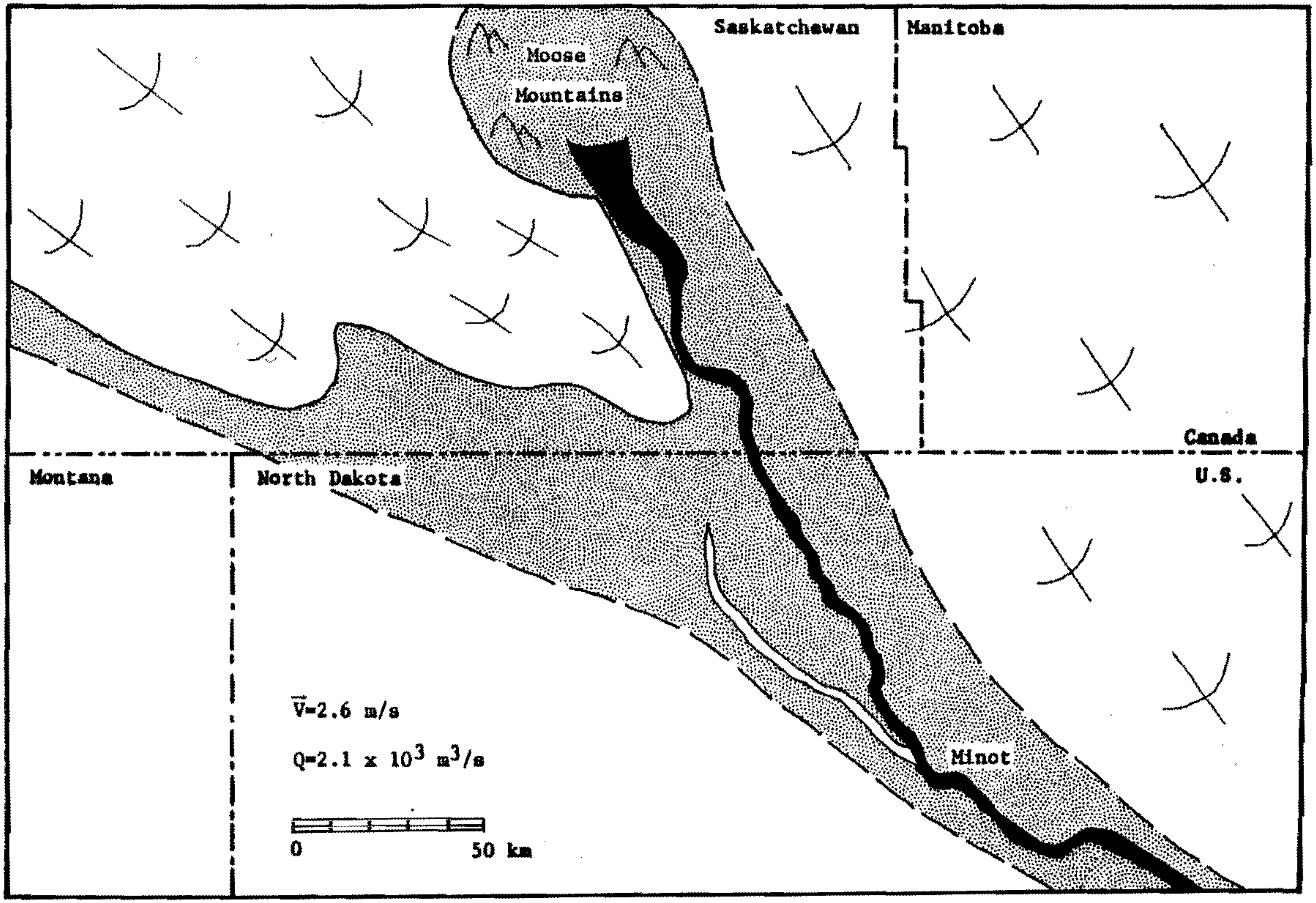
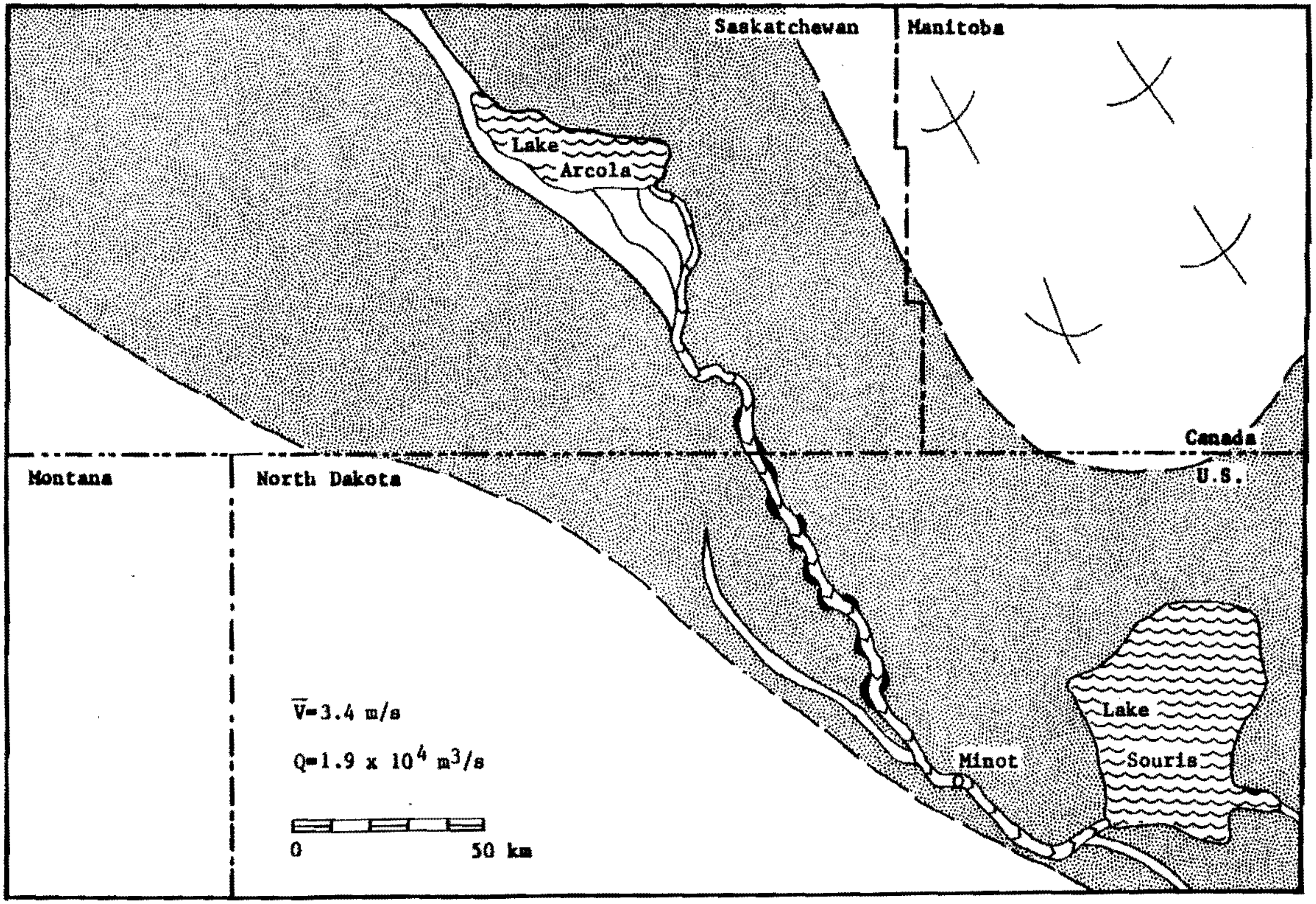
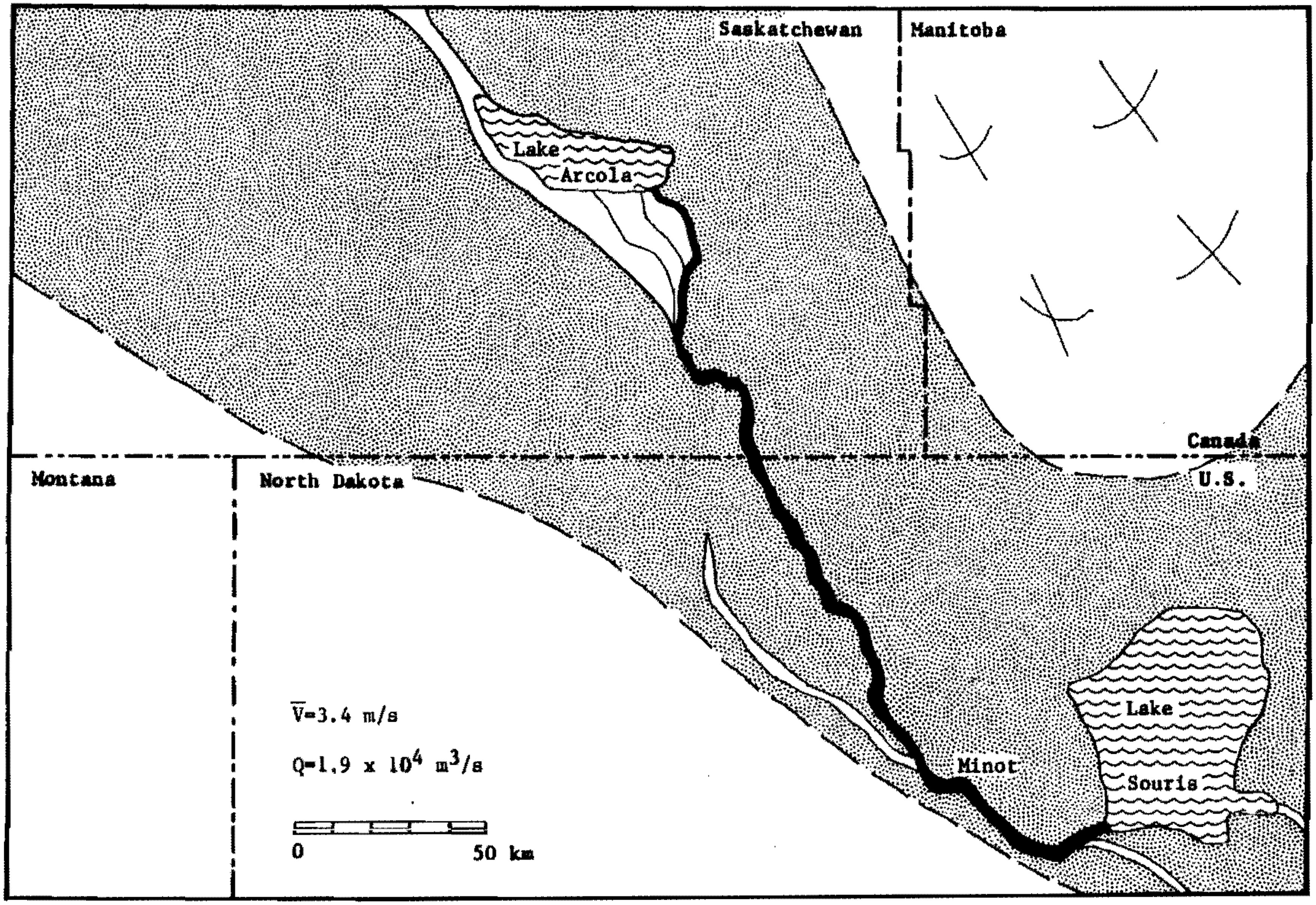

Saskatchewan
Manitoba
Moose
Mountains
Canada
U.S.
Montana
North Dakota
$\overline{V}$=2.6 m/s
Q=2.1 x 10^3 m^3/s
0
50 km
Minot

Figure 47. Paleogeographic map of phase 3.



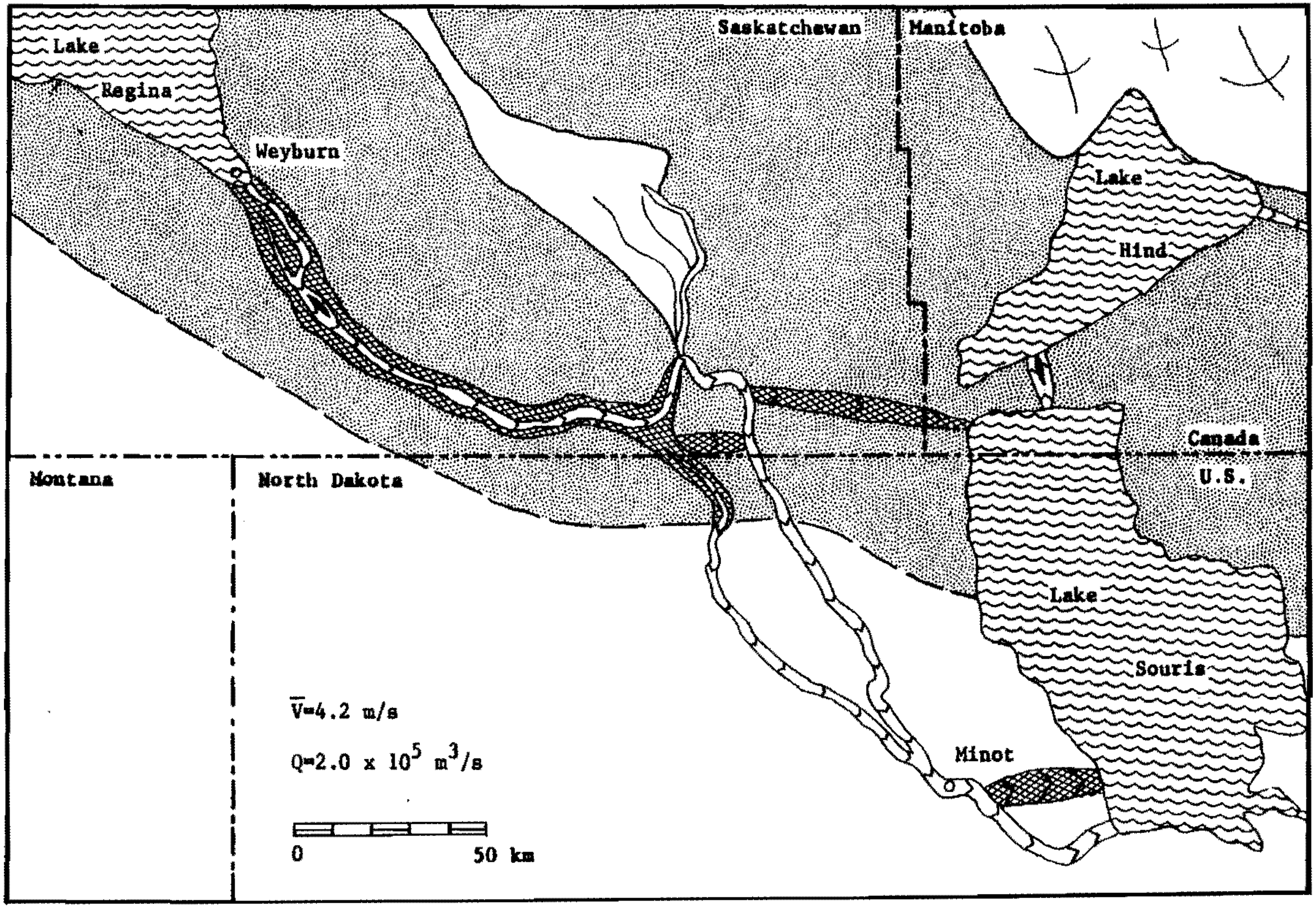
Saskatchewan
Manitoba
Lake
Arcola
Canada
U.S.
Montana
North Dakota
$\overline{V}$=3.4 m/s
Q=1.9 x 10^4 m^3/s
0
50 km
Minot
Lake
Souris

Figure 48. Paleogeographic map of phase 4.



Saskatchewan
Manitoba
Lake
Arcola
Canada
U.S.
Montana
North Dakota
$\overline{V}$=3.4 m/s
Q=1.9 x 10^4 m^3/s
0
50 km
Minot
Lake
Souris

Figure 49. Paleogeographic map of phase 5.


Lake
Regina
Weyburn
Saskatchewan
Manitoba
Lake
Hind
Canada
U.S.
Montana
North Dakota
Lake
Souris
Minot
$\overline{V}$=4.2 m/s
Q=2.0 x 10^5 m^3/s
0
50 km

that the drainage development described in this study took place within a period of 400 years.

Pleistocene Flooding

It was hypothesized by Kehew (1982) that Glacial Lake Regina flooded catastrophically and it has been proposed in this study that two other glacial lake floods occurred within the study area during the late Wisconsinan. These three floods, each of which is estimated to have lasted only days, appear to be almost solely responsible for the form of the Souris, Des Lacs, and Moose Mountain Valleys which still control the drainage today.

Within the Midcontinent region, Kehew and Clayton (1983) proposed that Glacial Lake Souris and Glacial Lake Hind drained catastrophically, causing floodwaters to enter Glacial Lake Agassiz. Teller (1981) hypothesized that Glacial Lake Agassiz also drained catastrophically. Outside of the Midcontinent region, catastrophic floods have been recognized from Glacial Lake Warren in New York (Wilson and Muller, 1981; Hand, 1975), a large series of supraglacial lakes in Indiana (Fraser and Bluer, 1983), and Glacial Lake Missoula in Montana (Bretz, 1929).

Common to most of the areas where catastrophic floods have occurred was the presence of stagnant ice. While active ice retreats by melting back, stagnant ice tends to melt down considerably before the stagnant ice margin makes significant retreats. Clayton and Moran (1974) estimated that the active ice margin retreated at a rate of 315 m/yr in areas where circular disintegration ridges are common (such as in the area of study). They also mention that stagnant ice may persist for hundreds, if not thousands, of years after the retreat of the active ice margin if there is sufficient sediment cover.

It seems likely that the presence of stagnant ice would inhibit the migration of drainage northward with the active ice margin. In the Northern Plains, where pre-Pleistocene drainage was to the northeast (Bluemle, 1972), glacial lakes would continue to grow until the lakes had enough head to drain opposite the land surface gradient. The inability of glacial lakes to drain through established drainage paths, coupled with the large influxes of water volume, would greatly increase the probability of catastrophic drainages. Kehew and Clayton (1983) have suggested that some floods could have been triggered by the flooding of an upstream glacial lake in a domino fashion. If the domino theory is correct, it would imply that not only one major valley developed, but a complete drainage network developed in days.

Ager (1973) has discussed "catastrophic uniformitarianism" to reflect his recognition of the importance of catastrophic events in the stratigraphic record. He notes that periodic catastrophic events have a much greater effect than gradual processes which occur every day. He calls this the "Phenomenon of Quantum Sedimentation"; perhaps a more appropriate title to recognize the significance of catastrophic Pleistocene drainage would be the 'Phenomenon of Quantum Drainage Development'.

SUMMARY AND CONCLUSIONS

Geomorphic interpretation of the Souris, Des Lacs, and Moose Mountain Valleys, and the gravel deposits in them, has resulted in the recognition of five phases in the development of the drainageways. The primary criterion used to distinguish the deposits was terrace correlation; this was facilitated by studying the differences in the morphologic relationships between the gravel deposits and the valley sides.

Characterization of the discharges associated with each phase of drainage development was accomplished by textural analysis of the sediments and the application of paleohydraulic methods. Textural analysis of the sediment involved a combination of sieving and photographic methods to determine the complete grain size distribution. Results of the size analysis indicate strong similarity in the sediments, and presumably depositional flows, of phases 1,4, and 5. The deposits associated with these phases aggraded rapidly in an unstructured manner.

Paleovelocity methods involved the averaging of six well-established approaches for the study of coarse sediments. Calculations, based on the size of the ten largest particles at each sample site show that velocities exceeding 4 m/s were achieved during phases 1,3,4, and 5. Other paleohydrologic calculations, including paleodischarge, paleodepth, lake volumes, and heights of ice dams, proved to be less consistent than paleovelocity estimates, but all indicate large volumes of water which discharged rapidly.

The following is a brief summary of the phases of drainage development proposed. Phase 1 discharges (3×10^4 m^3/s) initiated the development of the Des Lacs Valley, with a probable source of a supraglacial lake in the vicinity of Bowbells, North Dakota. Deposits, generally unstructured sandy gravel, occur relatively high on the valley walls and are confined to the lower Des Lacs and Souris spillways.

Phase 2 discharges (2.1×10^3 m^3/s) deposited outwash sediment, consisting of cross-bedded gravelly sand, in the Moose Mountain and lower Souris Valley. Glacial meltwater, with maximum velocities of 3 m/s, from the Moose Mountains was the source of these flows which commenced when the ice sheet divided around the Moose Mountains.

Phase 3 and 4 discharges (1.9×10^4 m^3/s) both probably resulted from the rapid draining of Glacial Lake Arcola. Phase 3 was an erosional stage that incised the Moose Mountain and lower Souris valley and induced landslides to occur along the valley walls. Deposition of unstructured pebble gravel throughout much of the lower Souris spillway occurred during phase 4. The deposits commonly are inset into the valley walls, indicating that they were truncated by erosion after deposition.

The upper Souris Valley was developed and the lower Souris and Des Lacs spillways were enlarged during the cataclysmic discharges of phase 5 (2.0×10^5 m^3/s) from Glacial Lake Regina. Huge bars composed of unstructured pebbly cobble gravel were deposited during this event. Several geomorphic features, including deeply incised channels, anastomosing channels, major drainage divide crossings, and streamlined hills are similar to features described in the Channeled Scabland. Phase 5 concluded the development of the spillways.

A glacial chronology proposed by Clayton and Moran (1982) indicates that the drainage described in this study occurred between 11,700 BP and 11,300 BP. Although active ice is generally assumed to have been northeast of the spillways during the time of drainage development, the presence of stagnant ice is indicated by streamlined collapse topography within a fluvially scoured zone, several inconsistencies in paleohydraulic calculations, and the abundance of landslide deposits in the lower Souris Valley. The information collected in this study was combined with data existing on the location of glacial margins to construct a series of paleogeographic maps. These maps depict the presumed location of the active and stagnant/dead ice-margins, and their relation to drainage development.

Information developed in this study has further revealed, along with numerous previous studies, that drainage development during the Pleistocene Epoch was very rapid. Drainageways developed by the short-lived, high discharge events still control much of present drainage.

APPENDICES

APPENDIX A

Average Dimensions and Shape Factor of Ten Largest Particles at Each Sample Site

Sample	dl	di	ds	CSF
RB 1	390	320	160	.45
	320	260	240	.83
	300	300	170	.57
	560	480	280	.54
	380	330	270	.76
	390	380	150	.39
ave	390	345	212	.57
EL 1	1830	1530	1360	.81
	1880	1060	500	.35
	1570	1470	105	.69
	1730	1280	900	.60
	2770	2040	1470	.61
	1400	1330	1100	.81
	1400	960	900	.78
	1480	1230	790	.58
	1370	1330	1140	.84
	1590	1300	1000	.69
ave	1702	1253	1021	.69
EL 2	680	500	380	.65
	610	400	280	.56
	600	440	390	.76
	890	680	350	.45
	760	480	280	.46
	720	350	270	.54
	690	360	210	.42
	590	490	390	.72
	580	390	340	.71
	890	540	350	.50
ave	701	463	324	.56

Sample	dl	di	ds	CSF
El 3	690	450	300	.53
	500	500	190	.38
	700	280	230	.52
	600	500	330	.60
	540	490	300	.58
	560	380	250	.54
	540	460	210	.42
	500	480	290	.59
	600	330	290	.53
	490	450	260	.55
ave	572	432	265	.53
OX 2	450	380	240	.76
	520	380	340	.76
	530	320	250	.61
	450	370	310	.76
	420	410	230	.55
	480	370	290	.69
	440	390	350	.84
	430	330	250	.66
	760	350	350	.68
	400	310	310	.88
ave	488	361	292	.69
OX 1	900	680	460	.59
	500	470	390	.80
	560	450	?	?
	540	410	280	.60
	550	460	390	.78
	590	430	380	.75
	460	?	330	?
	540	470	370	.73
	440	360	190	.48
	390	310	?	?
ave	547	446	340	.68

Sample	dl	di	ds	CSF
FD 1	270	260	140	.52
	190	160	130	.76
	220	190	150	.74
	280	150	140	.68
	340	240	220	.77
	190	160	130	.75
	230	190	190	.91
	290	160	130	.60
	300	240	130	.48
	330	230	190	.69
ave	264	198	155	.67
GE 3	550	280	250	.64
	620	470	200	.37
	510	350	270	.64
	460	280	270	.75
	420	380	350	.88
	400	300	230	.66
	500	320	300	.75
	520	300	260	.66
	440	350	200	.51
	480	310	220	.57
ave	491	337	259	.63
NG 3	420	270	230	.68
	390	250	180	.58
	370	230	230	.79
	330	300	180	.57
	350	250	180	.61
	350	290	200	.63
	370	270	250	.79
	320	250	150	.53
	450	320	290	.76
	360	230	200	.70
ave	371	266	209	.66

Sample	dl	di	ds	CSF
NG 4	680	680	380	.56
	590	530	430	.77
	600	440	270	.53
	480	470	230	.48
	490	390	260	.66
	640	520	220	.38
	500	480	300	.61
	600	370	330	.70
	460	380	220	.53
	460	380	220	.53
ave	565	470	292	.56
GN 2a	220	210	170	.79
	160	150	140	.90
	230	180	110	.54
	320	180	180	.75
	250	180	150	.71
	220	180	150	.75
	240	180	120	.58
	190	170	170	.95
	240	150	120	.80
	220	140	140	.74
ave	229	172	145	.73
GN 2b	480	420	310	.69
	430	340	210	.55
	340	260	180	.61
	450	270	270	.77
	380	260	250	.80
	450	270	190	.55
	460	300	260	.70
	350	290	150	.47
	380	280	200	.61
	480	310	220	.57
ave	420	300	224	.63

Sample	dl	di	ds	CSF
GN 1	550	290	250	.63
	410	340	240	.64
	470	290	280	.76
	450	290	200	.55
	470	280	270	.74
	420	400	270	.66
	560	340	200	.46
	360	340	230	.66
	540	360	290	.66
	500	400	230	.51
ave	473	333	246	.61

Sample	dl	di	ds	CSF
MN 2	900	560	450	.63
	820	490	330	.52
	750	520	390	.62
	960	750	600	.71
	990	800	480	.54
	770	670	?	?
	750	630	530	.77
	830	820	530	.64
	950	760	?	?
	770	490	390	.63
ave	849	649	469	.63

Sample	dl	di	ds	CSF
GR 1	190	180	120	.65
	190	180	80	.43
	200	170	120	.65
	160	120	120	.87
	150	110	100	.78
	220	130	110	.65
	160	140	130	.87
	120	100	80	.73
	190	100	80	.58
	100	80	80	.89
ave	168	131	102	.68

Sample	dl	di	ds	CSF
MN 3	570	400	220	.46
	560	300	260	.63
	440	320	280	.75
	390	240	230	.68
	480	320	290	.74
	490	410	210	.47
	530	440	330	.68
	430	360	240	.61
	520	310	300	.75
	590	290	230	.56
ave	500	347	260	.62

Sample	dl	di	ds	CSF
BS 5	690	520	320	.53
	510	350	290	.69
	520	360	260	.60
	650	380	240	.48
	990	310	250	.45
	740	570	310	.48
	480	400	280	.64
	530	390	290	.64
	470	360	190	.46
	470	390	330	.77
ave	605	403	276	.55

Sample	dl	di	ds	CSF
SY 1	400	240	210	.68
	390	230	210	.70
	290	210	210	.85
	520	240	220	.62
	410	240	230	.73
	480	280	200	.55
	500	330	300	.74
	340	330	320	.96
	390	280	250	.76
	360	310	280	.84
ave	408	269	243	.73

Sample	dl	di	ds	CSF
SW 1	470	300	210	.56
	560	520	430	.80
	400	300	90	.26
	440	330	110	.29
	400	330	190	.52
	540	370	290	.65
	380	250	200	.65
	500	450	370	.78
	460	450	370	.81
	630	500	420	.75
ave	478	380	268	.62

Sample	dl	di	ds	CSF
OX 4	640	390	210	.42
	970	570	360	.48
	430	270	240	.70
	410	330	120	.33
	390	290	220	.74
	420	290	250	.72
	410	280	240	.71
	440	280	260	.74
	400	300	280	.81
	430	410	230	.55
ave	494	341	241	.58

Sample	dl	di	ds	CSF
OX 5a	460	300	240	.65
	440	300	270	.74
	350	270	220	.72
	310	310	140	.45
	870	800	440	.53
	730	420	350	.63
	440	230	200	.63
	400	300	190	.55
	400	280	210	.63
	700	650	450	.67
ave	510	386	271	.61

Sample	dl	di	ds	CSF
OX 5b	670	420	340	.64
	430	280	260	.75
	450	330	280	.75
	450	280	190	.54
	330	270	220	.74
	370	260	220	.71
	420	320	290	.79
	440	280	230	.66
	430	240	190	.59
	400	270	250	.76
ave	439	292	250	.69

Sample	dl	di	ds	CSF
CP 1	270	240	140	.55
	280	170	170	.78
	300	280	220	.76
	420	270	170	.51
	390	240	200	.65
	260	220	190	.79
	320	210	180	.69
	550	210	210	.62
	210	200	200	.98
	330	180	130	.53
ave	333	222	182	.66

Sample	dl	di	ds	CSF
FX 1	440	390	390	.94
	480	370	330	.78
	680	500	360	.62
	620	470	320	.59
	600	470	380	.72
	490	380	360	.83
	640	470	460	.84
	600	420	400	.80
	720	490	460	.77
	720	440	360	.64
ave	599	440	382	.74

Sample	dl	di	ds	CSF
FB 1	510	230	230	.67
	380	230	230	.78
	400	340	330	.89
	800	580	400	.59
	400	300	260	.75
	420	230	180	.58
	450	260	160	.47
	480	280	150	.41
	580	320	300	.70
ave	491	301	249	.64

Sample	dl	di	ds	CSF
Bl 1	550	350	310	.71
	510	400	320	.71
	430	330	270	.72
	1070	980	570	.56
	590	440	320	.63
	400	330	210	.58
	520	360	190	.44
	590	390	290	.60
	640	340	260	.56
	670	340	330	.69
ave	597	426	307	.60

Sample	dl	di	ds	CSF
BP 2	460	430	300	.67
	380	310	290	.85
	600	350	300	.66
	680	420	300	.56
	500	480	300	.61
	500	380	300	.69
	460	440	300	.67
	380	330	320	.90
	400	330	330	.91
	840	510	270	.41
ave	520	398	301	.66

Sample	dl	di	ds	CSF
BP 3	380	190	180	.67
	420	210	180	.61
	340	210	170	.64
	240	230	170	.72
	290	180	140	.61
	320	200	170	.67
	290	180	160	.70
	340	250	170	.58
	320	180	170	.71
	340	180	170	.69
ave	328	201	168	.66

APPENDIX B

Textural Analysis Results

TABLE 9

Sieve Analysis Results for Phases 1 and 2

Percentages of sediments on the sieve screen of the stated phi size.

PHI SIZE	SAMPLE FX 1	BP 2	BL 1	MN 3	AB 1	AL 1	OX 3	FD 1
-7.5	-	-	13.4	-	-	-	-	-
-7.0	-	0.0	5.8	-	-	-	-	-
-6.5	11.0	2.9	14.4	11.1	-	-	-	-
-6.0	6.9	8.7	13.3	4.7	-	-	-	-
-5.5	13.6	8.0	6.9	0.0	-	-	-	-
-5.0	7.8	11.9	8.8	4.5	-	-	-	-
-4.5	5.2	11.4	9.2	6.7	-	-	1.3	-
-4.0	3.5	21.3	8.6	3.4	4.9	6.2	2.5	6.4
-3.5	5.3	3.6	3.1	5.2	5.9	11.3	4.5	18.1
-3.0	4.0	3.3	3.1	4.3	6.2	10.6	5.9	16.6
-2.5	2.7	2.3	2.1	3.8	6.6	11.6	5.2	13.9
-2.0	2.5	2.0	1.7	5.1	6.3	8.2	6.5	8.1
-1.5	2.7	1.6	1.4	5.5	8.6	8.5	6.7	6.1
-1.0	4.1	2.0	1.7	9.1	7.8	6.2	10.5	3.3
-0.5	4.6	2.2	1.7	9.4	7.7	6.4	11.9	2.8
0.0	6.4	3.1	2.0	7.3	5.7	4.6	14.4	2.9
0.5	5.1	2.6	1.6	3.3	7.0	3.7	10.8	5.0
1.0	3.9	3.3	1.5	2.9	11.2	6.9	9.4	6.4
1.5	2.7	3.5	1.4	2.4	9.7	7.0	5.8	4.0
2.0	1.6	2.5	1.1	1.6	8.1	5.7	2.2	1.9
2.5	1.0	1.7	0.8	1.3	2.5	1.9	1.0	1.1
3.0	1.8	0.8	0.6	0.7	0.8	0.9	0.5	0.9
3.5	0.7	0.5	0.4	0.5	0.3	0.2	0.4	0.6
4.0	0.5	0.2	0.2	0.3	0.8	0.1	0.2	2.0
4.5	2.8	0.8	0.9	1.0	0.0	0.0	0.4	0.0

TABLE 10

Sieve Analysis Results for Phase 4

Percentages of sediments on the sieve screen of the stated phi size.

PHI SIZE	SAMPLE OX 4	OX 2	NG 3	GN 2	GN 1	SY 1
-7.5	18.0	13.3	-	0.1	-	-
-7.0	14.6	7.3	-	0.0	13.0	6.7
-6.5	6.4	21.6	-	16.9	10.1	15.0
-6.0	4.2	8.6	8.0	4.5	8.4	5.6
-5.5	9.8	4.0	7.1	8.8	4.9	7.4
-5.0	7.8	5.1	15.4	7.6	5.7	7.6
-4.5	6.4	5.0	10.7	7.5	8.4	8.2
-4.0	3.4	4.4	8.0	9.4	8.4	12.3
-3.5	4.3	3.4	7.3	6.0	6.9	5.9
-3.0	3.5	2.5	9.0	4.8	4.8	6.1
-2.5	2.9	2.2	4.2	3.4	3.1	3.6
-2.0	1.9	2.1	3.8	3.0	3.0	3.3
-1.5	1.1	2.1	2.6	2.9	2.4	1.9
-1.0	0.9	3.3	2.9	5.2	2.8	1.6
-0.5	0.9	3.8	2.9	4.9	2.4	1.0
0.0	1.9	4.5	4.2	4.2	2.5	1.1
0.5	2.6	2.7	4.1	2.5	2.0	1.5
1.0	3.0	1.7	3.8	1.9	2.1	3.4
1.5	2.8	1.1	2.8	1.5	2.2	3.0
2.0	1.7	0.9	1.4	1.0	1.9	1.2
2.5	0.9	0.8	0.6	0.8	1.3	0.7
3.0	0.3	0.6	0.3	0.6	0.8	0.5
3.5	0.2	0.5	0.2	0.5	0.6	0.4
4.0	0.1	0.3	0.1	0.4	0.4	0.3
4.5	0.3	1.4	0.6	1.6	1.6	1.5

TABLE 11

Sieve Analysis Results for Phase 5

Percentages of sediments on the sieve screen of the stated phi size.

PHI SIZE	SAMPLE RB 1	EL 1	OX 1	BS 5	MN 2	SW 2
-7.5	-	4.9	18.4	-	12.0	-
-7.0	-	5.7	8.1	-	5.6	-
-6.5	16.8	22.7	19.0	3.8	8.7	4.5
-6.0	1.7	7.4	4.0	0.0	3.0	14.5
-5.5	8.7	4.6	1.8	7.0	5.9	5.7
-5.0	6.5	6.4	5.1	6.6	5.3	11.5
-4.5	10.5	7.5	2.4	7.3	8.2	11.5
-4.0	9.0	6.5	1.6	9.5	7.7	9.8
-3.5	4.1	4.5	1.3	6.1	7.5	6.3
-3.0	6.4	4.8	2.1	10.1	8.1	6.3
-2.5	4.7	2.7	1.8	9.2	4.9	5.0
-2.0	4.6	2.7	2.1	9.5	4.0	5.3
-1.5	3.2	2.8	2.1	6.1	2.1	3.3
-1.0	3.3	2.7	3.4	5.2	2.6	3.2
-0.5	2.4	2.1	4.8	3.1	2.1	2.3
0.0	3.0	2.0	7.7	2.7	2.5	2.1
0.5	3.7	1.3	4.6	2.2	2.4	1.5
1.0	4.3	1.1	2.3	2.5	1.6	1.4
1.5	2.8	1.0	1.2	2.1	1.3	1.4
2.0	1.4	1.1	0.8	1.3	0.8	1.1
2.5	0.9	1.3	0.7	1.1	0.6	0.9
3.0	0.5	0.9	0.7	1.1	0.6	0.6
3.5	0.4	0.8	0.8	0.7	0.5	0.5
4.0	0.3	0.6	0.6	0.5	0.4	0.3
4.5	0.8	2.3	2.7	2.3	1.6	1.3

TABLE 12

Phi Values of Textural Analysis

SAMPLE	MED	MEAN	SORT	SKEW	C 1%	MODE
FX 1	-3.81	-3.02	3.10	0.28	-6.71	-5.75
BP 2	-4.34	-3.46	2.62	0.52	-6.61	-4.25
BL 1	-5.39	-4.74	2.68	0.64	-7.72	-6.75
MN 3	-2.45	-2.81	2.83	0.03	-7.17	-6.75
AB 1	-0.76	-0.79	1.95	-0.04	-4.15	0.75
AL 1	-1.88	-1.50	1.95	0.21	-4.17	-2.75
OX 3	-0.71	-0.91	1.67	-0.08	-4.53	-0.25
FD 1	-2.68	-1.94	2.01	0.51	-4.17	-3.75
OX 4	-5.65	-4.87	2.90	0.53	-7.74	-7.75
OX 2	-6.05	-4.59	3.04	0.50	-7.72	-6.75
NG 3	-3.95	-3.29	2.39	0.42	-6.19	-5.25
GN 2	-4.23	-3.61	2.73	0.41	-6.82	-6.75
GN 1	-4.55	-3.95	2.93	0.46	-7.22	-7.25
SY 1	-4.53	-4.04	2.72	0.55	-7.18	-6.75
RB 1	-4.18	-3.49	2.69	0.35	-6.74	-6.75
EL 2	-5.13	-4.32	2.99	0.59	-7.65	-6.75
OX 1	-5.86	-4.19	3.52	0.35	-7.74	-6.75
BS 5	-3.02	-2.71	2.46	0.41	-6.63	-3.25
MN 2	-4.42	-4.13	2.83	0.41	-7.72	-7.75
SW 1	-4.38	-3.74	2.44	0.59	-6.65	-6.25

APPENDIX C

Coarse Sand Lithologies

TABLE 13

Coarse Sand Lithologies

PHASE	SAMPLE	CRYSTALLINE ROCK FRAGMENTS	CARBONATE	SHALE	OTHER SEDIMENTARY ROCK FRAGMENTS
1	FX 1	52.5	28.0	8.5	11.0
1	FB 1	56.5	28.0	13.5	2.0
1	BP 2	63.5	23.5	9.5	3.5
1	BL 1	55.0	26.5	8.5	10.0
1	BE 1	65.0	27.5	4.5	3.0
1	MN 3	71.5	24.5	4.0	0.0
2	AB 1	55.5	33.0	8.0	3.5
2	AL 1	71.5	19.5	6.0	3.0
2	OX 6	63.0	27.0	3.0	7.0
2	OX 3	61.0	26.5	7.5	5.0
2	FD 1	61.5	24.0	6.0	8.5
2	GE 3	62.5	30.0	5.0	2.5
4	OX 4	49.5	36.0	9.5	5.0
4	OX 2	58.0	29.5	5.5	7.0
4	MP 1	71.5	19.5	5.0	4.0
4	NG 3	66.5	27.0	4.5	2.0
4	GN 2	59.5	27.0	6.5	7.0
4	GN 1	59.5	25.5	6.5	7.5
4	SY 1	67.5	20.5	7.0	5.0
5	RB 1	53.0	20.5	12.0	14.5
5	EL 2	48.0	13.5	20.0	18.5
5	OX 1	62.5	27.0	1.0	9.5
5	BS 5	73.0	22.0	3.5	1.5
5	MN 2	55.5	32.0	4.5	8.0
5	SW 1	62.0	24.0	11.0	3.0

APPENDIX D

Conversion Factors of SI Units

to English Units

MULTIPLY	BY	TO OBTAIN
mm	3.937×10^{-2}	in
cm	3.937×10^{-1}	in
m	3.281	ft
km	6.215×10^{-1}	mi
km^2	3.861×10^{-1}	mi^2
m^3	3.53×10	ft^3
m/s	3.281	ft/s
m/s^2	3.281	ft/s^2
kg	2.205	lb
kg/m^2	2.048×10^{-1}	lb/ft^2
kg/m^3	6.242×10^{-2}	lb/ft^3
N/m^3	6.365×10^{-3}	lb/ft^3
m^3/s	3.531×10	ft^3/s

APPENDIX E

Sample Locations

Souris Valley Samples, Saskatchewan

HB 1	Goodwater 15'x30'	T.6N, R.13W, sec. 10
RB 1	Goodwater 15'x30'	T.5N, R.12W, sec. 18, SE, SE
EL 1	Oxbow 15'x30'	T.1N, R.3W, sec. 1
EL 2	Oxbow 15'x30'	T.1N, R.3W, sec. 36
EL 3	Oxbow 15'x30'	T.1N, R.3W, sec. 36
OX 3	Oxbow 15'x30'	T.3N, R.2W, sec. 15
OX 2	Oxbow 15'x30'	T.3N, R.2W, sec. 15
OX 1	Oxbow 15'x30'	T.3N, R.2W, sec. 15
FD 1	Oxbow 15'x30'	T.2N, R.1W, sec. 31, SW
GE 3	Oxbow 15'x30'	T.2N, R.1W, sec. 36, SW, NW

Moose Mountain Valley Samples, Saskatchewan

KB 1	Kisbey 15'x30'	T.8N, R.5W, sec. 21, SW
AC 1	Carlyle 15'x30'	T.7N, R.3W, sec. 29
AB 1	Alameda 15'x30'	T.6N, R.2W, sec. 13
AL 1	Alameda 15'x30'	T.4N, R.2W, sec. 16
OX 4	Oxbow 15'x30'	T.3N, R.2W, sec. 21, NE, NW
OX 5a	Oxbow 15'x30'	T.3N, R.2W, sec. 21, NE, NW
OX 5b	Oxbow 15'x30'	T.3N, R.2W, sec. 21, NE, SW
OX 6	Oxbow 15'x30'	T.3N, R.2W, sec. 21, SW, SE

Souris Valley Samples, North Dakota

MP 2	Moose River 7.5'x7.5'	T.161N, R.86W, sec. 2, SW
NG 3	Grano 15'x15'	T.161N, R.85W, sec. 31, NE
GN 2a	Grano 15'x15'	T.160N, R.85W, sec. 2, SE
GN 2b	Grano 15'x15'	T.160N, R.85W, sec. 11, NE
GN 1	Grano 15'x15'	T.160N, R.85W, sec. 19, SW, SE
GR 1	Grano SW 15'x15'	T.159N, R.85W, sec. 3
CE 1	Carpio NE 7.5'x7.5'	T.158N, R.84W, sec. 30, SW
BS 5	Burlington 7.5'x7.5'	T.156N, R.84W, sec. 22
BE 4	Burlington 7.5'x7.5'	T.155N, R.84W, sec. 12, NE
MN 2	Minot 7.5'x7.5'	T.155N, R.83W, sec. 21
MN 3	Minot 7.5'x7.5'	T.155N, R.83W, sec. 21
SY 1	Surrey 7.5'x7.5'	T.155N, R.82W, sec. 32
SW 1	Sawyer 7.5'x7.5'	T.153N, R.81W, sec. 4, NE, SW
VL 1	Velva 7.5'x7.5'	T.153N, R.80W, sec. 26, NW

Des Lacs Valley Samples, North Dakota

CP 1	Carpio 7.5'x7.5'	T.157N, R.85W, sec. 8, SW, NW
FX 1	Des Lacs 7.5'x7.5'	T.157N, R.85W, sec. 36, SE
FB 1	Des Lacs 7.5'x7.5'	T.156N, R.84W, sec. 18, NE
BP 2	Burlington 7.5'x7.5'	T.156N, R.84W, sec. 29, SW, NE
BP 3	Burlington 7.5'x7.5'	T.156N, R.84W, sec. 29, SE, NW
BL 1	Burlington 7.5'x7.5'	T.155N, R.84W, sec. 2, NE

REFERENCES

Armstrong, C.A., 1971, Ground-water resources of Burke and Mountrail Counties, Part 2--Ground-water basic data: North Dakota Geological Survey Bulletin 55 (North Dakota State Water Commission County Ground Water Studies 14), 86p.

Baker, V.B., 1973, Paleohydrology and sedimentology of Lake Missoula flooding in eastern Washington: Geological Society of America Special Paper 144, 79p.

Baker, V.B., 1983, Large-scale fluvial paleohydrology: in Gregory, K.J., ed., Background to paleohydrology, a perspective: John Wiley and Sons, Chichester, p. 453-478.

Baker, V.R., and Nummedal, Dag, 1978, The Channeled Scabland: National Aeronautics and Space Administration, Washington D.C., 186p.

Baker, V.R., and Ritter, D.F., 1975, Competence of rivers to transport coarse bedload material: Geological Society of America Bulletin, v. 86, p. 975-978.

Bluemle, J.P., 1971, Geology of McLean County, North Dakota: North Dakota Geological Survey Bulletin 60, Part 1 (North Dakota State Water Commission County Ground Water Studies 19), 65p.

Bluemle, J.P., 1972, Pleistocene drainage development in North Dakota: Geological Society of America Bulletin, v. 83, p. 2189-2194.

Bradley, W.C., and Mears, A.L., 1980, Calculations of flows needed to transport coarse fraction of Boulder Creek alluvium at Boulder, Colorado: Geological Society of America Bulletin, v. 91, p. 1057-1090.

Bretz, J H., 1929, Valley deposits east of Channeled Scabland: Journal of Geology, v.37, p. 393-427.

Christiansen, E.A., 1956, Glacial geology of the Moose Mountain area, Saskatchewan: Saskatchewan Department of Mineral Resources Report 21, 35p.

Christiansen, E.A., 1961, Geology and ground-water resources of the Regina area, Saskatchewan: Saskatchewan Research Council Geology Division Report 2, 72p.

Christiansen, E.A., 1979, The Wisconsinan deglaciation of southern Saskatchewan and adjacent areas: Canadian Journal of Earth Sciences, v. 16, p. 913-938.

Clague, J.J., and Mathews, W.H., 1973, The magnitudes of jökulhlaups: Journal of Glaciology, v. 12, p. 501-503.

Clayton, Lee, 1980, Geologic map of North Dakota: North Dakota Geological Survey.

Clayton, Lee, and Moran, S.R., 1974, A glacial process-form model: in Coates, D.R., ed., 1974, Glacial Geolmorphology: State University of New York, Binghamton, p. 88-119.

Clayton, Lee, and Moran, S.R., 1979, Oahe Formation: in Groenewold, G.H., Hemish, L.A., Cherry, J.A., Rehm, G.N., and Winczewski, L.M., 1979, Geology and Geohydrology of the Knife River basin and adjacent areas of west-central North Dakota: North Dakota Geological Survey Report of Investigations, no. 64, p.337-339.

Clayton, Lee, and Moran, S.R., 1982, Chronology of Late Wisconsinan glaciation in middle North America: Quaternary Science Reviews, v. 1, p. 55-82.

Clayton, Lee, Moran, S.R., and Bluemle, J.P., 1980, Explanatory text to accompany the geologic map of North Dakota: North Dakota Geological Survey Report of Investigations, no. 69, 93p.

Costa, J.E., 1983a, Paleohydraulic reconstruction of flash-flood peaks from boulder deposits in the Colorado Front Range: Geological Society of America Bulletin, v. 94. p.986-104.

Costa, J.E., 1983b, Hydrology of dam failures: Geological Society of America Abstracts with Programs 1983, v.15, p. 548.

Dury, G.H., 1964, Theoretical implications of underfit streams: United States Geological Survey Professional Paper 452-C, 43p.

Ethridge, F.G., and Schumm, S.A., 1978, Reconstructing paleochannel morphologic and flow characteristics: methodology, limitations, and assesment: in Miall, A.D., ed., Fluvial sedimentology: Canadian Society of Petroleum Geologists Memoir 5, p. 703-721.

Fahnestock, R.K., 1963, Morphology and hydrology of a glacial stream--White River, Mount Rainer, Washington: United States Geological Survey Professional Paper 422-A, 65p.

Folk, R.L., 1980, Petrology of Sedimentary Rocks: Hemphill Publishing Company, Austin, Texas, 185p.

Fraser, G.S., and Bleuer, N.K., 1983, Late Wisconsinan jokulhlaup at Delphi, Indiana: Geological Society of America Abstracts with Programs, 1983, v.15, p.226.

Freers, T.F., 1973, Geology of Burke County, North Dakota: North Dakota Geological Survey Bulletin 55--Part 1, 32p.

Friedman, G.M., 1958, Determination of sieve-size distribution from thin-section data for sedimentary petrological studies: Journal of Geology, v. 66, p. 394-416.

Hand, B.M., 1975, Paleohydraulics of catastrophic discharge from proglacial lake near Syracuse, New York: Geological Society of Americal Abstracts with Programs 1975, v.7, p.69-70.

Helley, E.J., 1969, Field measurement of the initiation of large bed particle motion in Blue Creek near Klamath, California: United States Geological Survey Professional Paper 562-G, 19p.

Kehew, A.E., 1979, Evidence for late Wisconsinan catastrophic flooding in the Souris River area, north-central North Dakota: Proceedings, North Dakota Academy of Science, v. 33, p. 32.

Kehew, A.E., 1982, Catastrophic flood hypothesis for the origin of the Souris spillway, Saskatchewan and North Dakota: Geological Society of America Bulletin, v. 93, p. 1051-1058.

Kehew, A.E., 1983, Geology and geotechnical conditions of the Minot area, North Dakota: North Dakota Geological Survey Report of Investigations, no. 73, 35p.

Kehew, A.E., 1984, Verbal communication: Geology Department, University of North Dakota, Grand Forks, North Dakota.

Kehew, A.E., and Clayton, Lee, 1983, Late Wisconsinan catastrophic floods and development of the Souris-Pembina spillway system: in Teller, J.T., and Clayton, Lee, eds., Glacial Lake Agassiz: Geological Association of Canada Special Paper 26, p.187-209.

Krumbein, W.C., 1942, Flood deposits of the Arroyo Seco, Los Angeles County, California: Geological Society of America Bulletin, v. 53, p. 1355-1402.

LeFever, R.D., 1984, Verbal communication: Geology Department, University of North Dakota, Grand Forks, North Dakota.

Lemke, R.W., 1960, Geology of the Souris River area, North Dakota: United States Geological Survey Professional Paper 325, 138p.

Maizels, J.K., 1983, Paleovelocity and paleodischarge determination for coarse gravel deposits: in Gregory, K.J., ed., Background to paleohydrology, a perspective: John Wiley and Sons, Chichester, p. 101-139.

Malde, H.E., 1968, The catastrophic late Pleistocene Bonneville flood in the Snake River Plain, Idaho: United States Geological Survey Professional Paper 596, 52p.

Marcus, M.G., 1960, Periodic drainage of glacier-dammed Tulsequah Lake, British Columbia: Geographical Review, v. 50, p. 89-106.

Mavis, F.T., and Laushey, L.M., 1949, Formula for velocity of beginning of bedload movement is reappraised: Civil Engineering, v. 19, p. 38-39.

Miall, A.D., 1977, A review of braided-river depositional environments: Earth-Science Reviews, v. 13, p. 1-62.

Novak, Ivan, 1973, Predicting coarse sediment transport: the Hjulstrom curve revisited: in Morisawa, Marie, ed., Fluvial geomorphology: State University of New York, Binghamton, p. 13-26.

Packham, G.H., 1955, Volume-, weight-, and number-frequency analysis of sediments from thin section data: Journal of Geology, v. 63, p. 50-58.

Pardee, J.T., 1942, Unusual currents in glacial Lake Missoula, Montana: Geological Society of America Bulletin, v. 53, p. 1569-1600.

Pettyjohn, W.A., and Hutchinson, R.D., 1971, Ground water resources of Renville and Ward Counties, Part III: North Dakota Geological Survey Bulletin 55, 100p.

Schnacke, A.W. Jr., 1982, Glacial geology and stratigraphy of southeastern Pierce and southwestern Benson Counties, North Dakota: M.S. Thesis, Grand Forks, North Dakota, University of North Dakota, 136p.

Schumm, S.A., 1965, Quaternary paleohydrology: in Wright, H.E., ed., The Quaternary of the United States: Princeton University Press, p. 783-794.

Scott, K.M., and Gravlee, G.C., 1968, Flood surge of the Rubicon River, California--hydrology, hydraulics, and boulder transport: United States Geological Survey Professional Paper 422-M, 40p.

Stone, K.H., 1963, Alaskan ice-dammed lakes: Association of American Geographers Annals, v. 53, p. 332-349.

Strand, R.I., 1973, Sedimentation: in Design of small dams: Washington, D.C., United States Bureau of Reclamation, p. 767-796.

Teller, J.T., 1981, The catastrophic draining of glacial Lake Agassiz: Geological Society of America Abstracts with Programs 1981, v. 13, p. 565.

Torpen, B.E., 1956, Large rocks in river control works: Civil Engineering, v. 26, p. 56-61.

Williams, P.F., and Rust, B.R., 1969, The sedimentology of a braided river: Journal of Sedimentary Petrology, v. 39, p. 649-679.

Wilson, M.P., and Muller, E.H., 1981, Catastrophic discharge of Lake Warren east of Batavia, N.Y.: Geological Society of America Abstracts with Programs 1981, p. 582.