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BAFFIN ISLAND SANDURS: A STUDY OF ARCTIC FLUVIAL PROCESSES

M. Church

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PREFACE

Between 1963 and 1965 the Geographical Branch, Department of Energy, Mines and Resources, carried out an investigation of the glacial hydrology of the Lewis Glacier and Lewis River at the north end of the Barnes Ice Cap in north-central Baffin Island. A study of the sandur plain in front of the glacier prompted further investigations of sandurs in the eastern mountains of Baffin Island, the result of which is this bulletin. The sandur plain at the head of Sarvalik arm of Ekalugad Fiord was chosen as the primary field area and the main field work was carried out in 1967 and 1968. This project was undertaken to study the environmental influences and particular physical processes which condition the development of a single landscape unit.

Through studies such as this the Geological Survey contributes to a program designed to facilitate the optimum utilization and conservation of Canada's physical environment for the maximum benefit of all the nation.

Y. O. Fortier,
Director,
Geological Survey of Canada.

Ottawa, November 25, 1971.

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BAFFIN ISLAND SANDURS:
A STUDY OF ARCTIC FLUVIAL PROCESSES

ABSTRACT

Valley sandurs form the major postglacial physiographic feature in much of the Canadian Arctic. The conditions favourable for sandur development: presence of an abundant supply of detrital material and the occurrence of relatively frequent floods competent to move the material, are characteristic of proglacial and some periglacial environments.

Nival floods, summer icemelt floods, and summer storm runoff generate significant sediment transport events. Storm runoff is particularly effective in the Arctic since there is little vegetation and the substrate is frozen, so that a high proportion of rainfall runs off immediately. The distribution of runoff events shows a disproportionate number of moderately high flows. Occasionally, extreme floods occur, produced by catastrophic drainage of ponded water from ice margins.

Sediment is transported mainly as bedload, though solution and suspended load are significant. Transport is much greater than the rate of primary detritus production as large volumes of glacially derived material are being redistributed.

Sandur streams are wide and shallow, a characteristic of channels in which most sediment moves as bedload. The major hydraulic adjustment to increasing discharge is made by stream velocity, indicating a change in bed form associated with the rapid increase of sediment transport that results in a sharp decline in flow resistance.

The long profiles of the streams are approximately parabolic: this form is probably conditioned by the nature of the aggradation on the surface as a whole. The sequence of pools and riffles along the channels constitutes a fundamental feature of the river. It appears to function as a regulator of energy expenditure in the stream and concomitantly influences flow resistance and the movement and storage of sediment.

The sandur surface is an aggradational feature. Erosion and deposition occur over wide areas during major floods. Change of stage is rapid so that deposition is often chaotic, and little internal structure occurs in many sediments. Significant pattern is found only in the mean size and variance of coarse materials (cobbles), indicating deposition under conditions of declining competence down sandur.

RÉSUMÉ

Les sandurs de vallée constituent l'élément géomorphologique le plus important du Postglaciaire de l'Arctique canadien. Les zones proglaciaires et certaines zones périglaciaires sont favorables à la formation de sandurs: abondance de matériaux détritiques et fréquence d'inondations capables de transporter ces matériaux.

Les crues liées à la fonte des neiges, celles provoquées par la fonte des glaces, de même que les eaux de ruissellement des orages d'été peuvent être de puissants agents de transport de sédiments. Dans les régions arctiques

l'écoulement des eaux de pluies représente une très grande capacité de transport car, en effet, il y a peu de végétation et le sous-sol est gelé, si bien qu'une bonne partie de la précipitation ruisselle dès qu'elle atteint la surface du sol. Le cycle hydrologique révèle un nombre disproportionné de débits relativement importants. De temps à autres des crues extrêmes sont engendrées par le vidangeage de lacs glaciaires.

La charge de fond constitue l'essentiel du matériel en transit, cependant, le matériel en suspension et en solution représente une quantité non négligeable. Etant donné que les cours d'eau remanie des volumes énormes de débris glaciaires, la quantité de matériel transporté dépasse de beaucoup la quantité de débris produit par altération.

Les chenaux des sandurs sont larges et peu profonds, une caractéristique des chenaux à charge de fond importante. A un accroissement de débit correspond une augmentation de la vélocité, ce qui indique un changement dans la forme du lit. Ce changement est associé à une augmentation rapide de la charge laquelle est liée à une réduction importante de la résistance à l'écoulement.

Le profil en long des cours d'eau est à peu près parabolique; cette forme est probablement liée à la nature de l'alluvionnement sur l'ensemble de la surface. La succession des seuils et des mouilles qui s'échelonnent le long des chenaux constitue la particularité la plus marquante des rivières. L'ensemble fonctionne comme un régulateur de l'énergie dissipée le long du cours d'eau et, par conséquent, exerce une influence sur la résistance à l'écoulement ainsi que sur le mouvement et la sédimentation du matériel.

La surface du sandur est une forme d'alluvionnement. Au cours des crues les plus importantes, l'érosion et la sédimentation affecte des superficies importantes. Les changements de niveau sont rapides et les dépôts sont souvent chaotiques, sans structure interne bien définie. La taille moyenne et la dispersion du matériel grossier sont les seules paramètres utiles; elles indiquent que la sédimentation se fait dans des conditions de compétence décroissante vers l'aval du sandur.

CHAPTER I

INTRODUCTION

Definition of the Study

Sandur, an Icelandic word signifying "sand" or "sand plain", has long been employed by Icelanders to refer to alluvial outwash plains formed by rivers carrying meltwater away from fronts of glaciers, and has come to be generally accepted to refer to glacial outwash. Sandurs are characteristically areas of rapid aggradation, crossed by braided streams that continually shift their pattern and course as local erosion and deposition occur.

Extensive outwash zones developed in front of retreating Pleistocene ice caps are found throughout northern Europe and North America on a scale, as befits association with continental ice, that is vast in comparison with presently active features. Present activity is mainly confined to valley deposits in the still-glacierized mountain regions of the world, though some aspects of relatively unrestricted outwash development can be seen in the broad valleys that open out to the Icelandic south coast from Vatnajökull. Here, the largest contemporary sandurs are found. In Canada, valley sandurs are developing in the Western Cordillera and in the mountains of eastern Arctic Canada.

This study of arctic fluvial environments in eastern and central Baffin Island, Canada, and of the alluvial processes currently active there, is focussed on this major class of glaciofluvial landforms.

Fluvial Processes in the Canadian Arctic

The Arctic has been recognized as a distinct "morphogenetic region", exhibiting a characteristic pattern of landscape development (Troll, 1944; Büdel, 1948; Peltier, 1950; and Tricart, 1950). These students, and others who have taken their lead, have suggested that the cold climate of the high latitudes, resulting in prominent frost activity and serving to inhibit the geomorphological processes commonly dominant in other climates, has produced a distinctive arctic landscape.

The suite of "periglacial" landforms: frequent talus and rock debris slopes, blockfields, frost-heaved terrain, patterned ground, frost cracks, ground ice phenomena, solifluction slopes and altiplanation terraces, doubtless lend an individuality to cold regions. Aside from the still inconclusive arguments about whether or not there are distinctive forms of periglacial slope development, however, these all represent merely details in the landscape.

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The large-scale organization of arctic landscapes remains not unlike that of other regions. Despite the recent exposure of much of the land from beneath Pleistocene ice, a reasonably well organized drainage network exists — much of it inherited from preglacial or interglacial time (cf. Summary for Northern Canada, in Bird, 1967).

Though arctic rivers flow only during the few summer months, and though there is no movement of surface water, or even of near-surface groundwater, in the Arctic for most of the year (exception being made for deep springs and a few large rivers whose drainage originates outside the Arctic), flowing water is still capable of great denudational and transporting activity when compared with other geomorphological agents. The spring snowmelt is remarkably rapid, and gives rise to a prominent freshet, so that whilst the moderate precipitation may be spread over much of the year, most of the derivative runoff is generated in a few days or weeks. In many parts of the high arctic much of the precipitation occurs as summer rainfall in any case. Because the ground is relatively bare, and because the presence of permafrost inhibits extensive infiltration of water into the ground, a high proportion of the rain runs off immediately on the surface. In many parts of the Arctic the continued presence of permanent snowfields or glaciers provides a source of significant runoff throughout the summer in response to seasonal warmth. Although Tarr (1897) emphasized the erosive power of summer meltwater streams after his expeditionary researches on Baffin Island and in West Greenland, and Büdel (1948), Peltier (1950) and Jenness (1952) discussed the significance of fluvial effects in the Arctic, fluvial action in the arctic regions has begun to be studied only in recent years. In Canada, Robitaille (1960) indicated the importance of streams in transporting materials, principally during the nival freshet, on Cornwallis Island in the central Canadian Arctic (75°N.). Mackay (1963) and Kennedy and Melton (1967) noted the efficacy of stream-waters in the Mackenzie delta region (69°N.). St-Onge (1965) emphasized the significance of nival fluvial activity on Ellef Ringnes Island (78°-79°N.), whilst Pissart (1967), discussing observations made on Prince Patrick Island (76°N.), pointed out the importance of the rare, heavy summer rainfall events in promoting fluvial activity. Rudberg (1963), working in the arid environment of western Axel Heiberg Island (79°N.) concluded that running water was the dominant landforming agent on deglacierized ground even here. The spring snowmelt did not appear to be particularly important in this environment, but heavy summer rains falling on nearly bare, impermeable ground affected extensive erosion. The importance of sheetwash and rillwash on exposed soil that is frozen below has been emphasized for middle latitudes by Tigerman and Rosa (1949), and has been speculated on in arctic context by Bird (1967). Finally, a posthumous note of F.A. Cook (1967), reporting stream velocity measurements in Mecham River at Resolute Bay on Cornwallis Island, explicitly suggested that stream runoff, principally during the spring freshet, is responsible for the bulk of all transport of weathered material.

In 1960, Rapp reported the startling result that running water provided the most effective transport process in Kärkevagge, a valley in northern Sweden, by way of transport of <u>salts in solution</u>. Since then, further measurements of actual sediment transport in arctic watersheds have begun to appear (Lamar, 1966; Arnborg, Walker and Peippo, 1967; Østrem, Bridge and Rannie, 1967) so that the importance of fluvial processes in polar regions can hardly remian in doubt.

Recent Fluvial Deposits in the Canadian Arctic

There has been little systematic mapping of surficial deposits carried out in the Canadian Arctic. Such maps as there are, however, indicate the presence of valley fill deposits in many river valleys. Maps by Ives and Andrews (1963), and by Sim (1964), covering an area of 100,000 sq km in central Baffin Island, indicate extensive fluvial deposits in all the major river valleys. Figure 1 consolidates these findings and extends the mapping to cover about 150,000 sq. km. St-Onge (1965) presented a geomorphological map of Ellef Ringnes Island which indicates similar conditions there. On Christie's (1967) map of northeasternmost Ellesmere Island (81°-83°N., 61°-79°W.) the fluvial deposits are subdivided as glaciofluvial or postglacial (meaning recent) deposits. While much of the deposit can be associated with glacial melt, active sedimentation is indicated in nonglacierized watersheds as well.

Whilst restricted to the valleys, fluvial deposits are unquestionably the most prominent recent features of the arctic landscape not directly due to ice activity. The range of such deposits runs in a continuum from alluvial cones, on which mass movements may also be important, and which may exhibit surface gradients in excess of 20°, through alluvial fans and valley alluvial plains to deltaic deposits in lakes or the sea (Allen, 1965). All are common in the Arctic in one place or another.

Almost all the valley fill deposits are composed of coarse sand and gravels, the frequent cobble beds. Braided streams characterize contemporary surfaces, and the channel scars on the surface of older deposits indicate similar hydraulic conditions prevailed on them as well. Most of the old deposits have been terraced. In many places this is at least partly the result of changing base level for the river. Isostatic uplift, which accompanied and followed melting of the Pleistocene ice mass, has caused general emergence of the coasts during the past 9,000 years. In other places, changes in hydraulic regime are the chief factors contributing to incision of older valley fills. Many of the older deposits appear to be of an extent and on a scale considerably greater than recent deposits, indicating that, at some time in the past, fluvial activity was more intense than at present. Such deposits may date from the period of higher meltwater discharge during rapid reduction of the late Wisconsin ice sometime after 9,000 years B.P., and extending through the Hypsithermal phase (see p. 12).

Alluvial plains are the most extensive fluvial deposits. Laid down as valley fills, or as broad plains in unrestricted country, these gently sloping surfaces (usually less than 5° gradient) of coarse clastic sediment are usually characterized by rapidly shifting, generally aggrading, braided streams with high competence and high bed-load sediment transport rate, rapidly changing sites of local erosion and deposition and, consequently, rapidly changing alluvial stratigraphy. Such deposits should be distinguished from those of a classical alluvial flood plain, which also include periodic overbank flood deposits of fine materials intercalated amongst the channel gravel deposits. Such material is carried in suspension by a river which in the short term, is normally relatively stable within a confined course. Studies of braided stream deposits have been largely carried out in the Alpine regions of the world, and extensively on proglacial outwash (Doeglas, 1962; Shumm and Lichty, 1963; Allen, 1965). Sandurs clearly belong to this group of deposits.

Major environmental factors that seem to be necessary for the deposition of coarse clastic valley fills are:

- an abundant source of coarse detrital material;
- sufficiently steep topography to sustain water-courses competent to move the material;
- hydraulic regime characterized by relatively frequent, high floods, so that significant quantities of material can be moved.

These characteristics are apparently present in possibly four environments: semiarid, periglacial, high mountain and proglacial.

In all these environments bare rock outcrop is common and physical weathering mechanisms produce coarse detritus. In many regions, an abundance of unconsolidated material was deposited during Pleistocene glaciations. Relatively sparse vegetative cover permits rapid erosion and movement of material to the water-courses. Frequent flooding is characteristic of all environments. In the semiarid regions the great concentration of runoff is due to the heavy rates of summer rainfall that are experienced in conventional storms characteristic of these regions; in periglacial regions the rapidity of snowmelt in the spring produced an annual nival flood, whilst summer rainfall onto largely frozen ground produces high proportions of surface runoff and subsidiary flooding. In the mountains a combination of heavy rainfalls and rapid spring snowmelt is found. Finally, in proglacial situations, the snowmelt floods continue throughout the summer as the glaciers melt in response to seasonal heating.

Sandurs

Types of Sandur

Two types of sandur have been conventionally recognized: "valley sandur" and "plain sandur" (Krigström, 1962). Krigström suggested the terms dalsandur and slättlandssandur for the two types. In English, the former has been conventionally referred to as a "valley train" (the term was apparently invented by Chamberlin in 1883), while the latter has been referred to as an "outwash fan" or "outwash plain". ("Outwash" is also used as an adjective in a much wider sense in reference to fluvial clastic deposits.)

Previous Studies of Sandurs

Studies of sandurs fall into two categories: stratigraphical and morphological studies of old deposits, mainly late Pleistocene, and studies on present-day sandurs. The widespread occurrence of late Pleistocene sandurs in regions subject to the last glaciation has prompted extensive study of them.

European interest in the valley fills of the Alps extends far back into the nineteenth century. Charpentier in 1841 identified the alpine schotters as "alluvium glaciaire". In 1857 Torell, after a journey to Iceland, asserted that the Heide of north Germany had been formed in front of the ice sheets of the Great Ice Age in the same way that the Icelandic sandurs are forming around Vatnajökull at the present day. Keilhack (1883) confirmed this insight, and appears to have been the first to introduce the Icelandic terminology (sandr, sander - see Spethmann (1911) for discussion of the etymology) into the mainstream European literature. Helland (1882) made the first measurements of sediment transport on the Hoffellssandur. Ahlmann and Thorarinsson (1937) summarized subsequent studies on the Icelandic sandurs down to the period of their own investigation beginning in 1936.

Jensen (1881, 1889) reported the existence of sandurs, or sioraq, in West Greenland, including a singular variety in some fiords where the glaciers apparently discharge only very fine silts and clays with the meltwater. The proglacial deposits in this case - most notably in Isortoq Fiord in West Greenland - are composed of silt plains or, in the extensive tidal zone, of mudbanks. The discoveries of subsequent explorers were summarized by Birket-Smith (1928). In the 1930's Bretz (in Boyd, 1935) and Flint (in Boyd, 1948) described outwash deposits in eastern Greenland fiords.

Extensive investigations of the sandur deposits of the north European plain and of the dalsandurs of the Alps are summarized by Charlesworth (1957). The Scandinavian sandurs have been mapped and described by Frödin (1925, 1954), by Mannerfelt (1945) and by Krigström (1960). Valley sandur deposits have been identified in most of the other alpine regions of the world (see the Summary in Charlesworth, 1957).

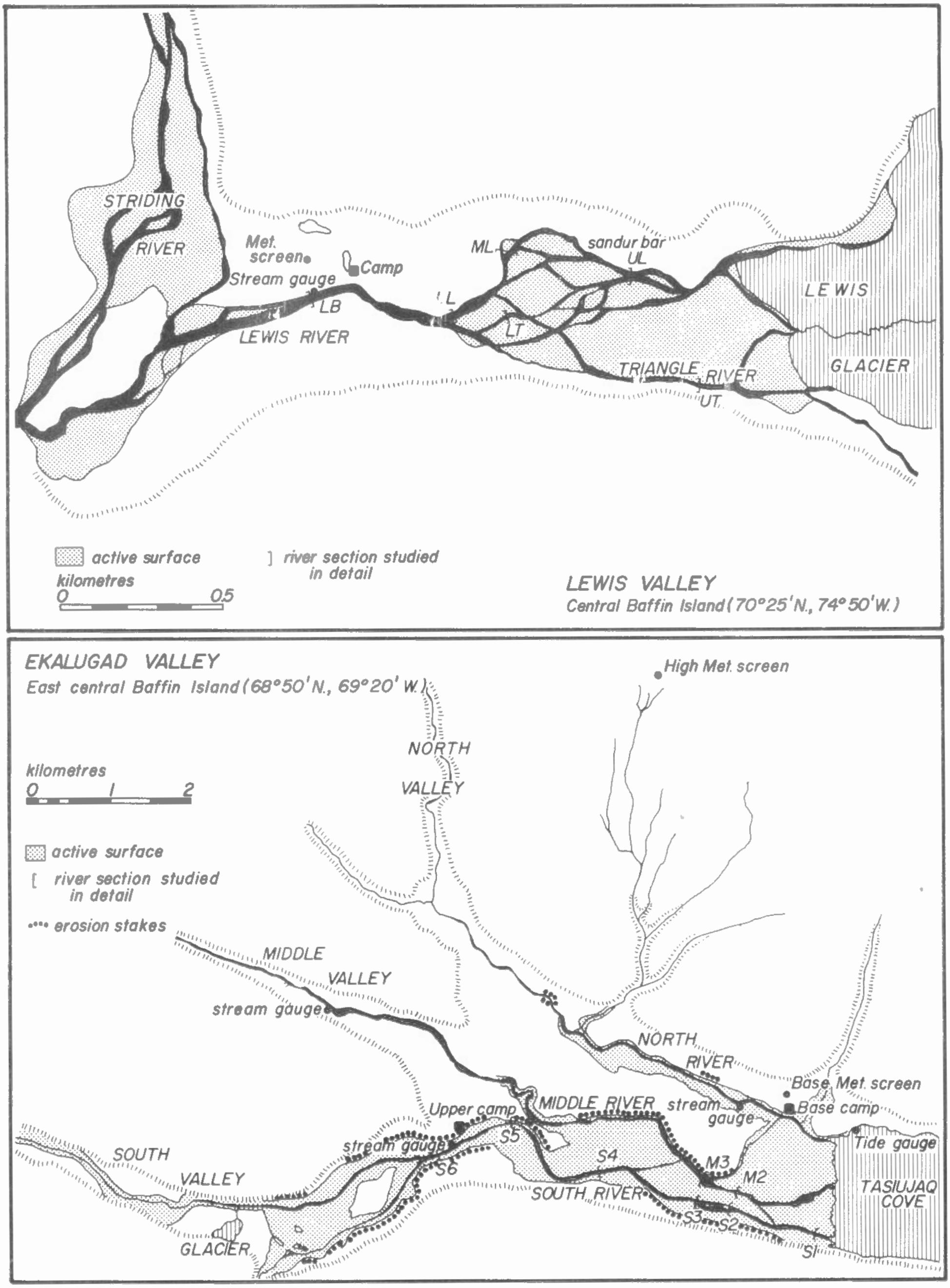
In North America, early work seems to have been largely independent of work in Europe. The existence of waterlaid deposits in association with the Ice Age glaciers was early recognized by Hitchcock (1857) and by Dana (1875). Chamberlin (1883) and Salisbury (1892) firmly established the nature and extent of sandurs in the northern United States. Major works on sandurs in the United States have been summarized by Flint (1971).

Early observations of sandur processes came from the Alaskan expeditions of Russell (1893), Gilbert (1904) and Tarr (Tarr and Martin, 1906, 1912, 1914; Tarr, 1908, 1912; Tarr and Butler, 1909). All of these works have been descriptive or mainly of reconnaissance nature. There are, in fact, very few studies of sandur processes. The first studies of sandur development which involved serious measurements were carried out on Icelandic sandurs by Thorarinsson (1939).

The 1951 and 1952 Uppsala University expeditions to the Hoffellssandur (in Hornafjordur, on the south coast of Iceland) made the first detailed study of the morphology and hydrology of a sandur. The results (Hjulström et al., 1954-57) include reports of the regional geology and geomorphology, climate, hydrology, and certain special hydrological aspects of the sandur. However, commentary on the mechanics of the alluvial processes was restricted to a brief summary by Hjulström (1952).

Krigström (1962) gave a descriptive analysis of the channel processes which contribute to channel anastomosing and bar building on the Icelandic sandurs.

In 1963 Fahnestock published a major study of the dalsandur of the Emmons Glacier (White River) on Mount Rainier, in Washington. This work is firmly based on the studies of river processes and morphology carried out in the United States Geological Survey during the preceding fifteen years, and presents the first detailed analysis of valley sandur processes. Recently, Fahnestock (1969) has continued his studies on the Slims River in front of Kaskawulsh Glacier, in Yukon Territory, Canada.



A. Lewis Valley.

B. Ekalugad Valley, at the head of Sarvalik arm of Ekalugad Fiord.

Figure 2. The field areas, showing installations and special studies.

The Field Program

Location of the Field Areas

Field studies of sandurs in Baffin Island were carried out in two locations, marked on Figure 1 by arrows. The first was a small sandur in Lewis Valley at the northwestern corner of Barnes Ice Cap (70°24'N., 74°53'W.). Field work was carried out between 1963 and 1965. The major objectives were hydrological, but extensive studies on the outwash and other ice marginal features were also carried out to provide detailed information on a sandur in a high arctic environment (Fig. 2A).

The primary field area was at the head of the Sarvalik arm of Ekalugad Fiord (68°50'N., 69°20'W.) on the Home Bay coast of east-central Baffin Island. Work at Ekalugad Fiord was carried on between 1966 and 1968, and attempted a comprehensive hydrological and sedimentological study of the presently active surface (Fig. 2B).

The two sites are in marked contrast to each other. Lewis Valley is an inland site on the margin of Barnes Ice Cap, a Pleistocene relict ice mass of the central Baffin plateau, and the watershed is about 90 per cent glacierized. Ekalugad Fiord is a coastal site in the eastern mountains of the island, where mountain snow and ice fields provide for much less than 50 per cent glacierization at present. Consequently, instructive contrasts occur in almost every aspect of the sedimentological environment.

The Plan of the Study

The objective of the field investigations was to study the sedimentary environment of sandurs in the eastern Canadian Arctic. A sedimentary or depositional environment is the pysical, chemical and biological system in which the sediments are formed (Scruton, 1960, p. 92-93). Such a complex system is best investigated within the framework of an environmental model. For the present purpose the most flexible model is the generalized "process and response" model of Krumbein and Sloss (1963). In the context of study of the terrestrial clastic sediments on Baffin Island, the biological element can be ignored. An operational framework for the study is presented in Figure 3, where the specific factors to be studied within each environmental element are noted.

The process elements are reviewed in Chapter II. In the following three chapters each aspect of the transfer process is studied. Since weathering processes were not directly studied, they are considered along with sediment transport in Chapter IV. Chapter VI presents an analysis of the response element; the morphology of the sandur deposits.

Inasmuch as they point to subjects that require major studies before an endeavour such as this could be completely successful, it is interesting to indicate here some of the shortcomings of the results. A great deal more must be learned about hydrology in the Arctic before its full significance in conditioning geomorphological events can be assessed. Even more generally, it appears that much more study must be given to the magnitudes and frequencies of various climatic and hydrological events before they can be firmly related to the pattern of geomorphologically significant events. That sequences of natural events are "drawn" from mixed populations is readily apparent in this study.

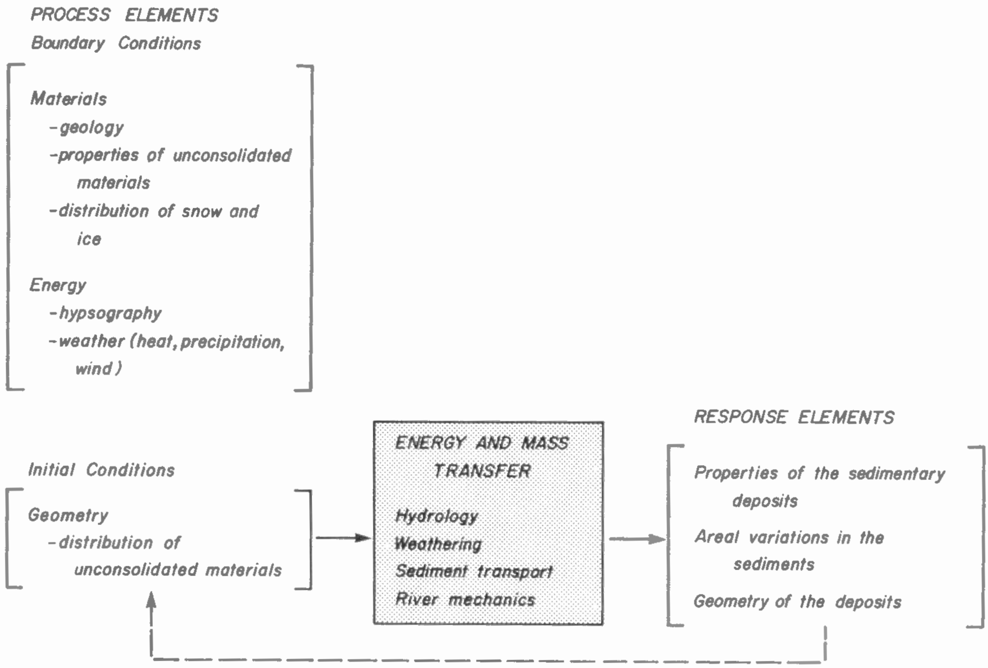


Figure 3. Operational process-response model (after Krumbein and Sloss, 1963) used in the design of this study.

The writer was aware, before the study was commenced, that knowledge of sediment transport mechanics (or, rather, lack of it) was likely to be problematic. Though several innovations were made, both in the field and in analysis, the results remain unsatisfactory.

The hydraulic studies produced singular results and, though a tentative interpretation of them has been made, it is clear that there is more than one major study to be done in this area alone.

Because of the potential value of much of the data for other workers, it has been reported in full in a data Appendix of 21 articles that accompanies the writer's doctoral dissertation, presented to the University of British Columbia (Church, 1970). Full descriptions of field methods and data reliability accompany the tables in that appendix. In order to reduce the length of this paper, only very brief discussions of these aspects are given.

Acknowledgments

The project was supported by the former Geographical Branch, Canada Department of Energy, Mines and Resources during the years 1963-67, and by the Division of Quaternary Research and Geomorphology, Geological Survey of Canada, thereafter.

A project as large as this has inevitably indebted itself to many people. Dr. J. D. Ives, Director of the Geographical Branch, initiated and encouraged the project. Dr. Olav Løken provided much help in his capacity as field supervisor of the Baffin Island project of the Branch, and Dr. J.G. Fyles, chief of the Division of Quaternary Research and Geomorphology (the present Terrain Sciences Division), Geological Survey of Canada, continued this support. Martin Barnett, Jane Buckley, Doug. Christian, George Falconer, and Doug. Hodgson provided administrative and logistic support. Special equipment was loaned for the project by the Inland Waters Branch. Laboratory work was carried out by Rolf Kihl and R.G. Kelly. Cartographical and computational work by Miss Linda McKim and Miss Lorna Procter was instrumental in readying the results for release. In 1967 the Arctic and Alpine Research Committee of the University of British Columbia awarded a grant to help the research. Computations were carried out at the Computing Centre, University of British Columbia.

The greatest contribution to the project was made by the field people. At one time or another substantial help was received from Angus Cherrington, Vincent Coulombe, Norman Gray, Peter Lewis, Erik Nielsen, John Richardson, Miss June Ryder and Miss Penny Crompton. Dr. J.T. Andrews, Dr. P.J. Webber, and R.B. Sagar, the leader of the Barnes Ice Cap parties during the Lewis River studies, extended assistance and collaboration in the field on many occasions.

The principal field assistants were Terry Day (1967), Alan Graves (1968), Robert Stock (1967), John Knight (1967-68), William Rannie (1963-64), and Barry Goodison (1965-67).

The manuscript has been reviewed and commented upon by Dr. R. Kellerhals, Dr. J.R. Mackay, Dr. B.C. McDonald, and Dr. O. Slaymaker. Whilst they provided much helpful advice that has been appreciated by incorporation into the paper, the many shortcomings that remain must be ascribed entirely to the writer.

CHAPTER II

THE ENVIRONMENT OF EAST-CENTRAL BAFFIN ISLAND

Physiography and Geology

Baffin Island forms the northeastern rim of the Precambrian Canadian Shield, with the surface rocks of the bulk of the island composed of granite gneisses.

Central Baffin Island may be subdivided into four major physiographic zones (Rand, 1963; Bird, 1967):

1. On the east coast a narrow coastal plain occurs in some regions (cf. especially the Clyde region), though it is absent along the Home Bay coast;
2. Behind this the mountains of eastern Baffin Island rise abruptly. This rugged region is about sixty to eighty kilometres wide, and has peaks rising to 2,000 m in the Penny Highland at the south-eastern end of the island. The Home Bay coast is, however, a relatively low "saddle zone" with the general summit level being at about 1,000 m. The mountains are extensively cut through by deep valleys -- probably many of them originally fluvial, but recently overdeepened by Pleistocene ice to form great, through-running fiords. The heads of the major fiords are found on the western side of the height of land in a region of lower, more rolling hills that give way to the central plateau of the island;
3. The central plateau forms the bulk of the island's breadth in its central portion, and falls away southwestward from around 900 m to about 300 m on its western side. Presently the rivers are entrenched in the plateau, so that about 300 m of relief may occur locally. The Barnes Ice Cap is located on the plateau and rises to a crest elevation of about 1,150 m. The southwesterly trending preglacial drainage was extensively disarranged during the Pleistocene and the ice cap still dams several large lakes on its eastern side which drain over cols into the fiords rather than to the west; and
4. Further west again are the Foxe Basin lowlands. They are narrow in places and are underlain predominantly by Paleozoic sediments which also underlie Foxe Basin.

The geology of much of the island is composed of the root zone rocks of a former major mountain range of Precambrian age. Two major suites of rocks occur, both derived from altered sediments. The true "basement" is a widespread migmatitic rock which exhibits highly complex internal structure. This is the country rock at Lewis River and forms the basement in the Home Bay sequence. Superimposed on this, and visible in belts of greater or lesser width, is a metasedimentary series composed, for the most part, of gneisses and schists. A major belt of such metasediments extends to the Home Bay coast and to the north in the Clyde region (Kranck, 1955).

The post-Precambrian history of the region has apparently been largely denudational (Martin, 1961). The major features of the dominantly southwesterly oriented drainage pattern (cf. Fig. 1) are pre-Pleistocene, and may be very ancient. In the early Tertiary, active tilting and block faulting occurred (Riley, 1956) and produced the presently uplifted eastern side of the island.

The present landscape details are overwhelmingly the product of the Pleistocene glaciations.

The Late Pleistocene Setting

Glacial Activity

The last major glaciation in Baffin Island, designated by Andrews (1965a) the Foxe Glaciation, probably commenced more than 85,000 years ago, but little evidence has been left of events prior to the moraine building phase in northern and eastern Baffin Island between 8,000 and 8,500 years ago (Falconer, Ives, Løken and Andrews, 1965; Blake, 1966; Andrews, Buckley and England, 1970), known as the "Cockburn" phase. Evidence summarized by Andrews (1965a, 1970), Løken (1965), Barnett (1966) and Andrews, Buckley and England (1970) indicates that subsequent major stillstands occurred at about 7,000 and 5,000 years B.P. After the stillstand of ca. 7,000 years ago (the "Isortoq" phase), deposition of extensive fiord-head sandurs commenced.

Once having retreated across the regional watershed in eastern Baffin Island the ice began to impound a series of proglacial lakes (see Generator Lake and Conn and Bieler Lakes today: Fig. 1). These have provided continued drainage to the fiords of large volumes of water capable of considerable fluvial activity. Most of the fluvial activity near the fiord-heads probably occurred between 7,000 and 3,000 years ago. Inland, sandur surfaces developed in many river valleys following the ice retreat.

It has been surmised that during the last glacial maximum the extent of the eastern mountain glaciers was probably not greatly different from today (Ives and Andrews, 1963). This probably began to change when more open water appeared in Baffin Bay after the late glacial maximum. After the retreat of the major outlet glaciers from the fiords, the local mountain snowfields pushed glacier tongues into the fiord valleys, breaching the lateral moraines left by the former valley glaciers (Andrew, Buckley and England, 1970; Andrews, Barry and Drapier, 1970).

Relatively little is known about the chronology of fluctuations of these local glaciers, though two major moraine systems appear to be associated with them in widely varying areas (Harrison, 1964, 1966). The first are moraine ridges whose age is 2,000 ± yrs. B.P. Inside these moraines, or occasionally overriding them, are a series of fresh moraines which mark a recent maximum. A variety of data, summarized by Harrison, leads to a reasonable inference that the ages of these moraines are A.D. 1750-1800, 1840-90, 1920-40, and 1950±. Since 1950, the glaciers have generally retreated, some by up to 5 km, from the outermost recent moraines. The advance which culminated between 1750 and 1800 also saw the development of extensive bodies of thin ice and permanent snow over the central Baffin plateau. Ives (1962) concluded that as recently as 200 years ago about 70 per cent of the upland was covered by permanent ice or snow. Today this value is only about 2 per cent, and the remnant ice bodies continue to melt rapidly (Falconer, 1966). These events have provided most of the runoff, along with seasonal melt of winter snow, to sustain the presently active alluvial surfaces on the island.

Stréam Regimen in Postglacial Time

Very little is known even of contemporary hydrologic regimen in the Canadian Arctic. Peak runoff rates measured in Baffin Island under the influence of heavy summer rainfall and warm temperatures are of the order of 1 to 1 1/4 $m^3 s^{-1} km^{-2}$. (These data were observed in watersheds of about 200 km^2 area: one watershed was mostly glacierized, while the other had about 10 per cent ice cover.) It is not likely that peak "normal" discharges should have been remarkably greater than this at any period in postglacial time. There are abundant signs, however, of extraordinary floods and of highly concentrated fluvial activity -- disproportionately large boulders are found within fluvial deposits; large canyons and gullies have been cut by the action of water. These features are the result of diversion of water by ice, and of sudden release of ponds or lakes which might have been held for long periods behind ice or snow dams (jökulhlaupe). Such events still occur along and near the margins of the mountain glaciers, and around Barnes Ice Cap.

It is likely that seasonal runoff and flow durations were somewhat greater through much of postglacial time (7,000-3,000 B.P.), particularly during the warm Hypsithermal phase. The scanty evidence suggests that a marginally warmer climate probably existed during much of the period (see summary in Løken, 1965). However, it is probable that runoff today is as great as it has been at any time during the past 3,000 years, which has generally been a cool period. Hence, it appears that the period of most active fluvial erosion and sedimentation during postglacial time in the eastern Arctic, as indeed in most of the northern lands, has been in the earlier period before about 3,000 years B. P.

The zone of greatest fluvial activity is closest to the margin of the ice. Here there is abundant material in the form of morainic debris available for reworking and transport, and the probability of experiencing large floods capable of doing great work is highest. Summer rainfall runs off immediately from bare ice, warm summer weather always finds more snow or ice to melt, and jökulhlaupe frequently occur.

The Climate

Regional Data

Baffin Island lies in the peripheral zone of the region of high arctic climate that covers most of the Canadian Arctic Archipelago. The general climate has been summarized by Thompson (1967). With the moderating influence of the North Atlantic Ocean felt from the southeast, it is neither so cold in winter nor so warm in summer on Baffin Island as at sites in similar latitudes across the rest of the Canadian Arctic. There is also considerably more precipitation here. It is nonetheless a fully Arctic climate and would be called "marine Arctic" in traditional classifications.

Winter circulation, which develops in late September and persists until May, is dominated by the major region of high pressure that develops over the western and central Arctic. During the early part of winter the "Baffin Bay Trough", a recurrent area of low pressure that persists over Baffin Bay and Davis Strait, brings much stormy weather to southern and eastern Baffin. Most of the winter's snow falls before Christmas.

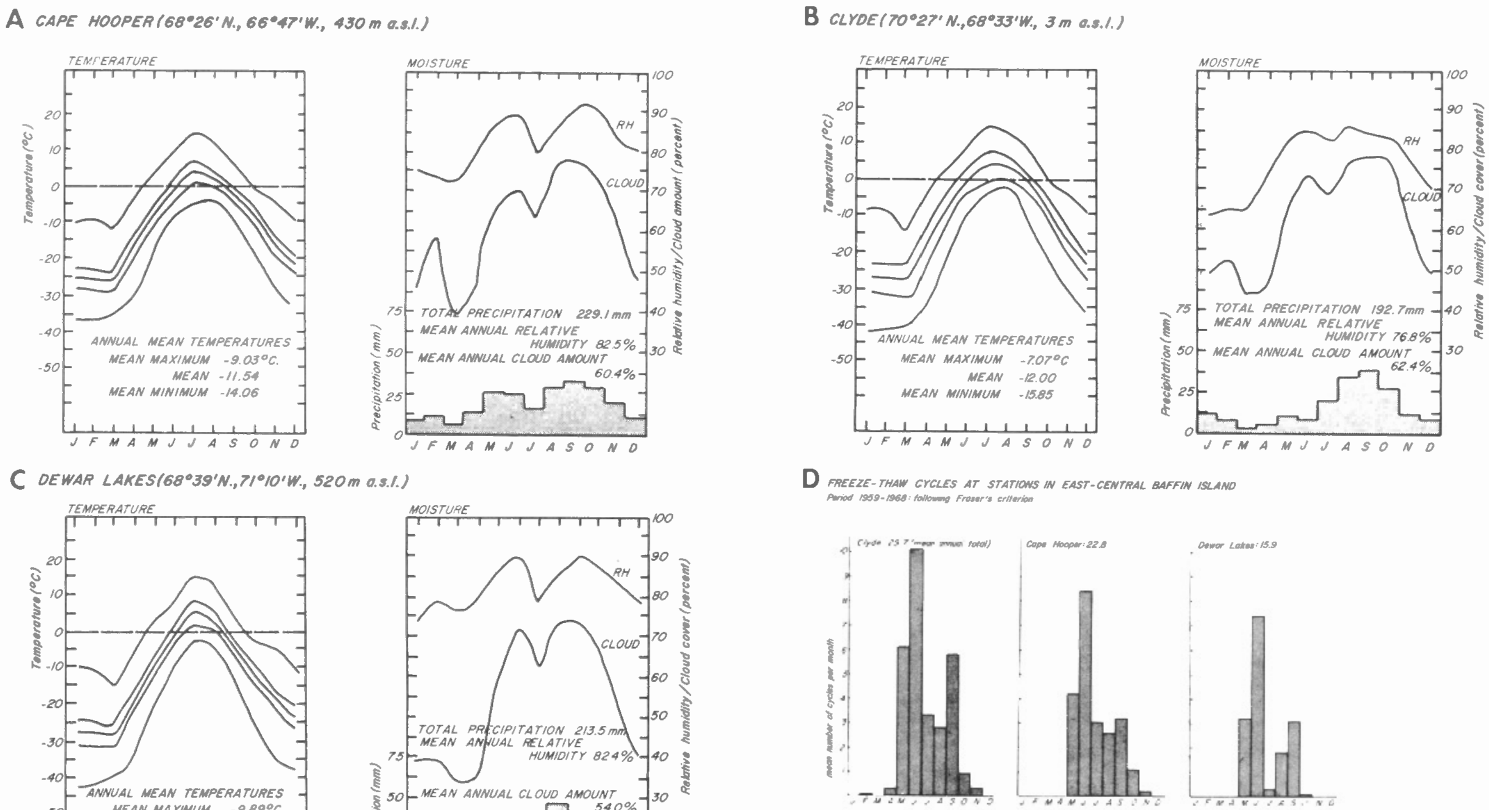
The summer is a period of weak circulation over the Arctic. Winter conditions usually break down in May with a series of disturbances that extend through the summer season. As the summer advances open water becomes more common along the coasts, and in these circumstances advection fog, extensive low cloud and drizzle are common around the coasts. July and August are usually the wettest months. Barry (1967) calculated that the median position of the Arctic front in July crosses central Baffin Island, confirming that this is a region of continued cyclonic activity throughout the summer season.

With the autumn, cold air outbreaks behind the frontal disturbances become more intense, storms become more severe, and the beginning of winter in later September is marked by the stormiest time of the year.

Three stations regularly report weather data in east-central Baffin Island; the climatological stations at Cape Hooper and Dewar Lakes, and the first-order station at Clyde (Fig. 1). Data from these stations for the 10-year period between 1959-68 are summarized in Figure 4, in order to summarize the regional climate. Dewar Lakes, on the central Baffin plateau, is a reasonably good indicator station for conditions in the Barnes Ice Cap region (but cf. Sagar, 1966, and Løken and Andrews, 1966, for a consideration of the apparently different conditions near the northern end of the ice cap). The other two stations represent the coastal regional climate.

A more germane indication of the range in runoff conditions that might be encountered on Baffin Island is provided by the variability of climate. Figure 5 displays various exceedance data for temperature and precipitation at Dewar Lakes. Summertime temperature variability appears to be rather small. However, in about 25 per cent of the years the mean temperature in June may exceed 0° C, and in about 15 per cent of the years, it may exceed 0° C in September. Then in about 4 per cent of the years, the melt season may approach four months in length. Similarly, in about 4 per cent of the years, the season may be as little as one month. It would appear, then, that the length of the season may be of greatest importance in determining total runoff from melt.

Summer precipitation is highly variable: in 5 per cent of the years, August precipitation may exceed 125 mm: in 20 per cent of the years it may exceed 100 mm, and it may be zero in 20 per cent of the years as well. The combined July-August precipitation may exceed 150 mm in 5 per cent of the years: and it may be less than 10 mm in a similar proportion of the years. Thus, storm runoff appears to be the most important contributor to runoff variability though its overall effect may be dampened somewhat by the normal coincidence of good melt years (clear and warm) with low precipitation. The reverse case is not so clear since cool, wet seasons that are very cool may be poor runoff seasons if much of the precipitation persists as snow.



All line data represent monthly means of daily values. Precipitation histogram represents means of monthly totals. The lines in the temperature diagram represent the following, in descending order:

- mean of monthly absolute maximum temperature
- mean of daily maximum temperatures
- mean daily temperature
- mean of daily minimum temperatures
- mean of monthly absolute minimum temperature

In the moisture diagram, RH designates the curve of mean daily relative humidity, and CLOUD designates the curve of mean daily cloud amount.

Figure 4. Climatic means for the period 1959-1968 at stations in east-central Baffin Island.

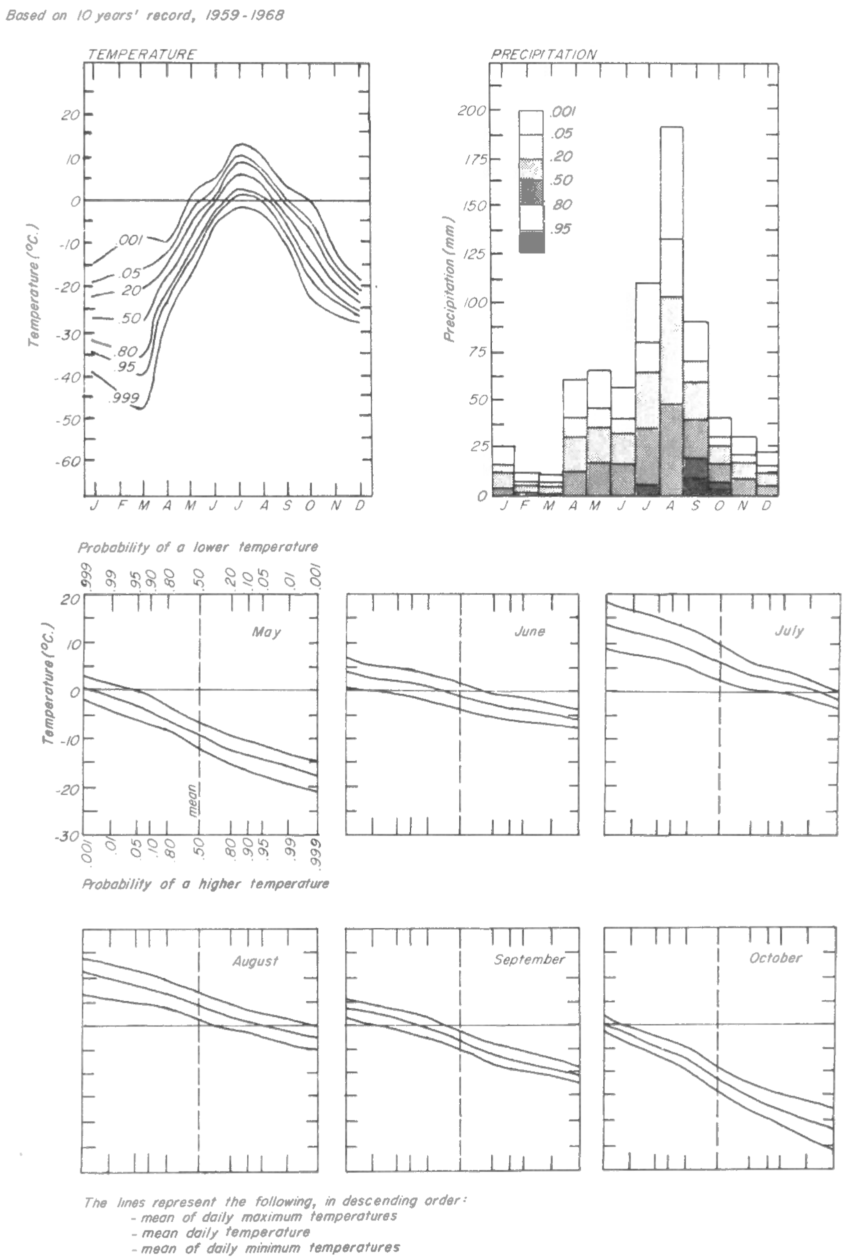


Figure 5. Exceedance values for mean temperature and precipitation at Dewar Lakes (based on the period 1959-1968) and derived probabilities for temperature exceedances.

Climate and Fluvial Processes

There are only two seasons in the Arctic: the winter, which normally lasts for nearly nine months of the year in eastern Baffin Island, and a short, cool summer season which extends from early June until mid-September.

Snow and ice melting in the spring contribute to a strong peak runoff in early July in most watersheds. During the rest of the summer, runoff steadily decreases as there is less and less snow remaining to be melted. However, in those watersheds where permanent snow or ice fields occur, renewed melting occurs throughout the summer during warm weather, so that the peak runoff is delayed until late July or early August, when the warmest weather usually occurs. The most effective melt occurs during strongly advective periods, since a good deal of direct radiation is reflected immediately from snow and ice surfaces.

The nival flood need not necessarily give rise to extensive sediment transport and associated fluvial activity. Pissart (1967) described a four-stage sequence of events during the snow-melt period as follows:

1. local melt of the snowpack (effect of solar radiation) whilst the temperature still remains below 0° C;
2. general melt and percolation of the meltwater to the base of the pack and into snow-choked stream courses, where it refreezes;
3. commencement of runoff over the snow and ice in the streambeds;
4. runoff on the streambed after melt of the bottom ice.

The third stage might persist through the major part of the flood period, effectively protecting sediment in the streambeds from erosion. A sequence somewhat like this was noted in the Baffin sandur streams during breakup (cf., p. 29), but the third stage was not observed to be important there.

Summer precipitation periodically punctuates the seasonal flow recession in contributions of storm peaks to runoff. During a cool, stormy summer, the seasonal hydrographs of streamflow consist mainly of a series of storm events. The Arctic is not a region of heavy precipitation indeed, most of the Arctic would qualify as a climatic desert, though it certainly is not a physiological one. Individual storm events associated with intense frontal precipitation, or occasionally prominent local convectional development, are capable of delivering rainfall at rates of up to 50 mm in 24 hours (49.5 mm was recorded in 24 hours on July 22, 1965 at Lewis River; Pissart (1967) reported 47 mm during one 24-hour period occurred at Mould Bay on Prince Patrick Island). This leads to heavy storm runoff.

Many of the primary processes of rock weathering, sediment production and preparation for erosion and associated with freeze-thaw activity (see also Chapt. IV, sub-section Periglacial Processes). Figure 4D summarizes freeze-thaw events indicated by the meteorological records for the three regional stations over the last 10 years. Fraser's (1959) criterion was used, defining a significant freeze-thaw cycle by a temperature amplitude extending at least from 34° F to 28° F and back. Cooke and Raiche (1962) have shown that a greater number of freeze-thaw cycles occur at the ground surface than occur in the free air (or in a screen), but that below about 1 cm only the annual cycle is important. Most of the freeze-thaw activity occurs in the spring when warming of the near ground environment is delayed near the freezing point for long periods by the presence of snow and ice. In the autumn, the return of freezing conditions is usually relatively abrupt.

From the point of view of sediment transport, the most productive floods are those caused by storm runoff, for in this case considerable overland flow occurs that is missing from the runoff pattern of meltwaters from permanent snowfields.

Lewis Valley

Introduction

The Lewis Valley, which drains the northwest corner of Barnes Ice Cap, is a relatively broad, open valley cut into the interior plateau to a depth of about 300 m. It represents the eastward extension of Isortoq Valley, whose headwater region is presently occupied by Conn Lake. Because Barnes Ice Cap still occupies the main valley, the upper Isortoq River is considered to be the stream that turns sharply north away from the ice cap at its northwestern margin. Lewis River is a tributary stream of the Isortoq (Fig. 6).

The bedrock, well exposed along the valleysides, consists of migmatic gneisses. Gneissic layering dips north at about 20° in the area, and rectangular jointing is superimposed upon this whose spacing is characteristically several feet except in the plane of bedding, where it is less than a foot. Brecciated material occurs in minor fault zones.

The recent history of the Lewis Valley has been treated by Andrews (1965a). The valley was occupied by ice until the early 18th century, after which Lewis Glacier began a more or less steady retreat that has continued to this day. There has been continual, but not morphologically continuous, deposition of glaciofluvial material throughout the retreat. Shortly after 1900 the glacier stood at the moraine that marks the outer limits of the present outwash deposits, and since about 1920 it has retreated over the ground now occupied by the sandur (Fig. 6).

The Present Sedimentary Environment

Lewis Glacier remains the only prominent valley outlet glacier of Barnes Ice Cap, and brings the ice to its lowest elevation around the entire ice cap margin (245 m a. s. l.). The glacier shows no sign of movement and is melting back relatively rapidly (Anonymous, 1967). Since it is a "cold" ice cap it is frozen to its bed, and so the ablation runs off entirely on the surface. Hence, the contributing watershed of Lewis River may be defined by the surface watershed of the glacier. The statistics for the watershed are presented in Table 1. They show that 88.8 per cent of the watershed area is ice.

An estimate made by Andrews (1965a) of the recent downwasting of Lewis Glacier provides an opportunity to estimate the runoff of Lewis River. If it is assumed that the drainage basin on the glacier has remained constant over the period since 1740, that ice movement has been relatively minor, and that the indicated downwasting has occurred equally over the entire basin, then the value of 130 m of downwasting since the mid-eighteenth century maximum deduced by Andrews for the area near Flitaway Lake may be used to estimate mean annual runoff. If it is also assumed that the mean ice area over the 225 year period since 1740 is 195 km^2 (it has in fact varied from 206 km^2 to 182 km^2) the following approximate calculation may be made:

$$\text{mean annual runoff} = \frac{0.9 \times 130 \times 195 \times 10^6}{225} = 100 \times 10^6 \text{m}^3 \text{ water}$$

due to glacier melt (0.9 is the ice density). The only available estimate of annual precipitation that might have some long-term relevance is 30-35 cm for Barnes Ice Cap (Sagar, 1966; from accumulation data). Hence, an average annual flowthrough of 60×10^6 to $70 \times 10^6 m^3$ of water will be assumed. This compares with a recently measured mean annual runoff of $148 \times 10^6 m^3$ water during the years 1963-1965 (Anonymous, 1967). All three years were considered to be cool years. At any rate, it is probable that the magnitude of runoff in the recent past has not been appreciably greater than it is at present. The measured runoff value implies a mean flow for a 70-day season of about 25 $m^3 s^{-1}$.

The outwash plain, though small, is highly active. It is about 1 km long, and is restricted in downvalley extent by the 1905 moraine of the glacier, through which Lewis River passes in a single channel and eventually drains to Isortoq River 1.9 km farther downstream. The sandur is about 1/2 km wide. Along its sides are a variety of terrace and kame deposits attesting to earlier glaciofluvial activities.

Most of the sediments available to the river are morainic materials melted out of the glacier, or reworked materials previously deposited in the river terraces. Though the upland is mantled with a thin veneer of ground moraine and solifluction occurs, the proportion of material delivered from this source is not great. Freeze-thaw weathering is important in the comminution of materials (cf. Fig. 7 - especially sample 6), but is relatively unimportant in primary delivery (from bedrock).

Vegetative colonization of the valley is recent and minimal, consisting mainly of lichens and scattered, pioneer species of higher plants.

The sublateral course of the river along the edge of the glacier has led to the glacier having a special influence on stream regimen. Snowbanks and ice falls along the glacier margin often block or divert the flow: most spectacularly at Flitaway Lake. Each year its outflow is impounded by snowbanks until late in the season. Then, when the dam is breached, drainage of the seasonal inflow is rapid. Hence, extraordinary floods are a normal feature of the river regime. Water storage and release under the snout of the glacier, or at slush ponds on the surface, similarly contribute frequent surges, or jökulhlaupe, at almost any time of the summer season. Hence, relatively high flood flows are fairly frequent on the river.

Ekalugad Fiord

Introduction

Ekalugad Fiord is located in the eastern mountains at a point where the peaks are relatively low (1,000 m±); Rand (1963) classified the region as the "Home Bay Upland". The fiords are straight and open, so that marine climatic influences penetrate easily right to the fiordheads. Consequently precipitation is fairly heavy and there are permanent snowfields on the mountaintops (Fig. 8) so that there is relatively great and well sustained summer runoff. Weathering processes, particularly physical rock weathering, are relatively active here, and so production of clastic sediment is quite rapid.

The fiordhead valleys cut obliquely across the trend of the folded rocks (see Fig. 9). In addition to the remanent bedding and textural foliation possessed by the rocks, open jointing with characteristic spacing of several feet, vertical and approximately rectangular, is common, especially in the more massive rocks. Hence in many places the rocks are susceptible to weathering.

Recent History of the Fiordhead

The postglacial history is outlined in Figure 10. An interesting feature of the fiordhead is the presence of several former outwash surfaces in the valley. Early deposits were ice-lateral or sublateral terraces developed while the ice was still present (T0 and T1). However, the T2 surface (visible in Figs. 8, 10, and 11) was a major sandur plain that was initiated after the ice had retreated to the head of the main valley. It is now terraced so that it stands approximately 12 m above the active surface. Reconnaissance has indicated that such old surfaces are common in the fiords of the mid-Baffin east coast. They mainly date from the period between 5,000 and 3,000 years B.P.

A sequence of terrace facets eroded in the T2 deposits after active construction of that surface ended has been designated collectively as "T3". The presently active sandur is designated T4. Its beginnings are obscure. Almost all of the sediment has been derived from the South Valley side, and a good deal of it is reworked T2 material. The development of the upland snow and ice fields has been most active in the South River watershed in recent times. Though no morphological evidence exists to provide information for a chronological assessment, it is likely that the present aggradational surface was initiated approximately 2,000 years ago, following the early maximum buildup of upland ice (cf., p. 10).
Development has continued sporadically ever since. There is no doubt that the evolution of the recent surface has been far slower than that of T2.

The redevelopment of upland ice in recent times had the effect of blocking drainage in upper South Valley so that a lake formed there which drained after eroding an outlet around the north side of the moraine. The ice pulled back from the moraine in about 1,800[1] A.D. and through drainage occurred almost immediately. At present, water ponds behind an icefall farther upvalley, and sporadically drains under it when the flow passage melts out. The resultant floods continue to maintain significant sediment transportation and redeposition on the surface.

The Present Sedimentary Environment

The present watershed of the Ekalugad sandur comprises three basins, one draining through each of the valleys at the head of the sandur (Fig. 9). Summary data on the basins is given in Table 2.

The upland snowfields, have decreased considerably in recent years; several easterly slopes and nivation hollows have ceased to carry permanent snow since 1948 (as determined from the aerial photography of that year). However, they still occupy about 20 per cent of the entire sandur

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1. Date derived from measurements of the diameters of the lichens Alectoria minuscula s.l. and Umbilicaria proboscidea (Andrews and Webber, 1964). Growth rates determined by Andrews and Webber were used in interpreting the results.

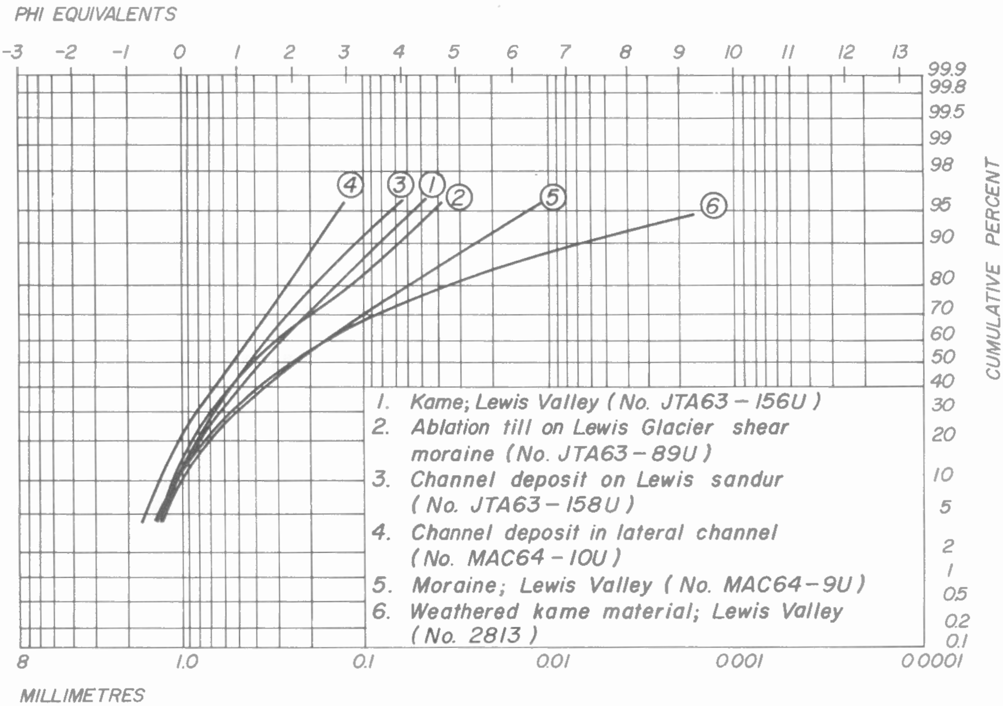


Figure 7. Sample materials from Lewis Valley (fraction finer than 2mm).

drainage area and can have a major influence on summer runoff patterns. Hence, the normal nival climate runoff pattern is somewhat tempered since warm summer weather can produce new peak flows as extraordinary melt occurs on the uplands. This effect is most strongly marked in South River basin where in a warm year the peak runoff will occur during late July or early August. North River, by contrast, exhibits the most nearly "normal" nival regime. In this, as well as in its morphological aspects, Middle River lies intermediate between the two.

South River basin, with virtually one-half of its area glacierized at present, has by far the largest area of upland ice and snow. It also has the most singular distribution of area with elevation of any basin: it possesses proportionally the largest valley area, below 400 m, and the largest area above 700 m, on the uplands (Fig. 12). Abrupt cliffs divide most of the upland from lowland areas of this basin, and virtually all of the upland area carries permanent snowfields.

The active sandur is being fed almost entirely from this tributary valley. The sandur is building directly into Tasiujaq Cove, at the head of the Sərvalik arm, so that it is developing on top of a delta. It is about 10 km long and, in its distal portion, 1.5 km wide. Periodic drainage of the icefall lake provides by far the largest runoff events across it.

Most of the drainage area is today mantled by only a thin surface cover of ground moraine, or by boulder fields (felsenmeer) from which fine materials have been washed away completely. Bedrock outcrop is very common on the uplands and steep slopes. Beyond the sandur, the only area of extensive recent deposits is in upper North River valley.

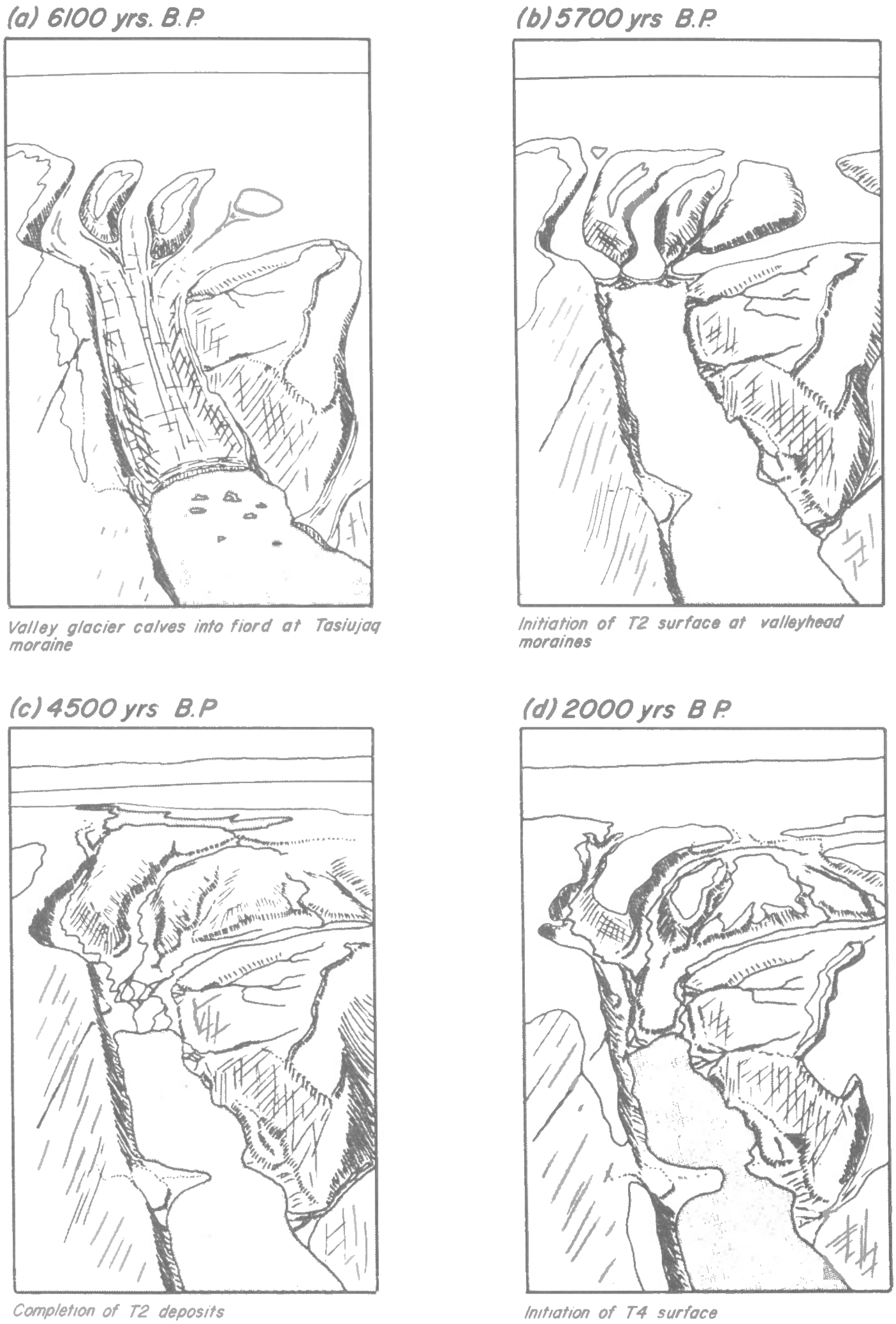
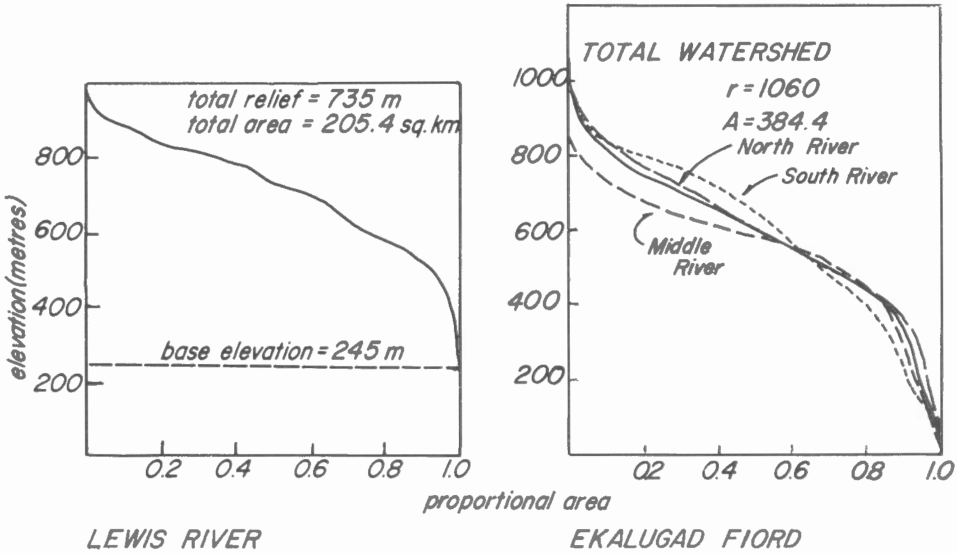


Figure 10. Stages in the postglacial history of Ekalugad Valley.



A. Lewis River
B. Ekalugad sandur watersheds

Figure 12. Hypsography of Baffin sandur watersheds.

Material is contributed to the rivers, and carried onto the sandur, from five sources:

- transport of fine materials from upland moraine by runoff;
- nival erosion and solifluction processes on the terraces and along the valley sides;
- transport of material supplied by recent ice (in South Valley);
- material from scree and rockfalls (Middle and South Valleys);
- reworking of earlier alluvial deposits.

The last-named source is most important today, though the moraine- and scree-derived materials from South Valley also remain important. Most of the materials are gravelly or cobbly. Within the fines, the alluvial materials are more or less well sorted (Fig. 13), but very poor sorting occurs in the kame and moraine materials.

Periglacial processes, including solifluction and patterned ground formation, do not appear to be highly active today. Most of the materials are very sandy, and hence not highly susceptible to frost processes. Solifluction lobes are active on some slopes, mostly within the valley where there is considerable unconsolidated material and especially where marine silts and clays have been exposed. However, simple frost heaving is probably far more effective in moving material downslope. The ground is wet through most of the summer season, and usually freezes wet. Frost-heaved boulders are ubiquitous on ground moraine areas.

Remarkable frost cracks have developed on the old T2 surface, but to judge by lichen growth on the ground these are generally inactive today. Similarly mud boils and stone sorting have occurred on restricted areas on the south side of the valley on the same terrace.

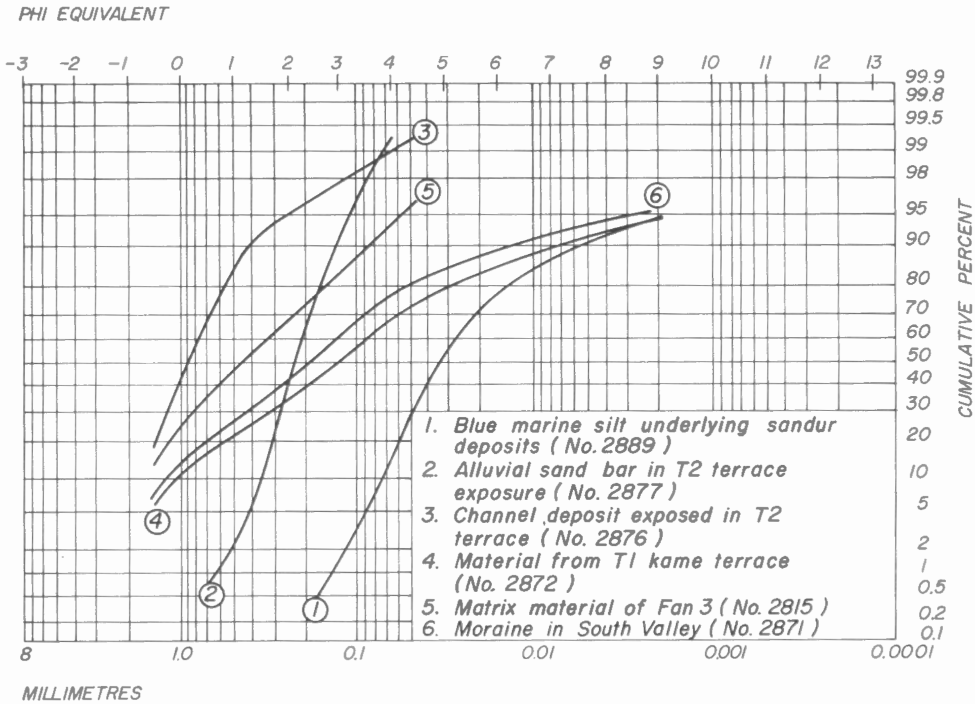


Figure 13. Sample materials from Ekalugad Valley (fraction finer than 2 mm).

The weathering activity of freeze-thaw cycles dominates the processes of both detritus production along the cliffs and attrition of material on the sandur.

The vegetative cover is influential in determining the relative stability of the various deposits in the valley. The oldest terraces are heavily vegetated with a Cassiope-Rhacomitrium heath. Large portions of the flat surfaces are marshy and colonized by various grasses and sedges. Most of the T2 surface is well drained gravel and is dry. Cassiope and Rhacomitrium are found in some frost cracks and old channel hollows. Silene acaulis is common on the gravelly surface and Hierochloe alpina populates the sandy areas. Mostly, the surface is a stable pebble pavement, however, bound together by the lichens Alectoria spp., Stereocaulon, Umbilicaria spp., and Cladonia. The lower T3 is more mesic than the T2 surface. Vegetation is not dissimilar to, but much poorer than, that of the old terraces. On recent surfaces there is blowing sand. Here, the pioneer colonizers appear to be Salix arctica and Salix herbacea, Papaver radicatum, Potentilla hypartica, and Hierochloe alpina. The active T4 surface is bare.

CHAPTER III

HYDROLOGY, AND ITS RELATIONSHIP WITH SUMMER WEATHER

Watershed Runoff

Total Runoff

Runoff measurements were made by normal stream-gauging techniques. The water budget of the Lewis River watershed for the seasons 1963-65 is given in Table 3 and the daily runoff hydrographs are presented in Figure 14. It should be noted that the data for 1964 and 1965 represent only partial budgets, since glaciological observations indicated that in both years significant ice ablation and runoff occurred after the end of the measurements. The reliability of the budget estimates has been assessed (Anonymous, 1967). The data may be used to derive the proportional contribution of each of the sources of runoff to the observed flow in the river (Table 4). The major variable factor in the table is the contribution of ice ablation. While 1963 was dominated by glacier melt, the stormy season of 1964 was affected mainly by snowmelt and seasonal precipitation. In 1965 ice ablation was more or less balanced by the contribution from annual precipitation (i. e., snowmelt plus seasonal precipitation). This factor shows up prominently in the runoff hydrographs, that for 1963 being much more regular under the influence of continued melt during warm weather throughout late July and August, than those of either 1964 or 1965.

The hydrographs of daily flows across Ekalugad sandur for the seasons 1967 and 1968 (South River only) are presented in Figure 15. Total runoff is given in Table 5. The sources of runoff are melt from the winter snowpack, direct runoff of summer precipitation, subtraction from watershed storage (in lakes and channels), and melt from the areas of permanent snow and ice. Since no direct information was gathered on the important first and last terms, no attempt has been made to construct a complete water budget.

The extremely low runoff in 1968 is evident. During the comparable period for 1967 the specific runoff from South River valley was 58.2 cm. The difference is ascribed to lower winter accumulation in 1967-68. Because of limited records, no total flow computations were carried out for either North or Middle Rivers in 1968.

Groundwater Flow

Coarse alluvial deposits normally admit considerable flow of water below the surface through the gravels themselves. In such a case there will be considerable exchange of water between channel flow and subsurface water. Where consistent abstraction of water occurs from channel flow to groundwater seepage, such losses will bear significantly on runoff estimates and can have a significant effect on sedimentation.

The extent of seepage can be estimated using Darcy's Law for flow in a porous medium. For horizontal flow we have (de Wiest, 1965)

(1) $$Q = (k/\mu)\, As$$

where k = the intrinsic permeability of the medium, measured in units of darcys, with dimension L^2;

μ = dynamic viscosity of the water;

A = cross-sectional area of flow;

s = hydraulic gradient.

Calculations based on field measurements have only ever been carried out once on a sandur (Hjulström, 1955), with the conclusion that groundwater discharge was only about 0.4 per cent of the total water discharge.

Because of the unfeasibility of accurately determining the permeability of the sediments at depth over the sandur as a whole, no field measurements were made to assess groundwater discharge on the Baffin sandurs. At Lewis River, the quantity is bound to be minimal at the gauging site, which is located at a moraine which presents a high frost barrier to discharge at any point except directly under the channel of the river.

Order of magnitude calculations may be made of the probable discharge at Ekalugad Fiord if an assumption can be made about the intrinsic permeability of the materials. From a table in de Wiest (1965, p. 171) we have the following data:

- for very fine sand (very well sorted): k = 9.9 darcys
- for medium sand (very well sorted): 2.6×10^2
- for coarse sand (very well sorted): 3.1×10^3

The matrix of sandur materials is only moderately well sorted, and consists of coarse to medium sands, but with some admixture of fine sand (cf., p. 124). For materials of mixed grain sizes the finer materials will control the intrinsic permeability since they will determine the frequency and magnitude of pore sizes. Hence, it may safely be concluded that intrinsic permeability lies in the range between about 10 darcys and 2.5×10^2 darcys over the sandur as a whole. This range of values of intrinsic permeability corresponds to a range of effective permeability between 1.6×10^{-1} cm sec^{-1} and 5.4×10^{-4} cm sec^{-1}, which corresponds well with the range of field values between 1.2×10^{-1} cm sec^{-1} and 9.0×10^{-3} cm sec^{-1} which was reported by Hjulström (1955) for the Hoffellssandur. The estimates appear, then, to be sufficiently good to justify calculations using equation (1) in lieu of field measurements. The further assumption was made that the depth to the frost table is one metre. This is a maximum that could be expected over the sandur as a whole and would only be reached fairly late in the season, so that the resulting discharge estimates are maximum values that could be expected. Also, the hydraulic gradient was reasonably assumed to be approximated by the surface slope of the sandur. The results of the calculations are given in Table 6. The best estimate of flow probably lies somewhere between tabulated values, though if open work gravels (old channels) dominate the groundwater flow system, the flow could be near the upper estimate. This indicates that for 1967 groundwater discharge was probably of the order of 0.4 per cent or less of the mean discharge of 39.1 m^3 sec^{-1} across the sandur.

Only occasionally were runnels found heading on the sandur, indicating that groundwater seepage was occurring to any extent. This tends to confirm the calculation that seepage is minimal, though on certain occasions it may be important. A high proportion of what recession flow there is (see p. 33) probably is derived from drainage of the sandur groundwater when cold weather sets in. Furthermore, during freezeback, the presence of the groundwater, which becomes trapped between permafrost below and the advancing seasonal frost above, leads to several special effects (Waller, 1966). The most spectacular is the growth of ice naleds (p. 101) in the channels, especially at the snout of Lewis Glacier, where a considerable hydrostatic head may develop. More significant are the production of highly saline residual waters from the freezing process (see p. 47) and the disturbance of channel beds caused by naled growth (see p. 103).

Characteristics of the Flow

Since runoff is dominated by snow and ice melt in all of the watersheds studied, hydrographs exhibit regular diurnal periodicity during normal weather periods as melting is influenced by the diurnal fluctuation of available heat. Since flow distances are not great, the periodicity is well marked, particularly in the proglacial Lewis River (see Fig. 21) and provides for a greatly exaggerated daily range of discharge and, concomitantly, of sediment transport rate.

The diurnal trend was investigated as a periodic function

$$Q_t = \bar{Q} + c_i \sin\left\{\frac{\theta_i t}{i}\right\} \tag{2}$$

where $\bar{Q}$ is the daily mean flow and θ_i are periodic phase angles. Hourly mean flow values were used as the basic data for analysis. The main diurnal harmonic, $c_1 \sin\theta_1 t$ accounts for more than 90 per cent of total diurnal variance in almost all cases, and the subsequent harmonics ($i > 1$) have the effect of adjusting the diurnal term to account for the skewed appearance of the usual daily hydrograph. Table 7 shows that after 4 harmonics have been removed, very little further variance remains, though the residual series were still oscillatory at approximately 5-hour intervals.

The phase angle of the diurnal flow cycle, θ_1, represents an approximation of peak flow time. It is graphed for each river in Figure 16. The most striking characteristic of the graph is the regularity of the Lewis River 1963 and 1965 series by comparison with the 1964 series and the Ekalugad series. These were seasons of relatively fine weather and extended periods of significant glacier melt, by comparison with the stormy seasons of 1964 and 1967. The regular recurrence of peak flow in late afternoon reflects this in the 1963 and 1965 series (though in 1965 the effect of several severe storms is clear). By contrast, frequent storms contributed the bulk of summer runoff and severely disturbed the patterns of the other seasons.

Interesting small scale variations in flow occur within the diurnal period. This was particularly apparent at Lewis River, as is seen on the example gauge level trace in Figure 17. The irregular perturbations that occur in the trace are apparently the result of frequent ice fall and slush avalanching on the glacier, that periodically block the channels and then release surges of water after a short period. Sometimes large amounts of ice or slush were carried down the river at these times (Fig. 18); the large volume of ice in the flow effectively damped out most of the river's turbulence and thereby exerted an obvious influence on suspended sediment transport in the stream. The water carried a much reduced load of sediment on such occasions. Such small-scale perturbations were not common on the Ekalugad rivers, whose watersheds are less heavily glacierized, and where the gauges were much farther from the upland ice.

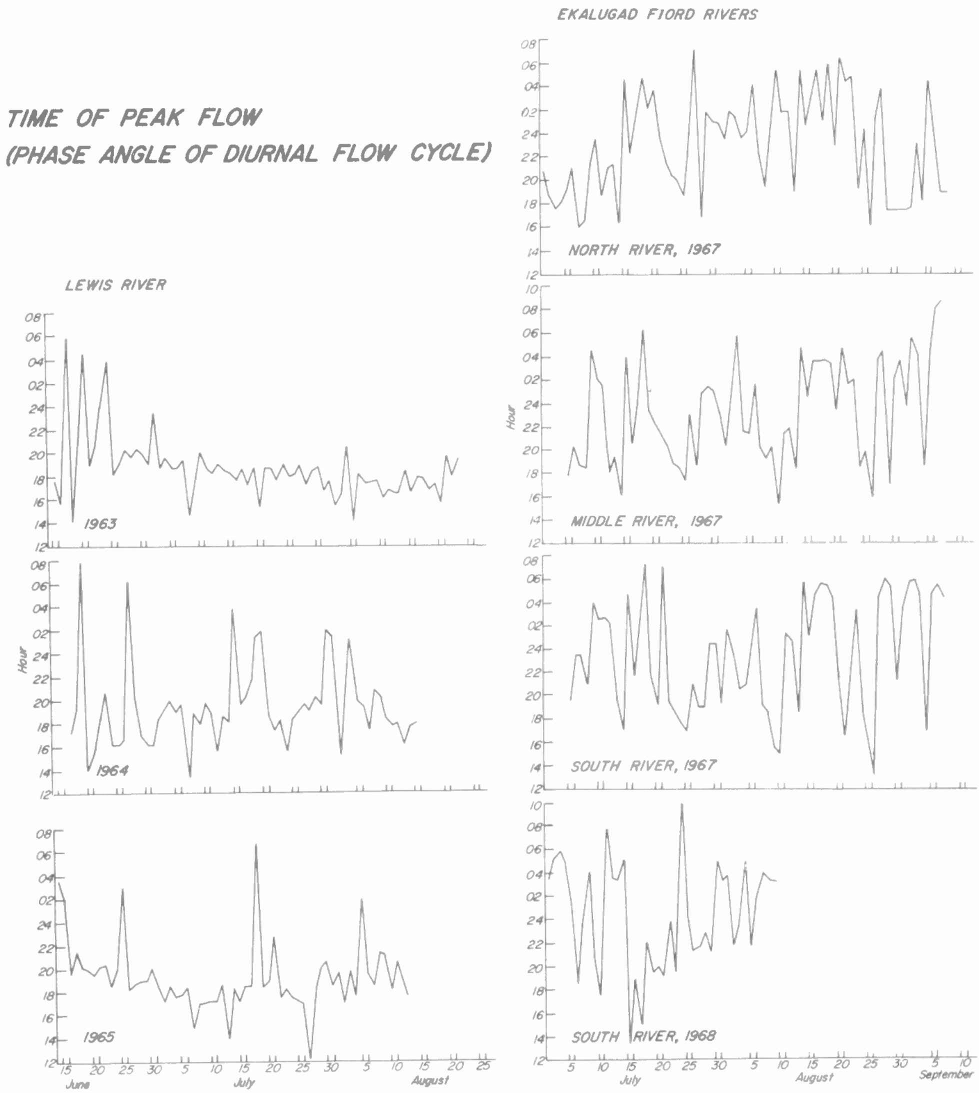


Figure 16. Time of peak flow (phase angle of diurnal flow cycle) for Baffin rivers.

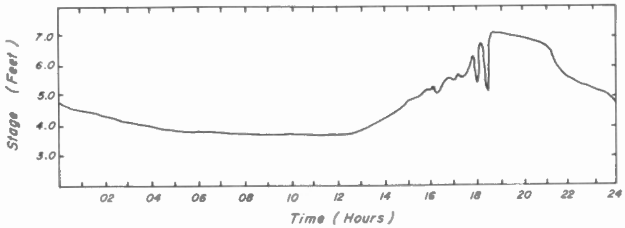


Figure 17.

Lewis River automatic gauge trace: 26 July 1964, showing minor surges and flow interruptions due to snow and ice falls and slush avalanching.

Seasonal Trend of Runoff

On a seasonal basis, the runoff exhibits a relatively regular pattern in response to summer melt processes just as it does on a diurnal basis. In a fair weather season, the peak of runoff at Lewis Glacier would be expected in late July or early August, when the warmest weather occurs, with meltwater runoff continuing at a fairly high rate throughout August (cf. p. 32). At Ekalugad Fiord a sharp peak of nival runoff would be expected in early July (see Pardé, 1933, on the contrast between glacial and nival runoff regimes). In order to investigate the seasonal structure of runoff, serial correlation was performed on the daily runoff totals for each river, before and after removal of a seven-day running mean which was intended to represent the seasonal effect. The one-step results are shown in Table 8. The 1963 Lewis River data exhibit a high degree of seasonal structure: a high proportion of the total variance is removed by the trend, and there is little remaining structure in the residual series (as indicated by the low serial correlation of residual series). By contrast, the other series generally show relatively less seasonal structure and a high degree of residual correlation, indicating strong structural influence of within-season events during these relatively stormy seasons. This is reported in the section entitled "Normal Weather Control" in Chapter III. (Though 1965 was a relatively high ablation year at Lewis Glacier several severe storms occurred that probably have great influence in the present analysis.) The South River 1967 series is anomalous: it exhibits no structure of any sort. This is possibly largely due to the effect of an ice dam burst in midseason (see, p. 38) which was preceded by runoff from only a small portion of the watershed. The series is not homogeneous, in that it does not represent runoff from a single, relatively stable watershed area.

Weak oscillatory effects at 6-day or shorter periods in the residual correlograms (Fig. 19) suggest that synoptic weather patterns are significant.

Runoff Periods and Summer Weather

There are four distinct periods during the arctic runoff season: breakup, the snowmelt period (the "nival" flood), late summer, and freezeback. Pissart (1967) divided the breakup period and snowmelt period into two phases each. In terms of total runoff, the end periods are more important for their time of occurrence than for actual runoff. During the summer season, more or less an entire year's precipitation runs off, so that for much of the season, streamflow bears little relationship to the precipitation inputs. In this respect arctic watersheds share many of the characteristics of high

Figure 18. Small jökulhlaup on Lewis River, carrying a heavy load of slush and ice: view upstream from the gauging section ($Q \simeq 20\ m^3 s^{-1}$).

alpine watersheds (Meier and Tangborn, 1961). In detail, the progress of the runoff periods is determined by the pattern of summer weather.

Summary of Summer Weather

Standard meteorological observations were maintained throughout the period of field investigations during each year of record. A summary of the record is presented in Figures 14 and 15, and mean climatic values are given in Table 9. Comparative aspects of the summer climate at the east Baffin stations of Cape Hooper, Clyde, and Dewar Lakes are also listed and it is seen that both the Lewis River and Ekalugad Fiord sites are consistently warmer and moister than any of the three regional stations. In both cases this is probably due to the valley location. This allows more effective warming during sunny periods, and often produces relatively high overnight temperatures with the appearance of a thin layer of cloud in the valley. Two of the regular stations are on the outer coast, which is an area of cool local climate in any case. Precipitation is increased at the field stations by local

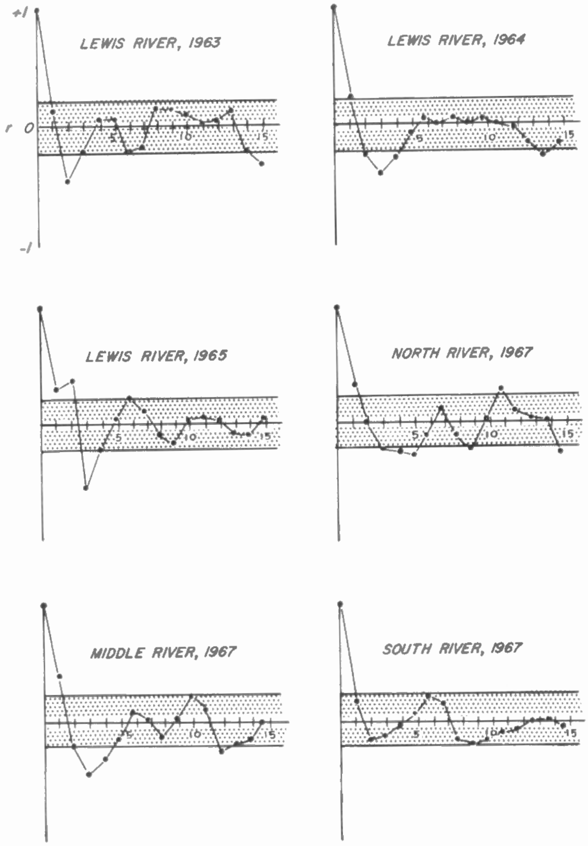


Figure 19.

Correlograms of the daily runoff (seasonal trend removed) for Baffin rivers. The shaded zone represents the region of nonsignificant correlations.

orographic effects: in the case of Lewis Valley, a quite disproportionate effect in terms of the actual elevation change is produced when air moving eastward along Isortoq Valley is forced to rise over the edge of the cold ice cap, so that often local showers persist here.

Exceedance probabilities (derived from Fig. 5) have been added to the Dewar Lakes summary results in Table 9B for complete months of record. Though not necessarily applicable at the field stations, they do confirm that the weather was generally cool and wet (with a few notable exceptions, as June 1965 and June 1968).

The summer weather patterns in Baffin Island may be conveniently divided into several dominant synoptic situations. At Lewis River four types emerged, as follows:

1. organized storms: cyclonic storms moving mainly from the southwest, but occasionally from the west or southeast, bringing heavy overcast and steady rain, sometimes with high winds;
2. overcast and drizzle: weak circulation periods, often associated with a persistent trough over central Baffin Island, produce a westerly drift of air bringing low cloud and showers from the west: often this issues in drizzle and fog around the margins of the Barnes Ice Cap;

3. partially overcast and showers: weak circulation condition, but with high overcast or broken skies, sometimes with limited convectional development: usually a transitional weather type, though it may be persistent (as in late June, 1965);
4. fair weather: high pressure ridges or cells produce clear weather, and either cool or warm conditions: at Lewis River local katabatic winds often occur to produce local cooling during such periods.

Similar weather patterns were noted also in Ekalugad Fiord, except that the types (2) and (3) of Lewis River were more or less indistinguishable as local cloud and drizzle in the fiords, often below an inversion, with easterly drift of air. Katabatic winds did not occur, either. Conditions are, in general, much wetter in the fiords than on the plateau, and precipitation is more common in all weather conditions. Each weather type produces a distincitve runoff pattern, superimposed on the seasonal progression of thermal conditions.

Breakup

While melt begins in early June, it is not until middle or late June that sufficient drainage has occurred from the snowpack to begin to find its way into channel flow. The normal course of breakup is for the river channel to turn to slush (Pissart's "standing water phase"), and then for flow to begin over the ice. It is only after flow is well established that the winter ice on the riverbed breaks up. Low flows may be supported by melt along the river banks for a period of days or weeks before the spring flood begins. During this period the length of connected channel is short and restricted to the valley bottom. At higher levels most of the melt is still absorbed into the ripening snowpack. An extended fair weather period or, more efficiently, a heavy rainstorm is necessary to flush an appreciable length of channel and initiate significant runoff. In Baffin rivers it was found that the first significant day or two of runoff were sufficient to break up bottom ice in that part of the channel in which flow occurred.

In 1963 at Lewis River, rain between 7 and 11 July cleared out a considerable length of the channelways on the glacier, so that afterward significant drainage commenced from the glacier. In 1964 and again in 1965 fine weather in early July produced extensive slush flows that cleared the channels. In 1967 at Ekalugad Fiord the fine weather beginning on June 29 had the effect of producing significant melt that began to clear the river channels from July 2nd. In 1968, no significant runoff occurred before the storm of July 11-12.

Detailed observations were made of water characteristics and of air temperatures on some days at Lewis River. Figure 21 shows too such days during the breakup period. The low range of water stage changes, high range of water temperatures and high water conductivities are all representative of this period. There is no significant clastic bedload sediment transport during this low flow period, although high concentrations of dissolved and very fine suspended material ("glacier flour") may occur.

Summer Flow

The majority of snowmelt occurs in late June and early July, except at the highest elevations, and most of the meltwater runs off in early July.

MacKay (1966) indicated July as the month of maximum runoff in arctic watersheds, as had been observed by St-Onge (1965) and Cook (1967). This is certainly true for nival watersheds, but the time of peak runoff in extensively glacierized watersheds may be delayed until later. The spring flood is well illustrated in the records for North and Middle Rivers at Ekalugad Fiord in 1967 (Fig. 15). Fine weather during the first two weeks of July produced a rapid runoff and, hence, a relatively high flood. At Lewis River, the similar effect is masked by the continuing influence of glacier ice melting.

During the later summer, runoff decreases in the Ekalugad rivers, and comes to be dominated by storm runoff. The warmest period in late July, 1967, did not produce peak flows as great as those in early July. By contrast, runoff continues unabated in the Lewis River, and in a reasonably warm season, such as 1963, the peak flow is reached in early August. This reflects several developments:

- the early season heat deficit in the near-surface zone of the glacier has been overcome;
- the albedo of the surface has been lowered by the melt of seasonal snow and superimposed ice so that the darker glacier ice is exposed, and melt carries on more efficiently for a given radiation heat input;
- the drainage network has extended over the entire glacier watershed.

During melt periods, either of snow or glacier ice, the daily runoff closely reflects heating inputs. Two such days are shown in Figure 21 from Lewis River. July 27, 1963, was a clear day, with high radiative heating producing a high runoff. By comparison, August 9, 1963, was cloudy and dull, and much cooler. Nevertheless the runoff, in late season, was higher than the runoff on the July day, partly because of the high overnight flow that was maintained.

By contrast, during stormy weather little diurnal periodicity occurs. Extended periods of cooler weather also affect total runoff, as is indicated by the greatly reduced flow during the cool period of August 1-8, 1964 at Lewis River, or that of July 15-22, 1967 on the Ekalugad rivers (see Figs. 14 and 15).

The bulk of sediment movement occurs during summer high flow events (Fig. 20). The most extreme daily ranges in flow, which condition the most radical changes from erosional to depositional conditions in the stream channels, occur at Lewis River during fair weather (type 4) conditions, when there is strong diurnal control on glacier ice melt. By contrast, damp, stormy weather usually exhibits relatively equable runoff conditions. At such times a large proportion of the watershed may be noncontributing because of freezing conditions, or meltwater runoff may continue at relatively high levels throughout the 24-hour period due to the persistence of advective thermal conditions (Fig. 14). At Ekalugad Fiord, the reverse is the case, with storm events, especially late in the season, providing the most extreme fluctuations in discharge in North and Middle Rivers (Fig. 15). This is typical of arctic rivers without permanent snow and ice fields in their watersheds. South River, which does have a significant proportion of permanent snowfield, presents no clear pattern.

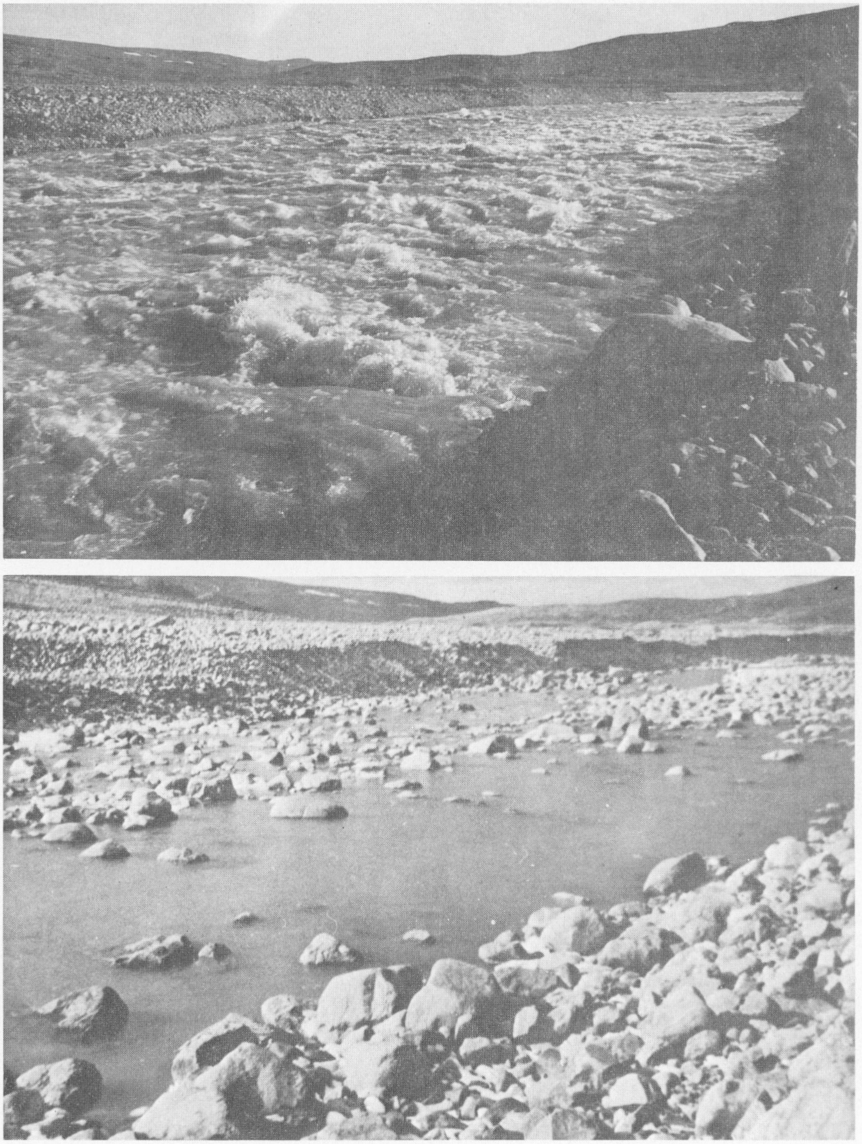


Figure 20. Lewis River at high and low stage (about 2 $m^3 s^{-1}$ and about 250 $m^3 s^{-1}$): view downstream from the gauging section.

Freezeback

During the freezeback period at the end of the season there is little new melt for runoff. Streamflow is maintained almost entirely by recession flow from groundwater discharge. Because of the restricted active layer zone that is contributary, flow very quickly declines to a low level. Smaller streams become dry before freezeup. At Lewis River, where most of the watershed is glacial, residual flow from lateral pondage and some subglacial passageways near the glacier snout provide some flow, but the decline in flow is remarkably rapid once melt ceases (Fig. 20).

We may tentatively conclude that, whereas glacier melt produces the most extreme ranges of runoff and "flashiest" runoff patterns in the highly glacierized Lewis watershed, the more normal nival watersheds at Ekalugad Fiord produce extreme ranges under the influence of storm runoff in types (1) and (2) weather. The runoff regimen of the Lewis River will be termed "proglacial", while that of the Ekalugad rivers (except South River) is "nival". South River represents a complex runoff type affected by both influences.

Magnitude and Frequency of Flow Events

Sources of Flow

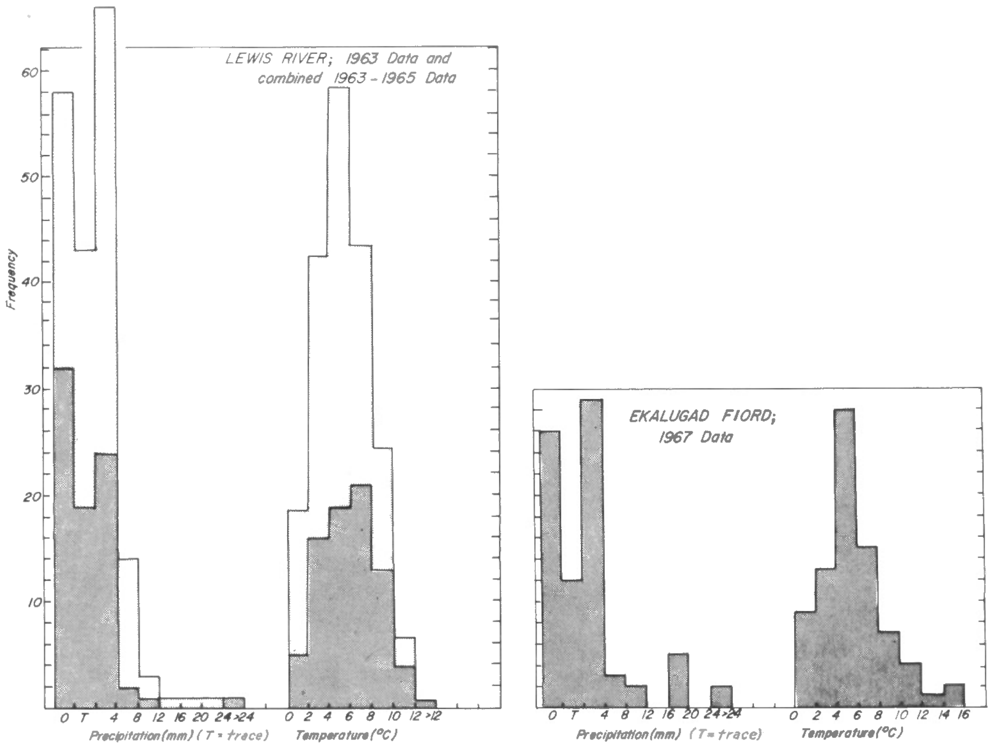
The physical process of runoff formation in an arctic watershed possesses several peculiarities that are to a greater or lesser extent connected with thermal conditions. The chief contributions to runoff are direct precipitation and direct runoff from seasonal snowmelt. The question of contributions to runoff via groundwater flow is complicated by frost phenomena. The active layer is restricted by the frost table, and so little deep percolation occurs. Over large areas of rocky terrain, such as most of Baffin Island, the water retaining capacity of the near-surface, seasonally active layer will be very low, and exchange with deep groundwater minimal. Seasonal net exchange of water with permafrost storage will be small. Evaporation is a minimal factor in the hydrology of the eastern Arctic because of the low ambient temperatures and relatively high humidities. It appears, then, that runoff is overwhelmingly a surface phenomenon, and hence it can be monitored adequately by techniques for surface runoff measurement.

Considering runoff on a watershed scale, seasonal precipitation is simply additive to water derived from melt processes, though estimating the runoff from a particular storm is complicated by the varying effectiveness of the snowpack in retaining precipitation water.

Normal Weather Control

Flow exceedances are rather difficult to relate to the controlling factors, since the distributions of the contributory components must be known, and they can only be combined tractably if they are all independent. In the present case, the contributing event distributions pertain to precipitation, temperature, and antecedent runoff. The last factor expresses the historical component in the determination of flow level.

The frequency distributions of weather elements (Fig. 22) indicate that, whereas precipitation is highly skewed, temperature is only moderately so. The daily duration series for each variate is plotted in Figures 23 and 24. (It is important to note that these plots do not constitute frequency curves of themselves; the degree of serial correlation present in the data for each series, except precipitation, precludes the assignment of unconditional probabilities to the indicated recurrences. The plots, do how-



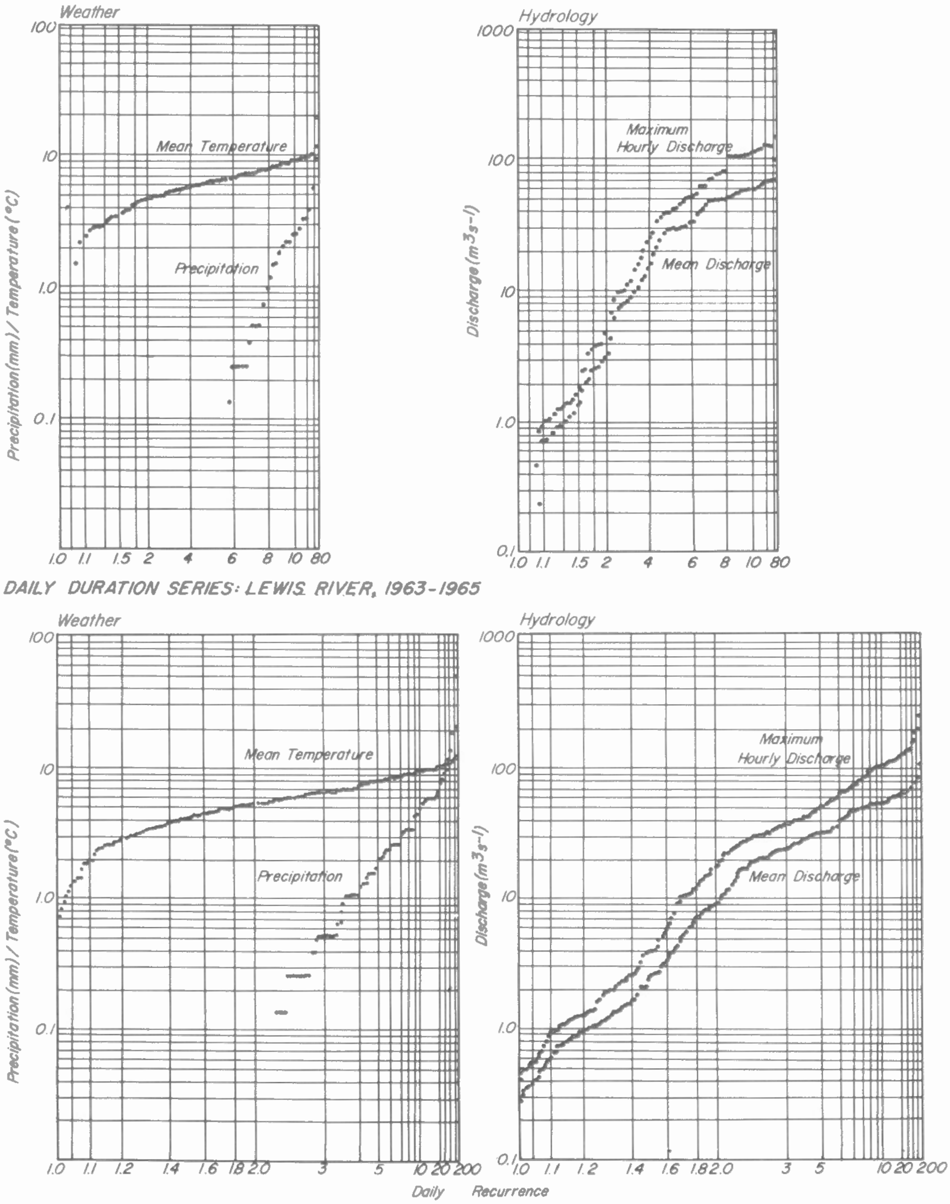
A. Lewis River: 1963 data and combined 1963-1965 data
B. Ekalugad Fiord: 1967 data

Figure 22. Frequency distributions of temperature and precipitation data during the runoff season at Lewis River and Ekalugad Fiord.

ever indicate the form of seasonal distribution of the events controlling runoff under "normal" weather control.)

The problem of covariance between runoff and antecedent temperature terms make it relatively pointless to pursue the discussion analytically. Furthermore, (p. 63) it becomes apparent that the transformation of flow frequency to the frequencies of geomorphologically significant sediment movement events comes under such rigorous constraints as could not be met by the approximations possible with the available data. Suffice it to note that the resultant runoff exhibits "piecewise" regular behaviour, the inflection points probably corresponding to changes in the relative dominance of the various controlling factors.

In the matter of exceedances it is, however, extreme values which are of primary concern. As a sequence of extremes they may be distributed in entirely different fashion from "normal" events, since the controlling factors are quite probably different (see p. 38). There are not nearly enough available data to allow any analytical comment. It is pertinent to point out, however, that notable irregularities appear in the tails of the duration curves. In the lower end of the distribution of runoff, it is probable that a second distribution is present. Most events here belong to the first or



A. Weather, 1963
B. Hydrology, 1963
C. Weather, 1963-1965
D. Hydrology, 1963-1965

Figure 23. Daily duration series: Lewis River.

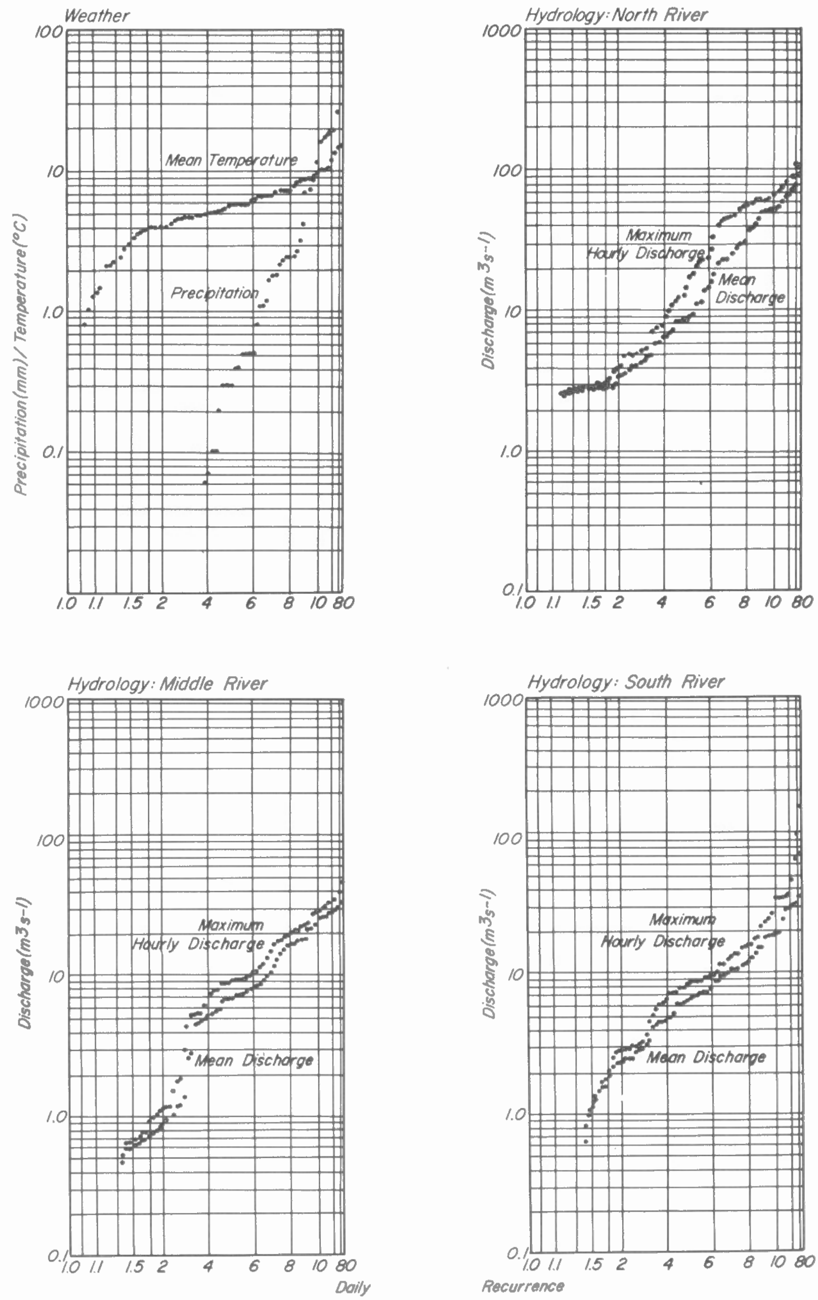


Figure 24. Daily duration series: Ekalugad Fiord, 1967.

last few days of the season, when runoff is very restricted, from the lowest reaches of the watershed only, and when frequent freezing interferes. Hence, they do not represent runoff under the same environmental conditions as the other events. Likes (1966) has similarly pointed out that the flood frequency series of a small river in Alaska is not homogeneous due to the changing source and controls of runoff through the season.

Extreme Events

At the upper end of the distribution one or two very extreme events appear (cf. especially, South River, 1967). These events do not occur under normal weather control and are not part of the "normal" runoff sequence. In the case of the South River flood, an ice dam burst, causing catastrophic lake drainage. The form of the runoff hydrograph during the jökulhlaup (Fig. 25) was typical of such drainage events: rising steadily, then irregularly, to a peak flow, and then falling abruptly. Various workers have speculated that this is due to the gradual enlargement of the ice-walled drainage ways (Arnborg, 1955) using heat energy generated by the flow (Gilbert, 1971; Mathews, in press) so that the capacity of the channel is greatest when the water at the source becomes exhausted. Altogether, between 1000 hours on 20 July and 0800 hours on 22 July, 5,936,400 m^3 water passed the South River gauge, of which about 4,800,000 m^3 during 30 hours can definitely be ascribed to the lake drainage. The peak discharge of 200 m^3s^{-1} flooded a large portion of the lower sandur. Jökulhlaupe were frequent at Lewis River, though no outstanding one occurred in 1963-1965. Drainage of ponds and cavities near the glacier snout was responsible for small events. The peak runoff observed there was caused by a combination of very heavy rain and rapid melting occurring together on July 21-24, 1965; this was not an abnormal event, but appears so in the short term observations. An ice-dammed lake continues to exist in the Lewis River watershed.

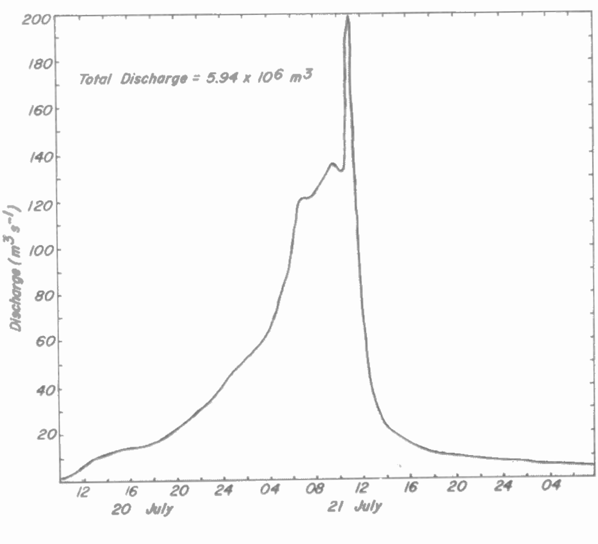


Figure 25.

The South River flood of July 20-22, 1967, caused by the draining of a small lake trapped behind the ice-fall in South Valley.

CHAPTER IV

MOVEMENT OF SEDIMENT

Sources of Sediment

Periglacial Processes

Primary production of material for fluvial transport today is largely the work of frost-shatter erosion of material on the cliffs and within older deposits. The primacy of frost wedging processes in rock weathering in the Arctic has been asserted by many workers, most recently by Washburn (1965, 1969).

Freezing and thawing relies for its geomorphological effectiveness on the presence of water; rock-bursting due to the freezing and thawing of water in the rock crevices is the dominant mode of weathering. Attrition of unconsolidated materials is carried on in a similar manner, and occurs prominently in the active layer. Parting of rocks on foliation planes (microgélivation, so-called by Tricart, 1953) and flaking of mineral grains from the surface of the rock are the usual results. The ground usually freezes damp in September, adding to the efficacy of the phenomenon. Along the valleyside cliffs, icing is a common occurrence, with large volumes of ice often building up. Most of the shattered material is released by the main spring thaw in early June.

The materials produced in east Baffin Island are overwhelmingly of sand size or larger. The general paucity of very fine materials produced by freeze-thaw processes would appear to be a function of the relatively coarse-grained rocks (cf. Hopkins and Sigafoos, 1957). Wiman (1963) reported grain size distributions of the weathered products of experimental freeze-thaw processes on several rock types, amongst which a gneiss and mica-schist probably roughly represent the extremes of the rock types found in Baffin Island (his gneiss was probably considerably more rotten than most Baffin gneisses). His results confirm the paucity of very fine material that can be expected: generally much less than 10 per cent of the weathered material was finer than the lower sand limit. Feldspar weathering is not generally sufficiently advanced to yield easily disaggregated clay minerals (see p. 44).

Delivery to the streams is accomplished through three mechanisms: direct fall in rockfall and avalanche events, solifluction, and nival erosion.

Rockfall and avalanche events are common on the scree slopes and debris slopes along the valleysides at Ekalugad Fiord (Figs. 26 and 27). Stock (1968) has carried out a study of the processes that occur on the scree slopes at Ekalugad Fiord, which indicated that the majority of significant activity occurs in the spring (June and early July) during the snowmelt period. Nevertheless, a significant level of rockfall activity apparently continues through summer and into autumn due to summer rains and delayed seasonal thaw in sheltered places. Delivery of material from the scree slopes to the

Figure 26. Gullying in bedrock and detrital cover on the south side of South Valley below the upland snowfield: a contemporary source of fluvially transported materials.

Figure 27. Debris slopes on the south side of Ekalugad Valley: some material rolls onto the sandur surface from talus and debris slopes. Avalanching occurs in the gullys below the upland snowfield.

streams occurs directly only in Middle valley, where talus slopes are built into the river. Elsewhere fine materials may be washed down directly into the rivers by snow meltwater or by rainwater.

Solifluction is not highly active at either Lewis River or Ekalugad Fiord but does occur on some valley side slopes or terraces. Such activity is promoted by over-saturation of the active layer during the melt period, and so belongs to the "flowage" rather than "frost-heave" type of solifluction. Solifluction is conspicuously confined to places where there is a considerable admixture of silts in the predominantly sandy regolith.

"Nival erosion" indicates processes of material entrainment and transport that are carried on under the snowpack, and under and around snow banks during snowmelt. The entrainment of particulate matter in such situations accounts for a moderate proportion of the suspended material moved by the streams. Chemical dissolution and removal of materials probably occurs mainly in this environment. The chemical purity of meltwaters (see Fig. 30), combined with the high carbon dioxide content of the snowpack atmosphere (Williams, 1949) make this a very corrosive environment. The question of chemical weathering and solution transport in arctic environments is discussed on page 44.

Nival hollows, some carrying permanent snowbanks, are particularly common on the hills around Lewis River. The margins of the upland snowfields at Ekalugad Fiord present a similar environment.

Glacial Processes

The bulk of unconsolidated material available at the present day is derived from glacial activity during Pleistocene times. It appears, furthermore, that each new glacial advance derives only a modest amount of completely new clastic material from bedrock and that, for the most part, the ice moves about a great deal of previously deposited material. Nevertheless, by the close of the last major glaciation, a large amount of detrital material had accumulated.

It is difficult to assess the sediment "yield" of active glaciers on any scale to permit comparison with fluvial transport of material away from the ice front. Various investigators (Rainwater and Guy, 1961; Østrem, Bridge, and Rannie, 1967) have implied that measurements of sediment transport in proglacial streams should, at least over a long period, give a fair representation of the erosive capacity of a glacier. However, most material is deposited immediately at the edge of the glacier and though fines may be washed out fairly promptly, coarse material may be reworked or removed only over many decades in the course of events which have little or nothing to do with the concurrent erosional or sediment transporting activity of the glacier.

In summary, whereas glacial activity ultimately provided the bulk of clastic materials in the valleys, and may directly provide a portion of fine materials in transport in the rivers, relatively little material is directly contributed from glaciers. Contemporary rates of sediment yield from glaciers would be difficult to establish and this was not attempted in this study.

Fluvial Erosion of Older Glacial Deposits

The major process of sediment uptake into the rivers is not related to primary rock weathering or to glacial activity, but is the work of direct fluvial

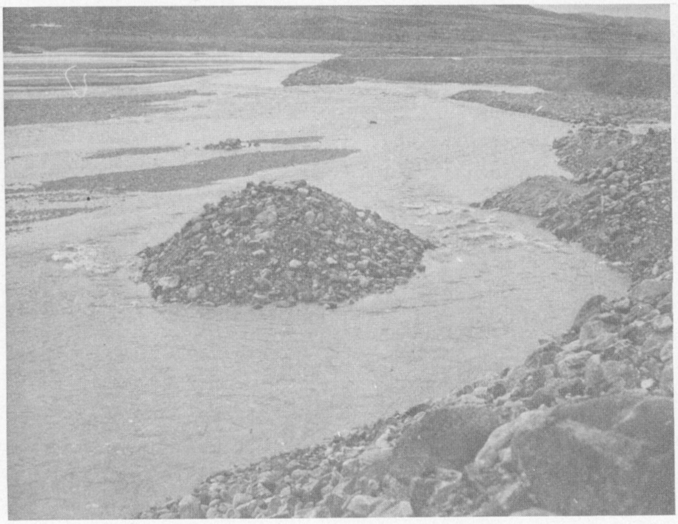


Figure 28.

Erosion of moraine material on Lewis sandur: lag detritus from previously eroded moraine is visible in the background.

activity. Most glacial material is ultimately picked up by stream attack on older deposits that may be some distance from the present glacier. At Lewis River, this is the major means of sediment entrainment (Fig. 28). Along the side of the Lewis Glacier, kame terrace and morainic debris is quickly carried away as the river erodes lower and lower on the hillside, while in proglacial position, the river actively erodes the old moraines where it flows along or through them. At Lewis River, as much as 90 per cent of the transported material is picked from such "proglacial debris".

Channel Erosion and Reworking of Older Fluvial Deposits

Material derived from reworking of older fluvial deposits contributes an important part of the total sediment carried by the rivers and, at Ekalugad Fiord, comprises the bulk of sediment supply. Here, the erosion by the rivers of the T2 terrace edges provides material. In order to determine the volume supplied from this source, a series of 117 stakes was placed along the terrace edges (Fig. 2) and the distance from the stake to the edge measured on several occasions to determine scarp retreat in the intervening period. The measurements are summarized on Figure 29, and in Table 10 as weight of material removed. A surprisingly large amount of material slumped off the scarps during the winter, probably a result of frost activity, and was removed during the following summer. Most material slumps incoherently as the base of the slope is undercut, but in some places block slumping occurs where a deep frost wedge has weakened material some distance from the edge. At some places along North and South Rivers solifluction material moves directly over the scarp edge and slumps into the channel.

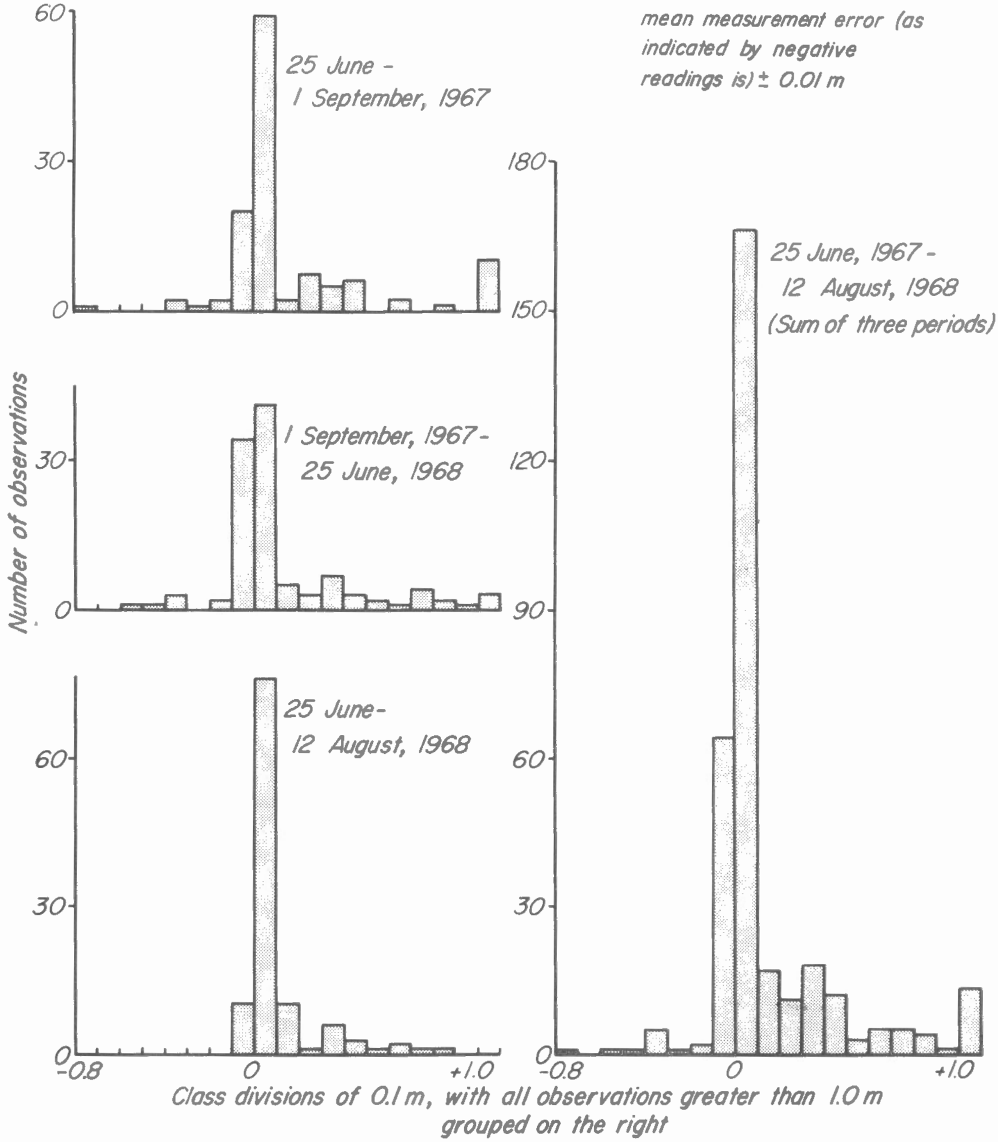


Figure 29. Distribution of erosion stake measurements along the T2 terrace, Ekalugad Fiord.

Solution Transport

The Importance of Chemical Weathering in the Arctic

Most investigators have tended to minimize the importance of chemical activity in the Arctic. Beyond the obvious dominance of frost-shatter mechanical weathering, two factors contribute to this impression:

- the extremely low temperatures, which inhibit many reactions;
- the fact that most water during the summer season runs off the land relatively quickly, and does not spend a great deal of time in the ground environment, so that it picks up very little dissolved material.

MacDougall and Harriss (1969) reported a "lack of large-scale chemical weathering" in sediments from northern Melville Island.

Nevertheless, certain processes do occur even in granite terranes. Tarr (1897) noted that iron-staining was common in Baffin Island and West Greenland. Hill and Tedrow (1961) emphasized the acid character of soil environments, which leads to translocation of iron, manganese and aluminum, iron oxidation, and solution of carbonates. Dahl (1966) and Roberts (1968) have recently pointed out the occurrence of "weathering pits" in granites and gneisses on the arctic uplands of northern Norway which they tentatively ascribe to a combination of microgelivation and chemical weathering. Williams (1949) pointed out that snowpack atmospheres are usually enriched in carbon dioxide so that snow meltwaters are highly corrosive (he indicated that this was of major importance for nivation hollow development). Tyutyunov (1964) indicated that certain chemical processes continue at temperatures below freezing. Washburn (1969) concluded that chemical weathering is locally important in eastern Greenland, citing oxidation, solution and deposition of calcium carbonate, chemically aided exfoliation, cavernous weathering and granular disintegration, and the occurrence of case hardening and desert varnish.

In the granite-gneiss of central Baffin Island, the visibly outstanding chemical activity is iron oxidation. Iron stain, almost certainly developed since the last glaciation, is common on rock outcrops, and visible weathering rinds of iron oxides occur in many exposed rocks. In extreme cases a violet or black casing may develop (also reported by Dahl, 1966). Magnetite, augite and biotite are all susceptible to iron oxidation. Clay mineral formation does occur in feldspars but it is severely limited. Hydration is active to a limited extent in schistose rocks and, by contrast, isolated instances of effluorescence have been observed, particularly in well-drained terrace gravels. Silica dissolution also occurs noticeably on Baffin Island (see next sub-section).

Characteristics of Sampled Waters

Samples were regularly taken at Lewis River during 1964 and at the three Ekalugad rivers in 1967. Sampling was done at 4-day intervals, and the samples were normally taken in late afternoon near daily high stage. Approximately one litre of water was dipped from the river each time and stored in an airtight polyethylene container for shipment to the laboratory. The number of samples was 14 for Lewis River, and between 9 and 11 for each of the Ekalugad rivers. An anomalous result, obtained from runoff

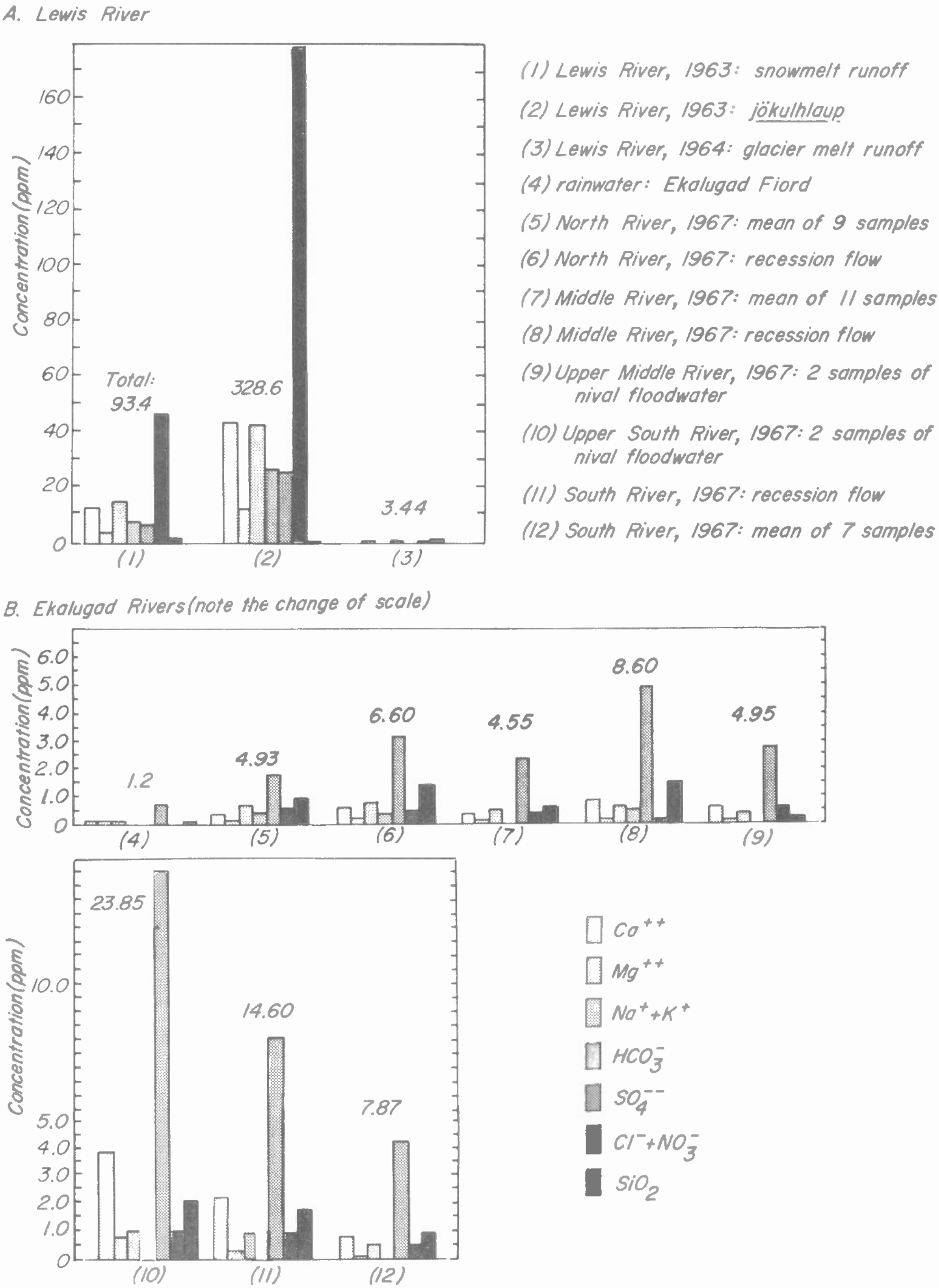


Figure 30. Concentration of dissolved materials in stream runoff, Baffin Island rivers.

during a very heavy storm on July 14, 1967, was removed from the North River (Ekalugad Fiord) records for some analyses (see below).

On the whole, waters were very soft, though exceptions were noted, as at the Lewis River during a small jökulhlaup (see Fig. 30A). Waters draining igneous or metamorphic terranes are normally soft, the value of 50 ppm total dissolved solids having been quoted as a representative upper limit (Gorham, 1961). In nival basins with rapid runoff of meltwaters one would expect far lower values.

Waters at Ekalugad Fiord are softer than those at Lewis River. The higher values of solution load evidenced at Lewis River appear to be the result of ion exclusion during freezing processes on the glacier and in the stream channels during the winter, and the "flushing out" of disproportionate amounts of winter weathering products in early spring. The relatively high proportion of "glacier milk" carried in the Lewis River at this time is probably associated with a good deal of free ionic material.

Table 11 indicates the proportions of major ions in the Baffin streams, along with comparative results from other sources. The waters are not unlike those sampled elsewhere in the Arctic (Thomas, 1964), in that alkali-chloride and calcium-sulphate types are prominent. However, the comparative absence of bicarbonate is notable.

Amongst the cations the proportions are relatively stable at each field area. Lewis River is somewhat magnesium-poor and potassium-poor with respect to Ekalugad Fiord. Probably most of the potassium, and much of the magnesium are derived from precipitation: thus the relatively greater importance of current precipitation in the hydrology of the Ekalugad rivers probably accounts for the difference. Sodium is probably the chief ion supplied from the country rocks: sodium and calcium are probably delivered from incipient weathering of plagioclase feldspars (Reiche, 1950; Dennen and Anderson, 1962).

By comparison with glacier meltwater in granite-gneiss terranes elsewhere in the world, the relative sodium-richness of Baffin waters stands out, as it does relative to the average composition of igneous rocks. This probably reflects the very elementary stage reached by chemical weathering processes on Baffin Island: the removal of the most highly mobile soluble ions – such as Na^+ and Ca^{++} – is virtually all that is achieved.

The anions present a remarkable picture. Whereas at Ekalugad Fiord the major anions in the waters almost perfectly reflect the composition of rainfall, at Lewis River a radical change appears to occur, with HCO_3^- and especially Cl^- appearing in considerably quantity. The latter ion was the major single constituent of the jökulhlaup water of July 5, 1963, and it varies between 10 and 50 per cent of the total dissolved ions even of glacial meltwaters. For this reason, the source of the chloride appears to be in the glacier, and ultimately atmospheric. No rainwater samples were taken at Lewis River, but one from Ekalugad Fiord showed far lower chloride content than found in Lewis River waters. Two hypotheses for the source of the chloride arise:

1. that the rainwater at Lewis River contains relatively high chloride content – implying a fairly local source;
2. that chloride enrichment occurs during melting and refreezing processes on the glacier.

For the first hypothesis it may be pointed out that local drizzle is common at Lewis River brought by low cloud moving off Foxe Basin about 80 km to the west. The relative abundance of Na^+ ions could be related to an accompanying transfer of sea salt. This sea is largely ice choked for much of the season, however, and is not likely an unusual source. On the other hand, the remarkably high chloride content of the jökulhlaup waters argues for the "enrichment" hypothesis. This water had subsisted under the glacier, probably throughout the winter, and had become steadily more saline as water was "frozen out of solution". The proportion of chlorides in snowmelt waters is also remarkably high, possibly for a similar reason. However, by this hypothesis, some of the other ions must have been precipitated from solution in order to alter the balance so much. Possibly the very low temperatures would effect this.

The bicarbonate at Lewis River was probably derived either by atmospheric dust transport from the limestone lowlands of Foxe Basin or from tills with that provenance. Limestone erratics have been found scattered across central Baffin Island east of the Foxe Basin. Some of the Ekalugad sulphate no doubt derives from weathering in the schists: sulphide stains are common on the cliffs below schist outcroppings.

Of passing interest is the role of silica solution, which is normally important in the overall reduction of granite-gneiss terranes. Concentration of SiO_2 in sampled waters was consistently of order 0.1-1.0 ppm. This amounts to 10 to 20 per cent of total solution transport, or 15 to 25 per cent of the material actually dissolved from the land.

The concentration of dissolved solids is significantly correlated with discharge of the streams (Table 12). The correlations are all inverse, indicating that supply of dissolved materials is, on the whole, more conservative than the water supply. At Lewis River, variance reduction is greater than 90 per cent in correlation with discharge, but on the Ekalugad rivers the results are much poorer, becoming increasingly poor as the extent of glacierization becomes less. Since the main source of supply of dissolved materials is the ground environment, it appears that the variability in ion supply is relatively reduced as the proportional dilution of ground environment water with glacial meltwaters becomes greater and hence overall correlations are much improved in more highly glacierized basins.

In light of the above some variables were sought which might explain a portion of the variability in supply of dissolved materials from the ground environment. At Lewis River, estimates of groundwater discharge are available which were entered into the analysis. It was found that the variance explanation was raised to 97 per cent (see Table 12).

At Ekalugad Fiord, no such data were available. The period of time since the last rain was entered into the analysis. While the result was significant in all cases except Lewis River, only at North River, Ekalugad Fiord, was predictive ability notably increased (see Table 12). It appears that the effect of rain is to saturate the ground and mobilize a "fresh supply" of soluble materials, so that higher concentrations of dissolved materials are experienced soon after rain, than after a longer period.

Discharge of Dissolved Materials

In order to obtain a solution transport rating curve, discharge of dissolved materials may be related to stream discharge. This represents a multiplication by Q of both sides of the equation relating concentration of dissolved materials with Q. The resulting relationship is statistically spurious, since it involves mainly a correlation between discharge and discharge (Benson, 1965). Nevertheless, for predictive purposes the equations $Q_c = f(Q)$ (Table 12), provide reliable estimating equations: apparent variance reduction is uniformly high, though of course the standard error of estimate is not reduced from that obtained for the relationships between concentration and discharge. Figure 31 illustrates the relationship derived for Lewis River.

Summary results are given in Table 13. Specific losses are remarkably comparable in all cases. At Ekalugad Fiord, the most highly glacierized basin has the highest specific losses of material, so that it is possible that the presence of glacier meltwater has some effect on the rate of solution. Furthermore, there may be a disproportionate amount of chemical activity associated with the production of glacier milk (though there is not a great deal apparent in any of the rivers studied). The extent of solution activity along the ice margin and under the ice is not known at all.

Suspended Sediment

The Nature of the Suspended Sediment Load

The general prevalence of mechanically weathered materials means that in arctic rivers most particles are of coarse sand grade and larger. Consequently, bed-load sediment transport dominates. Nevertheless, suspended sediment transport is considerable. A considerable amount of fine material does exist. McDowall (1960) has shown that freeze-thaw weathering rapidly reduces particle size of certain clay minerals to as little as 0.001 mm equivalent particle diameter. At this scale, the strength of molecular bonding is important, and so the process is selective, favouring minerals that possess planes of weakness. It is probable that mechanical weathering of this type is the dominant process in very fine particle reduction in the Baffin watersheds. At Lewis River, a moderate amount of very fine material, produced by glacial attrition, is also delivered directly from the glacier.

Grain size analysis of suspended sediment materials indicates that particles as small as 0.001 mm (1μ) are found, and that the range extends, normally, as high as 1 mm (Fig. 32). Amongst the coarser materials the

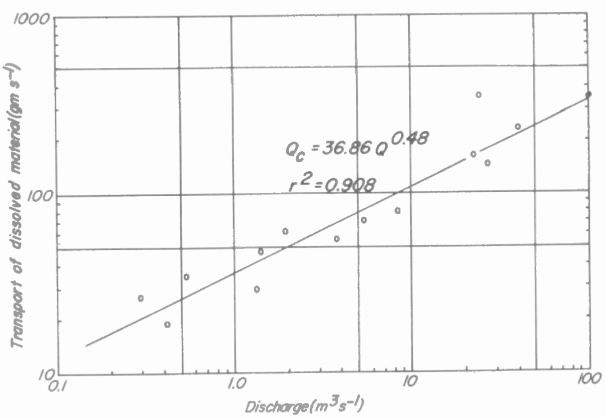


Figure 31.

Relationship between transport of dissolved materials (Q_c) and discharge (Q) at Lewis River.

distinction between bed and suspended load becomes very arbitrary since particles may move on the bed for some time at high flow, and then be swept into suspended motion at some point where stream turbulence is particularly intense, to be again deposited on the bed farther on.

Water samples for determining the suspended sediment concentration were generally taken with a United States Geological Survey Type DH-48 suspended sediment sampler. Normally two bottles (about 800 cc) of water were taken, each bottle representing a completely integrated sample in the vertical. The samples were both taken at the same location in the stream, as far out as it was feasible to wade. At high flow sampling was often restricted to near one bank, and the results can be considered to be representative only if it is assumed that turbulent exchange was sufficiently active to maintain a reasonably uniform concentration of sediment across the channel. The condition appeared generally to be met in the streams that were regularly sampled.

The water samples were filtered in the field, and the filter papers returned to the laboratory for ashing and weighing in the usual manner. A relatively fast paper was used in the field and, in controlled tests, it appears that a small, relatively constant amount of very fine material (in the colloid range) was passing through the paper. This amounted to about 3 ppm or less.

Samples were taken four times daily at Lewis River in 1963, and twice daily in 1964 and 1965 (0700 and 1900 hrs. local time). On the Ekalugad rivers samples were generally taken between 1900 and 1930 hours daily, with some 0700 samples. On several occasions samples were taken at hourly intervals for extended periods of time so that changes in sediment transport could be compared in detail with changes in other hydrological variables.

The distribution of observed suspended sediment concentrations is shown for each river in Figure 33. The highest sediment concentration recorded was 4794 ppm on North River at Ekalugad Fiord, during the storm of July 14, 1967. Discharge at the time when the sample was taken was 101.2 m^3s^{-1} (1735 hrs.). At 1930 hours, a sample taken on upper North River, with a flow of 106.5 m^3s^{-1} had a concentration of 997 ppm. It is probable that sediment concentrations of 1000 ppm or above were maintained for about 16 hours during this flood. No other sample closely approaches the North River 14 July value however, the next highest being 1741 ppm on Lewis

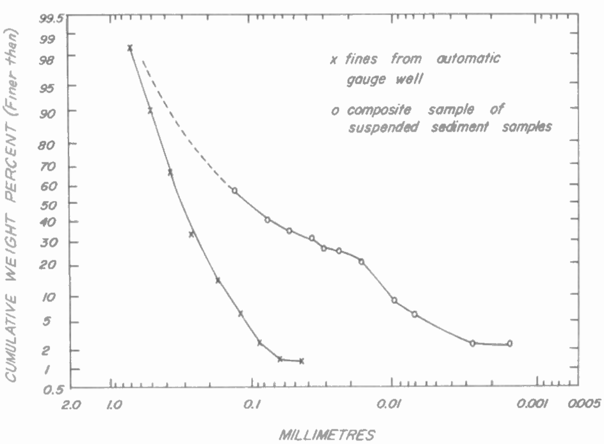


Figure 32.

Grain size distributions of suspended sediment at Lewis River. The guage well material is biased toward the coarsest sizes.

River at 1730 hours on 31 July, 1963. Discharge was 132 $m^3 s^{-1}$, derived from glacier melt (there had been a very small amount of rain during the preceding 24 hours).

Only half a dozen samples exceeded 1000 ppm, values of 500-1000 ppm being most common on all the rivers during high flow events. These values are far lower than maximum concentrations that have been recorded in glacial meltwater streams. Mathews (1964) infers that concentrations of 8000-10,000 ppm occur in the meltwater stream of the Athabasca Glacier in the Canadian Rockies, and Jensen (1881) measured suspended sediment concentrations of 9750 ppm in the West Greenland Isortoq Fiord. Fahnestock (1963) measured 17,200 ppm in the White River at Mount Rainier, Washington. Available results indicate, however, that in igneous or metamorphic terranes, the peak values are much lower, and of the same general magnitude as those observed in the Baffin streams. Pytte (1969) lists suspended sediment concentrations of up to 2600 ppm in Norwegian glacial streams; Rainwater and Guy (1961) report values up to 2800 ppm from arctic Alaska.

Examples of the relationship between concentration of suspended sediment in the rivers and discharge are shown in Figure 34. A portion of the scatter shown in the figures can undoubtedly be ascribed to errors and inconsistencies in the sampling procedures. Nevertheless, a clear measure of covariation remains.

Suspended sediment discharge relationships of the form $Q_s = pQ^j$ (suggested in Fig. 34) were used by Straub (1935) and Campbell and Bauder (1940) and were included by Leopold and Maddock (1953) as part of their "hydraulic geometry". Though relatively successful relationships have been found for some alluvial rivers, there is no reason to suppose _a priori_ that the suspended sediment discharge _should_ follow such a simple relationship. Suspended sediment is generally formed in part by the "wash load" of the stream; that is, material once entrained, can be moved right through the channel system at almost all stages of flow. The discharge of wash load is limited by supply, rather than by the competence or capacity of the stream. Gottschalk (1958), after distinguishing between sheet erosion and channel (linear) erosion as sources of sediment supply, asserted that better relationships are obtained between sediment concentration and stream discharge when the load is composed primarily of coarse-grained materials derived from channel erosion, as opposed to very fine materials of either source. Suspended sediment concentration might be expected, then, to be closely related to discharge alone only when there is a more or less unlimited supply of sandy material available along the channel, in the banks and in river channel bars. This condition may be approached in the Baffin outwash rivers that flow through moraine and older glaciofluvial deposits along much of their course, but there remains a wide degree of variance in the data.

In order to investigate this variation, correlation was made between sediment concentration and the following factors:

- rate of change of discharge during the hour preceding sampling (dQ);
- relative rate of change of discharge (dQ/Q);
- conductivity of the water ($\mathbb{C}$);
- temperature of the water (T);
- number of days since the last rain (P).

Their effect was investigated by their role in additional variance reduction in multiple regression. The most significant secondary factor is the length of time since the last rain, which appears in six of the eight equations (Table 14). It may be concluded then, that rainfall directly influences sediment yield in most situations. Of the two cases where it does not appear, one is the immediately proglacial Lewis River, and the other (upper Middle River at Ekalugad Fiord) contains several lakes and pools in its upper course which would probably act to dampen out secondary effects.

Conductivity appears in these relations, but it has variable sign, and so its effect is not clear. Since it is highly correlated with discharge in any event, it is probably not important physically. Temperature and relative rate of change of discharge each appear once. The general absence of the rate of change of discharge from the equations was surprising: it is possible that in small watersheds the effect is not detected in the same way as on larger rivers, where rising stage is commonly associated with greatly increased sediment concentration.

Short term channel scour is not significant in normal runoff on these rivers, either, so there would be no connection between changing discharge and scour. Figure 35 presents detailed data on stream discharge and suspended sediment transport for three 24-hour periods at Lewis River, showing that the concentration of suspended sediment follows very closely the variation in discharge itself. Essentially, it appears that suspended sediment concentration follows the relationship $S = f(Q, P)$.

As an interesting aside, $Q_s \propto Q^2$, more or less, for most of the streams, with the exponent rising to 2.5 for upper Middle and South Rivers. This approximately second power relationship for suspended sediment discharge (implying near linearity of the response of suspended sediment concentration to changes in discharge) has been found by many investigators, including Campbell and Bauder (1940), Mathews (1964) and Lustig and Busch (1967). Leopold and Maddock (1953) suggested the value should commonly be between 2.0 and 3.0. Axelsson (1967) found considerably higher values in a deltaic environment with a high proportion of very fine materials.

For each river the relationship between suspended sediment transport and discharge was investigated at two places. At Lewis River, the study reaches were on the upper Lewis River, along the north edge of the ice cap, and at the base station on the proglacial main river. The relationship between sediment concentration and discharge, which is good in both locations, indicates that the load is relatively larger and increases more rapidly with discharge at the upstream site. In ice lateral position, continuous melting of material out from the ice, slumping of recent moraines, and downslope relocation of the channelway as the glacier retreats, produce a continuously abundant supply of sediment. In the proglacial position, the dilution of runoff by clean supraglacial meltwater serves to lower the concentration of material in the water for the same discharge. Nevertheless, since the discharge at the base site is always far greater than that at the lateral site, the sediment discharge, and very often the absolute concentration are greater here.

In the three cases at Ekalugad Fiord, no significant change occurs in discharge between the two measurement sites, so the differences in relationships represent real differences in the sediment transport conditions through the respective reaches. The upper measurement sites are located above the active outwash plain on North and Middle Rivers; in the former case, at the base of the "canyon" that the river has cut through T2 deposits; in the latter case, at the lower end of a series of pools in Middle Valley.

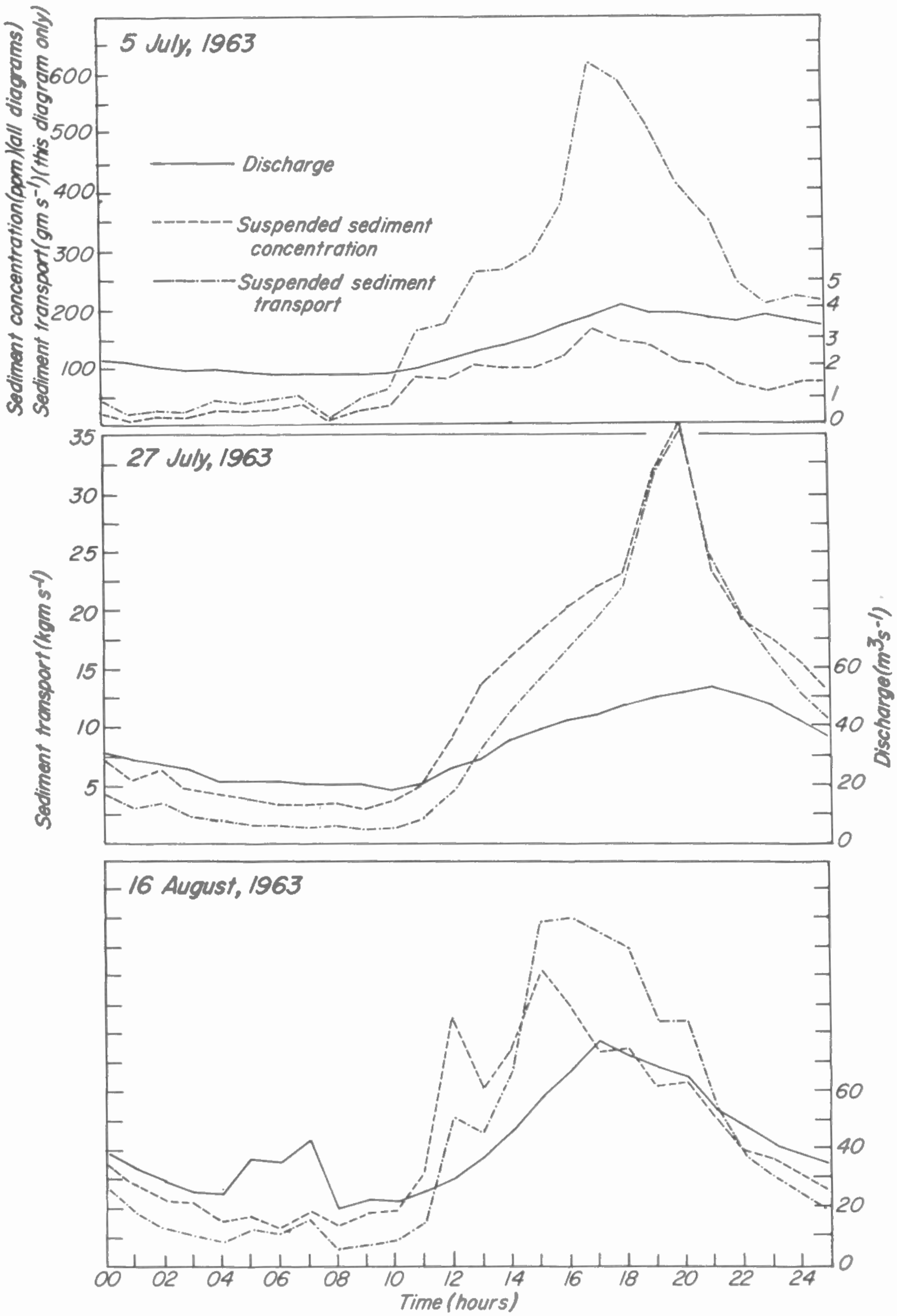


Figure 35. Detailed observations of suspended sediment transport at Lewis River.

On South River, the upper measurement site is at the point where the two main branches join. The distal sites are all near the sea. Between the two sites are the main active sedimentation areas on all three rivers. In all three cases, the relationship between suspended sediment concentration and discharge shows that the lower sites have higher concentration at low discharges, but that concentration increases more rapidly at the upper sites, so that at high flows, higher sediment concentrations are experienced there. This indicates that whilst net scour takes place downstream at lower flows, at high flow there must be considerable loss of material from suspended load down the length of the sandur. The same argument applies to net transport of sediment. The ratios of sediment concentration (or transport) between the upper and lower sites are as follows:

North River: $0.38Q^{0.18}$

Middle River: $0.21Q^{0.72}$

South River: $0.10Q^{0.64}$

The effect may in part be due to losses from channel flow to groundwater in the upper portions of the sandur and its reappearance at the distal end free of material load (but cf. p. 25). The discharge at which sediment concentration (hence transport) is the same in both sections is, for the three rivers, $210 m^3 s^{-1}$; $8.7 m^3 s^{-1}$; $38.3 m^3 s^{-1}$. The value for North River is very high, and probably is only rarely exceeded, but on Middle and South Rivers suspended sediment concentrations are frequently higher at the upper station. It appears, from this evidence, that North River is capable of transporting the imposed load of suspended sediment, but that Middle and South Rivers are not.

Suspended Sediment Discharge

Discharge of suspended sediment in the rivers was computed from operational equations given in Table 14. Daily totals of sediment transport are shown in Figure 40. Summary data are contained in Table 15.

Calculations of specific sediment losses from the watersheds must be interpreted with great caution. Aside from the inherent unreliability of only a few seasons' data, it is not clear what the data may mean in view of the uncertain sources of material. At Lewis River, for example, the high rates of sediment yield are the result of extensive erosion of glacial rubble. They do not represent the product of normal erosional activity, nor even the yield of materials from glacial activity (cf. p. 41). Hence, it is uncertain whether or not they may be referred to erosion over the entire watershed. It is certainly invalid to refer the yield to the unglacierized area alone, though the material has most recently been derived from a small portion of this area.

At Ekalugad Fiord, watershed sediment yields have been inferred from the upper sediment stations, since the lower stations are affected by changes on the present outwash plain. The difference in sediment yield between 1967 and 1968 at South River is striking evidence of the different geomorphological significance of the two seasons. The data for Middle River may be low since the station was located below a series of shallow ponds which probably trap considerable sediment in the short term. Similarly, much of

Figure 36.
Illustrations of storage of fine materials in the sandur channels.

A. Riverbank sandbar in Lewis River. Material carried in suspension at high flow is dropped along the riverbanks on declining stage.

B. Sand and gravel in a channel area that is abandoned at low flow (Lewis River).

C. Veneer of sand on a gravel bar: North River, Ekalugad Fiord.

the sediment yield from the upper South River watershed is probably lost in the ice-dammed lake in upper South Valley.

The difference in sediment yield between the upper and lower stations reflects the net sediment gain/loss on the sandur between. These data are summarized in Table 16. While North River produced a net deficit (as predicted from the equations), Middle and South Rivers both deposited material in 1967. South River, however, lost some in 1968. The eroded material is derived mainly from bank erosion in channel migration, though net degradation of the channel may also occur to some extent. Deposited material is mostly left as channel bars and channel fill in migrating channels (Fig. 36). However, overbank deposits occur in high floods, when a veneer of material may be left over a wide area.

Movement of Sediment as Bedload

The Nature of Bedload

The bedload sediment discharge of a river is normally defined as that portion of the total load which moves in almost continuous contact with the bed, by rolling, sliding, by mass movement in a "dense disperse layer", or by saltation close to the bed.

The theory and measurement of bedload sediment transport is less well developed than almost any other aspect of open channel hydraulics. There is no completely valid theoretical approach as yet, and there exists no satisfactory method for field sampling (Colby, 1963). While no field program of bedload sediment transport observations was planned for Lewis River, it was intended to carry out such a program at Ekalugad Fiord in order to establish empirical bedload sediment transport rating curves. Unfortunately, the river flow conditions during the period available for such studies were uniformly low, and no satisfactory sets of observations were achieved. A small group of measurements did throw some additional light on "threshold" criteria. However, for the main computations for both Lewis River and the Ekalugad Rivers, recourse was had to computational formulae.

Most bedload transport functions are given in terms of a relationship between a bedload function, ϕ, representing particle inertia, and an entrainment function , $1/\Psi$, representing stream power. The experiments which were carried out to establish the latter function by Shields (1936) may fairly be said to apply to "normally loose" boundaries only – that is to streambed materials resting in a nondispersed state, without imbrication, and with random packing. Such is almost never the case in nature, where by contrast materials are usually either "overloose" (open-packed, due to the presence of large volumes of water in the sediment, and often with a paucity of matrix materials), or "underloose" (close-packed or imbricated, and usually with pores solidly packed with matrix materials).

Observations made at Ekalugad Fiord indicate that the river gravels usually occur in an underloose state. This is common in gravel bed streams. In such cases higher tractive forces will be required to move materials in a certain size range than would be indicated by Shields' criterion. Measurements were made in the Ekalugad rivers by two methods. Series of size-graded pebbles, cobbles and boulders were marked and set out in groups of nine in various places in the rivers. They were set on the bed and hence

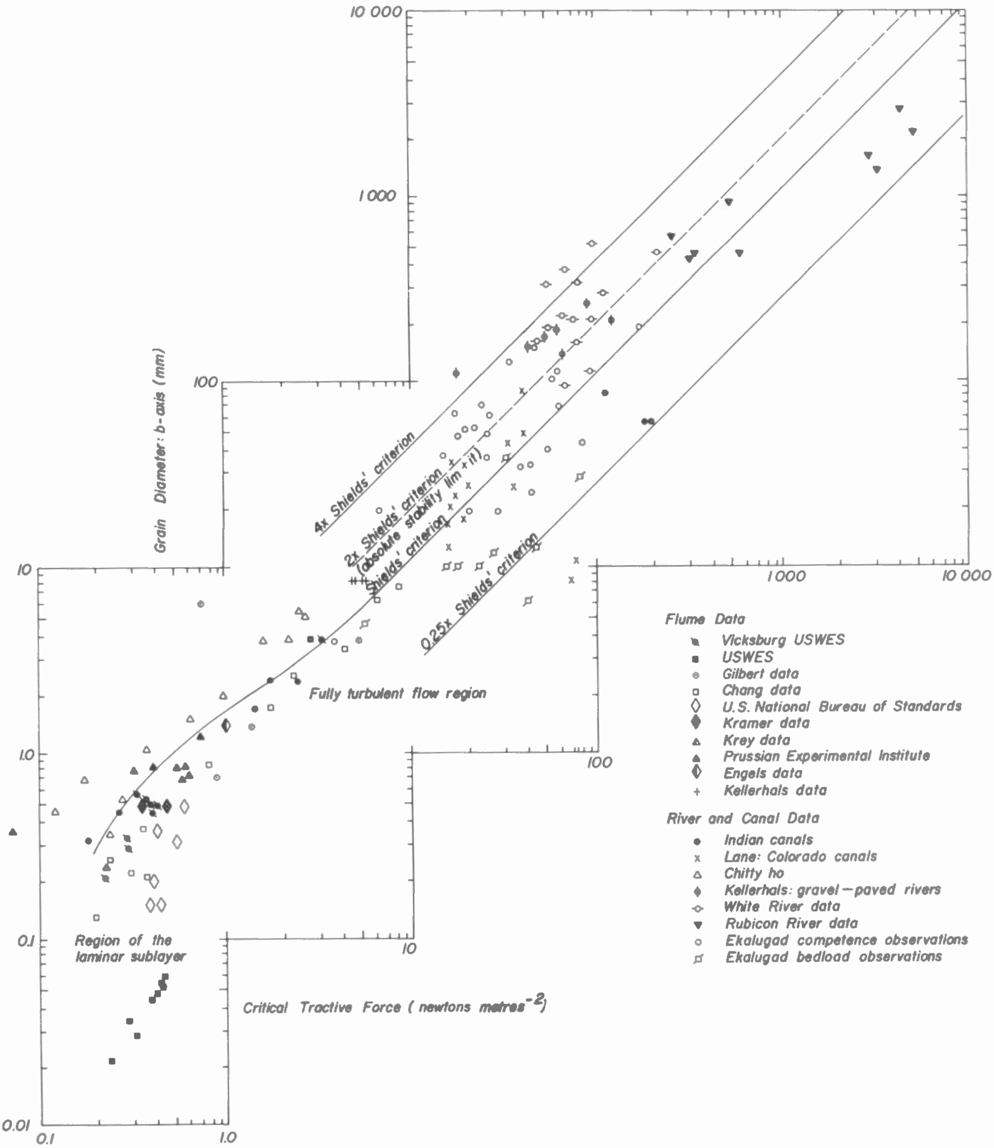


Figure 37. Laboratory and field data on critical tractive force.

were not protected in any way by the bed. The weights ranged from the order of 20 g up to the order of 15,000 g; b-axis diameters varied from order 20 mm up to order 200 mm. At each set, a single stage water level recorder was set up to record maximum stage.

Each site was visited periodically, to note the crest stage and to examine the cobble sets for movement. The largest stone to have moved was assumed to represent the competence limit for the observed highest stage. Stage values were converted to tractive force estimates by the equation $\tau = \rho g d \theta$, where d was calculated from the crest stage recording, and θ was surveyed for the reach. The results are plotted in Figure 37.

The second group of critical tractive force estimates arose from measurements of bedload sediment transport. A VúV-type bedload sediment sampler (Novak, 1957) was used. Because of low flows during the sampling period the overall program was a failure, but 9 successful samples were recovered at near-threshold conditions. For each sample, the largest cobble was determined (or the D_{90} parameter, in the cases where all the material consisted of fines). Tractive force was estimated as $\tau = \rho v_*^2$, where v_* was measured directly from the velocity profile of the streamflow at the sample site. Results are listed in Table 17 and are plotted in Figure 37. These data, representing transport of natural bed material, more accurately reflect the true threshold conditions in the Ekalugad rivers than do those derived from the boulder sets.

It is interesting that, while the theoretical (Shields') competence of the stream in two cases exceeded the size of the largest bed materials, in no case did the size of the largest material in motion approach the size of the largest bed material, and only in two cases did it approach the Shields, competence. The difference between the plotting positions of the two sets of data strikingly show differences in competence that arise when material is set in exposed positions on top of the bed, as opposed to when it has been incorporated within streambed deposits.

Apparent critical threshold values are tabulated in Table 18 for a variety of data appearing in Figure 37 via the equation

$$(6) \qquad 1/\Psi_c = \frac{\tau_c}{\Gamma (s_s - 1) D_m}$$

where τ_c is the observed critical tractive force and $s_s - 1 = 1.65$. Unfortunately, the data are not truly comparable from one case to another, since it is not always made clear in the original sources whether the τ values refer to the stress at absolute bed stability (i.e. no movement), or at Shields' threshold. Most of the river and canal data are probably referred to absolute stability, however. Flume data provided by Kellerhals, which indicate only slightly underloose conditions, are instructive. He achieved the gravel pavements in the flume by a process of flushing fines out of the bed, so that the pavement is a lag pavement. Hence, the degree of imbrication and structural strengthening exhibited by the gravels was probably low.

Bedload transport computations were carried out using a modified transport function, $\emptyset = f(1/\Psi - 1/\Psi_c)$ where $1/\Psi_c$ was empirically determined for the river bed conditions encountered.

In attempting to assess bedload transport by formula, a second consideration arises, that of the availability of material. Whilst material comprising the stream bed normally moves only sporadically during high flows, much smaller material can normally be handled by the stream at all normal flow stages. In the sand range particularly, much material that is suspended at high flow will move near the bed at lower flow levels. Since the stream is almost always competent to move such material, the discharge of this "wash" load will be limited by the material supply rather than by the stream power. Where active riverbank erosion is going on, or after a storm, such material is in plentiful supply, but after a long period of recession flow and slight stream activity, all such material may be washed out and the river bed lag material, which is too coarse for the stream to move at such stages, may be all that remains available. In such a period, the stream will flow "underloaded". This factor cannot be accounted for in strictly computational equations. Hence, the bedload sediment discharge can only be looked upon as "potential sediment transport" at full supply. Actual sediment discharge will be somewhat less.

Bedload Discharge Computations

Of the several bedload sediment transport equations available, the Schoklitsch (1934; and in Shulits, 1935) and Meyer-Peter and Müller (1948) equations have been widely applied in gravel-bed streams. Because it is amenable to adjustment of the critical tractive force, the Meyer-Peter and Müller formula was adopted for use. The full form of the equation, for channels having negligible bank resistance, is:

$$(7a)\ \gamma (k_t/k_r)^{3/2}\ d_* \theta = 1/\Psi_c (\gamma_s - \gamma) D_m + B\ (\gamma/g)^{1/3}\ g_s'^{2/3}$$

(the notations are as defined in the list of notations). The critical function limit, $1/\Psi_c$, was increased from Meyer-Peter and Müller's value of 0.047, to 0.094 (2 times as large). This is somewhat less than the average value of 0.118 deduced from sediment transport measurements at Ekalugad Fiord (see Table 18). The rather less conservative figure was adopted because all the transport measurements which led to the value of 0.118 were made at very low flow when the problem of supply of material may have impinged upon the observed results.

By transposing one term, and substituting for all the constants, this is reduced to

$$(7b)\quad g'_s = \left[0.646\theta^{1/4} D_{90}{}^{1/4} \bar{v}^{3/2} - 0.665 D_m\right]^{3/2}$$

This equation was operationalized as given in Table 19.

The formulae are represented in terms of discharge, where the hydraulic geometry of the stream sections (Fig. 38) for which computations were made was used to transform the equations.

D_m is the mean size of the bedload. All the sections used for calculations in the Baffin rivers are very rough and highly turbulent. Thus most of the fine materials passing through would be thrown into suspension locally at almost all stages. It may be expected, then, that the suspended sediment sampling encompassed virtually all of this material. The particle-size distribution data entered into the computational equations were accordingly biased toward the coarser fractions by making the computational mean size of materials equal to the actual D_{65} size. Therefore D_m is the D_{65} size of

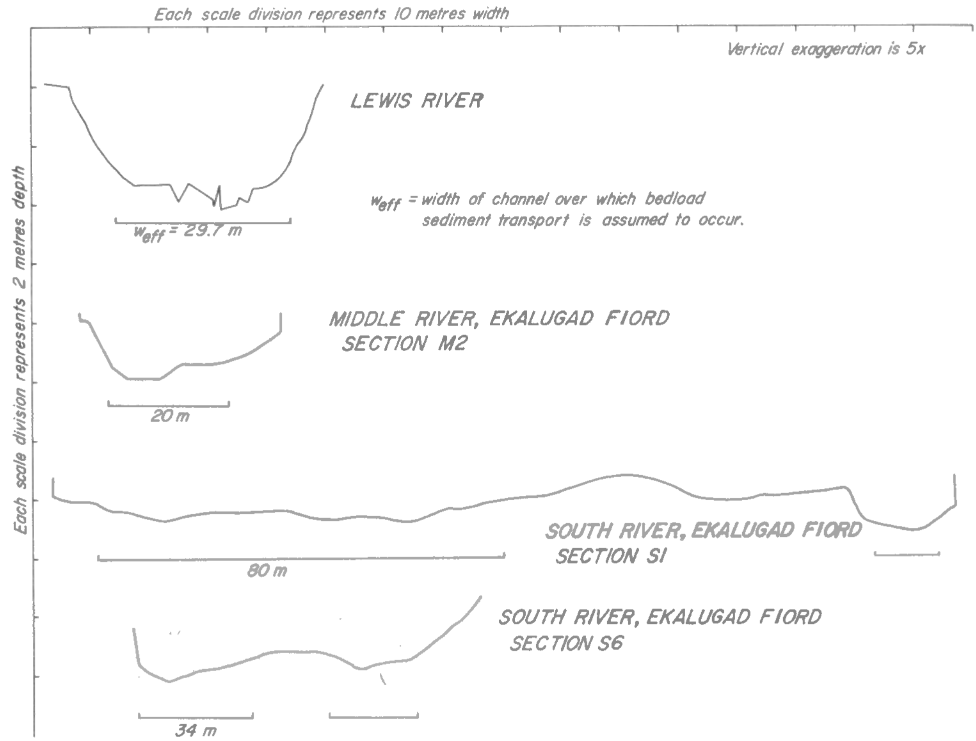


Figure 38. Geometry of channel sections used for bedload sediment transport calculations.

a composite curve derived from several samples of bedload material grab-sampled in each stream: the position and morphology of the sampled deposits clearly indicated that the material had moved during the most recent high flow. In the case of the Ekalugad rivers the composite curve was referred to all three sections. The data for these particle-size curves are shown in Figure 39.

D_{90} represents the D_{90} size of bed material sampled in each section. This material normally formed a bed "pavement", indicating that it did not move in normal flow. Nevertheless, it is the appropriate particle-roughness parameter. In each section a fixed "effective width" was chosen as the zone in which bedload sediment occurred. This is reasonable in these wide, trapezoidal sections, which are already flowing in the entire width of the section when transport begins. No calculations were made for North River, since there was no section for which the hydraulic geometry was satisfactorily determined. The values of Q_{crit} in Table 19 give the discharge at which sediment motion begins according to the formula. The values are in reasonable accord with observed movements in the streams, except that upper South River may be somewhat low.

The difference in threshold value between lower and upper South River stations is entirely due to greater velocities recorded at the upper site. Transport remains higher at the upper site at all stages, indicating that net

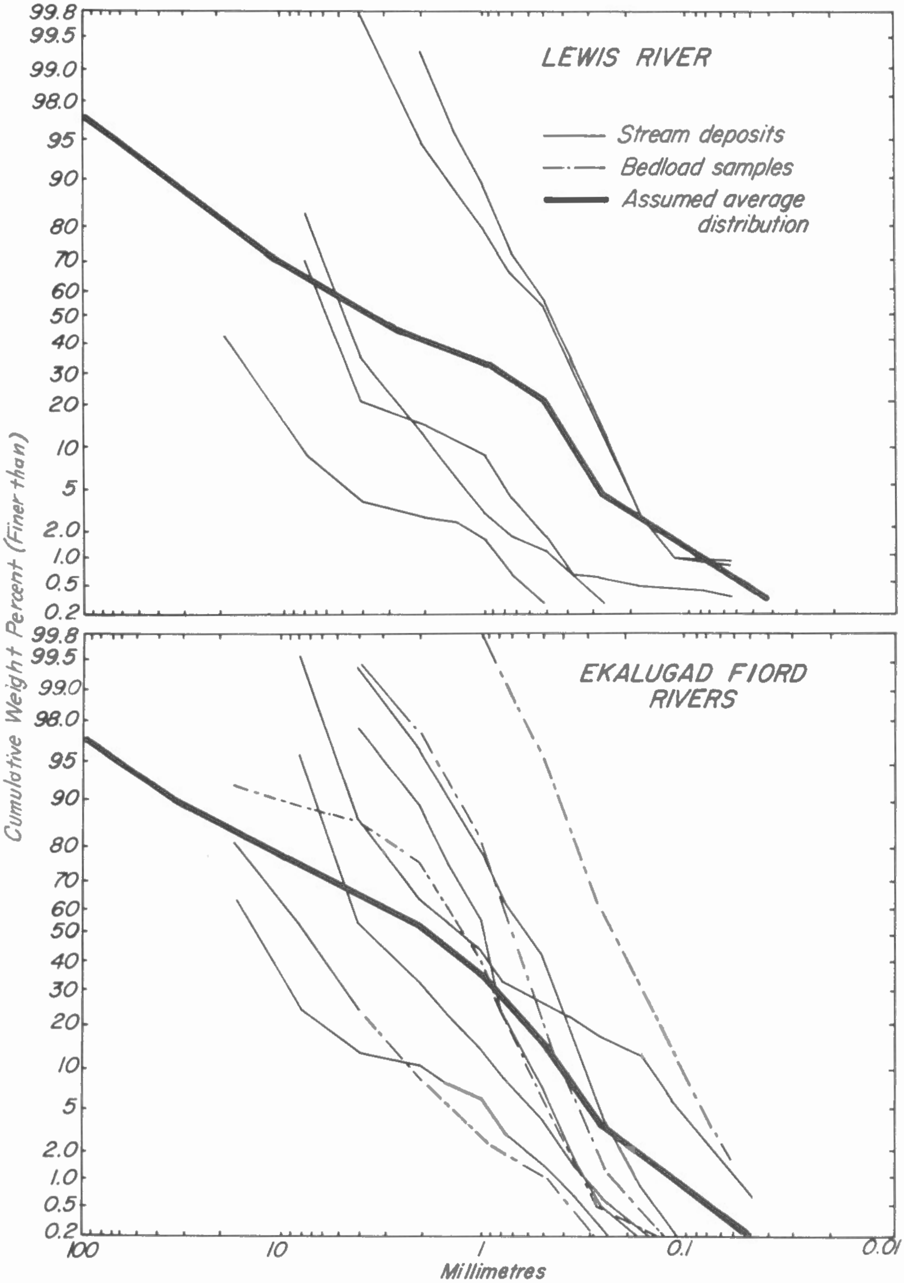


Figure 39. Distributions of bedload material size, and accepted mean distributions for computational purposes.

deposition of material must occur between the two sites. Bedload transport through the pools on upper Middle River is very low, so that net transport of bedload is certainly higher in the lower course. Nevertheless, most of the sediment is gained from erosion of the T2 terrace, so that streambed aggradation may still occur and this in fact appears to be the case.

Bedload sediment transport data is shown on a daily basis in Figure 40 and is summarized in Table 20. Sediment yield data for Middle River basin should be interpreted cautiously, since transport at the measurement site reflects conditions on the outwash plain rather than in the upstream basin. As with suspended sediment data, the area to which to refer sediment yield at Lewis River remains a very doubtful proposition.

The data indicate that on South River, 65,533 metric tons were lost from bedload transport in 1967, and 2,510 metric tons in 1968, between the upper and lower measurement sites.

The Pattern of Sediment Movement

Total Sediment Discharge: The Dominance of Bedload Transport

The total sediment discharge, by solution, suspension and bedload transport, is summarized in Table 21. Distribution of the load is relatively conservative at Lewis River, despite a wide variation in flow amongst the three years, with about 20 per cent passing in suspension and 80 per cent as bedload. Solution accounted for less than 1 per cent every year. At Ekalugad Fiord bedload was overwhelmingly important in 1967, apparently accounting for 90 per cent of the load. In the very low flow year of 1968, however, bedload was relatively unimportant at the lower South section, with 38 per cent passing as solution load, and 49 per cent suspension (total sediment transport was almost nil). At the upper section, where about 6 times as much sediment is indicated to have passed, over 90 per cent was bedload – this accounts for all the difference between the two stations.

Even discounting the probability that the computed bedload transport is higher than the actual bedload transport achieved these data show dramatically the dominance of bedload sediment transport in the sandur rivers, in marked contrast to most rivers, in which bedload transport is supposed to be relatively minor. Even mountain rivers in the Alps apparently move very little bedload (Moosbrugger, 1958; Bogardi, 1958). By contrast, in nival and proglacial environments and in rapidly eroding mountain regions, bedload transport dominates sediment movement (Pardé, 1958).

The Relation of Sediment Movement to Discharge

The distribution of sediment transport is progressively less equable from solution to suspended to bedload, indicating that extreme flow events become more important in the movement of each part of the load. This can be deduced from the sediment transport equations. For solution load, the equations take the form

(8a) $Q_c = c_1 Q^{c_2}; c_2 < 1.0$

from which

$$\text{(8b)} \quad \frac{dQ_c}{dQ} = c\ bQ^{c_2-1}, \text{ and } c_2-1<0$$

Hence the rate of increase of solution load decreases with increasing discharge, which dampens the effect of large flows. The distribution of solution transport will be more equable than that of discharge itself. For suspended load (Q_s), the form of the equation is the same, except that $c_2>1.0$, so that in the derivative equation $c_2-1>0$. In fact $1.9<c_2<2.5$, so that the rate of increase of suspended load will increase more or less linearly with discharge. (Both of the foregoing equations also contain a negatively powered term for the effect of time since the last rainfall in their operational form.)

The bedload equation is of the form

$$\text{(9a)} \quad Q_b = ws'\left[c_1Q^{c_2}-c_3\right]^{3/2};\ c_1\sim 0(0.1), 0.75\leq c_2\leq 1.20,\ C_3\sim 0(1)$$

At discharges considerably above the initial value at which bedload movement begins, this equation is nearly equivalent to

$$\text{(9b)} \quad Q_b = ws'c\ Q^{3/2^{c2}}$$

which is now of the same form as those above. The derivative equation has powers between 0.1 and 0.8 so that the rate of increase of Q_b increases slowly with Q. The equation is more conservative than that for suspended load, but the presence of a finite threshold for the commencement of transport makes the bedload appear to be more variable at moderate flow levels. Since the values of the constants are much greater (when all the equations are expressed in gram units) actual bedload transport is also much greater than suspended load transport at all realistic flow levels.

Table 22 shows the proportion of total seasonal runoff and sediment transport of each type achieved cumulatively on the highest 1, 2, 3, 4 and 5 days of flow/transport (most of the time the days coincided for each type of discharge, though minor discrepancies are found, depending on how peaked or equally distributed the given day's discharge hydrograph may have been). Interesting features of the table include the very high concentration of sediment movement events achieved by storm dominated hydrographs as opposed to melt dominated ones (see Lewis River 1964 as opposed to Lewis River 1963); the extremely high concentration of transport in single remarkable events (cf. The South River, 1967, records, within which the jökulhlaup of 20-22 July stands out) and the increasing sensitivity (as seen above) of suspended and bedload movement to the extreme events.

Between 30 per cent (Lewis River, 1963) and 75 per cent (South River, 1967) of total sediment movement (excepting the unusual record of 1968 in upper South River) occurred in any one season in less than 8 per cent of the total runoff time of record, and over three seasons at Lewis River, 25 per cent of the total sediment transport occurred in about 2.5 per cent of the time. These are far from the most extreme results on record but they do point out the achievement of a major proportion of total sediment transport in a very restricted period of time.

In determining the total sediment transport, flow duration (hence, sediment transport duration) is important. Figure 41 presents Wolman diagrams for certain of the rivers to illustrate the distribution of sediment transport work. The distributions are, for the most part, bimodal, with transport-frequency peaks occurring at moderate flows, near the mean flow, and again at very high (but not the highest) flows. This distribution of transport-frequency indicates the great sensitivity of sediment transport to detailed departures of the stream-flow frequency from a smooth function. So sensitive is the transformation of frequency through the sediment transport equations that the approximation of discharge frequency by a smooth function (see p. 35) seems rather futile unless it can be shown that in the long term limit the distribution of events must conform (one is tempted to say a priori) to that particular function. This degree of sensitivity has not generally been noted.

Erosion and the "Geological Norm"

The total denudation represented by the sediment transport is difficult to assess. Data from single seasons are meaningless, except insofar as they indicate the order of magnitude of erosion. In the present instance additional complications arise from the large proportion of glacierization (especially at Lewis River), so that one cannot clearly assign an effective area of erosive activity, and the fact that much of the present sediment yield is derived from removal of unconsolidated glacial sediments. Hence sediment yield bears no relationship to present primary rates of sediment production in these watersheds. Sediment yield values indicate denudation rates varying from order 10 mm/1000 years (South Valley, 1968) to order 500 mm/ 1000 years (Lewis River, 1963). The three year mean at Lewis River is 277 mm/1000 years averaged over the entire present watershed. Almost all of this material is derived from earlier deposits at ice margins, and may originally have been derived from a much larger area than that of the present watershed.

This state of affairs will presumably last only so long as Pleistocene materials remain easily accessible to the rivers. In light of the importance of fluvial activity in shaping the landscape even in the Arctic, and in view of the comparative rarity of glacial events, one may reasonably define as the geological "norm" of denudational processes (Lowdermilk, 1934) that condition in which the river system is just capable of transporting and disposing of an amount of material equivalent to that readied for movement by the weathering processes of the ambient environment. The Baffin watersheds are currently in a state of heightened activity with respect to such a "norm"; that is, they are actively redistributing a mass of detritus left behind by the recent glacial events, which can in this context be looked upon as a serious perturbation in the geological-scale fluvial evolution of the landscape. The landscape is still going through a process of "relaxation" after this perturbation, and the appearance of the major alluvial deposits of the postglacial period is the main consequence of this.

Exactly what constitutes the "geological norm" for erosion in such an environment as Baffin Island is difficult to say. Denudation by solution processes on Baffin Island, which appears to be of order 1 mm in a thousand years, is in good agreement with the range of 1 to 10 mm reported for arctic and subarctic regions. It is reasonable to expect relatively low rates on massive igneous terranes. Dahl's (1967) data on granite weathering, which he concludes to be due to a combination of solution and microgelivation, are particularly instructive in this regard. Summary data of Corbel (1959) on erosion of silica rocks confirms the result as well: it appears that 1 mm ± represents the long term "norm" for solution weathering on Baffin Island in the present environment.

The movement of clastic materials from the Baffin Island watersheds appears to be of an order that would lead to 300-500 mm of denudation per millenium. This value is in good agreement with data derived from other glacierized watersheds (Thorarinsson, 1939; Lanser, 1958; Corbel, 1959; Pytte, 1969): it is considerably higher than results from most nonglacierized mountains (Geiger, 1958; Lanser, 1958) and from most arctic and subarctic basins (Corbel, 1959; Rapp (including results of Eriksson), 1960; Axelsson, 1967; Arnborg et al., 1967; Østrem, Bridge and Rannie, 1967). However, there is sufficient variation in the results to discourage any categorical statements. Hence, whilst the Baffin watersheds appear, from measurements of present rates of sediment transport, to behave like remarkably high energy environments, it is not clear that they would continue to exhibit such behaviour after substantial completion of postglacial redistribution of materials. Data from otherwise comparable watersheds in the Austrian Alps, where only one has any current glacierization, show a difference in yield approaching 50 times (Lanser, 1958). Consideration of the present primary weathering processes in the Baffin watersheds makes it clear that "normal" sediment production is at least an order-of-magnitude lower than the present total yield rate. Probably the data of Rapp (1960) from Kärkevagge represent upper limits of what could be expected in Baffin Island (10 mm per annum).

Hence, the development of the major valley fill deposits in most of the Canadian Arctic since Pleistocene times is an expression of sedimentation under the abnormal conditions left after the Pleistocene glaciation, and still extant.

Aeolian Sediment Movement

While there are no major deposits of windblown materials in the sandur areas on Baffin Island, the occurrence of wind-rippled sands and the development of lag pavements of gravel or cobbles in some instances where water action was not responsible, indicate the activity of the wind. Dust-covered snowbanks in spring further attest to the movement of material by the wind.

In order to determine the relative magnitude of the movement of materials by the wind, a tower was erected on a small sand area at Lewis River in 1965 and samples of wind-borne material were caught at various levels up to 100 cm above the surface. Since the results were intended to be only indicative, they are not reported in detail. Over the one-month period of observation, about 1600 gram/metre width of "live" surface were transported to the west away from the sandur; over 40 per cent of it in 1 day and over 70 per cent of it in just 3 days. The bulk of the material moved within 50 cm of the ground, and probably was in saltation, though considerable quantities of dust were found at 100 cm above the ground.

The significance of this result for deflation of materials from the sandur must be qualified by several factors. First of all, the extent of "live" surface must be considered, that is surface which has loose sand upon it that can be entrained. Surfaces stabilized by lichens or by cobble lag deposits will not permit the movement of significant quantities of material: for considerable transport to occur there must be material continuously available for entrainment. At Lewis River the live surface in the 1900 moraine area is not very wide (only a few tens of metres along the river) and it is unlikely that much material escapes other than very fine silt and clay which can be borne in suspension to considerable heights. This material is not quantitatively significant. It appears that windspeeds of about 12 km hr^{-1} are necessary to initiate measurable movement, but that the increase in movement is rapid after that. During 1964 and 1965, winds in excess of 12 km hr^{-1} were observed for about 20 per cent of the time at Lewis River (Fig. 14).

Fristrup (1952) claimed that a significant amount of wind deflation occurs in winter, when very strong winds occur. Though some movement occurs (as evidenced by dirty snow in spring), most of the valley floors in Baffin Island are snow covered, and in any event, they often freeze when wet and so become hard. Hence, winter deflation is not thought to be appreciably more important than summer season deflation.

The measurements indicate that the volume of material deflated is not important by comparison with fluvial transport. Nevertheless, the process remains significant for the detailed evolution of parts of the sandur surfaces.

CHAPTER V

HYDRAULICS: RIVER CHANNELS ON THE SANDUR

Introduction

Coarse, clastic materials are noncohesive. Hence, resistance to erosion by flowing water is invested almost entirely in the weight of individual particles (in natural channels tight packing of materials with a wide variety of grain sizes, and imbrication become important). In naturally occurring materials exhibiting a wide range of grain sizes channel bed and bank erosion are selective, so that gradually a "lag" deposit of the coarsest materials comes to armour the channel bed and sides. So long as sediment transport into the reach from upstream is sufficiently small that continual aggradation does not occur, the channel will become stable for flows below the competence level of the lag material. Imbrication and packing may further raise the threshold of competence for flows capable of altering the channel shape. The majority of channel adjustment is by bank erosion. Hence, channels in noncohesive material tend to be very wide with respect to depth (Schumm, 1960). Henderson (1966) has emphasized that a wide channel, with more or less sinusoidal banks and a flat central section represents a stable section in noncohesive material on the point of entrainment. The tendency for such channels to widen excessively is an importnat aspect of the development of braiding.

Channels can be dichotomized into single and multiple channelways. Single channels exhibit a variety of forms ranging from the pool and riffle sequences of small, steep streams, through varying degrees of sinuosity to fully developed classical meanders. Various writers have pointed out that pools and riffles, or thalweg meanders can be detected in almost all "straight" reaches (but see Miller, 1958). Leopold and his coworkers (see Leopold, Wolman and Miller, 1964) and Henderson (1966) have concluded that the quasi-regular spacing of pools and riffles or of flow deflections represent the morphological consequences of essentially similar flow mechanics through the entire range of channels and that they can all be treated similarly.

The braiding that characterizes multiple channelways indicates a regimen different from that found in single channels. Multiple channels are found in two distinct situations. High energy streams flowing in noncohesive sediments and carrying heavy sediment ' ads commonly exhibit braiding. By contrast, channel division often occurs in deltaic situations where the energy available to the stream is rapidly diminishing. This circumstance often occurs also on alluvial fans where the specific energy available to the stream is often less than in the single channel upstream from the apex of the fan. Price (1947) and Rubey (1952) explicitly designated channel anastomosis as "braiding", in the classical sense of the term, in the presence of an appreciably reduced energy gradient. Lane (1957) indicated two major categories of braiding; that associated with steep slope, and that associated with aggradation. Dury (1969) has set forward a six-fold division of multiple channels, but has advanced no hydraulic reasons for his distinctions. Howard et al. (1970) have provided a description of braided channel topology.

Meandering, pool-and-riffle sequence, and braiding may be found superimposed upon each other in one channel. Taken collectively they all represent channel adjustments to the imposed water and sediment load, and to the strength of the materials in which the channel is excavated, so that a "graded" profile results (in the sense of Mackin (1948)). Meandering is normally accepted as the reaction of an initially overcompetent or underloaded stream: the sinuosity developed through erosive activity increases form resistance to flow, and more or less reduces the rate of specific energy expenditure by lengthening the channel course, until a stable state is reached. By contrast a variety of underlying reasons has been cited for the development of channel braiding, only part of which are undercompetence or overloading with sediment (Lane, 1957).

Bank erodibility per se has been cited as a cause of braiding (Fisk, 1944; Mackin, 1956; Brice, 1964): where the banks are very weak, excessive lateral erosion may lead to very wide channels in which shoaling occurs on central bars, and so to the development of multiple channels. Doeglas (1951) ascribed braiding in some Pleistocene rivers to rapid, large variations in runoff. Whilst this phenomenon is undoubtedly a part of the regimen of many braided rivers, it seems insufficient as a cause by itself. Steep slope (Lane, 1957) or change in slope have also been suggested as causes of braiding, but seem to be associated consequences in some instances, an indirect causal factor in others.

Common to almost all discussions of the cause of braiding is a heavy sediment load. In particular, appreciable bed load transport seems to be a necessary part of the mechanics for anastomosis in high energy channels (Hjulström, 1952; Fahnestock,1963). In these circumstances local overloading (Stebbings, 1964) or decreased competence (Leopold and Wolman, 1957), determined by variation in slope, channel depth and width, discharge fluctuation, or irregularity in the channel, lead to bar development and possibly eventually to braiding.

Braiding has most often been associated with environments of rapid aggradation. Leopold and Wolman (1957) asserted, however, that braided channels can exist in equilibrium in response to particular combinations of hydraulic variables. Lane (1957) and Brice (1964) indicated that braided channels also occur under degrading conditions.

Leopold and Wolman presented a discriminant function which divided meandering from braided channels according to slope. Henderson (1961) refined this criterion by taking grain size of bed material into account. His equation was such that most of the single channels followed it, while a few had substantially lower slopes than indicated. Braided reaches all exhibited substantially higher slopes. Chien et al. (1961) have suggested a more complex functional criterion which takes into account flow variability as well.

At Lewis River and Ekalugad Fiord, braided channels are common, but they are localized on certain sectors of the sandur. Elsewhere single channels occur. All the channelways exhibit more or less regular pool-and-riffle sequences associated with major sedimentary bars. Well developed meanders do not occur, though North River at Ekalugad Fiord, which flows in a single channel over almost its entire course, exhibits discernably regular deflections of the channel along its course.

Sectional Characteristics

Geometry

Sandur channels are characteristically wide and shallow. Their shape varies from trapezoidal form, to that of an extreme scalene triangle. The latter form occurs where the thalweg is near one bank. Midchannel bars are common, even in single channels (i.e. the bar remains submerged at all normal flow levels). This illustrates the extent to which the braiding state is a function of stage.

Width-depth ratios, w_s/d_* (Fig. 42), were computed for a variety of sandur channels, grouped into five classes. The last group includes minor channels on Ekalugad sandur and some valley side tributary channels for comparative purposes. Summary statistics of the channels are given in Table 23. Analysis of variance amongst the restricted groups indicated that they do not differ significantly from each other ($F = 1.89$, with df 4 and 82) on the whole, though clearly there is considerable variation between the group of "small channels" and the others; this probably represents a scale difference. Some remarkably wide, shallow channels occur in upper Middle River (w_s/d_* ranges as high as 400). These channels are stabilized by cobble beds and vegetated banks in the old T2 materials.

The channels reported here have relatively much greater width than channels on the White River outwash (Fahnestock, 1963). The difference is again largely a matter of scale; flow there was of the order of 1/10 that observed in the sandur channels here. The summary statistics of those channels suggest a distribution not unlike that observed in the present case, however (see Fig. 43).

Channel geometry is closely associated with the nature of bed and banks (Schumm, 1960); channels in noncohesive sands and gravels are commonly very wide. For an increasingly wide channel, the shear on the bed is increased at the expense of that on the bank (Leopold and Wolman, 1957). Increasing width/depth ratio is associated with a more uniform distribution of shear between bed and banks (Davidian and Cahal, 1963), a point of particular importance if channel stability is to be attained. In noncohesive channelways such adjustments are readily made. Wolman and Brush (1961) pointed out that to pass a given fluid discharge along a given slope requires a specified cross-sectional area for a given particle roughness determined by the nature of the bed material. A given specific discharge will require a certain velocity and depth of flow in order to maintain bed stability (or to pass the imposed bed load). If stresses exceed bank strength, then channel widening will carry on until stability is reached. In the same study, the authors found that for coarse materials in flumes, channel widening tends to continue, in the presence of weak banks, with increasing flow, so that shear stress on the bed is always distributed so as to maintain values just above the threshold for material transport. In natural channels, bed and/or bank armouring may confuse this picture however.

Stebbings (1964) has found that a limiting geometry exists beyond which channels transporting high amounts of bed load begin to braid. Of course, when a channel splits into two new ones, the mean geometry for the two new channels becomes much more conservative than it was for the single original channel - other things being equal the value of w_s/d_* should, initially, be approximately halved. Amongst all the data collected, however, the mean

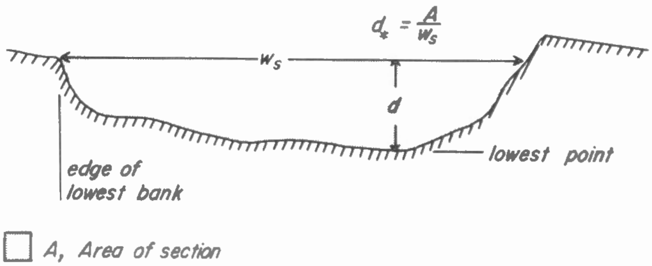


Figure 42.

Definition sketch for channel geometry.

width/depth ratio for nonbraided channels is not appreciably greater than for braided channels (Table 23B). It seems that anabranches rapidly adopt a sectional mean geometry that is very similar to that of the undivided parent channel. To some extent, differences that appear could be explained by scale effects.

Hydraulic Geometry

The hydraulic geometry was computed for a group of sections at Lewis River, and on Middle and South Rivers at Ekalugad Fiord (Fig. 2), in the manner established by Leopold and Maddock (1953). The relationships are similar in all sections, especially for velocity and area. The derived relationships are displayed in Figure 44.

The exponents of the equations of hydraulic geometry indicate the variability of each parameter with discharge. Amongst the sections studied, those on Middle River at Ekalugad Fiord may be somewhat atypical in that they are both located in an area of current aggradation and rapid channel changes. The remaining sites are remarkably uniform in their characteristics. The values of exponents measured in these sections are listed in Table 24, with some results from other studies in Table 25. By comparison, width, depth and area are all relatively conservative, and a proportionately greater amount of the adjustment to changing discharge is taken up by changing velocity than is the case in most rivers. In fact, in only one case (Upper Triangle River) is velocity not the most variable "dimension". Hydraulic data reported by Arnborg (1955) for the gauge section on the Hoffellssandur, in Iceland, indicate a similar result, through, suggesting that this may not be an unusual case for sandur streams. Fahnestock's (1963) data reveal considerable departure from this situation, but unfortunately, because of the instability of the channels, he was constrained to make successive measurements in many different channels, so that the interpretation and comparative value of the results remain unclear.

The high rate of change of velocity is associated with the relatively large and rapidly increasing sediment load that the streams carry at high flows: the maintenance of high competence to move the large fractions of the bed load requires rapid increase of velocity, and width and depth changes are relatively small by comparison.

The high rate of increase of velocity in the sandur channels indicates that resistance decreases more rapidly as discharge increases than is the case in most channels. If it is assumed that flow resistance is similar on bed and banks, then the Darcy-Weisbach flow resistance number

(10)
$$ff = \frac{8gR\theta}{v^2} \simeq \frac{8gd_*}{v^2}$$

is proportional to Q^{f-2m} (Leopold, Wolman and Miller, 1964), where f is the exponent in the relation $d_* = c\,Q^f$, m is the exponent of $\bar{v} = k\,Q^m$, and θ is considered to be constant. For the sandur channels $ff \propto Q^{-0.65}$, which is approximately twice as steep a relationship as has been encountered in most channels in noncohesive material. Three main reasons may contribute to this:

- rapidly increasing sediment discharge at higher flows may contribute significantly to damping of turbulence and increased mass transfer in the section (Vanoni and Nomicos, 1960);
- at progressively higher flows an increasing proportion of channel flows "straight through" the pattern of pools and riffles that exists at low flow, greatly reducing the form resistance associated with these elements;
- at high flows, a "live" bed presents lower boundary resistance than occurs with a stable bed at low flows.

In open channels with rough boundaries flow normally conforms either to a logarithmic friction law (Keulegan, 1938) or to a power equation, commonly the Manning formula, of the form

(11)
$$\bar{v} = c_1 R^{c_2} \theta^{1/2}$$

where R is the hydraulic radius, and is here taken to be equivalent to d_*. This equation can be represented as

(12)
$$\frac{\bar{v}}{v} = \sqrt{\frac{8}{ff}} = C_3 \left\{\frac{d_*}{D}\right\}^{c_4}$$

where d_*/D is relative roughness, as in the logarithmic equations. D is some representative bed particle size, here taken to be D_{90} (the choice of D will not affect the power of the relationship so long as it remains constant in the section).

By comparing the Manning formula with the formula for ff it is seen that

(13)
$$\sqrt{\frac{8}{ff}} = \frac{d_*^{1/6}}{\sqrt{g}\, n}$$

Now, if we accept Strickler's relation for the roughness due to particle size in gravel streams, viz. $n = 0.038D^{1/6}$ (metric units), we find that the variation in flow resistance due to the boundary resistance generated by particle size is expressed in

(14)
$$\sqrt{\frac{8}{ff}} = 8.4 \left\{\frac{d_*}{D}\right\}^{1/6}$$

This provides a criterion for flow resistance due to particle resistance alone which is valid in the range $7 < d_*/D < 130$ (see Ackers, 1958, p. 25).

It appears that not only is the size of bed particles important in determining boundary resistance, but so is the spacing. Koloseus and Davidian (1966) have shown that a relationship exists between roughness element concentration and the resistance coefficient which is positive up to quite dense concentrations, and then becomes sharply negative (presumably as the tops of the roughness elements begin to present a "uniform bed" to the flow). White's (1940) contention that "dominant" roughness elements (the largest particles) are distributed with somewhat regular spacing in channel beds would suggest that this effect is present in stable channels. Significant bed load transport might sharply reduce resistance by eliminating this effect as the sediment movement "smooths" the bed.

Once the bed of the stream becomes live, the standard formulae cannot be appealed to in order to partition ff between form resistance and particle resistance (ASCE Task Force, 1963), since ff no longer conforms to these laws. It is not known whether the bed is actually moved in the sandur streams, or whether the boulder pavements present at low flow are merely covered with moving material, mainly sand. For the more stable sections, the latter is probably the case. As a check on whether or not live bed conditions might be expected to occur, the bed load transport according to the bed load transport equations (see p. 58) was computed for their respective sections at very high flow with the results shown in Table 26. If it is assumed that material moves approximately at the friction velocity there is a layer of about 3 cm depth moving in each section except lower South River. Since many of the cobbles in motion are considerably larger than that, it seems that general motion of the bed may not occur in these sections. But some of the less stable sections (in particular Middle Lewis River, and M3 and S4 at Ekalugad Fiord) definitely undergo total bed removal during high flows.

In an attempt to determine directly the form of the resistance equation, values of $\frac{\bar{v}}{v_*}$ were regressed on corresponding values of $\frac{d_*}{D_{90}}$ for each of the sections studied in order to determine the parameters c_3 and c_4 in equation (12), assuming that D_{90} remains constant. The results are shown graphically in Figure 45. In almost all cases flow resistance is initially much higher and decreases far more rapidly than would be indicated by either the Manning formula or the usual logarithmic law. The intercept c_3, is less than 8.4 in every case (in only three cases is that value even encompassed within the $\alpha = 0.05$ confidence interval of the regression estimate). The exponent, c_4, is much higher than 1/6 (=0.16), indicated by equation 14, in all but three cases -- significantly so in eight cases. In five cases, it is significantly greater than the mean value of 0.5 determined by Wolman (1955) and Leopold and Wolman (1957) for a variety of natural channels (including anabranches of sand-bed rivers). The mean exponent for the South River sections was 0.67.

Thomas (1949) found that Beas River, a gravel river in the Punjab, had a similarly high exponent in the friction law, of about 0.70 (determined from Thomas' plot of Beas River data).

Several attempts to analyze the departure of the friction coefficient from that indicated by a boundary particle resistance law have been made, normally involving the partition of some term in the friction formula. Einstein and Barbarossa (1952) proposed to divide the hydraulic radius, R, whilst Brooks and Taylor (reported in ASCE Task Force, 1963) proposed to partition the energy gradient, θ. Inglis (in discussion of Einstein and Barbarossa, 1952) proposed a threefold division of friction as follows:

- due to boundary particle roughness;
- due to ripples, dunes, bars and other transitory bed forms;
- due to channel curvature, projections into the flow, structures, etc.

The last group of roughness elements are quasi-permanent, or change only very slowly, as opposed to the second group, whose existence changes with the flow, and which may even depend in detail upon recent flow history. It is the combined variation in the first two effects that is the principal cause of the well known change in friction factor with discharge. Hence, it should in principle be possible to isolate particle and bed form resistance. However, the changing nature of the boundary as sediment transport increases makes this a frustrating exercise, so far as physical meaning goes.

Leopold et al. (1960) have indicated that at Froude numbers in excess of 0.4 - 0.55, a new source of flow resistance becomes apparent (over and above particle and form resistance) which they call "spill resistance". It is associated with excessive deformation of the surface when flow discontinuities occur, with considerable energy losses. Flow resistance increases rapidly in response to this effect. In the sandur, streams flow around isolated, very large boulders for which $D \sim d$, and flow around bank obstructions and irregularities contribute considerably to "spill" resistance. Locally, flow about such obstacles becomes supercritical, and this is followed by energy dissipation in a hydraulic jump downstream. At Lewis River $\mathbb{F}$ approaches 1.0 in the channel as a whole at peak discharge in two sections, both of which contain large roughness elements. The variation of flow resistance with discharge is shown in Figure 46 for the sandur sections. There is considerable scatter in the relations, at least some of which can be explained in terms of changing form resistance as water levels rise in the irregular channel sections.

Change in form resistance cannot, however, explain entirely the behaviour of the sandur channels, and it is clear that live bed conditions must occur. The effect of live bed conditions is almost always to decrease the effective particle diameter at the bed (the bulk of the sediment transported is sand). Hence, values of $\frac{d_*}{D_{90}}$ will increase more rapidly than indicated under the assumptions by which Figure 45 was constructed. The relationships for sections with live bed conditions will have much lower slopes than those indicated, and hence might conform more closely to the Manning or logarithmic equations. When a combination of decreasing form resistance and decreasing particle resistance operates over a range of flow conditions the results may be very complex.

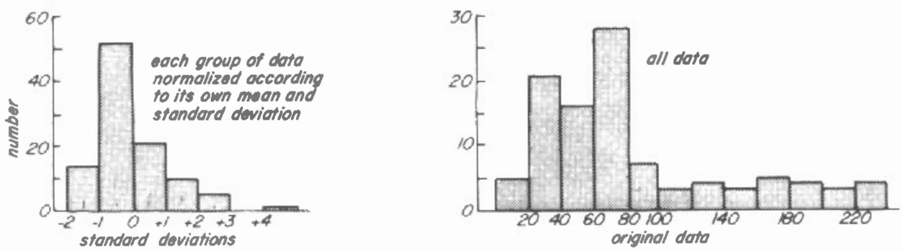


Figure 43. Distribution of w_s/d_* (all channels).

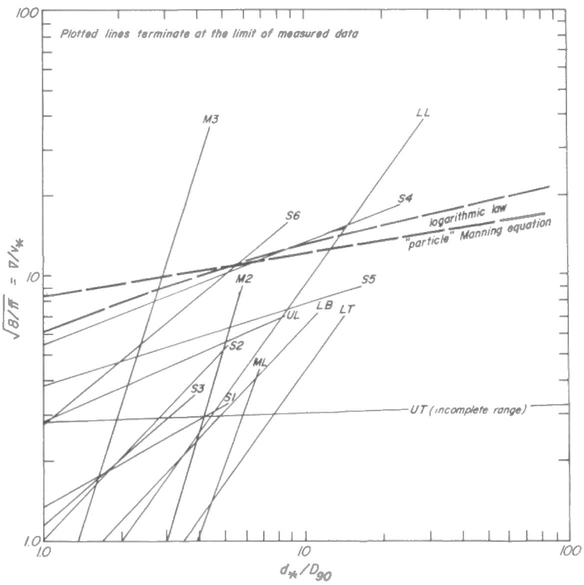


Figure 45.
The friction relationship in sandur channels.

It is nevertheless clear that, in some channels, extention of the indicated hydraulic geometry to high, but still feasible, flow levels results in flow resistance requirements smaller than those which can be attained in the surveyed channel. The "live bed" condition which pertains at high flow may allow the apparent limit set by equation 14 to be exceeded: nevertheless, friction cannot decrease below a smooth boundary limit. In order to consider likely conditions at very high flow levels in the sandur sections, Table 27 was prepared, showing the extropolated value of $\sqrt{8/ff}$ based on the observed hydraulic geometry, and limit values based on equation 14 using various assumptions about effective particle size. Seven of the fourteen channels are unstable according to the observed, low-flow criterion, four according to "D_{65} of transported material" criterion (which is reasonably conservative), and two according to the "sand" criterion (which appears to be unrealistically conservative). The results for the first three South River sections are interesting in this respect. All three are braided sections, and the channel-ways at all three sections are cobble lined and well defined; these do appear to be stable sections. This is in marked contrast to say, Middle 3, or the upper three South River sections, where very wide, shallow sections and the presence of actively moving gravel on the bed indicate a propensity for braiding. In those channels which are indicated as being neutral or unstable, bed changes occur at very high flow in order that the channel can continue to convey the imposed load.

In such a situation it is necessary to increase depth and/or reduce the velocity of flow, or to decrease the size of roughness elements at the boundary. This may be achieved in several ways:

- scour in the channel bed, so that greater depths are achieved;
- backwater development; usually this leads to overbank flooding;
- development of a smooth, "live" bed, with coarse materials moving in a dilatant state on a carpet of sand: the sand determines the effective particle roughness:

- selective scouring at certain points in the channel so that a deeper and narrower channel is created with the ability to sustain higher flow velocities; this normally occurs when the sediment load of the stream is too great to permit general scour and is accompanied by general deposition of material.

The last two adjustments are the common ones in rivers carrying very high sediment loads in noncohesive material. The last one results in the development of a wide, shallow section, with flow chutes (see p. 90), or channel braiding where a wide, shallow channel is divided into two (or more) steeper ones able to convey the water and sediment load in the face of the imposed boundary resistance. Experiments carried out by Shen and Vedula (1969) have shown that a channel undergoing aggradation in this manner may shoal and widen in such a way that the shear stress at the bed decreases with time. In such a situation the channel develops a tendency to braid. These morphological developments often emerge and become sharpened only on falling stage; at very high stage the channel continues to operate as a single channelway.

Brice (1964) in his study of braided sections of the Loup Rivers in Nebraska arrived at the same conclusion:

> "The effect of braiding . . . is to decrease the width of flowing water and . . . the changes in effective area and mean velocity accompanying braiding are similar to the changes accompanying flow from a wide single channel to a narrower single channel.
>
> ".... the effect of braiding on a wide channel is to render it hydraulically similar, at least in area and mean velocity, to a narrower single channel."

Maddock (1969) has arrived at similar conclusions from a general consideration of hydraulic geometries:

> "In natural channels . . . for a given discharge and with slope essentially a constant, within the range of adjustment of hydraulic roughness through changes in bed forms, filling will occur in high velocities and scour with low velocities . . . An increase in concentration of sediment requires a higher velocity to move the increased load. At constant discharge and width, this means a lower depth. If the slope is unchanged, the friction factor must decrease, and this decrease is associated with sediment deposition on the bed. A decrease in sediment concentration yields the converse of the previous situation. Velocity must decrease; hence, depth must increase. This is accompanied by scour which is associated with a greater factor.
>
> "If the friction factor is already at the lowest possible value, an increase in sediment concentration can only be accommodated by a continuous deposition of material, on the bed and an increase in the slope of the stream. It is practically impossible to increase velocity and depth quickly by decreasing width by natural processes. If the friction factor is already great, a decrease in concentration may result in a decrease in slope, but more likely the result will be an increase in width of the stream, braiding channels, or meandering (p. A67-A68)."

So long as the slope is fixed in a section, changes in discharge and sediment load must be accompanied by adjustments in the friction factor or geometry. Because of the limitation placed on the range of the friction factor, ff, a channel in noncohesive material which must pass a wide range of discharges of water and sediment cannot remain stable. Adjustments of slope and geometry occurring with high discharge commonly lead to braiding in such channels.

Downstream Characteristics

Downstream Hydraulic Geometry

The sandur rivers have no major tributaries within the reaches across the sandur surfaces. Hence, a "downstream" hydraulic geometry in the sense of Leopold and Maddock (1953) cannot be developed. However, for the six sections on South River, at Ekalugad Fiord, it is possible to study the mutual variation of the other parameters at a constant flow level. For this purpose a high flow of 100 m^3s^{-1} was chosen --- close to the maximum that could be passed through the normal channelways. The following groups of relationships were derived:[1]

(15a) $\bar{v} = 1.3q^{0.8}$ or (15b) $\bar{v} = 56.2w_s^{-0.8}$; $r^2 = 0.96$

(16) $\bar{v} = 3.38\ d_*^{2.1}$; $r^2 = 0.41$

(17a) $d_* = 0.65q^{0.2}$

(17b) $w_s = 26.6d_*^{-3.0}$ $\Big]$ $r^2 = 0.61$

These indicate that relatively narrow deep sections have the higher mean velocities. Wide, shallow sections, with lower mean velocity, are presumably zones of relatively high flow resistance, because of the greater wetted perimiter.

(18) $\theta = 0.0016\ d_*^{-4.1}$ $r^2 = 0.19$

(19) $\theta = 0.15\ w_s^{-1.7}d_*^{-9.1}$ $r^2 = 0.37$

(20) $\theta = 45.0\ D_{90}^{5.0}$ $r^2 = 0.72$

Slope tends to be greater in the shallow, low velocity zones (though the relationships are very poor); more definitely, greater slope is associated with larger bed material size.

(21) $\bar{v} = 11.5\ d_*^{2.6}\ \theta^{0.10}\ D_{90}^{0.2}$ $r^2 = 0.53$

(22) $\bar{v} = 2.2q^{0.8}\ D_{90}^{0.2}$ $r^2 = 0.97$

(23) $\dfrac{\bar{v}}{v_*} = 0.19\left\{\dfrac{d_*}{D_{90}}\right\}^{2.0}$ $r^2 = 0.68$

The resistance equation maintains a high exponent, just as in the sectional case, indicating its great sensitivity to changing relative roughness throughout the system.

---

1. Since $q=Q/w_s$, and since Q is constant, relationships involving w_s and q as the only different quantities are merely reciprocals of the same result.

Adopting an alternate approach of examining the correlations amongst the primary variables, the structure indicated by Figure 47 emerges, wherein the strongest relationship by far is between q and $\overline{v}$, confirming that most of the adjustment of the stream to changing conditions lies in velocity. The connection between the "conveyance" set (q, v, d_*, w_s) and the "resistance" set (θ, D_{90}) is made between depth and slope, though probably the underlying physical relationship is more properly between bed particle size and specific discharge, or velocity.

Long Profile Analysis

Detailed long profile surveys were carried out with stations at 30-or 45-metre (actually, 100-or 150-foot) intervals on both the Lewis-Triangle Rivers and on the three Ekalugad rivers. For the Lewis-Traingle streams, only streambed elevations were determined, but at Ekalugad Fiord both bed and water surface elevations, and a variety of additional data, were obtained.

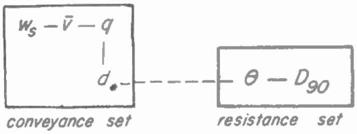
Initially the long profiles (Fig. 48) were analyzed to determine the trend of stream elevations. The Triangle stream and part of Middle River turned out to be essentially rectilinear, but all the other streams are concave in profile. An investigation of various fitting equations determined that satisfactory least squares fits were provided by piecewise fitting of a second order parabola, whose general form is

$$z = c_1 + c_2x + c_3x^2 \qquad (24)$$

At Lewis River, one equation satisfactorily described the entire reach. On the Ekalugad rivers, two reaches appear to exist with clear morphological distinction on the ground. In the case of Middle and South Rivers this occurs in the form of prominent zones of sedimentation and very wide, shallow channelways: upstream the rivers generally flow in single channels 1 or 2 metres below the sandur surface, whereas downstream the rivers flow, often braided, with the channels showing no incision at all beyond live channel depth. On North River the break occurs in a similar zone of sediment deposition just below the bottom of an entrenched section of the river where it cuts down through T2 terrace deposits.

In all cases the quadratic term in the fitting equation is small, indicating that the river sections are almost rectilinear (Table 28).

The problem of concavity in long profiles has been broached by many workers. The earliest functional forms of the profile to gain widespread attention were exponential (Shulits, 1941) and logarithmic (Jones, 1924) forms. The logarithmic form was entirely an empirical invention in its early application, whereas the derivation of the exponential form was based on Sternberg's (1875) observations of the exponential decrease of particle size downstream and its relation to long profile form. Yatsu (1955) gave a rather convincing demonstration of Sternberg's effect in homogeneous reaches (i.e., reaches with no major tributaries or other major hydraulic changes). A significant difference in interpretation underlies the two forms: the exponential function is asymptotic with respect to the distance; the logarithmic, with respect to elevation. The former implies restricted range of elevation controls the profile; the latter, restricted distance.



Correlation Matrix

	w_s	d_*	$\bar{v}$	q	D_{90}	θ
w_s	1.00					
d_*	-0.78	1.00				
$\bar{v}$	-0.98	0.64	1.00			
q	*	0.78	0.98	1.00		
D_{90}	0.07	-0.43	0.03	-0.07	1.00	
θ	0.02	-0.33	0.08	-0.02	0.85	1.00

* $q = Q/w_s$ and, since Q = constant, q and w_s are reciprocal measures of the same variable. Correlation coefficients involving q and w_s in the table are complements of each other.

Figure 47. South River correlation set of hydraulic variables.

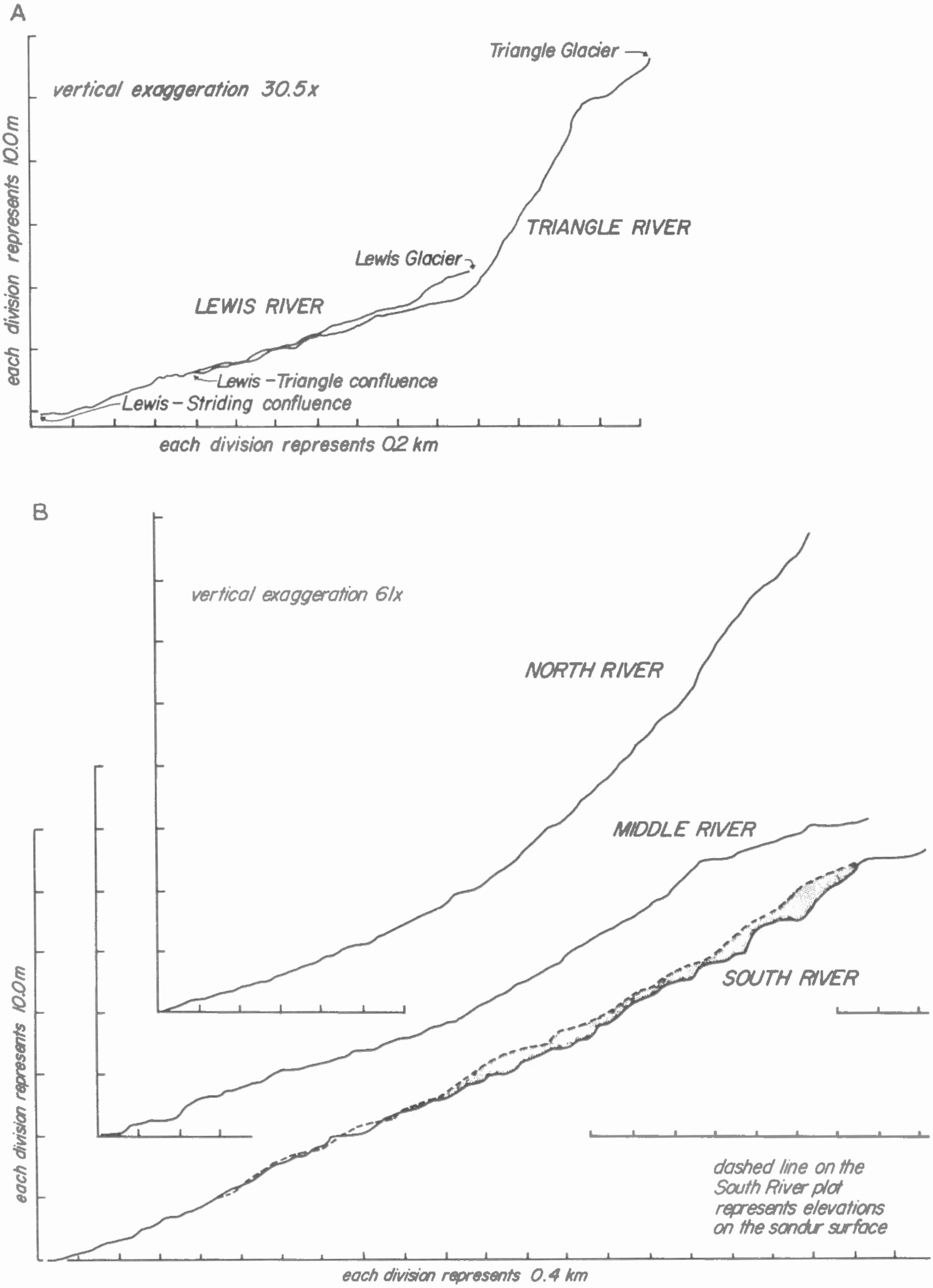
Mackin (1948) made the observation, long obvious to engineers, that a graded channel should develop a straight profile through a homogeneous reach, since all sizes of sediment will be moved right through, albeit at different rates. "Sorting" of bed sediment size, which might appear over very long distances, would be restricted to the Sternberg effect. On the other hand, in an aggrading stream, coarse materials would be continuously deposited in each reach, and would remain there, so that real sorting could occur downstream and a concave profile would result. Work by Culling (1960) and Devdariani (1967) produced models for long profiles that were dominated by exponential terms.

The present situations are particularly simple, since the rivers are hydraulically homogeneous throughout the reaches of interest. However, constraints are placed on the development of the profile in the form of imposed end points. The upper end point is determined by bedrock outcrop in the floor of the channels, or by the location of a profile discontinuity. The lower end point, though it may be moved by sedimentation (or by erosion, in the case of the profile discontinuity) over a period of time, is probably essentially fixed by sea level (Ekalugad) as far as short term adjustment to the water-sediment load is concerned.

Consider the stationary case of a stream reach moving a water-sediment load from point A(x, z) to point B(x + dx, z + dz) under steady flow conditions through a prismatic reach. The total load in the reach (water plus sediment) is w(x). The slope in the reach will adjust to the applied load in such a way that $\frac{dz}{dx} \propto w(x)$, if we suppose that total resistance does not change drastically (it is, in any case, a conservative value in the resistance equations). More particularly

$$(25) \qquad \frac{d}{dx}\left(\frac{dz}{dx}\right) \propto \frac{dw\,(x)}{dx}$$

if w(x) is a constant (i.e., all the material carried in the river passes through the reach), then the right-hand side of the equation is zero and the solution is



A. Lewis and Triangle Rivers B. Ekalugad sandur rivers

Figure 48. Longitudinal profiles of sandur rivers.

a straight line profile. However, if w(x) decreases downstream in proportion to $e^{-\beta x}$, which will fit the Sternberg effect, or Mackin's proposal for an aggrading stream, then

(26) $$\frac{d^2z}{dx^2} = Ae^{-\beta x};$$ A being a constant of proportion.

The solution of (26) is

(27) $$z = \frac{A}{2} e^{-\beta x} + C_1x + C_2$$

Under the conditions $z = 0$ at $x = x_1$ and $z = z_1$ at $x = 0$,

(28) $$C_2 = z_1 - 1$$

(29) $$C_1 = \left[1 - z_1 - \frac{A}{2} e^{-\beta x_1}\right] \Big/ x_1$$

This was the solution arrived at by Culling (1960), beginning from more general assumptions. Equation (27) represents an exponential decay function superimposed upon a straight line.

If we consider the Taylor's series expansion of $e^{-\beta x}$, we have

(30) $$e^{-\beta x} = 1 - \frac{\beta x}{1!} + \frac{\beta^2x^2}{2!} - \frac{\beta^3x^3}{3!} + \ldots\ldots$$

and substitution into the solution reduces it to the form

(31) $$y = A + Bx + Cx^2$$

if we ignore terms in the expansion above that for x^2, which is justified so long as β is small.

The straight profiles of Triangle and lower Middle and lower North rivers indicate that they are competent to move all the sediment delivered to them, generally, and that they are not, on the whole, aggrading. The other profiles are apparently not steady profiles. The upper profile segments on the Ekalugad rivers show no signs of general aggradation today: they are well defined, cobble paved channels, slightly incised below their "flood plain" (though areas of instability occur, as S4 reach on South River). North and Middle Rivers, at least, are certainly stable in these reaches. The present form may be the result of armouring the original aggradational profile with slight degradation involved in the process. In these cases, regrading toward a rectilinear profile has been arrested by the heavy armouring.

Entrechment in the proximal regions appears not to be an unusual. situation on active depositional surfaces. Arnborg (1955) reported a similar morphology on the Icelandic sandurs, McDonald and Banerjee (1971) have described the feature on sandurs in the Canadian Rocky Mountains, and Hooke (1967) has noted its occurrence on alluvial fans. Whilst this may reflect changes in stream regimen or sediment regimen (and such changes could certainly be argued for the Baffin sandurs within the past 2 or 3 centuries) it is also possible that the effect is entirely normal. Near the proximal end of the depositional surface, the general level may be determined by the concentrated effects of large floods which flow outside the channels. Farther down, divergence of flows onto the depositional surface dissipates the concentrated volume of large floods by comparison with the still largely channeled flow of lesser events, so that they come to the same level. At this point the more normal floods may also diverge from the channelways. Only direct observation of a wide range of floods will discriminate amongst these hypotheses.

The persistence of two profile segments in the Ekalugad rivers also indicates a relative stability of profile form. In a highly active depositional environment, one might expect such discontinuities to be rapidly eliminated. However, the transition from one reach to another, occurring in zones of apparent active sedimentation in all three rivers, indicates that current sediment transport mechanics is of importance in maintaining the distinctness of the two sections.

The residuals from trend fitting (Fig. 49; Lewis River characteristics not illustrated) were submitted to serial correlation analysis: the correlograms (Fig. 50) suggest Markov-type dependence, though none of them are particularly neat. In respect of "information transfer" along a river it appears reasonable to expect Markov relations. Information may be passed downstream, via the water flow, or it may be passed upstream via waterlevels, provided that the flow remains subcritical.

Upstream control may also be termed "channel control", while downstream control is familiar as "base-level control". Because of the existence of rapids in these rivers with locally supercritical flow at high levels, it is not likely that downstream control can be truly effective for a great distance. If we consider the measurement stations to be numbered consecutively upstream from the bottom of the surveyed reach (the sea at Ekalugad Fiord), then for upstream control.

(32) $$x_i = r_1 x_{i+1} + \overline{x} + \xi$$

where $\overline{x} \sim 0$ for series made up of fluctuations about the mean. Ignoring $\overline{x}$, we see, by iteration, that

(33) $$x_i = r_1^{\,2} x_{i\pm2} + \xi;\; x_i = r_1^{\,3} x_{i\pm3} + \xi;\; \ldots\; x_i = r_1^{\,n} x_{i\pm n} + \xi$$

As $r_1^{\,n} \rightarrow 0$, the effective transfer of information from $i\pm n$ to i also approaches 0, representing the limit of information transfer. Relationships are given in Table 29. In all cases, the regressions are highly significant and the residuals from regression are not significantly correlated.

The same procedure was carried out with slopes, and with changes in width and thalweg direction (Fig. 49 and 50). The serial correlation of the residuals remained significant, in general, for slope and thalweg direction, indicating <u>independent</u> influence on their value over distances of greater than 1 station length (Table 29).

The extent of information transfer was ultimately estimated by the number of steps necessary to reduce $r_1^n < 0.1$. Most effects show effective transfer lengths of 5 to 9 steps (225 to 425 m or 750 to 1,350 ft) at Ekalugad Fiord, though some effects, especially width and thalweg deflection, tend to be far more conservative. Width is the most conservative aspect of channel geometry, while the range of thalweg deflection is restriced by the direction of valley slope so that it will appear as a relatively conservative variable.

The correlograms of residuals after Markov fitting are relatively irregular. Periodicity in the residual series is suggested in most of the cases however; there is a significant number of significant coefficients in the correlograms for slopes and thalweg deflections, and in the case of North River at Ekalugad Fiord, for bed and surface elevations. Power spectra were generated (Fig. 51) in an attempt to detect the significant frequencies in

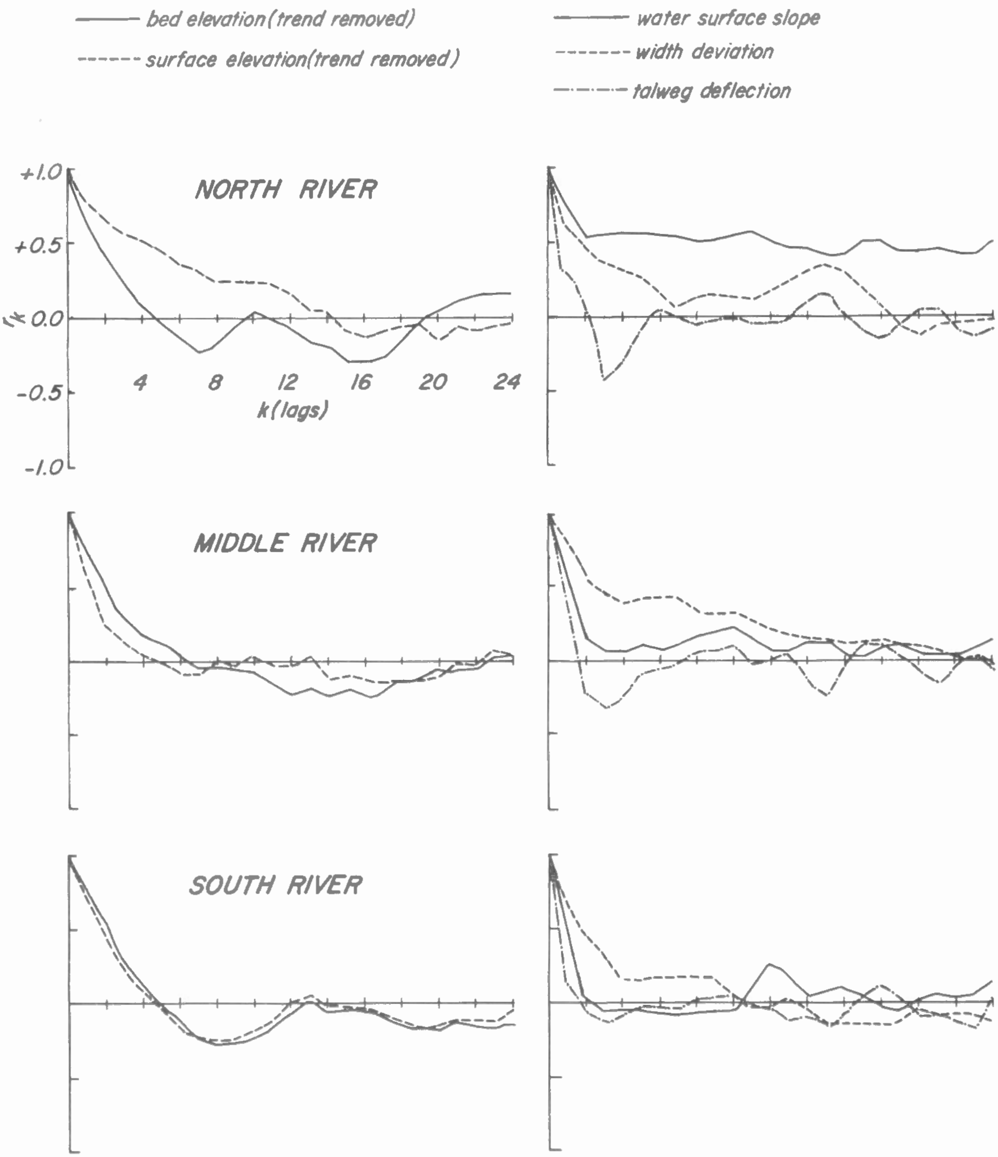


Figure 50. Correlograms of long profile characteristics of Ekalugad sandur rivers.

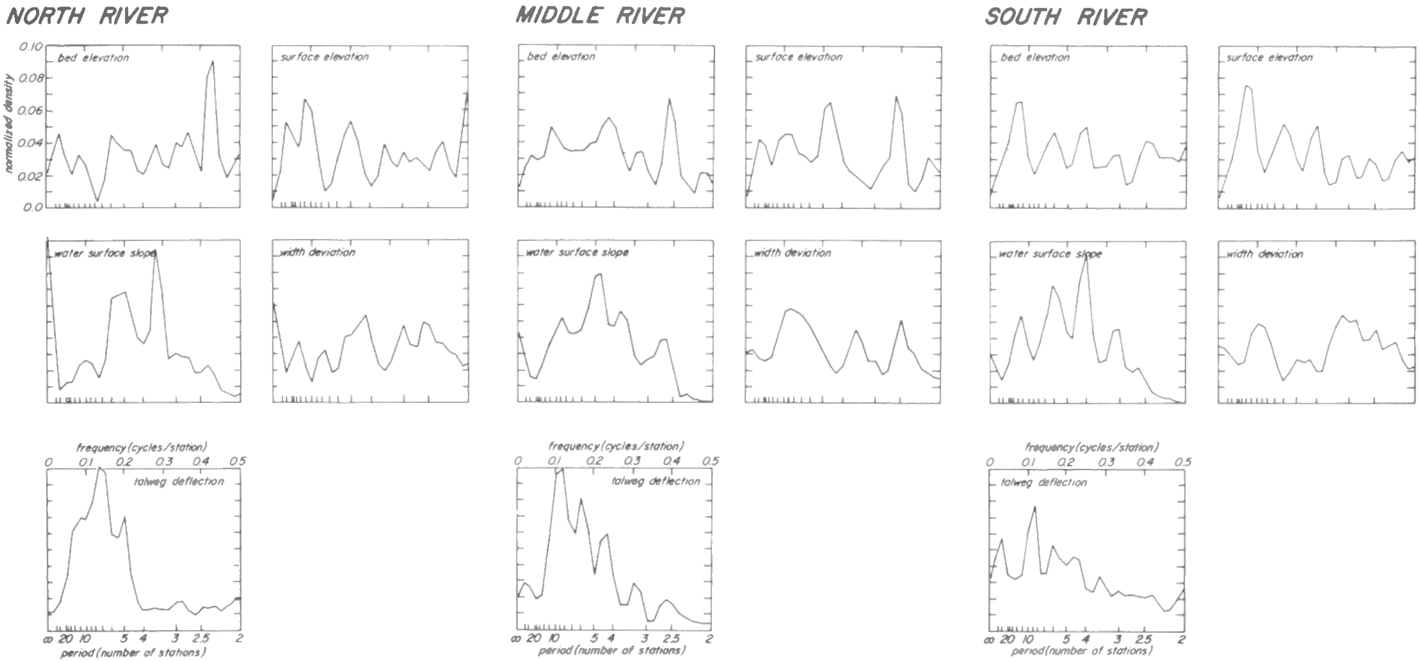


Figure 51. Spectral characteristics of rivers at Ekalugad Fiord.

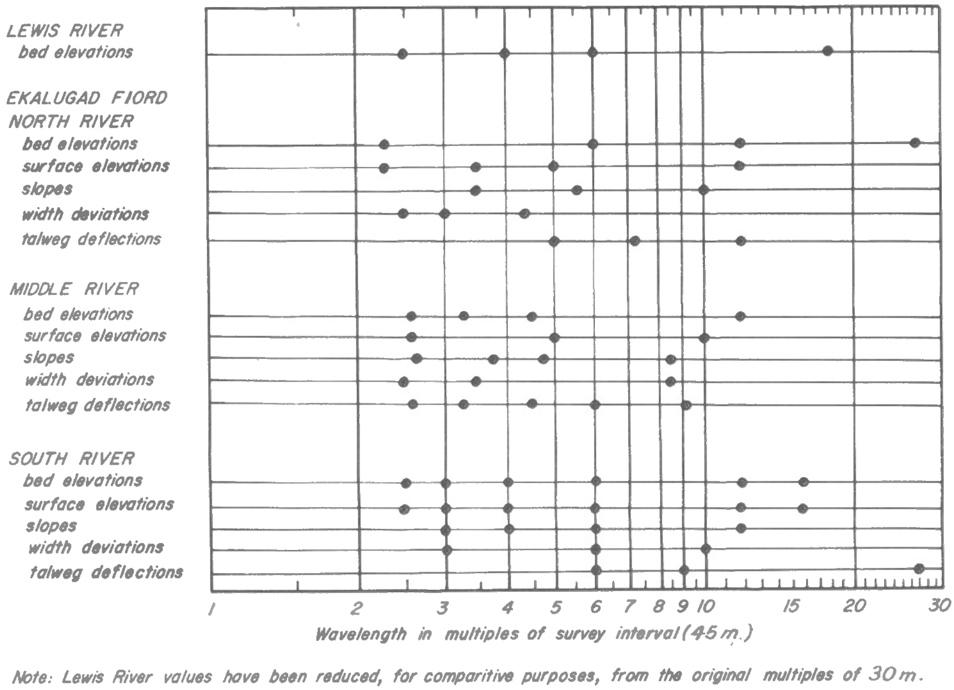


Figure 52. Location of variance density peaks in longitudinal profile spectra.

these residual series. Variance density peaks are shown in Figure 52, in terms of wavelength (stations). A dominant pattern of peaks occurs in every case, corresponding to recurrent phenomena at short, intermediate and long wavelengths. On the Ekalugad Fiord area rivers a peak in spectral densities occurs for profile elevations, slope and width at between 2.3 and 3.0, and for thalweg between 5.0 and 6.0. (Thalweg deflection peaks would be expected to be only half as frequent as other factors, if they are associated, since a half-wave in the thalweg pattern will correspond to a full oscillation in the rest.) An intermediate peak occurs at between 3.5 and 5.0, with thalweg peak at 7 to 10. On North River a new peak occurs at 5 to 6 and 12. These intermediate peaks are usually the most prominent. On South River a long wavelength peak occurs at 12 to 16, with thalweg peak at 27. Whilst this periodicity is not clearly developed in the other rivers, minor peaks in bed elevation suggest that it might be present.

Distances corresponding to these station-number wavelengths are given in Table 30, along with the ratios of wavelength to mean channel width for the Ekalugad Fiord area rivers. The strong intermediate wavelengths appear to represent the dominant relationships (5-7 widths) noted by other investigators (see Leopold, Wolman and Miller, 1964).

In order to investigate the cross-correlations amongst the various characteristics of long profile more closely, cross-spectral computations were carried out. The cospectra indicated that in all cases except those involving thalweg deflections virtually all of the simultaneous covariance of the variables occurs at very long wavelengths: six stations length, or about 275 m seems to be the lower limit of such interaction. In fact, most of the covariance is contained in "periodicities" approaching the entire range and there are no well defined finite periodicities. This is especially so in North River. The obviously positive relationships amongst the elevation variables and slope were the only ones whose coherence (measure of coordinate variance reduction at corresponding wavelengths between the two series) was significant (see Panofsky and Brier, 1968, for methods).

Cross-spectral phases (indicating the lag between corresponding wavelengths of two series) showed, as expected, that bed and surface elevations maintain good phase relationships. Slope exhibits a lag of just over 1/4 cycle behind the elevation data (i.e., -0.3 of a circle).

Relationships involving width and thalweg deflection were generally ill-structured and not significant.

Bed Material and River Morphology

The character of the bed material should bear important relationship to other aspects of the river morphology. In order to investigate this proposal, a special set of observations was carried out on South River at Ekalugad Fiord, with stations every (137 m) 450 feet, where, in addition to the other sample measurements, a sample of 50 cobbles was taken from the streambed (Fig. 49c) using Wolman's (1954) method.

Cross-spectral characteristics of mean cobble size with respect to the other long profile data are plotted in Figure 53. The cospectra indicate that positive correlation occurs with the elevation series and with slope, becoming most pronounced in the short wavelengths. This presumably reflects the tendency for large stones to be found on riffles, where the residual elevations are positive and the energy grade is steep. A negative relationship is indicated with width - narrow parts of the river are paved with large cobbles since such sections tend to have high velocities and high specific discharge. The relationship with thalweg deflections is not clear, though there is a fairly large positive covariance at intermediate wavelengths, suggesting that areas of more extreme flow curvature are ones of high bed stress. These are presumably areas of high energy dissipation in all respects.

The cross-spectral phase varies widely. At long wavelengths the cobble series appears up to 1/4 cycle in advance of elevation fluctuations, but at short wavelengths it falls behind. The relationship with slope is in good phase correspondence over the entire range. For width, no clear relationship appears, though in the area of greatest covariance the phase relation is good. For thalweg the relationship is similar to that for elevations.

The lower limit of statistically significant coherence is 0.57. This value is exceeded in the middle wavelengths for the elevation relationships, and in the short wavelengths for slope.

Channel Form and Evolution

Form and Occurrence of Channel Bars

Spectral analysis of the channel long profiles has established the presence of a suite of oscillatory elements in the channel beds of sandur rivers. Previous workers (cf. Leopold, Wolman and Miller, 1964) have indicated that meander lengths tend to vary between 11 and 16 channel widths, providing a spacing of 5 to 7 widths between crossover points. Similarly, the distance between successive riffles in straight or irregular channels has been reported as 5 to 7 widths, as has the spacing of gravel concentrations in sand bed arroyos of the U. S. Southwest (Leopold and Miller, 1956). Maddock (1969) has demonstrated remarkably clear sets of alternating sand banks in rectified sections of the Colorado and Rio Grande Rivers, with a spacing of $2\pi w_s$. These spacings correspond to that of the "intermediate elements" which account for the greatest portion of variance in the Ekalugad and Lewis channels.

A wide range of miscellaneous observations has been made on the spacing of macroform elements and their relative sedimentary properties. Wertz (1963) reported finding alternating zones of boulder accumulation and sand fill in the mid-course sections of some Arizona mountain-front arroyos. Neill (1969) has described large "waves" or bars in the bed of Red Deer River, with wavelength approximately $2w_s$ (he took all major forms on echo sounder traces into account). Wolman and Brush (1961) and Stebbings (1964) have described large "sheet" waves of similar (scaled) dimensions in flume experiments with coarse sands. Znamenskaya (1962) has repeated the description for natural channels.

Speight (1965, 1967) carried out spectral analyses of meander wavelengths on Australian rivers and found that a great variety of spectral peaks occurred. The dominant one (in terms of variance reduction) occurred at a wavelength of about $28w_s$ corresponding to $14w_s$ for the riffle series), and was not that which probably would have been chosen by visual inspection. On the basis of this evidence, and that from the Baffin sandur streams, it is safe to conclude that a range of macroform wavelengths exists in natural channels.

The dominant elements in the sandur channels are comprised of a sequence of pools and riffles which become particularly obvious in the upper, steeper portions of the channels. The riffles comprise bars which take one of several characteristic forms:

- right diagonal bars (flow is diverted to the right across it) (Figs. 54 and 56);
- left diagonal bars (flow is diverted to the left across it) (Figs. 54 and 57);
- "spool" bars (or diamond bars) (so-named by Krigström, 1962) (Fig. 60).

The former types appear in various configurations in straight channels and in channel bends (Krigström, 1962), and contribute significantly to lateral erosive activity in the stream by deflecting the flow from a straight, down stream course. The latter type, which contains superposed elements of both the preceding kinds, is particularly associated with braiding, since down-cutting around the bar results in the development of two channels and leaves a central island (see Brice, 1964). Leopold and Wolman (1957), in providing excellent descriptions of the development of such bars both in the field and in the laboratory flume, designated them as the basic unit of channel anastomosis. Krigström (1962) showed how complex braiding patterns can be built up by superposed spool bars on successively larger scales and Williams and Rust (1969) suggested a hierarchical division of bar components. This is typified by the braiding pattern of the Lewis River (Fig. 63.)

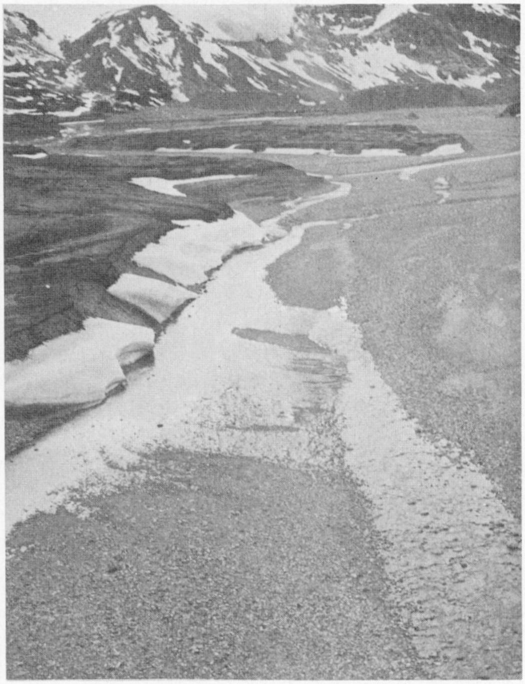
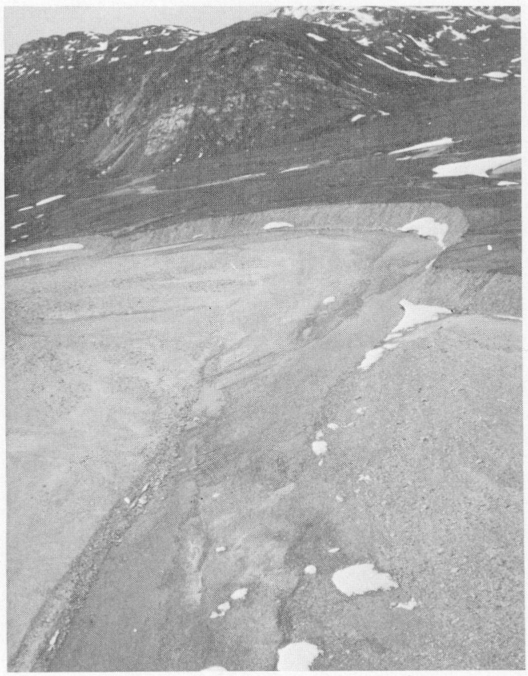


Figure 54. Detail of Figure 56

A. Right diagonal bar; this is the "upper South River diagonal bar" surveyed in detail (Fig. 56). Cross-section S6 is just upstream from the bar. Very low flow.



B. Left diagonal bar; right diagonal bar behind. The bars normally occur in alternating order. The angle of the bar across the channel suggests that reflection of standing waves at high flow may be involved in their form and maintenance. Very low flow.

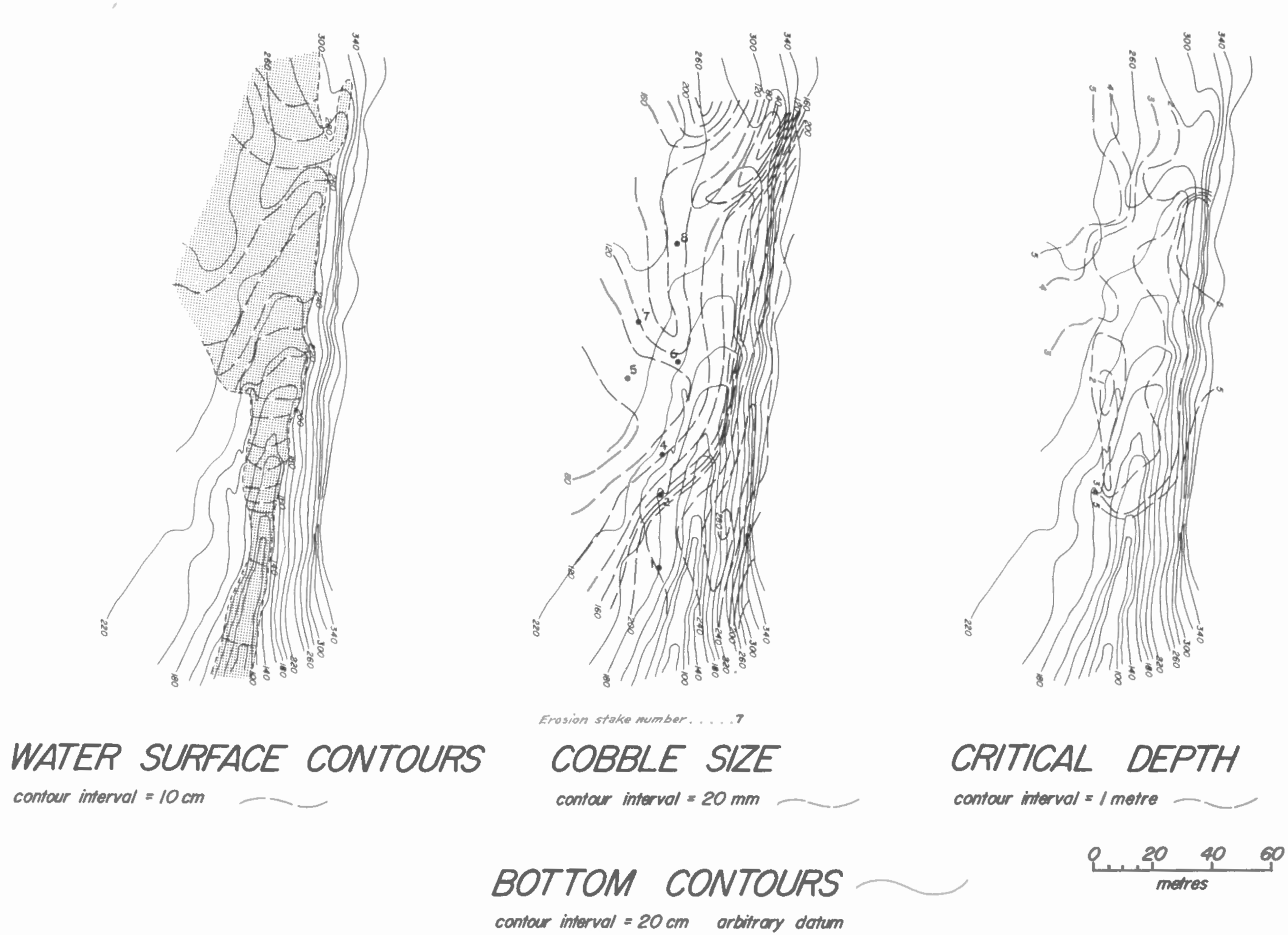


Figure 56. Diagonal bar in Upper South River: survey results. (Positions of erosion-deposition stakes are shown on the first bottom contour map - <u>see</u> Table 32 for results.)

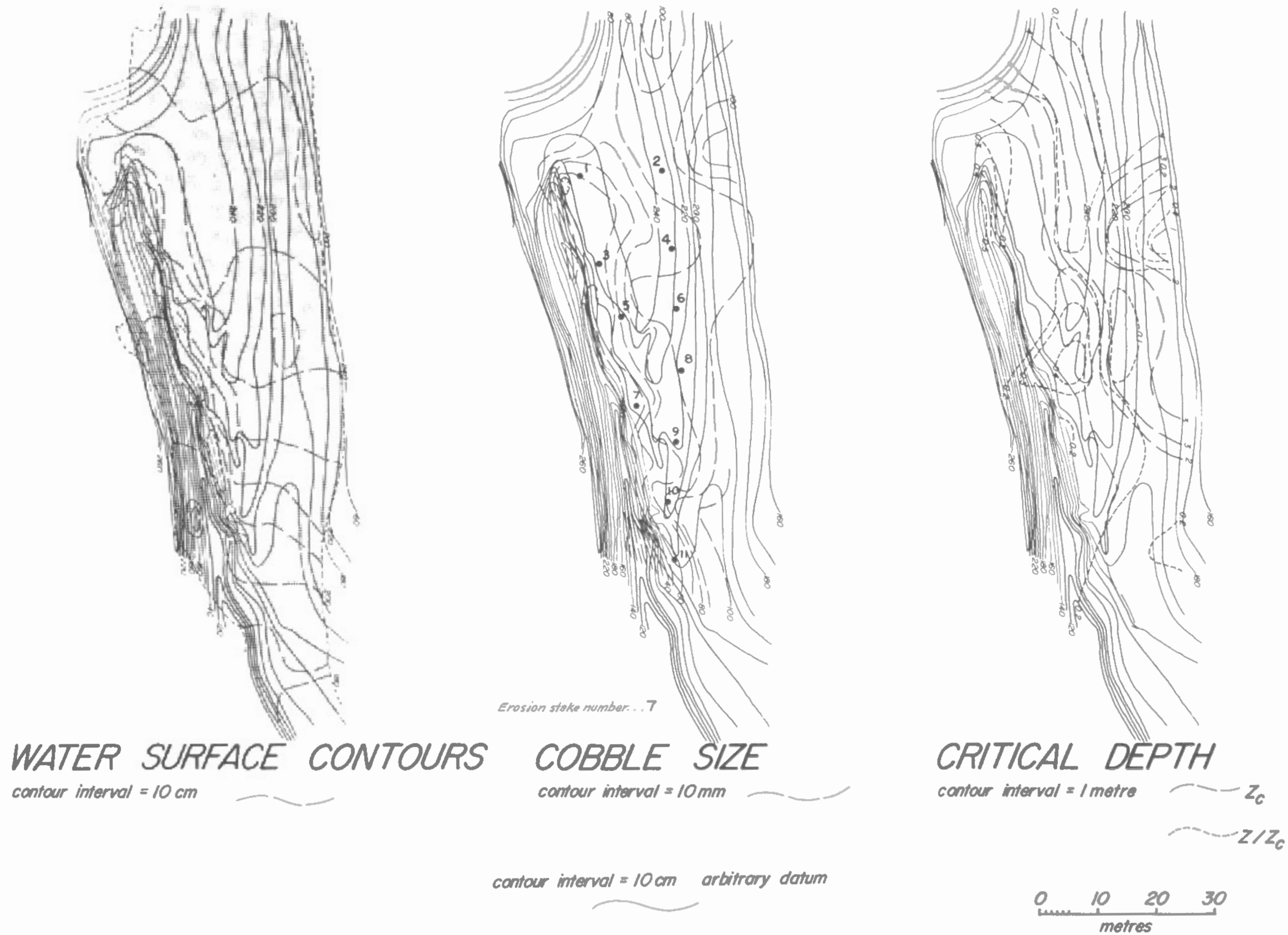


Figure 57. Diagonal bar in Middle River: survey results. (Positions of erosion-deposition stakes are shown on the first bottom contour map - <u>see</u> Table 32 for results.)

The diagonal bars, which are the more usual riverine form, may develop from the spool bar when one of the branches comes to completely dominate the other (see Figs. 34 and 35 in Leopold and Wolman, 1957), though development in the opposite direction can occur as well (Fig. 34 ibid.). Diagonal bars may be associated with the development of meanders when they are found, as is often the case, in successive alternating form downstream. In the sandur channels, however, they provide much more abrupt flow deflections than would be associated with true meandering.

Occasionally, a riffle is situated squarely across a stream as well, most often where a pool spills into a major set of rapids. Such locations are usually controlled by the presence of very large boulders and are probably not determined by normal fluvial processes, as are the other riffle types.

The occurrence of the major bars is mapped for South and Middle Rivers at Ekalugad Fiord from large scale airphotos (Fig. 55): there are 43 such features in 8.70 km on South River, and 27 in 5.94 km (up to the major pools at the entrance to Middle Valley proper) on Middle River. The mean intervals of recurrence are 200 m and 220 m respectively on the two rivers, which corresponds to the recurrence of the strong oscillations detected in the spectral analysis. The succession of right and left diagonal bars is obviously not random.

Whereas the material comprising the bars moves during floods, the bars themselves appear to be relatively stable. In Figures 54 to 60 a variety of bars are shown. In addition to these major "channel element" bars, several other types are found along the rivers which are more strictly depositional in form, including point bars in channel bends, lateral bars and triangular bars at channel confluences (Krigström, 1962). These are slack water locations where sediment load tends to be deposited during the falling stage. Lateral bars, in particular, (a small one is shown in Fig. 36) are sites of temporary storage of fine material that moves in suspension. These bar forms may have local significance in hydraulic conditions (large lateral or point bars may serve to reduce flow width), but they are largely scoured out during high flows.

Hydraulic Conditions in the Pool and Riffle Sequence

Among the survey maps of channel bars presented in Figures 56 to 61, particularly interesting are the maps showing critical erosion depth $z_c = \tau_c / \rho g \theta$, from tractive force criteria. To compute τ_c, the relationship

$$\tau = 1.8\overline{b} \tag{34}$$

was used, where $\overline{b}$ is the mean b-axis diameter (mm) derived from samples of 10 (sometimes 25) large cobbles picked up at each sample site (the largest cobbles present on the stream bed were sought for sampling). The relationship derives from equation 6, with the entrainment function $1/\Psi$ set equal to 0.118 (see Table 18) and so is not the same as that normally derived from the Shields' relation. z_c is characteristically lowest along the bar edges where θ (slope) is greatest. The ratio z/z_c is also greatest here (z being the flow depth at the time of survey; the particular stage has no physical significance, but it provides an arbitrary uniform flow datum). It appears that, as stage rises, erosion occurs here first, and that in flood flows the lip of the bar is largely scoured away, so that the water moves across the bar on a more uniform slope. With declining flow material is again deposited along the bar edge. The overloose bed state encountered along the leading edge of many channel bars after flood flows confirms this situation.

Sedimentary deposits in the channel, such as the riffles, act as flow "controls". Small deposits have little effect, since the contraction in flow area is small: however, as they grow in size the contraction in flow area increases until the available specific energy ($E = z + \frac{Q^2}{2gA}2$, where z is the depth of flow referred to the lowest point in the section) is just sufficient to pass the flow through the contracted section: flow will be critical. If the deposit develops beyond this, specific energy must increase further: this is done by increasing $z = z_c$ and produces an upstream backwater. Flow across the riffle at the contraction is supercritical: energy is released in flow resistance across the riffle, and in a plunge pool at the lower side of the riffle, or in a hydraulic jump downstream after a period of supercritical flow.

Of the cross-sections studied in the sandur rivers (see p. 69), 8 were on riffles and 6 were considered to be chutes below riffles, or in pools. Table 31 gives the mean values of exponents in each situation. On the riffles depth and width tend to increase relatively more quickly than in the pools, at the expense of the relative rate of increase in velocity. The downstream control at the front of the riffle, and associated backwater curve, as opposed to the general channel control below the bars, probably leads to this result. Resistance appears to decrease less rapidly on the riffles than in the pools. Keller (1971) noted the same effects, and used them to forward an explanation of bed material size sorting. However, the entire measure of differences in the present case is provided by the two extreme Middle River data.

The important implication of flow over a riffle or bar is the effect on sediment transport. As flow approaches a bar from upstream, the channel becomes shallow. Accordingly relative roughness increases; flow velocity may also decrease in the area behind the riffle (cf. equations 15 to 17). Sediment being carried as bedload will be deposited, then. At the same time the channel usually becomes considerably wider in an attempt to maintain flow area. This is a zone of flow divergence and decreasing rate of specific energy expenditure. In this situation boundary resistance becomes very great. From the Darcy-Weisbach equation we have

(35) $$v = \sqrt{\frac{8gd_*\theta}{ff}}$$

i.e., for a given ff, the most efficient channel is a deep one. It appears, then, that the deeper channel below a bar is more efficient than the wide, shallow one above, so that once the bar is established, the channel will "draw down" water across it, into a zone of flow convergence where the rate of specific energy expenditure increases greatly. On the bar face, where flow becomes supercritical, erosion may occur and a chute, or trench will begin to work upstream along the bar. In this way characteristic "bar and chute" structures form (see, also, the conclusions on p. 74). At higher flows the critical point will move upstream across the front of the bar, and general erosion of the bar lip will occur, as suggested above.

The growth of a chute below or beside the bar (Wertz, 1963) is the key feature of the development of a distinctive morphological form. At a diagonal bar the chute develops toward one side of the channel, whereas at a

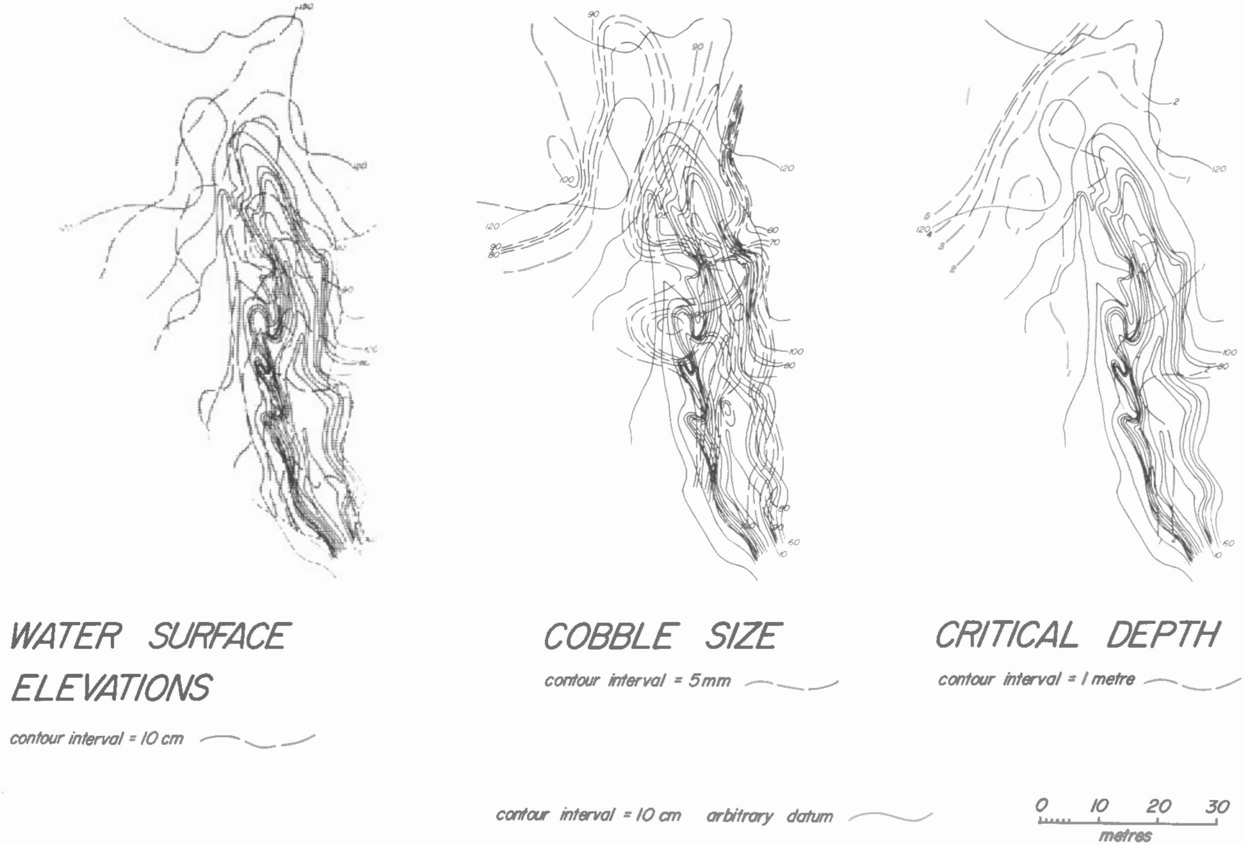


Figure 58. Chute in Middle River, photographed at very low flow.

Figure 59. Chute in Middle River, photographed at very low flow.

spool bar a chute develops on either side. Minor chutes then develop through the edge of the bar, expedited by the drawdown phenomenon, some of which may later become major channelways. It is often not clear whether the chutes are erosional, or merely nonaggraded channel areas downstream of a depositional bar front (Stebbings, 1964). To judge by the cobble pavement that floors them, the latter is the likely case in many of the instances on the Baffin sandurs. However, these areas maintain themselves by active scouring of any sediment that moves into them.

In late 1967 and 1968 an erosional chute began to develop in the large "sedimentation zone" of Middle River and turned the entire area into a right diagonal bar (Fig. 59). This system has been mapped (Fig. 58) and shows the same characteristics as the other bars.

Bars as Storage Areas of Material

Channel bars store sediment in transit. The material found on bar surfaces was usually smaller than the bed pavement found in the channel locally (see Nordin and Beverage, 1964, for comments on similar conditions in a sand bed river). The stability of the bars in many channels suggests that they are in general equilibrium with the flow conditions of water and sediment.

Several bars were observed at Ekalugad Fiord by means of iron stakes consisting of 1 1/4 cm diameter concrete reinforcing rods, driven about 75 cm into the bar, so that changes in the bar surface could be measured. The results of these studies are presented in Table 32. While different periods are characterized by aggradation or degradation, the overall form of the bars over a relatively long period has not changed greatly. Nevertheless, the surface may change completely in detail in some cases, as in the Middle River "sedimentation zone" between 1966 and 1967 (the <u>mean</u> change in surface elevation between the two surveys was only -2.74 cm; see also Fig. 61). The short term record from the Middle River left diagonal bar is particularly interesting: the apparently great degradation between 1967 and 1968 nearly all occurred on the lip of the bar, whereas there was no net change in the main channel area behind and across the bar.

An indication of the magnitude of the stress brought to bear on the bed is provided in the fact that at this site nearly all the rods were severely bent over. Whether the rods were deformed by the stress of fluid transport, or by impulsive blows from moving cobbles near the bed, is not known.

Although large volumes of material are set in motion by high flows, it is significant that individual cobbles in the bed load do not appear to move very far at any one time. This fact, recently emphasized by Tricart and Vogt (1967), is demonstrated by the high recovery rate of marked cobbles in the competency tests carried out on the sandur. Similarly Emmett and Leopold (1965) pointed out that, although scour seemed to be a general rule on a channel bed during flood, material moved only short distances, so that sediment discharge was far less than the generality of movement would suggest. Hence, much material is stored for long periods in the channel.

The nature of individual cobble movements during flood flows might lead to some understanding of the rate of material movement through the bars (and, incidentally, to an estimate of bed load sediment discharge). For the Baffin rivers, however, the results of observing the movements of marked cobbles were very irregular (Fig. 62). It appears that once dislodged, individual cobbles maintain their progress over the bed until caught

Spool bar at very low flow.

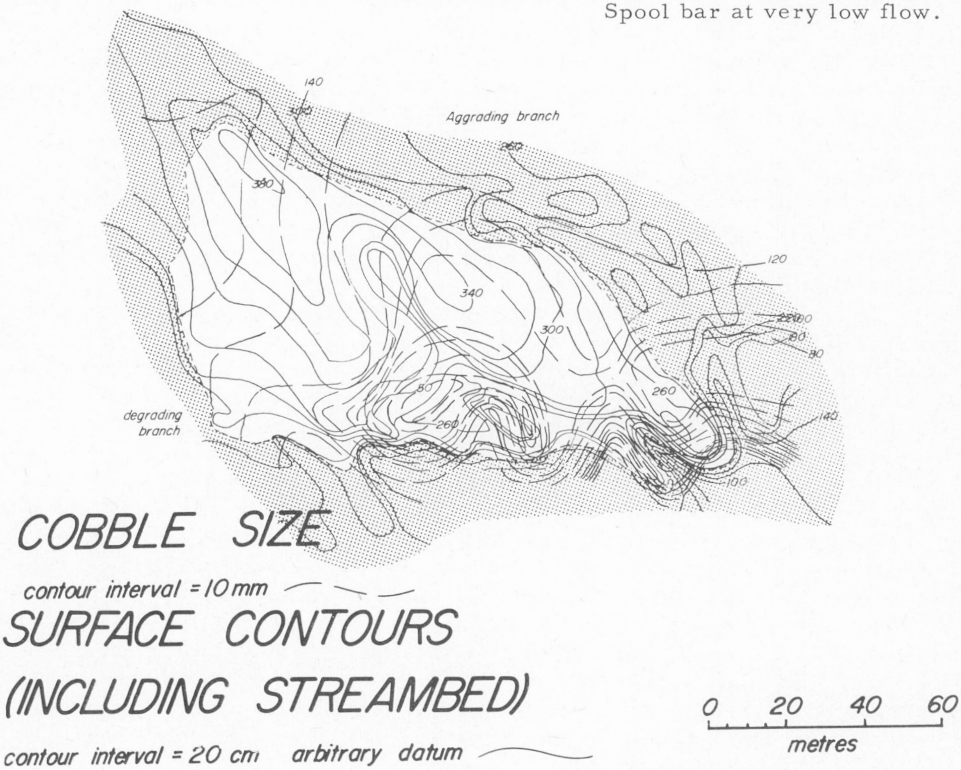


Figure 60. Spool bar in North River: survey results and photograph. The north side (near) channel is smaller than the other, and may eventually be captured. The bar would then be converted into a left diagonal bar.

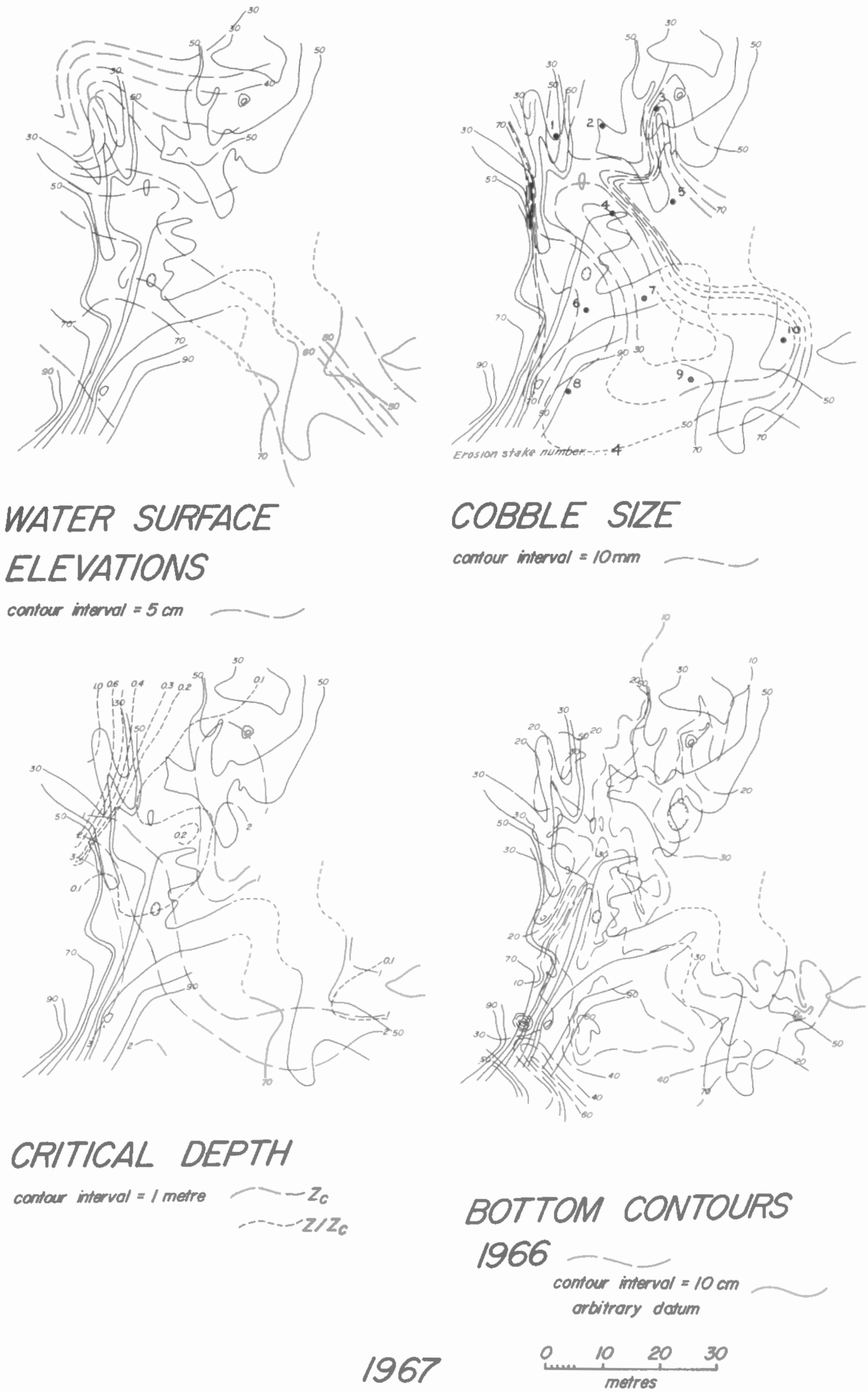


Figure 61. Sedimentation area in Middle River, survey results. (Positions of erosion-deposition stakes are shown on the first bottom contour map – <u>see</u> Table 32 for results.)

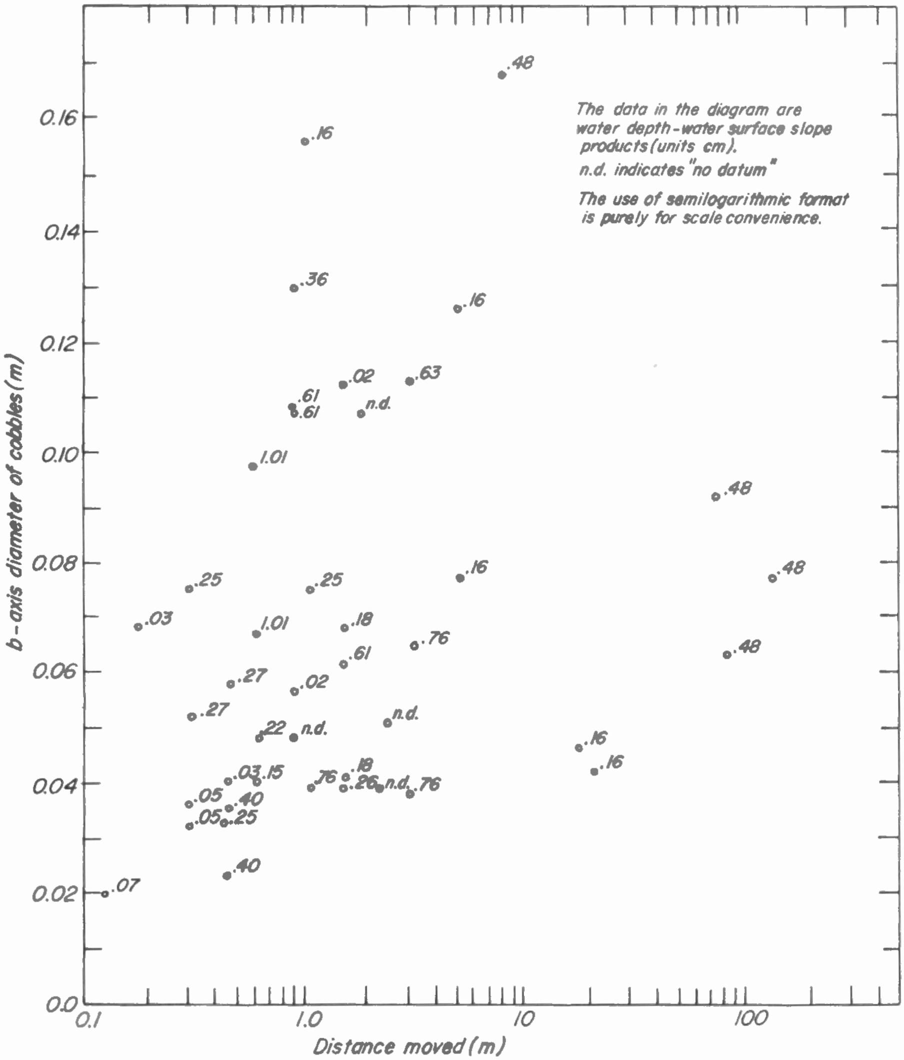


Figure 62. Scattergram of grain size versus distance moved in sandur streams: data derived from competence measurements made at Ekalugad Fiord. The water depth-water surface slope products given in the diagram are directly proportional to tractive force.

by a protruding cobble. The result is a highly imbricated bed structure. Such interception appears to be random and the pattern of individual movements is similarly random. Nir (1964) also reported inconclusive results from a study of cobble movements in ouadis. Apparently it requires a very large experiment to determine if any meaningful pattern does exist in cobble movements, and only one such experiment has ever been done (Leopold, Emmett and Myrick, 1966; Langbein and Leopold, 1968), in a desert arroyo. Two interesting results appeared:

- there is an apparent tendency for cobbles to be less mobile when they are densely distributed on the bed, as they are in riffles;
- movements tend to be from riffle to riffle

The Ekalugad data show that individual movement lengths are much less than the distance between the dominant bars, so that several displacements probably occur in each bar-to-bar journey. However, this is not important: the second conclusion is more or less implied (over the long term) from the first. It is possible, too, that greater regularity of cobble movements would show up in high floods, when a heavy sediment charge moves in a dispersed state along the bed. Such events were not discriminated from lesser ones in the Ekalugad data.

Langbein and Leopold (1968) advocated kinematic waves to account for this behaviour, based on their observations of cobble mobility. Whilst this may indicate the nature of the bar accumulation it does not explain the spacing. It is probable that the spacing is related to some characteristic flow parameter, there being no apparent morphological reason for its regularity. It has been established that a hierarchy of channel perturbations, or disturbances, occurs, amongst which the series at about 5 to 8 widths recurrence is dominant. It is probable, in this context, that the occurrence of these forms is related to energy distribution in eddies.

Braiding and the Channel Pattern

On an established sandur surface, braiding can be initiated in two ways.

The first, primary anastomosis, describes the division of a channel due to sedimentary processes within the channel. In hydraulic terms, this occurs when flow resistance becomes too high to pass the water-sediment load, and is not reducible by single channel adjustments. In such a case sediment is deposited and local aggradation occurs to increase the available specific energy of flow. The usual morphological result is the development of a spool bar and the division of the channel into two chutes around the bar which are steeper, relatively much deeper, and hence more efficient, even though the total resistance in these two channels may be considerably higher than in the upstream reach (see p. 74 and 90).

In order to investigate the hydraulic conditions associated with primary anastomosis, the mean exponents were computed from the sections studied according to the channel state (Table 33). Note that divided reaches were treated in terms of the total channel region in the original calculations. Width increases most quickly in the braided reaches; depth least quickly there. In both cases, reaches with a "propensity" toward braiding are in intermediate position, but in the latter case they are significantly closer to braiding than to straight reaches in their behaviour. As a result, increase in velocity is greater in the braiding and "near-braiding" reaches than in the stable, single channels, confirming the conclusions reached on pages 70 to 76 that braiding sections are wide, shallow ones where velocity is increased most rapidly in the attempt to cope with heavy sediment load. Resistance declines much more rapidly in these sections as well (from much higher initial values due to the large relative roughness and relatively great wetted perimeter).

The second form of braiding is secondary anastomosis. On an established sandur surface there are many old channel scars and dry channels, left behind when the river has shifted radically. Such channels are often reoccupied at high flows when the present main channels overflow (Krigström, 1962; Williams and Rust, 1969). At Ekalugad Fiord, for example, there is a considerable interchange of water between Middle and South Rivers on the distal portion of the sandur at high flow through former main channelways. Similarly water moves from South or Middle Rivers northward across the sandur, toward North River at high flow. This mechanism is probably the main vehicle for braiding at any one time on a sandur surface.

The evolution and ultimate stabilization of a channel division will depend on how it handles the water and sediment load carried into it. In natural situations conditions are very variable (Hjulström, 1952; Axelsson, 1967). Often aggradation occurs at the head of the less efficient branch, and degradation at the head of the more efficient one, perhaps abetted by unbalanced division of the sediment load. Efficiency will probably depend on the nature of flow conditions downstream in each branch. If the process becomes sufficiently unbalanced, one channel may ultimately capture the entire flow (as appears to be happening at the head of the illustrated North River spool bar at present (Fig. 60) and transform the division point into a diagonal bar. Flow division angles do not seem to be important (Axelsson, 1967).

The channel pattern of braiding rivers appears to be highly unstable (Fahnestock, 1963; Chien, 1961), but in fact the main channel may remain relatively stable for a number of years, and shift much more slowly than individual anabranches (Krigström, 1962). This is particularly so at Ekalugad Fiord, where active anastomosis occurs only at very high stage. At Lewis River on the other hand, the sandur plain becomes water-covered relatively often (Fig. 63), yet the main stems of both the upper Lewis and Triangle Rivers have remained in their present position for at least 10 years, and aerial photographs taken in 1948 reveal a similar pattern.

In order to investigate the stability of the system of channels at Lewis River, the principal channel network on the sandur was redrawn in "graph" form from aerial photography for several years between 1952 and 1967 (see Fig. 64). Each channel division point or confluence was determined (sometimes more or less arbitrarily) as being a $1 \rightarrow 2$ or $2 \rightarrow 1$ joining, i.e., multiple branchings or confluences were resolved into sequences of simple articulations. There are two input streams, viz. upper Lewis River, and Triangle River (which includes water coming from the South side of Lewis Glacier) and one outlet stream. In this situation, the number of links, (channel segments between articulation points), number of confluences, and total number of articulation points (or nodes) are functions of the number of division points, p, as follows:

$$\text{number of nodes, } h = 2p + 1$$

$$\text{number of links, } L = 3\ (p + 1)$$

The number of division points observed in the Lewis River channel network over several years is given in Table 34. Though the number is variable, it is difficult to assign significance to the pattern. The source photographs were taken at various flow stages, and though an attempt was made to include major dry "channels" on low stage photographs, and to exclude minor "bar surface" riffles at high stage, it is not certain that the graphs of "principal channels" represent a homogeneous population by any means. The 1961 flow condition (Fig. 65), when most of the sandur was water-covered, presents a particularly acute problem in interpretation. Furthermore, since the most

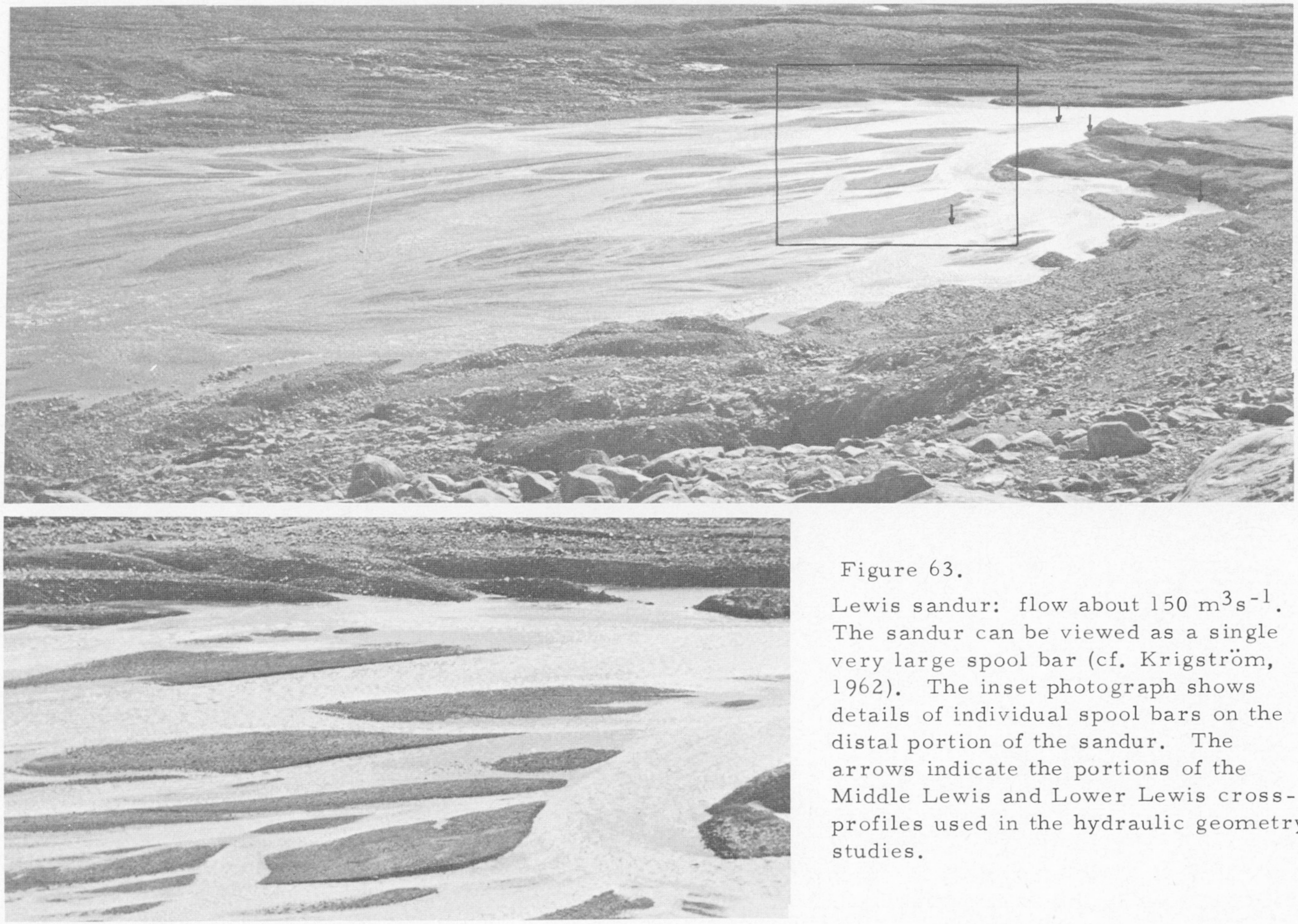


Figure 63.

Lewis sandur: flow about 150 $m^3 s^{-1}$. The sandur can be viewed as a single very large spool bar (cf. Krigström, 1962). The inset photograph shows details of individual spool bars on the distal portion of the sandur. The arrows indicate the portions of the Middle Lewis and Lower Lewis cross-profiles used in the hydraulic geometry studies.

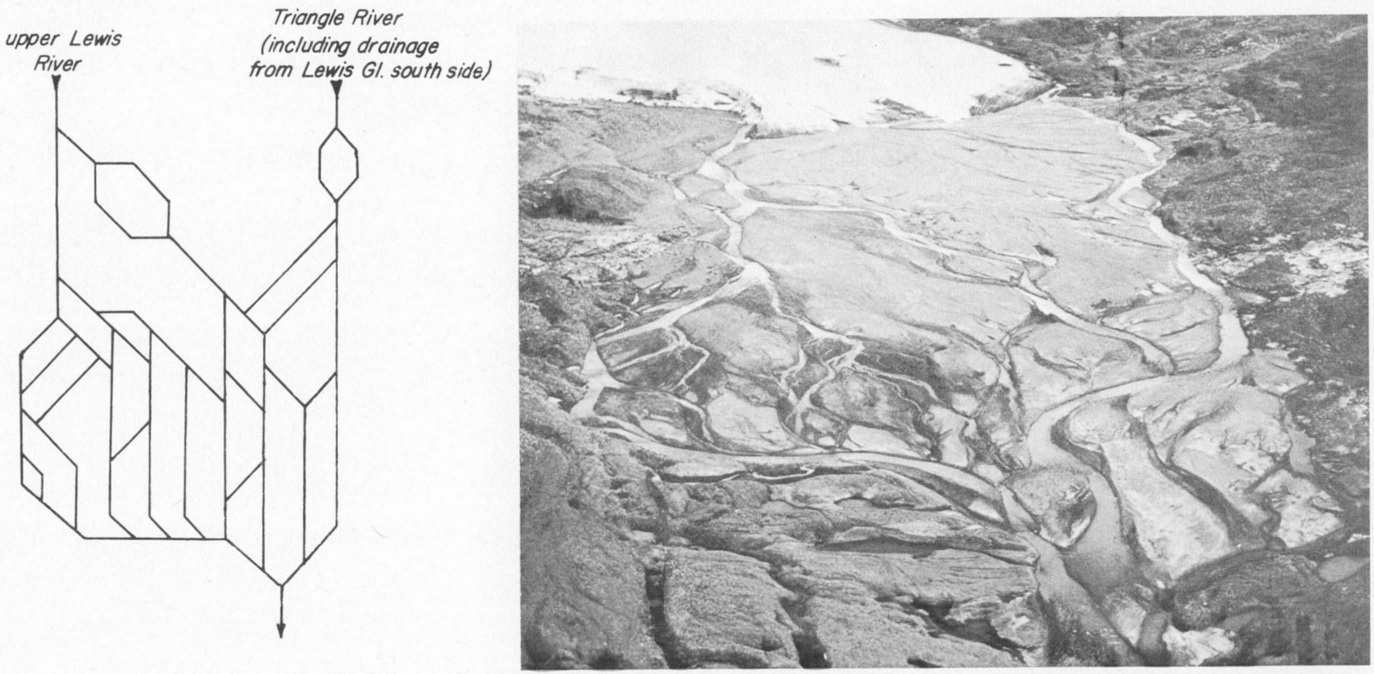


Figure 64. Network of "principal" channels on Lewis sandur, August 17, 1967.

radical changes in channel pattern occur at such high stages, the hydraulic significance of the numerous channel divisions that only appear at lower stages remains obscure.

One may imagine that a greater number of division points would signify the presence of a larger number of smaller channels in the network, with a greater total resistance to flow than would obtain with a smaller number of larger channels. Since the total cross-section of flow would also be larger, however, such a channel network may remain just as viable as one with lower resistance. Hence, a variety of channel networks seem to be equally feasible as equilibrium networks, and the pattern that actually occurs probably reflects the fortuitous outcome of sediment transport and flow conditions at very high stages, when anastomosis is generated.

By contrast to the major channels, minor channels may exist only very ephemerally. Their pattern of existence, as observed at Ekalugad Fiord, appears to follow a sequence of initial scour when overflow occurs onto the sandur surface from a major channel; establishment of a more or less stable channel with a lag-gravel bed and sides; and lastly infilling with sand (see Fig. 66). Since these channels are usually established at a higher level than the bed of the main channel, little bed load may be diverted (this is particularly true of channels that persist due to secondary anastomosis) and so the channel fill that appears during waning floods may consist of relatively well sorted sands that were carried in suspension down the main channel. The channels may go through an entire cycle of existence in one flood, but are usually reoccupied during many subsequent high flows. Small flows derived by seepage from the main channel may persist for a long time. The result is a ribbon deposit of sand across the sandur. Such features are rarely found stratigraphically since major flood aggradation events usually destroy them before they are buried.

Naleds

A singular form of channel disturbance which affects the bed stability of sandur channels is the occurrence of naleds. These "ice mounds" develop beneath the beds of the channel after freezeback in the autumn. Water left in the channels and the channel bed freezes downward, and the freezing plane ultimately encounters the frost table below. Closure of the frozen zones is achieved in some sections of the channel bed before others, and some zones ultimately become isolated cells of unfrozen water and sediment below the stream bed. As freezing continues static pressure grows in the remaining zone of unfrozen material. If pressure is great enough the frozen "roof" is forced upward, or may be ruptured completely so that water flows out and freezes above. Clean ice in the structure consists of vertically oriented "candles" that may be as much as 50 cm in length.

The structures may lie transversely in the channel, or may be oriented along the channel (the most common situation), or they may be round. They very from 2 or 3 in diameter and a few cm in elevation to about 50 m in length and 2 in height (much higher forms have been observed on the snout of the Lewis Glacier, but not in the river channels). Some of the structures are composed of completely clean ice, probably developed from a reservoir of water trapped between river ice above and the frozen bed below. Other forms appear to have lifted individual large cobbles or isolated masses of material from the bed (see Fig. 67A and B), suggesting that the freezing

Figure 66.

Minor channels on the sandur.

A.

Cobbed paved, scoured channels.

B.

Channel largely choked with sand deposited from suspended sediment load. Seepage of water through the gravels from the main channels maintains some flow.

C.

Sand fill in a former minor channel.

plane was near the bed itself. In still other cases the entire river bed has been lifted up (Fig. 67C), indicating that the freezing plane was below the bed.

The hydraulic and geomorphological importance of these structures is that they may severely derange a section of channel bed which was quite stable before. Their persistence after the commencement of flow in the spring may lead directly to a rearrangement of the channel pattern. They also pick up a great deal of material, some of it very large, that must be moved at least a small distance if the naled ultimately melts into flowing water. In some channel sections naleds reappear from year to year, in other places they occur randomly.

Summary of Stream Flow, Sediment Transport and Channel Form

The stream flow across Baffin Island sandurs is the product either of a nival or a mixed nival-proglacial regimen. Nival regimen is characterized by a peak snowmelt runoff in early July, and generally lower flow levels later in the season punctuated sharply by summer storm-induced floods. In the proglacial regimen continued high flows can occur into mid-August under the impetus of continued glacier or upland snowfield melt. In addition extraordinary floods may occur in the proglacial regimen as jökulhlaup. Jökulhlaup aside, the most severe floods appear to be storm induced, though melt processes are capable of sustaining continuous high flows for much longer periods. The distribution of flows is highly skewed, but the recurrence of "flood events" is probably rather more frequent than in more normal flow regimens because of the very high proportion of water that immediately runs off the land on the surface.

Sediment transport is dominated by bed load movement, though solution and suspended sediment transport are by no means negligible. Depending on the nature of the season, from 25 to 75 per cent of total sediment transport may occur during the 4 or 5 peak flow days. This is not, however, so severe an inbalance as may occur in many rivers in more temperate regions, and the relatively high frequency of significant transport events is an important feature of the sandur environment. The total work of sediment movement appears to be relatively evenly distributed amongst normal flows (near the mean) and high flows.

The sediment supply is abundant, much more so than would be allowed by current rates of detrital production. The redistribution of the abundant sediment left behind by the Pleistocene glaciers is the main reason for the prominence of the alluvial valley fills today. The actual movement of sediment through the river system is very discontinuous and large volumes of material are stored along the channelways. On rising stages material is picked up from the stream bed and banks and moved ahead, only to be abruptly dropped on the falling stage. The strong diurnal influence on melt processes means that this characteristic sequence of movements is not only typical of short-lived storm events, but also of melt-induced runoff on a day-to-day basis in relatively small watersheds. Each day's runoff during active melt periods, and each storm runoff event, constitutes a flood wave in the channels.

The river channels have the wide, shallow form typical of streams flowing in noncoherent, coarse clastic alluvium. Significantly, velocity and rate of velocity adjustment are both unusually high in the sandur

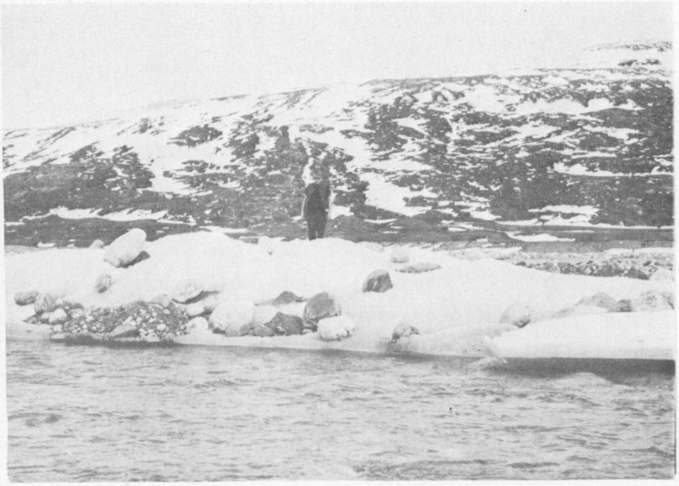
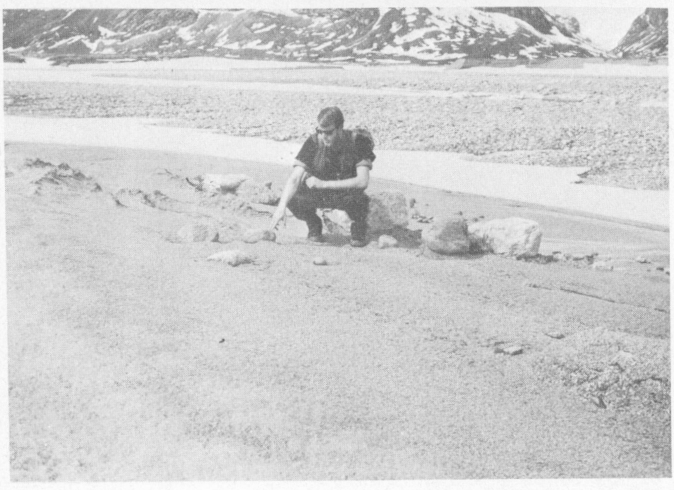


Figure 67.

Illustrations of river naleds:

A. Large, longitudinally oriented naled in South River, Ekalugad Fiord, which has picked up large cobbles from the bed.

B. Material picked up from the bed during naled growth, and candle ice, which forms the ice structures.



C. Section of South River bed which has been heaved up in its entirety by about 1.5 m. The man is pointing to a sequence of marked cobbles which had been left on the bed in the previous summer and which have not been appreciably disturbed.

rivers, reflecting the sharp decline in boundary resistance with the heavy sediment transport that occurs at high stages. The profiles of the sandur rivers are generally concave, indicating that general aggradation has occurred recently. Current sedimentation appears to be concentrated on the distal portions of the sandurs (along the lower profile segments at Ekalugad Fiord), whilst the channels on the proximal portions are slightly entrenched.

In detail, sediment in the channels forms a spectrum of channel bed bars, among which a dominant group occurs with a recurrence length of about 5 to 8 channel widths. The initial formation of the bar series is obscure. Although any number of morphological irregularities might account for the initial development of sediment deposits in the channel, the regularity of the sequence must be based on flow phenomena. The bars are an important element of the resistance of the channel to flow.

Several considerations are relevant to the adjustment of channel resistance to changing flows. The size of the material in the bed and banks will determine the boundary resistance. A measure of covariance between channel slope (representing energy expenditure rate) and material size has been demonstrated. The channel width to depth ratio has an important bearing on the distribution of shear stress at the channel boundary, and on the total resistance offered by the channel (in terms of the total length of wetted perimeter). There is a clear tendency for channels in weak, noncohesive material to adjust to increasing flows by excessive widening, so as to distribute the flow stress as widely as possible. The result is rising flow resistance due to the increasing boundary length. The characteristic alternation of sections possessing such characteristics with deeper, narrower sections of lower resistance throughout the channelway, suggests that the entire stream is acting as an oscillating system (in this case with compressional characteristics; Hjulström (1942) made the same suggestion for meandering rivers, where transverse waveforms became apparent). The implication is that sediment, particularly bed sediment, can be most efficiently passed through the system in discrete steps, with energy expenditure accordingly concentrated so as to initiate movement at each stage. The effects of fluctuating discharge may be responsible for this – if sediment cannot be passed right through the system at steady discharge, it must be stored enroute – and so concentration of storage and mechanisms for initiating motion become efficient means of moving material. At the same time, channel bars are the most conveniently adjusted aspect of channel form when it becomes necessary to make changes in resistance characteristics to accommodate different flow levels (Norgaard, 1968), so that the great variability of flow in sandur channels may be associated with the wide range of resistance conditions, volume and frequency of sediment movement through the dual activity of channel bars as resistance elements and sediment storage areas.

When the calibre or load of sediment becomes too great for normal channel processes to adjust to them, extraordinary changes take place; local aggradation, channel division and braiding may occur, so that flow is diverted to new areas of the sandur. The propensity of channels in noncohesive materials to widen by bank erosion during high flow periods, with subsequent deposition downstream on the channel floor, redistributes large volumes of material on the sandur and continually reworks the surface sediment. Such a sequence of erosion-deposition is characteristic of braided channels (Popov, 1962).

CHAPTER VI

DEPOSITION: MORPHOLOGY OF THE SANDUR

The Sandur Surface: General Morphology

Surface Form

As the name implies, sandur surfaces comprise extensive areas of sand and gravel, and locally boulder fields, and are completely devoid of vegetation in their recently active parts. The surface appearance (Fig. 68) of the sandur is that of a gravel plain, with frequent ribbons or "islands" of sand where former channels or quiet water bars existed. Large boulders are common, especially on the headward portions of the plain where the surface often becomes a rock desert (Fig. 69). Much fine material has been deflated by wind, but immediately under the surface there is always an abundant sand matrix. Traces of former channels are prominent on the surface, and may lend up to 1 metre of local relief. The pattern of the former channels indicates the general direction of sediment movement.

Krigström (1962) identified three zones on Icelandic sandur surfaces, characterized by rather different river behaviour and depositional character. The proximal zone is crossed by a few main streams flowing in well-defined, possibly somewhat incised, channels which tend to be relatively deep and narrow. Farther downstream, usually beyond a prominent zone of sedimentation, the rivers break up into many anastomosing channels which shift rapidly on the sandur. This intermediate zone possesses the classical attributes of sandurs, the channels becoming wide and shallow. In many sandurs that drain to lakes or the sea a third, distal zone, appears where the rivers merge into what is virtually a single sheet of flowing water that runs to the delta front of the sandur. Very shallow depths prevail, though deeper, major channels persist. This zone is hardly present on the Baffin sandurs, even those in the fiords, though at very high flow characteristics of the distal zone may appear in that nearly the entire surface becomes drowned even on an inland, moraine-delimited sandur like the Lewis sandur (cf. Fig. 65). Material size may be important in determining whether or not the distal zone occurs in that it appears to be associated with sand deposits. The Baffin sandurs are commonly composed of gravels and very coarse sand right to the distal end so that channel paving and maintenance of channel banks provide sufficient local relief to prevent general drowning of the surface except during very high floods.

The long profile of the sandur is similar to that of the rivers, usually concave (see p. 76). Figure 70A presents the profiles of several sandurs. The Hoffellssandur (data from Sundborg, 1954) and White River dalsandur (Fahnestock, 1963) appear to be relatively simple surfaces. Lewis Valley sandur presents an interesting case where the proximal end of the sandur (outside the active river channel) passes into little altered ablation moraine which continues right up onto the glacier (cf. Fig. 71). At Skeidarárjökull

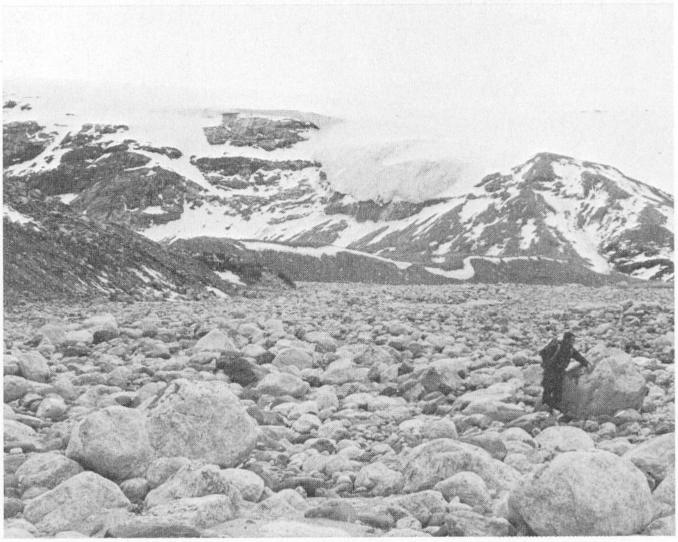


Figure 69. Large cobbles on the proximal portion of Ekalugad sandur. Some of the largest ones were delivered by rockfall from cliffs on the right (out of view) and have been rounded in situ by flood waters. Many of the rocks have been transported, however, as proved by the presence of lithologies alien to the cliffs.

(Thorarinsson, 1939) and Breidamerkurjökull (Price, 1969) in Iceland, the active sandur surface begins directly on the glacier. Ekalugad sandur provides an example of a compound profile with two regular profile segments occurring.

Analysis of the long profile form (Fig. 70B) revealed that four of the six surfaces examined were reasonably well approximated by simple exponential equations of the form

$$(36) \qquad y = c_2 e^{-c_1 x}$$

The two east Baffin sandurs, Ekalugad and Tingin, are represented by the sum of linear plus exponential functions

$$(37) \qquad y = c_4 + c_3 x + c_2 e^{-c_1 x}$$

The fit is not ideal, a slight convexity remaining in some cases which may indicate tendencies toward purely linear form. (In the case of Ekalugad sandur, the existence of two segments in detail produce this appearance when the entire surface is being considered as a simple form.) Another east Baffin sandur that was examined in part (Confederation Fiord sandur) was purely linear, while McDonald and Banerjee (1971) reported that two small outwash plains examined by them in the Rocky Mountains of southern Alberta each possessed few distinct linear segments. This appears to be unusual for an active surface, however.

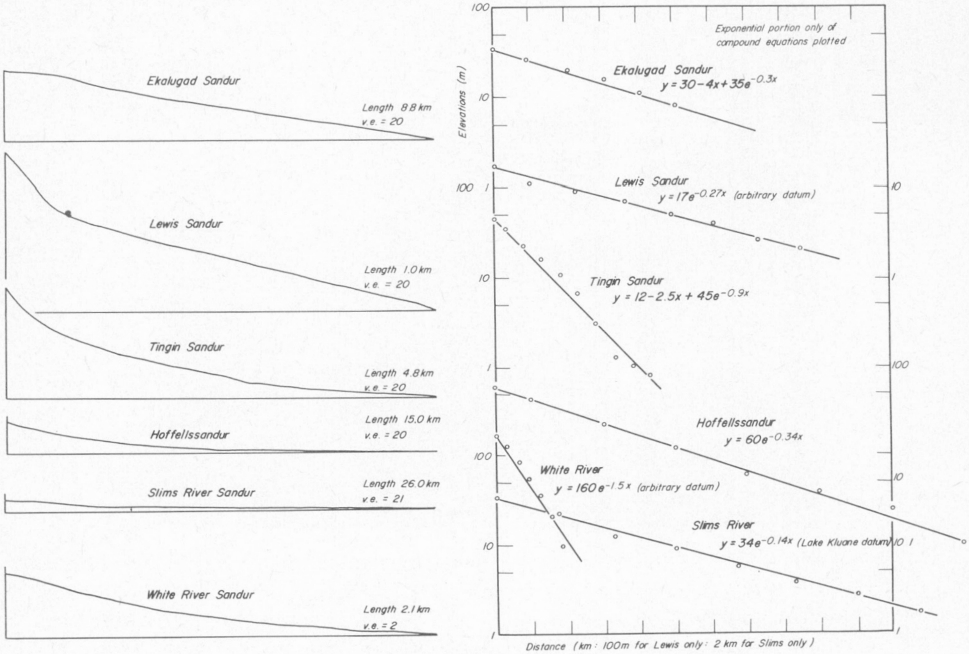


Figure 70. Longitudinal profiles of sandur surfaces:
A. Survey results
B. Exponential fitting equations

Figure 71. Glacier ice preserved under ablation moraine at the snout of Lewis Glacier. Preserved ice has been found up to several hundred metres beyond the present snout.

An interesting feature of many sandur plains already alluded to (see p. 79) is that the proximal portion rises above the river. This has been noted on Hoffellssander, at Markarfljót and some other Icelandic sandurs (Krigström, 1962), on the Slims River (Fahnestock, 1969) and on the Baffin sandurs. Various reasons can be ascribed for this behaviour. The most obvious is that the rivers are regrading in their proximal sections in response to changed or changing conditions of runoff or sediment supply. This view is developed by Fahnestock (1969). On the other hand, the suggestion that modest incision of the rivers below the sandur surface may be a normal circumstance in the proximal zone was made on p. 79. Morphological support for this suggestion is that while the surface in the middle and distal portions of the Baffin sandurs is formed of channel-related features almost entirely (Fig. 68A) coherent channel traces are lacking in the proximal zone where, nevertheless, the surface appears to consist of equally recent gravels (Fig. 68B). It appears, then, that a "high flood" surface, elevated a few metres above the river channels, may be quite normal. Another consideration, exemplified by the Lewis Valley and Icelandic sandurs (see Price, 1969; 1971), is that if sandur development may occur supraglacially, then when the ice melts back a considerable amount of material may be left elevated above the new level of the river issuing from the ice "front". Such a circumstance is usually clearly evident by the development of kettles and general subsidence on the formerly supraglacial surface.[1]

The cross-profile of the sandurs in the proximal and intermediate zones may be quite irregular. Though often upwardly convex in form, especially where the sandur has considerable lateral freedom, valley sandurs often display a steady cross-valley slope, as does the Lewis sandur (Fig. 72), or an irregular form. Commonly, the major channel zone is built considerably above adjacent areas on its own sediment deposits, as is the case at Lewis River, and consequent avulsion is one of the main causes for periodic radical shifts in the channel pattern. Several metres of relief may be present across the sandur.

Flood Deposits

Although the great majority of flows across the east Baffin sandurs are contained within the channels, with sediment movement essentially continuous through the channel system, periodic flooding distributes material over a wide area. Major shifts in channel position, which are of great importance in determining the locus of sedimentation, are themselves usually associated with floods. The incidence of flooding is variable: on a very active sandur such as that at Lewis Glacier, a large area of the surface may be flooded for several days each summer season. On the other hand, the large and relatively inactive surface at Ekalugad Fiord may experience only one or two such events in a season, or perhaps none, and then only a small portion of the sandur is actually flooded. Over much of the surface the water is too shallow and moves too slowly to effect notable bedload movement, although considerable movement and redeposition of fines may go on. However, in the region close to major active channels large volumes of gravel and sand are moved. It was estimated that, during the jökulhlaup of July 20-22, 1967, on

[1] It is not proposed that these arguments apply to Fahnestock's case. The field relations clearly substantiate his view at Slims River.

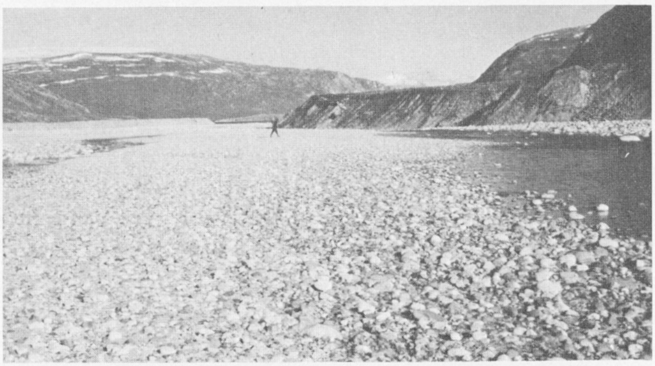
Figure 72. Cross-section of Lewis sandur.

South River at Ekalugad Fiord, between 20 and 60 per cent of the active surface was directly affected at various points down sandur. Since the channels may shift quite rapidly, any position on the sandur surface (except, perhaps, in the proximal zone) may be reworked and receive fresh sediment within a period of a few tens of years.

The sedimentary deposits associated with floods take three major forms: channel fill, "levees", and surface veneer. Channel fill deposits (Figs. 73 and 74) occur normally as the last stage of scour-and-fill processes that accompany flooding. It was not possible to study the progress of scour and fill in detail because of the impossibility of working in the sandur rivers during flood events. The depth of scour may not be great in many sections as the main channel beds are heavily armoured. However previously existing fill deposits and channel bars are certainly extensively modified or removed. On falling stage large volumes of material carried near the bed are dumped in the channel once again as the river loses its capacity to transport it. Deposits develop in areas of flow divergence which, because of changes in the flow pattern that may have occurred during the flood, may not be the same as they were before the flood commenced. A channel fill area will continue to act as a zone of flow divergence and sediment deposition until the river successfully cuts a channel through or around it (see section on Channel Form and Evolution). Locally, "fans" of material occur where the river has developed across a major node of sedimentation and a large amount of material has been deposited in a zone of general flow divergence that may be tens or a few hundred metres in diameter. Such zones may develop into spool bars.

"Levees" are not well defined on the sandur rivers. However, where overbank flow occurs greatly elongated areas of riverbank deposit occur (Figs. 73 and 75). They are often laterally graded (in the manner of conventional levees), with very coarse gravel and cobbles being dumped at the channel edge and finer materials occurring farther away. These long gravel bars along the river might, then, be termed "sandur levees". These overbank deposits are often associated with channel fill and may result in superelevation of the river above the general sandur surface. Often, as well, the channel fill and levee deposits become the agency for a shift in the channel location. If a major portion of the river is diverted behind the levee, the entire channel may ultimately be deflected from its old course in this fashion.

Surface veneer appears across wide areas after inundation. Material deposited commonly varies from fine gravel to sand. Often the sand is rippled. Subsequently wind may deflate the deposit so that a gravelly surface is most common.



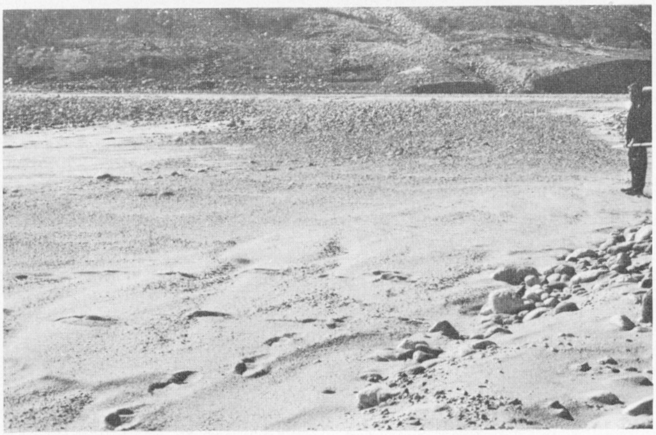


Figure 73.

A. Channel fill and "levee" deposits along South River resulting from the flood of July 20-21, 1967. (Photo 1968)

B. View on the ground of the channel fill deposit in the centre of Figure 73A. View downstream. Note the coarseness of the surficial lag materials: matrix of finer materials is shown in photo below.

Figure 74.

Major channel fill deposit on the distal portion of Ekalugad sandur. The edges of the former channel can be seen by the coarse materials in the right-centre and on the far left. The channel may have been as much as 2 m. deep.

A good deal of sediment sorting occurs locally on such deposits, the downstream sides of the bars characteristically having finer, better sorted material (see Fig. 75). It is probable that this feature, and in fact much of the "structure" possessed by the deposits, is a result of winnowing of materials and adjustment of the surface to flow on the declining stage, and that it does not represent conditions during the height of the flood when sediment movement is general.

Cobble orientations were measured at various sites on the Lewis sandur in order to determine the extent to which preferred orientation appears in the surficial materials. Figure 76 indicates the range of orientations observed in samples of 100 cobbles. No clear pattern emerges: some orientation patterns show weak maxima across the direction of flow, whilst some show maxima parallel to the direction of water flow. Principal axis planes (a-b planes) normally had very low dips. Andrews (1965b) found that outwash-plain stone-orientation patterns showed far less organization than glacially derived ones and in fact in a majority of the samples from the sandur surface the pattern is not significantly different from random. Lane and Carlson (1954) noted that the bed cobbles of irrigation canals in Colorado had a preferred long axis orientation normal to the flow direction; Doeglas (1962) observed the same effect in torrential river gravels. On the other hand, Krumbein (1940) noted that flood gravels in the San Gabriel canyon showed a preferred orientation parallel to the flow direction. Material that moves by being rotated or rolled along will take up a transverse orientation with respect to the flow, so that it rotates about the two lesser axes, whereas material that is shoved or pushed over the bed, or within a disperse layer, will take up a position parallel to the flow, so as to present a minimal drag profile to the moving medium. The former type of movement is probably typical of low rates of transport, and the latter typical of more general movement. In the rapidly fluctuating flows of sandur rivers both forms of cobble motion are probably common, and so bimodal orientation patterns, or sometimes random patterns, may reasonably be expected. An odd feature of the sandur surface samples is that flood surface materials seem to be more highly organized than channel materials.

Stratigraphy

Sedimentary structures are poorly developed in Baffin sandur sediments. A wide range of grain sizes is present (cf. following section) and bedding remains rudimentary (Figs. 77 to 78) except in certain backwater locations where relatively good sorting and well-developed bedding may be found: such zones are usually destroyed by renewed erosion later on as the main channels swing across the sandur. Channel side bars, which may be preserved, are often the sites of the best developed bedding. Crossbedding and laminated fine-grained sand may be preserved in such locations. Sometimes festoon bedding is found, though most sandur bedding is parallel.

Jewtuchowicz (1953) has described extensive lamination, crossbedding and channel dunes preserved in sands in Polish Pleistocene sandurs. Similarly Doeglas (1962) noted festoon structures and foreset laminations in sand and fine gravel deposits of the torrential Ardèche River in France. The nature of sediment movement would seem to be closely associated with these sedimentary patterns, with dune, antidune and moving bar forms common. Ore (1964) has described foreset bedding in both contemporary and ancient

Figure 75. "Sandur levee" on South River in the area immediately downstream from that shown in Figure 73. Note the gradation of material across the deposit, development of a minor "back channel", and the small chutes which indicate that flow diverged out of the main channel and across the bar during flood.

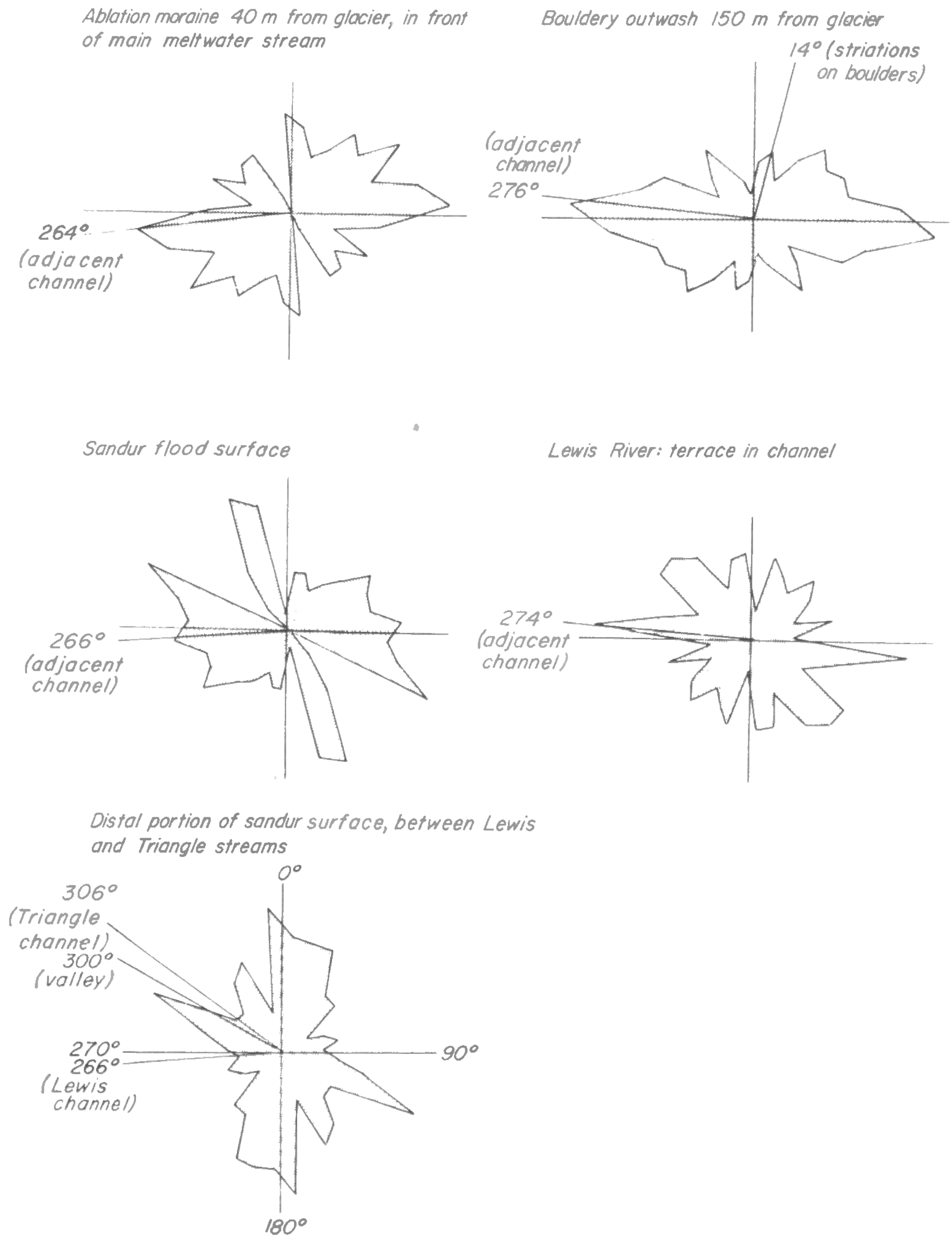


Figure 76. Surface boulder orientation diagrams: Lewis sandur (magnetic bearings). Plots give distribution of a-axes for samples of 100 stones with a/b> 1-5; a> 5 cm. Samples were collected on the sandur surface immediately adjacent to the present channels.

Figure 77.

Rudimentary bedding in an exposed section of the T2 surface at Ekalugad Fiord. Sorting is very poor, the "beds" standing out chiefly by virtue of improved moisture retention in horizons where there is a greater portion of fine materials.

Figure 78.

Closeup view of materials in a section on the Lewis sandur. "Bedding" is chiefly evident by variation in matric properties. Note the prominent normal imbrication.

braided channel deposits. It was more typically associated with "transverse" (diagonal) bars than with spool bars and was quite rare at sites where rapid aggradation was occurring. Hence, foreset bedding appears to be characteristic of tabular deposits of channel material reworked at flows lower than those necessary for completely live-bed conditions. Sorting is relatively good in such an environment. Similarly, small trough-fill structures (Harms and Fahnestock, 1965) are typical of such conditions rather than of rapidly aggrading sites. Smith (1970) confirmed the association of "longitudinal" (spool) bars, coarse and poorly sorted materials and horizontal bedding, as against cross-stratified, well-sorted sands, in the Platte River, and Boothroyd (1970) indicated similar conditions on the Scott Glacier outwash fan in Alaska. Williams and Rust (1969) gave a detailed account of bedding forms and structures, mostly in fine materials, in the deposits of the Donjek River, Yukon Territory. Waechter (1970), studying the Texas Red River, reconfirmed the occurrence of a wide range of structures in fine (sand) materials.

Anderton (1970) has recently noted large scale cross-stratification in cobble-gravels of paleo-outwash deposits. In all such cases, however, the material forming the deposits is far smaller than the scale of flow which produced them (e.g., material size much smaller than flow depth). The calibre of the gravels composing the Baffin sandurs is generally larger than would be expected to develop such forms in the small sandur rivers, and so the rarity of such forms is not surprising. Graded bedding is also rare.

The poor sorting and rudimentary bedding are the results of deposition of material from transport during rapidly falling water stage (cf. Stewart and LaMarche, 1967). The initial deposition of very coarse materials leads to discontinuous, horizontal bedding. Entrapment of fines later amongst the coarse materials tends to obscure whatever structure the original deposits may have had (cf. Ore, 1964).

In order to study the fabric of the deposits, orientation and dip studies were made on samples of 100 stones collected in pits dug in the sides of recent bar deposits at Lewis River. Samples were taken from about 30 to 60 cm below the present surface: frost disturbance was judged not to be significant. The fabric patterns are shown in Figure 79. The orientation patterns are not dissimilar to those possessed by the surface cobbles, with strong parallel and transverse orientations appearing. Often, peak orientation strengths occur obliquely to the flow direction inferred from present channels, suggesting that flow directions, or local currents in the channel, may have been different at the time of deposition of the sampled material than now. Ore (1964) found similar results, and showed that divergence can occur between the deposits and the flow. The largest discrepancies occur on channel-side bars where the water often flows across the channel locally.

Orientation and dip patterns in all of the subsurface samples are highly significant ($\alpha \leq 0.01$ in the χ^2 test applied to sample distributions), lending support to an earlier suggestion that the flood deposits are relatively well organized. Possibly the movement of large volumes of sediment imposes an order on the ultimate deposits that channel cobbles, moving individually at lower stages, need not conform to.

Dip patterns show that normal imbrication (i.e., upstream dipping) occurs in all the deposits. This would be the commonly expected pattern, except perhaps in the tone of flow separation on lee faces of prograding channel bars. The general absence of any such phenomenon, or of foreset bedding, suggests that the bulk of sedimentation on the Baffin sandurs is achieved

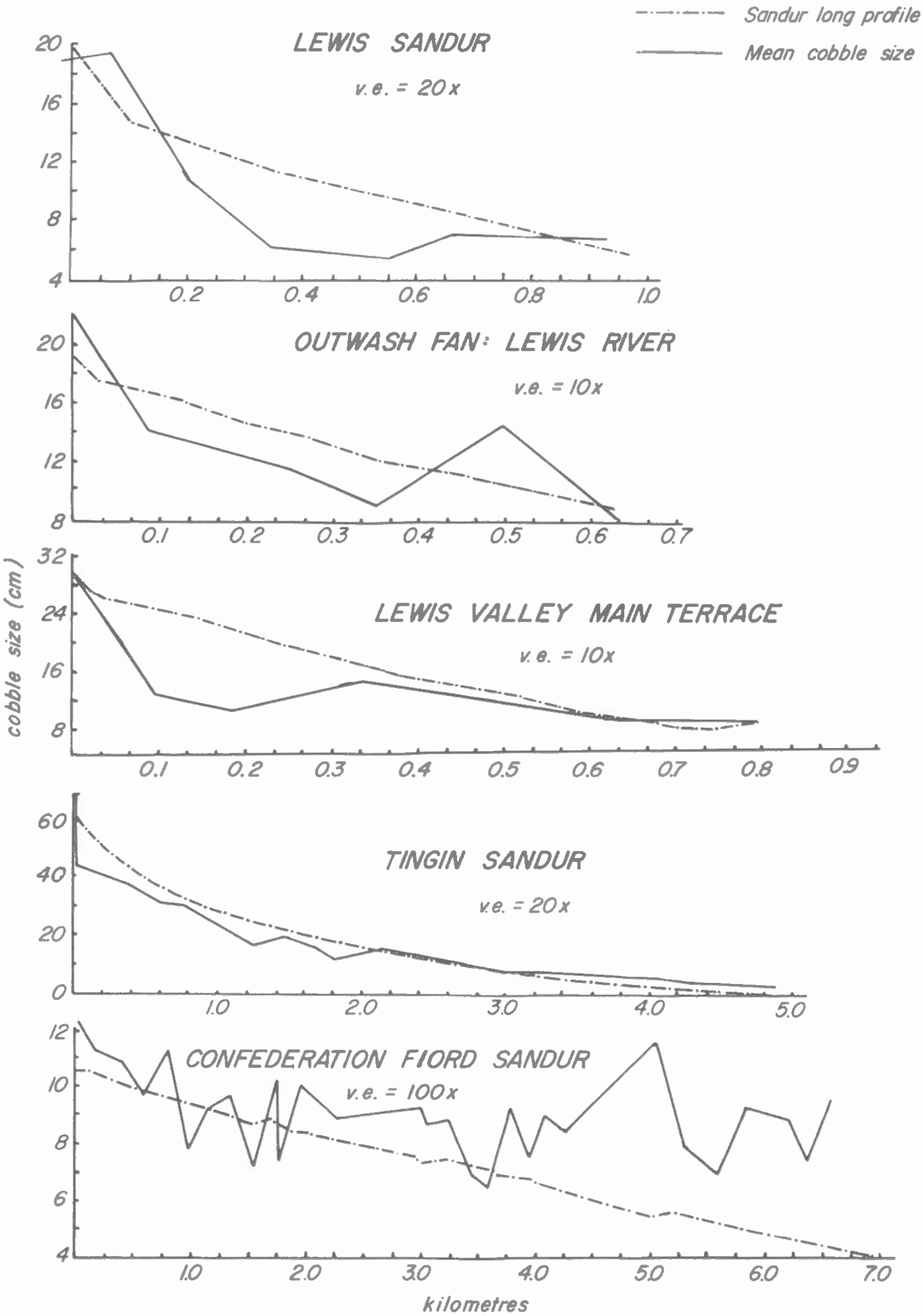


Figure 80. Variation of mean cobble size downsandur.

as massive, essentially structureless sheets of material laid down parallel to the channel or sandur surface, as seen in Figure 77. Bar structures that may develop in such deposits appear, in their details, to be erosional (see p. 90).

The Sandur Surface: Characteristics of the Materials

It seems reasonable to expect sediment characteristics to change down sandur. In particular one would suppose that the size of materials found on a surface would decrease downstream, and that sorting would increase, in response to abrasion and selective transport. One expects size to be related to the change in gradient downstream (through competence considerations). Reconnaissance traverses of several sandurs on Baffin Island, and of several of the terrace levels at Lewis River, support this concept in general (Fig. 80). Boothroyd (1970) reported similar results from an Alaskan outwash fan. It is apparent, however, that other factors, such as lateral proximity to the rivers, relative topographic elevation (locally), extent of water sorting or wind deflation that has occurred since the main depositional events, and so on, will affect the detailed pattern of distribution and character of surficial deposits. In the succeeding sections the distribution of the materials on the Ekalugad sandur surface will be taken up in some detail.

Krumbein (1941b) stated that the fundamental attributes of clastic particles were size, shape, roundness, surface texture, petrology, and orientation in situ. In considering the mean character of deposits, the variance of these characteristics also becomes important. Sneed and Folk (1958) attributed the geometrical characteristics to the initial shape of the particles as liberated from the parent rock; internal structural characteristics of the rock; rock size (i.e. several of the form parameters appear to depend on the size of the clast); distance of transport; the transporting agent; and finally to chance events in the course of transport. In the following sections, surface texture will be dealt with only as an adjunct of petrology, and orientation will not be considered.

At Ekalugad Fiord the surface materials were sampled systematically at 146 sites laid out in a square grid pattern (cf. Fig. 81 for the sample positions). For sampling purposes, the division between "coarse" and "fine" materials was arbitrarily set at 8 mm b-axis diameter. A sample of 96 "coarse" clasts was collected at each site, using a radial grid to locate the individual stones. Each stone was classified petrologically and measurements were taken of the principal axis lengths and minimum radius of curvature in the principal axis plane. A sample of fine material was collected as well by making four random grabs. A separate series of samples of coarse materials was taken from the bed of South River for comparative purposes. These samples were collected according to Wolman's (1954) method, and sample size was 48 at each of the 45 stations. No petrological information was gathered. (The samples have previously been referred to on p. 84).

Petrology of the coarse materials

Rock types were classified in order to investigate distinctions in depositional behaviour that might be ascribed to the petrology of the materials.

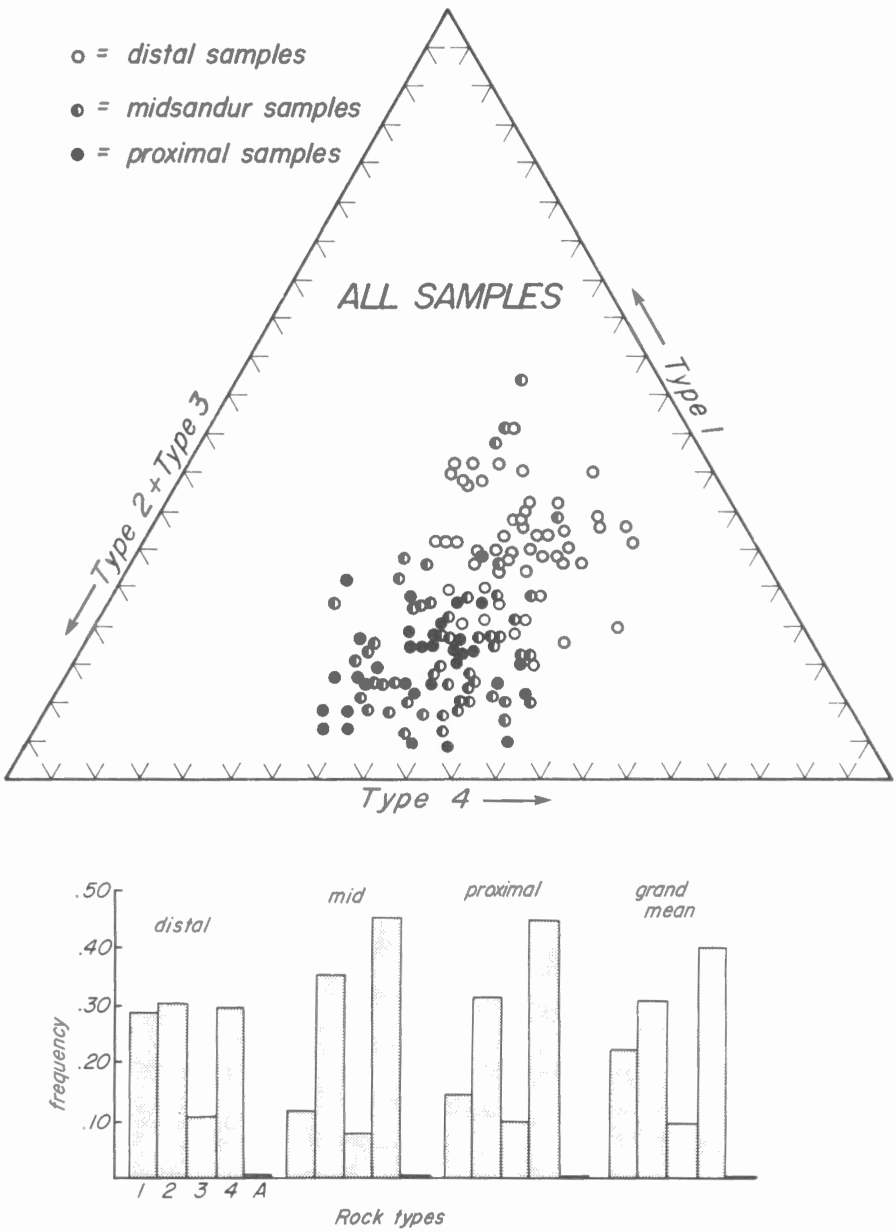


Figure 82. Rock type distribution in sandur samples (types are classified in the text). See Figure 81 for the sandur sections summarized in the histograms.

Four main categories were recognized (see Fig. 9 for distribution of source rocks):

Type 1 -biotite-quartz-plagioclase migmatite, ±muscovite (biotite < 5%):
Type 2 -plagioclase-biotite-quartz gneiss, ±muscovite (biotite < 25%):
Type 3 -biotite schist or plagioclase-biotite schist (biotite > 25%), medium-textured:
Type 4 -biotite-plagioclase gneiss or muscovite-orthoclase-gneiss (micas > 25%), characteristically coarse-textured and commonly rotten.

Three additional classes (pegmatites, ferromagnesian rocks, and "others") provided less than 0.5% of the rocks (collectively designated as type A in Fig. 82). The main distinctions which might have relevance for the sedimentological pattern would be expected to be found amongst those rocks dominated by quartz, micas, and by feldspars. The proportional distribution of the types is given in Table 37 and in Figure 82. The chief distinction apparent is the decline of Type 4 gneisses down sandur, which is compensated almost exactly by the rise in frequency of quartz-bearing rocks, particularly the Type 1 migmatite. Most of this change occurs on the distal portion of the sandur. Micaceous and feldspar-bearing rocks must be being preferentially deposited or comminuted more rapidly on the distal portion of the sandur relative to quartz-bearing rocks. The relative decline in Type 4 rocks by comparison with the other types is probably mainly due to attrition abetted by mechanical weakness caused by the presence of coarse-grained micas, but is possibly also occasioned by the susceptibility of the feldspars to incipient chemical weathering, which may aid materially in comminution processes.

On the whole, the rocks are mineralogically similar types, and no great differences were detected in their appearance on the sandur. In the following, the behaviour of the different types will not generally be distinguished, though summary statistics will be reported by type in figures and tables, so that comparisons may be made.

Size characteristics of the coarse materials

The size distributions of coarse materials, based on b-axis diameter, are summarized on Figure 83. Apart from the wide variation between the upper and lower envelopes, the curves indicate generally similar distributions. Some samples appear to be impoverished in small material. This may either be the result of biased sampling in very rocky areas, or may reflect actual paucity of pebbles in areas of coarse lag deposits. Materials sampled on the bed of South River showed far less variation, both within and among samples, than those from the sandur surface: the mean Trask sorting coefficient was statistically significantly higher for the sandur surface samples (1.69, as against 1.42 for the South River samples - significantly different at the $\alpha = 0.01$ level), and the variance of the Trask coefficients was significantly greater for the sandur surface.

The mean distribution of sizes for all samples taken together is shown in Figure 84. Even though plotted on phi logarithmic scale, moderate skewness persists in the results. The mean of mean sample sizes, taken over the entire sandur, is 69.2 mm, with a standard deviation of ±47.1 mm (Table 35). The standard deviations are similar in magnitude to the corresponding mean sizes, so that the coefficient of variation of the grain size distribution is usually close to 1.0.

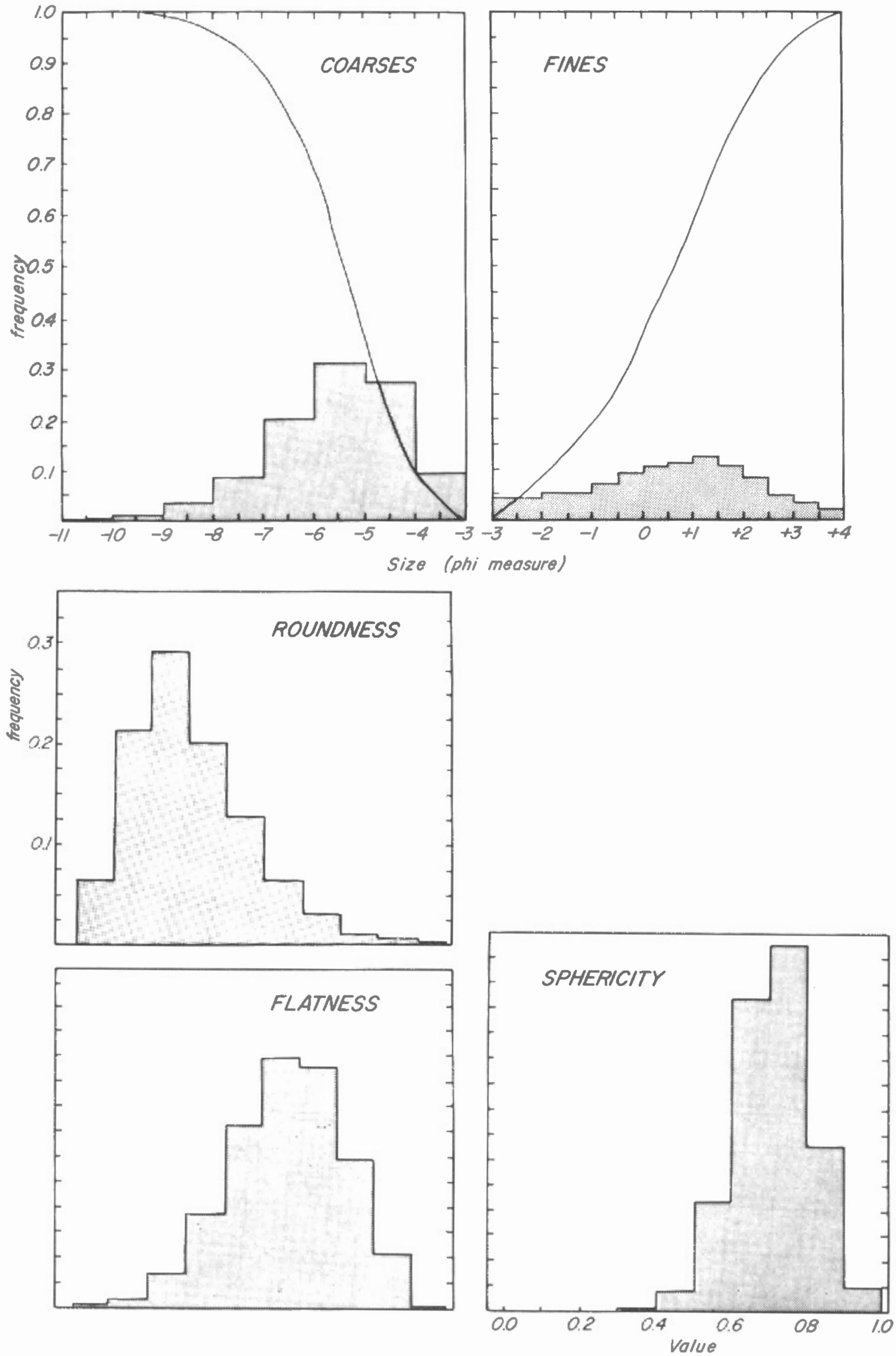


Figure 84. Mean distributions of characteristics of sandur surface materials (all samples pooled).

The distributions of coarse materials are unimodal and are nearly all positively skewed, reflecting the presence of a few relatively large cobbles in each sample. Mean skewness for all samples is +1.983, with a standard deviation of ± 1.047. The mean reduced kurtosis of the grain size distributions is +3.205, which is leptokurtic. However, the standard deviation of ± 7.878 associated with that figure indicates that a wide range of kurtosis values is present. Kurtosis serves as a measure of sediment sorting that is complementary to the variance; whereas the latter reflects the absolute range of sizes, the former reflects the relative range of sizes, by comparison with the range expected in a normal distribution. Whilst some samples are somewhat platykurtic, some extremely leptokurtic ones occur, indicating a wide range in sorting conditions.

The mean of the 10 largest cobbles found in each sample was computed and its value was very closely correlated with mean size over the entire sandur ($r = +0.94$). The mean of means is 198.5 mm which is 2.9 times the mean cobble size. Median grain size and D_{85} size were similarly very closely correlated with mean size everywhere.

Form and Shape of the coarse materials

The form of the coarse materials was investigated by three standard measures

(38) $R = \frac{2r}{a}$ (Cailleux roundness)

(39) $F = \frac{2c}{a+b}$ (reciprocal of Cailleux flatness)

(40) $S = \sqrt[3]{\frac{bc}{a^2}}$ (Krumbein's intercept sphericity)

All the measures vary from 0.0 to 1.0. The index R is a measure of surface regularity of the rock which generally reflects the extent of abrasion it has undergone. Since it contains the parameter r, the minimum radius of curvature in the principal axis plane, it is more or less independent of the other two measures, which represent more properly measures of the shape of the rock. The indices F and S are closely related to each other: i.e.

(41) $$\frac{F}{S^3} = \frac{2}{b/a + 1}$$

The only degree of variation is provided in the axis ratio b/a. Since this is a relatively conservative quantity, the two measures are highly correlated on the sandur.

The grand distributions of the roundness, flatness and sphericity for all sampled rocks are shown in Figure 84. Whilst F and S are approximately normally distributed, the roundness, R, shows a decided positive skew. Whereas most rocks are of relatively low roundness, a few very round cobbles occurred, mainly on the distal portion of the sandur. Table 35 gives the mean of sample means over the entire sandur, and indicates that whereas roundness is relatively low (indicating that the rocks are poorly abraded and tend to have sharp corners and rough surfaces), the sphericity is quite high, indicating that the rocks are blocky or compact.

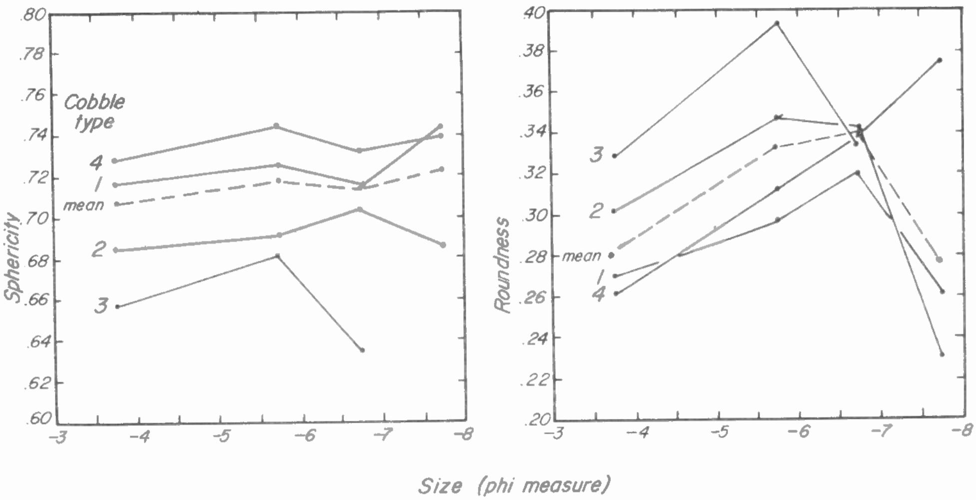


Figure 85. Variation of form characteristics of cobbles with size.

Sneed and Folk (1958) have presented evidence to show that sphericity is dependent upon the size of the rock: this effect was investigated by examining the shape parameters of rocks in four separate 1/2 ∅ size ranges. The degree of selectivity involved (there are 16 categories, but even these do not cover the entire range of possibilities) precluded analysis on a sample-by-sample basis because of limited sample size. Hence ten groups of nine samples were chosen, giving 864 stones per group from which to segregate stones in each category, using most of the samples on the South River side of the sandur. (The ten groups thus permitted down sandur changes to be investigated at the same time.) Results are given in Table 36 and are displayed in Figure 85 (missing data indicate that there were too few stones in the class for reliable computation; 5 stones was arbitrarily set as the lower limit of acceptable group size).

In three of the four rock types a break occurred in roundness values. The increased probability of larger clasts being relatively fresh products of freeze-thaw weathering on the cliffs probably conspired to reduce the roundness of very large rocks (rocks that obviously had been frost shattered on the sandur were avoided in sampling). No significant pattern of varying sphericity appeared over the size range studied.

The shape of the coarse materials was investigated in terms of the Zingg classification. Spheroidal and disc-like shapes are overwhelmingly dominant (Fig. 86). Proportional distribution of all the rocks by petrology and form (Table 37) shows that all the major rock types except schistose rocks are preferentially spheroidal or disc-like; the latter are preferentially disc-like or bladed. These distributions reflect different cleavage and jointing in the source rocks, which is blocky in the gneisses and migmatites and sheet-like in the schistose rocks. The proportions of each Zingg group do not change significantly down sandur, though the number of spheroids and blades increase slightly at the expense of the numbers of discs and rollers.

In summary, the character of the coarse materials seems to be strongly influenced by their primary condition. Skewed size distribution, low roundness, and shape conforming to that expected from their cleavage or jointing in bedrock all indicate that fluvial activity has not greatly modified the rocks.

The fine materials

Analysis of the fine material samples was restricted to the material finer than 8 mm. Materials finer than sand size (lower limit 63μ) contributed a significant proportion of the total materials present in only one sample (0.197 in No. 2952), which was a site where there was obvious contamination of the sandur materials by marine silts being washed out from beneath the T2 terrace along South Valley. The sandur deposits are generally restricted to the sand sizes and coarser, very fine materials normally being transported directly to the sea.

A composite grain size distribution made up from all 146 samples is shown in Figure 84, and indicates a relatively uniform distribution. The means of sample moment measures are given in Table 35.

Individual curves have been plotted (Fig. 83) as "cumulative per cent coarser" in semilogarithmic form. Four types of curve can be detected.

- convex up: indicating a relative paucity of fine materials;
- concave up: indicating a relative paucity of coarse materials;
- sigmoidal: indicating a logarithmic-normal distribution of sizes;
- inverted sigmoid: indicating a relative paucity of sizes near the middle of the range (bimodal distribution).

The logarithmic-normal distribution of sizes is that which would be expected to occur purely by chance in a deposit abstracted randomly from a population of materials with a finite size limit (Rogers et al., 1963). Processes of erosion and rock disintegration presumably act in such a manner. Basal moraine or earth-flow deposits probably preserve such a "primary distribution" of materials since no sorting processes occur (Elson, 1961; Tricart and Vogt, 1967). It is also possible that on the sandur certain high flood deposits show a total lack of selective processes. Sigmoidal grain size distributions are the product of unreworked "free deposition" (Tricart and Vogt, 1967) of materials; that is, deposition under moderately changing transport conditions, so that there is neither an influx of very large materials, nor wholesale deposition of fines. Such "declining stage" deposits are characteristic on the sandur.

Convex and concave distributions indicate deposition in various special environments. Convex distributions lack fine materials: such deposits indicate selective sorting in an "active" environment, being lag deposits

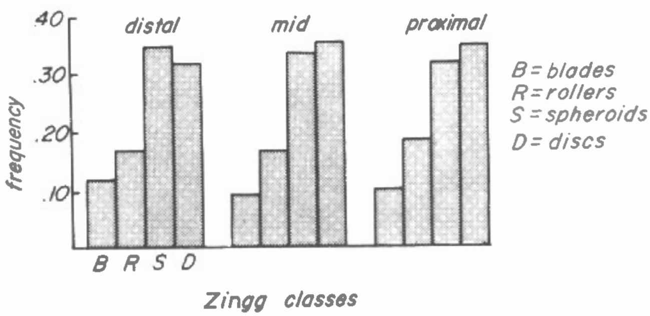


Figure 86.

Frequency of Zingg shape classes.

from which fines have been winnowed out (the sample usually represents the monotonic tail of a unified size distribution which peaks in the coarses). Such a process is common in channels and on channel bars, and in fine deposits may also be a consequence of aeolian action. Concave distribution of materials indicates deposition of fine materials in a "passive" environment i.e. in quiet water. Very little coarse material finds its way to such a spot. Such areas are by no means restricted to the distal end of the sandur, but can occur in channel backwaters or sandur surface areas away from active currents.

Inverted sigmoidal, or bimodal, curves are mainly the result of mixed depositional environments on the sandur. The admixture of fines washed out of the T2 terrace along the south side is a case in point. The possibility for finer materials to contaminate previously deposited coarser materials is particularly strong in areas where coarse and fine materials were deposited at different times.

As with the coarse materials, standard deviation of the samples is highly positively correlated with mean size. However, skewness and kurtosis are also highly negatively correlated with the first two moments. Hence, as mean grain size tends to become smaller, standard deviation becomes less, but skewness and kurtosis increase. Such effects are the result of a shift (as the mean size becomes smaller) towards an "independent" distribution of fines, positively skewed, and away from distributions that are merely an appendage of unitary grain size distributions in the coarse materials.

In considering the entire range of grain sizes, it is clear that a relative paucity of materials occurs in the 0.5 to 8 mm range. Slatt and Hoskin (1968) found a similar situation on the Norris glacier outwash, and this is a common feature of fluvial sediments (Pettijohn, 1957). (No unified quantitative distribution of sizes can be shown due to the different means of measuring fine and coarse materials: by weight and by number respectively.)

Several explanations have been attempted to account for the paucity of very coarse sands and granules. Sundborg (1956) and Meland and Norrman (1969) have put forward hydraulic arguments based on preferential erodibility (cf. Hjulström, 1939) or most rapid transport velocity for materials in the coarse sand and granule range. Other workers, exemplified by Yatsu (1959) have preferred an attritional argument based on the thought that rock clasts will not easily disintegrate to sizes less than the characteristic mineral grain size. On Ekalugad sandur, the bimodal effect appears mainly as a result of combining samples into a composite distribution, and is probably due to mixed depositional processes in the cases of real bimodality, so that no distinction can be made between the two arguments on the sediment characteristics alone. Since rock disintegration by abrasion or by chemical action appears to be relatively minor (cf. p. 44, 123 and 129), the chief cause of the paucity of granules appears to be hydraulic. The mechanism does not appear to work strictly in Sundborg's sense, however, as most individual deposits are unimodal.

Material distribution

Figure 87 displays the distribution of materials and of some of their characteristics over the sandur. Although strong trends are readily apparent in grain size and standard deviation in the coarse materials, the remainder of the parameters showed no clear patterns. (An apparently artificial character in some of the maps is the result of machine contouring.)

Spatially distributed data may exhibit structure in terms of areas of homogeneous character more or less sharply delineated by transitional zones, in terms of trends extended over considerable area, or in terms of oscillatory structure. Only the second possibility revealed any coherent patterns of sediment distribution.

Trends

Trend surface analysis was carried out from first to sixth order for each of the characteristics of grain size distribution and particle shape in the coarse materials, and for selected characteristics of the fines. Results are summarized in Table 38, and Table 39 shows the detailed analysis of variance for one subject as an example of the derivation of the summary results.

Consistently significant trends are shown only by the mean size and standard deviation of size, which possess highly significant linear trends and generally significant increments to variance reduction through the third order. Grain sizes decrease downvalley and north across the sandur: i.e. away from South River, which is the primary source of recent sediments. The coefficients of the first order expression indicate how rapidly each variable changes in each direction.

The rate of decrease of the standard deviation closely parallels that of mean size for all effects, and the absolute magnitude is also very similar. Complete first order equations for the two variables for the combined samples are as follows:

(42a) $\bar{b} = 158.9 - 2.6x - 2.9y$

(42b) $\sigma_b = 151.2 - 2.8x - 2.8y$

Hence, the coefficient of variation of grain size is near 1.0 everywhere on the sandur.

The third order fits shown in Figure 88 provide some elaboration on the pattern. The decreasing grain size downvalley reflects the decreasing competence of the river as it spreads onto the less restricted area of the lower sandur, and as it approaches the sea. Residuals from the third order surface have positive values at the proximal end of the sandur on the north side. This area is just below high cliffs with active talus slopes on the face: the very large material on the surface here (Figs. 70 and 87A) was derived from rockfall rather than by fluvial action. The large negative residuals immediately downvalley suggest that the large rock sizes found under the cliffs have distorted the whole surface upward and that the true mean size of fluviatile materials that should be expected there is not even so great as indicated by the third degree trend surface. The only other area of notable positive anomalies occurs on the south side near the distal end of the sandur. Once again, debris slopes contribute a few boulders to the surface here which will have "contaminated" the true fluviatile material locally. The residuals from the third-order trend of standard deviation appear as though they may contain a mild oscillatory structure downvalley.

Skewness and kurtosis exhibit almost no variance reduction, only in one or two isolated cases is total variance reduction significant at the third or fourth order level. These two variables have substantially homogeneous fields over the entire sandur. The map of kurtosis (Fig. 87D) exhibits isolated peak values. The fact that skewness shows no discernible trend indicates that as the entire distribution of materials shifts towards finer sizes and smaller variance down sandur, a proportion of relatively large materials remains so that the symmetry of the distributions does not improve.

Roundness and sphericity display weak patterns. Total variance reduction is generally significant for roundness by the second order and the first order coefficients indicate that roundness increases down sandur (except for Type 4 rocks, this observation suggests that these coarse-grained mica-feldspar rocks are subject to disaggregation on the sandur). Sphericity, on the other hand, does not normally show aggregately significant variance reduction until the high orders, and the trends are variable.

The variation of roundness and sphericity down sandur by size classes was analyzed by searching for trends in the mean data for the ten groups of samples previously mentioned (p. 123). No consistent trends were found (Fig. 89). A similar analysis was carried out for the 44 samples of bed materials from South River, grouped into 11 sets of 4 consecutive samples. Only the aggregate variation over all sizes was investigated because of the lack of sufficient cobbles for examining characteristics in specific size ranges. The results were remarkably uniform (Fig. 89) and it is concluded that no pattern exists in the behaviour of sphericity.

The fine materials similarly exhibit no clear trends. This is not surprising inasmuch as high floods are presumably competent to carry all sizes in the fines range at any place on the sandur. Hence the fines represent "chaos" deposits, dumped as a result of declining transport capacity on falling stage rather than as a result of failing stream competence. Variability in the characteristics of the fines will be local rather than systematic (p. 124 to 125), as determined by local depositional conditions.

The proportion of heavy minerals in the 125 to 177 μ range shows a moderate decrease down sandur (the range of all samples lay between 0.031 and 0.132). The first order surface of the proportion of heavy minerals is

(43) $H = 0.0996 - 0.0008x - 0.0016y$

All coefficients are significant. It is not clear whether this result is due to hydraulic conditions or to weathering effects.

Implications for the fluvial processes

The alluvial surface of the Ekalugad sandur is aggradational. The only sedimentary parameters which exhibit a significant degree of spatial organization are the mean and variance of the coarse materials (fraction > 8 mm). The higher order moments do not present consistently significant patterns, nor do the form parameters, with the exception of a slight apparent increase down sandur in roundness.

All of these observations are accounted for by simple selective deposition of the coarsest materials during floods as diverging flood waters lose competence down sandur. The slight increase in roundness is an expected consequence of the abrasion sustained by the stones in moving down sandur, and, in any case, it is reasonable to expect that rounded stones may move farther down sandur than angular ones of similar size. Krumbein (1941 a) showed that roundness initially increases very rapidly, so that it is not surprising that a change should be detected even over a few kilometres.

The lack of pattern in the sphericity suggests that particle form does not affect transport selectivity over the short distance involved. Pashinskiy (1964) and Meland and Norrman (1969) have presented evidence that particle shape effects transportability in normal fluvial environments. Failure to detect the effect here implies either that the distance is too short, or that significant transport mainly occurs in high floods when selectivity of movements is minimized. A slight change in the proportions of the various petrologies down sandur, related to relative susceptibility to weathering or comminution, is not a major influence on form.

The down sandur variation of grain size and surface form is approximately exponential. This is in accord with the form of the river long-profiles (p. 76), and corresponds with the findings of other investigations of purely aggradational surfaces (Krumbein, 1937; Blissenbach, 1952 and 1954; Bluck, 1964, amongst others).

The sediment transport data given on p. 61 indicate that as much as one-half of the sediment load of South River may be lost on the sandur. The indicated losses for 1967 would amount to about 1 cm of aggradation if the material were uniformly distributed over the surface (most was concentrated in the flood deposits of the July 20-21 flood, however).

An interesting contrast appears between the character of the materials in the river and on the sandur surface. Grain size in South River does not decrease down sandur nearly as much nor as regularly as on the surface (cf. Figs. 49c and 83 for South River grain size). The long slope of the river conforms to that of the surface, yet the actual cobble sizes on the bed do not conform to the proposed relationship between slope and grain size in an aggradational situation. This is because the distribution of grain sizes on the river bed is arrived at by a different process than that on the surface: the channel bed consists largely of lag material from which finer materials have been washed away. The efficacy of the washing process will be determined by local channel slope, and in particular by the pool and riffle sequence, rather than by the general slope of the river. Hence, the end result is quite different from the end result on the adjacent, flood aggraded sandur surface.

The Distal End

Deltaic processes in Ekalugad Fiord

At its distal end, Ekalugad sandur is prograding directly into Tasiujaq Cove. A high-angle, "classical" delta is developing at the end of the sandur. The bulk of the total volume of sediments underlying the sandur is composed of deltaic or marine bottom sediments.

Tasiujaq Cove is almost isolated from the sea by a moraine (Fig. 90 in pocket). The "thalweg" of the 100-metre-wide passage which connects the cove with the open fiord is only 3 m deep in its shallowest portion and is floored by large boulders that are lag deposits from the eroded moraine. The mean depth of the cove is 39.7 m and the surface area is approximately 3.6 km^2. It is ice-covered for between 9 and 10 months, opening up only between late July and the end of September. Ice significantly influences the character of sedimentation in the cove.

The average daily tidal range in the cove, determined from automatic tide gauge records maintained during 1967, is 0.84 m. Without considering freshwater discharge, this would require a daily movement into and out of the cove of 3.0×10^6 m^3 of water. The tide appeared to be basically diurnal, but with a double high tide often occurring. High tide characteristically occurred early in the morning, and was probably influenced by the time of peak river inflow. The average daily influx of fresh water from the rivers during 1967 was 3.38×10^6 m^3 of water, with a maximum daily influx of 12.84×10^6 m^3 on July 14. In view of these data, it must be concluded that stream discharge may affect water levels during periods of significant runoff and that at such times water continued to flow out through the passage to the open fiord during the entire tidal period. The freshwater discharge from the rivers normally flows directly out over the saline bay water, so that stratification of the water occurs in summer. The contrast between the surface water and the deeper cove water is marked by a pronounced thermocline at 3 to 5 m depth, and by a halocline above about 12 to 15 m. Below that the temperature and salinity are both remarkably uniform in the regions of -1.4°C 20‰ to 30‰ respectively.

A program of bathymetric sounding, water sampling and bottom sediment sampling was carried out in the cove in 1968. The main purpose was to study the environment and morphology of the prograding delta front that is being deposited at the end of the sandur. Methods and results are given in Knight (1971). Knight has treated the sedimentology and sedimentological environment. The present remarks will emphasize the morphological features of the delta.

The consequence of the stratified character of the cove waters is that the rivers normally discharge into the bay essentially as two-dimensional free jets. This is not always true, however. When the water is carrying a particularly heavy sediment load, it may dive at least below the surface water of the cove. Whether the flow ever penetrates directly into the salt water was not studied. Suspended material settling out of a plume dispersing on the surface should be relatively well sorted, since the size of sediment, which directly determines settling time (other things being equal), will determine the distance that it can be carried before it falls into the quiet water below. By contrast, bed load material and material carried down by diving currents will slump and slide along the bottom before coming to rest.

In order to limit investigation to materials that could have moved in suspension, Knight (1971) confined his analysis to material <2 mm in diameter (this comprises 100 per cent of 30 of 50 samples). Furthermore Knight presented his sedimentological analyses in terms of Folk and Ward's (1957) phi graphic measures. The first two moment measures are usually highly correlated with the graphic measures; the last two are not necessarily. Since the mean and standard deviation will be of main interest here, Knight's results will be directly illustrated and discussed.

Figure 91 shows the distributions of mean and standard deviation of the fines (< 2 mm diameter) at the distal end of the sandur and across the delta face. It is apparent that an abrupt decline in sorting occurs on the delta (all 12 of the samples with greater than 5 per cent of material larger than 2 mm occurred in water of less than 21 m in depth in this zone: cf. Fig. 91). Offshore sediments again show somewhat improved sorting. A major implication of this pattern is that a large proportion of the material arrives at the delta front as bedload and immediately slides down the delta front. The bottom morphology amply illustrates the effects (Fig. 90). Prominent lobes of sediment extend

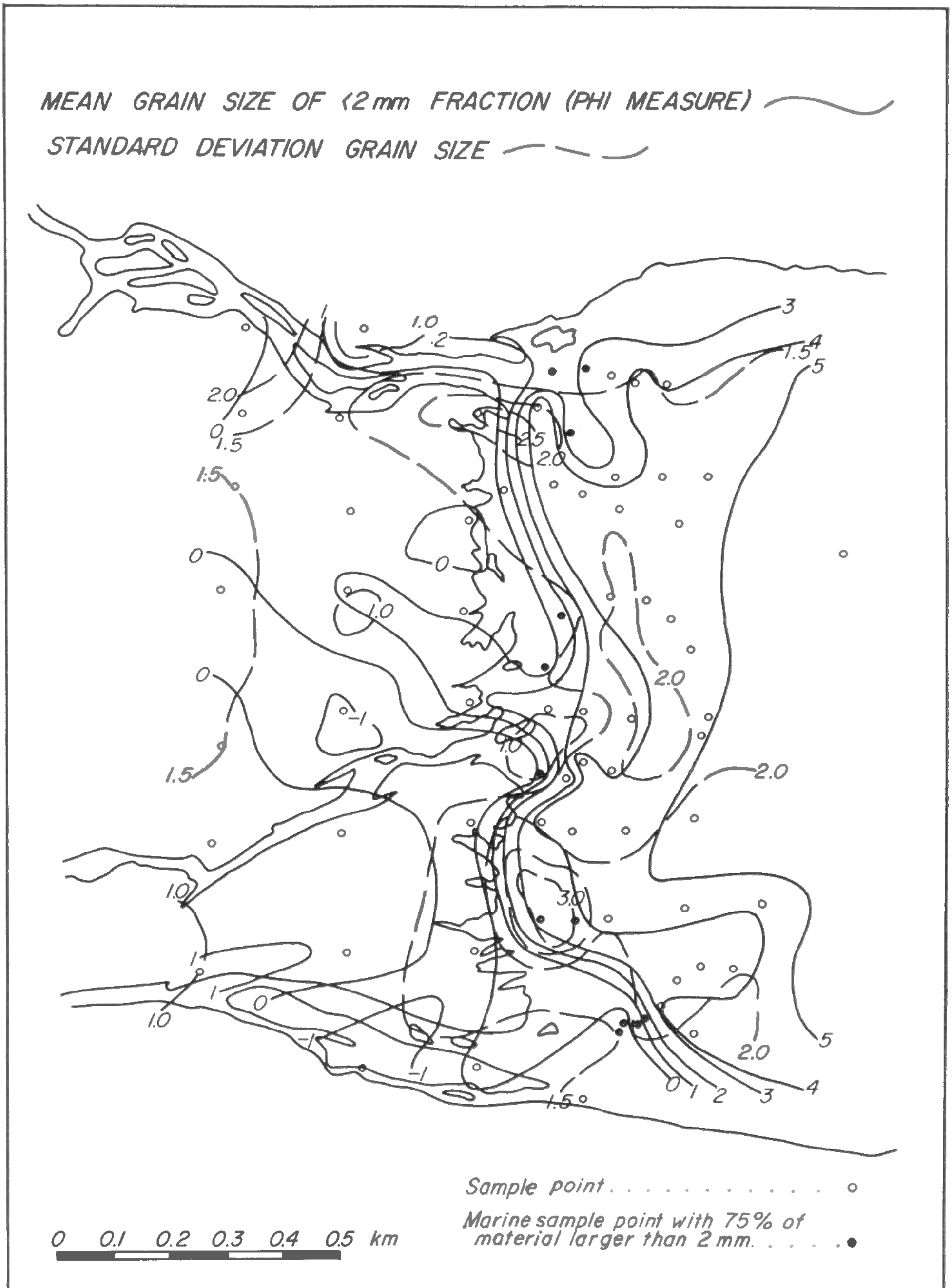


Figure 91. Delta front in Tasiujaq Cove: distribution of mean grain size (<2 mm fraction: phi measure) and of standard deviation grain size.

basinward from the mouths of the three rivers, and off Middle River levees are apparent (an additional lobe here marks a recently abandoned course of the river). Whilst material stands at angles of up to 35 degrees on the delta front in the quiet water zones, in the areas off the river mouths slopes hardly exceed 10 degrees even on the upper, steepest portion of the face: frequent sliding and slumping of materials probably occurs in these zones (Fig. 95). Such submarine action would account for the poor sorting on the delta face.

Knight pointed out that the generally low available energy in this restricted, estuarine situation would also maintain poor organization of the sediments, since secondary sorting by current and wave action would be small. The great variability of inflows would lead to the superposition of a variety of disparate sedimentation events. He also stressed the importance of ice rafting and the presence of river sediment on the bay ice each spring (placed there by early floods discharging onto the ice) as contributors to the poor degree of sediment sorting through their leading to dumping of large materials, particularly gravels, far from shore.

The bottom samples from the delta are mainly positively skewed (cf. Fig. 92) whereas those from the distal end of the sandur surface are largely negatively skewed (< 2 mm fractions); this is the best distinguishing criterion between the two environments (Fig. 93). The condition of the bay materials is produced by the admixture of coarse materials by rafting or slumping into the dominantly fine sedimentation environment. The general coarseness of materials on the South River side attests to the dominance of that river in sediment delivery.

In the river channels coarse material is found right to the edge of the delta face. Yet the steep face appears to consist largely of sand. Only the upper metre or two could be directly examined; since the edge is sharp (Fig. 94) it is not surprising that large material does not lodge there. It is probable that a good deal of very coarse material (cobbles) is found farther down the foresets: the occurrence of coarse material in the samples taken here was noted above. The nature of the sampling procedure (a Dietz-Lafond grab sampler was used) prejudiced the sampling procedure toward the fines. (Near the shore it was not unusual to lose a sample when a stone was caught in the jaws of the sampler.) The mean slope of the delta front (0-50 m depth) varies from 10 degrees on the north side to about 7.5 degrees on the south side (Fig. 95), but the upper part of the foreset zone has slopes ranging from 20 degrees to 35 degrees (except off the mouths of Middle and South Rivers, as noted). There is a striking morphological similarity between this situation and the experimental model of Johansson (1963).

The orientation of the sediment lobes in the bay suggests that the river "plumes" in the bay are drawn into a counter-clockwise general circulation. Knight (1971) confirmed this from miscellaneous surface observations. The pattern of grain sizes also follows this trend, again suggesting a relatively simple primary pattern of sedimentation.

Most of the sediments on the delta foreslope can be classified as gravelly sand or sandy gravel (Knight, 1971). However, in the bottomset zone beyond about 50 m depth silty sand and sandy silt are found. Farther from shore the hypothesized sediment pattern controlled by settlement of suspended sediments becomes evident. Nevertheless, ice-rafted material may still be present to produce apparent anomalies in the pattern.

Downvalley growth of the sandur

The sandur extends downvalley by growing over its delta. Deposition at the delta front is localized to the areas around the river mouths at any particular time, but when one area has prograded forward for a time, the unstable sandur channel inevitably shifts to another location, so that the entire delta front progrades relatively uniformly.

Tasiujaq Cove, being virtually isolated from the sea, is an efficient sediment trap. For order-of-magnitude calculations one may take trap efficiency to be 100 per cent. In 1967 the Ekalugad rivers delivered the amount of clastic material shown in Table 40 (based on sediment transport observations in the lower sections). Open packing is assumed to occur in the newly deposited sediment dumped into the bay so that the gross density of the deposits is assumed to be 1.55. The calculation based on close packing is

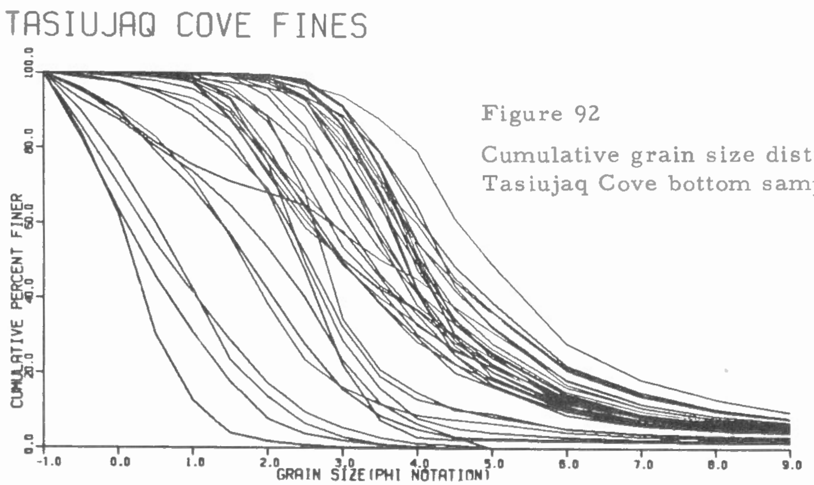


Figure 92

Cumulative grain size distribution of Tasiujaq Cove bottom samples.

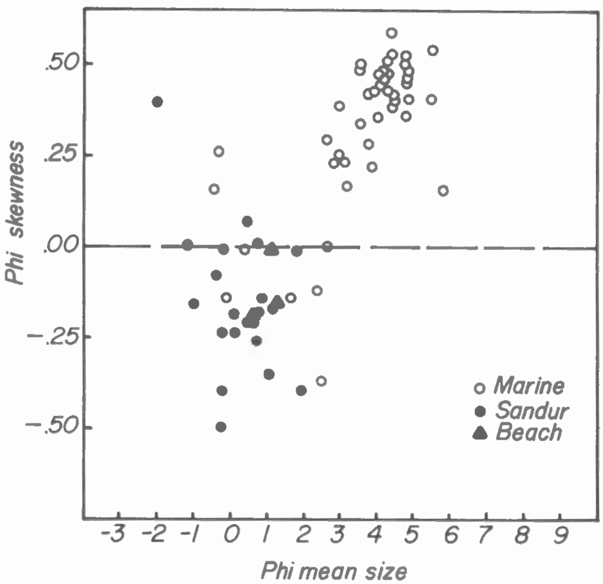


Figure 93

Tasiujaq Cove sediments: mean versus skewness (after Knight, 1969).

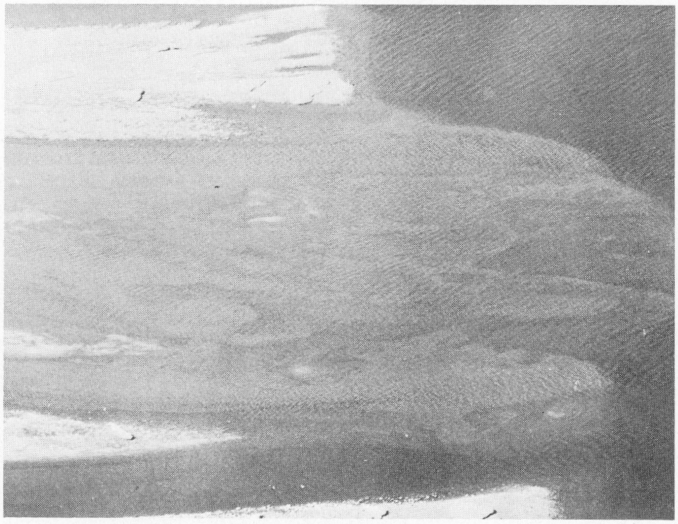


Figure 94. Delta edge in Tasiujaq Cove at the mouth of South River. The edge is sharply defined and drops abruptly at the maximum angle.

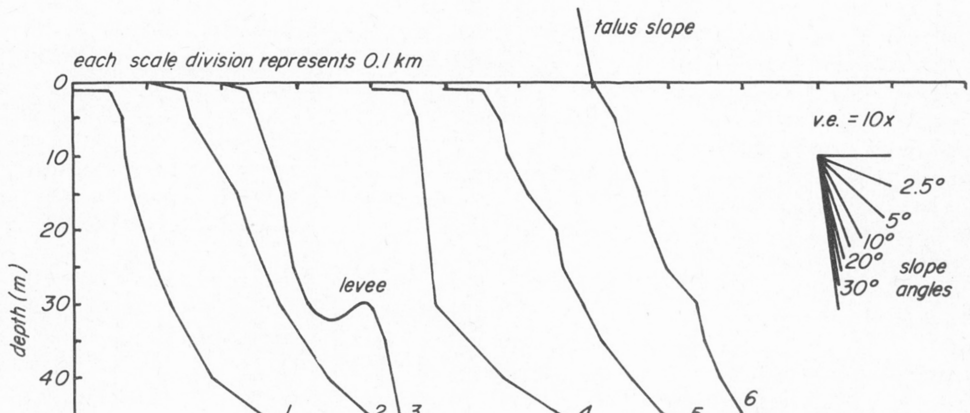


Figure 95. Profiles across the delta front in Tasiujaq Cove. Positions of the profiles are indicated on Figure 91.

assigned a density of 2.5. If one makes the further simplifying assumptions that the bedload material is added uniformly to the 1.45-km-long foreset wedge of the delta down to 55 m depth, whilst the suspended material is uniformly distributed over the area of the bottom below 55 m, then the tabulated results are arrived at. The increment to the delta front is expressed as bed thickness on a 9 degree dip: this represents 1.5 m of progradation at the front for open packing. Whilst the computations are obviously unrealistic at several points (not least of which is the erection of such a calculation on one year's data), the result is nevertheless useful in indicating the order of magnitude of the sedimentation effects.

By another yardstick, the T4 surface has advanced approximately 1,300 m into the bay since its inception 2,000 or more years ago. This represents progression at the rate of somewhat less than 0.65 m per year. The discrepancy between the two estimates is reduced when one considers that the older deposits will have become reasonably well consolidated: assumption of close packing would reduce the first estimate to 0.95 m. The remaining discrepancy must be judged in the light of the appearance that much of the period has been cooler than the present (p. 12), and the probable non-representativeness of the single year's sediment transport observations. At all events, the rate of progression of the surface appears to be of the order 1 m per year. At this rate the cove has another 2,000 years of existence left. At current rates of isostatic recovery (Andrews, 1969) the cove should become isolated from the sea in 1,000 years or so, so that it will probably finish its existence as a lake.

Conclusions on the Mechanics of Sandur Development

Sediment transport observations (section 4) indicated that in an aggrading river such as South River, Ekalugad Fiord, net sediment deposition occurs on the sandur surface at all significant flow levels (flows in excess of about 6 $m^3 s^{-1}$ in the present case). Nevertheless, the great bulk of aggradation is achieved during a few high floods. This indication is borne out by the nature of the sandur sediment deposits. The surface of the sandur is a montage of flood deposited sediments: near the channels, river bars and "sandur levees" occur, whilst farther away the sheet deposits of very high floods veneer the surface. Poor sorting and normal imbrication characterize the gravelly deposits.

Though high floods promote a great deal of scouring along channelways during rising stage, there is even greater aggradation associated with the peak and waning stages, so that extreme events contribute the greatest amount of aggradation (see also Scott and Gravlee, 1968).

Sandur sediments are sedimentologically immature. Although the texture of the coarse materials changes rapidly down sandur, with size decreasing and sorting increasing accordingly, there is a complete lack of pattern in the fines. Form parameters show little or no change down sandur, and the bedding is crude. No systematic variation in textural or form parameters can be detected between different rock types, which differs from the situation described by Ehrlich and Davies (1968) on Alaskan outwash. The mechanism of textural differentiation, that of selective deposition under changing flood competence, is different from that which produces increasing maturity downstream in normal fluvial deposits (see Plumley, 1948; Sneed and Folk, 1958). Much larger transport distances than are available on Baffin sandurs are normally required to achieve maturity via selective transport and communication in fluvially transported materials.

The contrast between sediment characteristics in the river thalweg and those on the sandur surface emphasizes the derivation of surface characteristics from flood events. Sediment size changes very little downstream on the streambed, indicating that relatively homogeneous conditions pertain throughout the channel length during normal flow. It appears that sediment sizes below those found in the bed can be transported right through the system at normal flow levels: that is the river is purely a transporting feature, rather than a feature of aggradation, and sediment sorting is achieved on the river bed by selective erosion rather than by selective deposition.

Nevertheless, the river's profile is constrained to take up generally the form of the aggrading surface over which it flows, so that on an active sandur, the river profile is concave up. The downstream change in gradient probably conditions the small change in grain size that does occur in the thalweg, though other hydraulic adjustments, such as the shift toward wider, shallower channels with high velocity of flow, probably compensate this effect somewhat. By contrast, on an inactive sandur, the regrading river is free to impose its preferred form on the sandur surface, so that the sandur becomes a rectilinear "transport surface". Degradation in the upper course and aggradation in the middle zone of the profile are the normal means by which the change is initially achieved. The Confederation Fiord sandur (Fig. 80) provides a good example, from amongst those studied, of this effect.

CHAPTER VII

CONCLUSIONS

Most of the major valleys on Baffin Island possess some form of valley fill, either presently active or the product of fluvial activity in earlier postglacial times. It is evident that nearly all such valley deposits represent aggradational surfaces of the same nature as the Lewis and Ekalugad sandurs. The diagnostic characteristics of braided rivers, concave-up long profile and decreasing grain size downvalley commonly appear on these surfaces. Only in rather special circumstances are mature transporting surfaces found, usually where there has been no glacial activity for a considerable time.

In watersheds with no permanent snow or ice, valley filling processes are very restricted or nonexistent. Ekalugad Fiord represents a situation where permanent ice and snow does occur on the uplands. Activity in such "nival" cases varies from slight to moderately active. The greatest rates of sedimentary activity occur in directly proglacial positions. Valley

Figure 96.

Active sandur in the north arm of Tingin Fiord, east-central Baffin Island.

glacier tongues exist in many of the East Baffin valleys, though their direct influence on the major valley fills is often strongly attenuated by the recent moraines which characteristically surround their snouts. Ponding inside the moraine often severely affects the direct glacial sediment supply. Around the margins of the Barnes Ice Cap, activity is somewhat restricted at present since a large proportion of the total runoff from the ice cap passes into proglacial lakes.

It is difficult to judge relative "rates of activity" in absence of measurements. Condition of the rivers and of the surfaces appears to be the most useful guide. At Ekalugad Fiord, the entire T4 surface is currently active: that is, all areas are inundated sufficiently frequently to prevent the growth of any considerable vegetative cover or development of major non-fluvial features.

The valley sandur in the north arm of Tingin Fiord (Fig. 96) was also examined in some detail. This sandur is fed directly from a valley glacier which is, however, encircled by recent moraines with pond water inside. Although, to judge by the extent of braiding and long profile, the river seems to be much more active there than the ones at Ekalugad Fiord, in fact on portions of the distal end of the plain frost cracks appear, and bunch grass and other pioneer plants appear on the surface most of the way up its length. Similarly, the sandur in Itirbilung Fiord appears active (Fig. 97), but here again the presence of moraines and ponds inside the moraines does much to dampen the direct effects of glacier melt.

Sandurs in the south arm of Tingin Fiord (Fig. 98) and in Confederation Fiord (Fig. 99) are no longer subject to glacial runoff as there is no longer any significant quantity of ice in their watersheds. Here the rivers have begun to degrade, leaving the recent surfaces terraced. The rivers have nearly straight profiles over long distances and flow in restricted channel tones. A quasi-regular sinuosity is clearly evident in the river pattern at Confederation Fiord. These sandurs are inactive.

Although not as active as some sandurs in direct proglacial positions, the surface at Ekalugad Fiord represents a median level of activity for East Baffin Island. The Lewis Valley, though it represents a very small area, provides an example of the most active fluvial environment one is likely to find on Baffin Island today. The presence or absence of a mechanism for generating jökulhlaup type floods may very well mark the difference between surfaces that appear to be very active today and ones that do not.

In the introduction it was asserted that fluvial activity must be the dominant agent of physiographic change in nonglacierized areas in the Arctic. It remains to list the outstanding characteristics of fluvial activity:

1. The rivers flow for only about one-quarter of the year and in areas without permanent snow or ice fields the normal high water period is restricted to the spring freshet lasting two weeks to a month only. This may seem more extreme a situation than it is. In many regions with sharply seasonal precipitation regimes, the period of effective runoff is as small as or smaller than it is in the Arctic. In the Arctic the effective control is thermal.
2. Runoff is extremely flashy, since snowmelt runoff is controlled by the temperature, and often snowbanks in the water courses provide a major source of water. Summer storm runoff occurs virtually entirely on the surface due to the presence of permafrost in the ground.

Figure 98. Old sandurs in the south arm of Tingin Fiord.

3. The land is largely bare of vegetation, so that overland flow or runnels may be highly effective in initiating erosion and river bank erosion is easily accomplished. Thermal effects associated with frost melting often abet the process.
4. The relative effect of chemical dissolution is minimized (but is by no means minimal) with respect to the transport of clastic material. This is the characteristic by which arctic fluvial activity is distinguished from that of all other regions.

None of these characteristics is necessarily associated with the proximity of glacial events. Hence, it seems reasonable to assume that, even should relaxation from the last glacial episode be completed, a considerable tempo of fluvial activity will be maintained. However, the visible effects of fluvial activity will decrease with time. The major valley fills have required, for their construction, a rate of material supply that primary weathering processes will not support, so that depositional activity in a mature fluvial landscape in the Arctic would be much more localized than it recently has been. The sandurs remain the most impressive product of the response to an abrupt change in the nature of geomorphological processes at work in the landscape.

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APPENDIX A

Notation and Units

Metric units are employed throughout the text. Some field measurements were carried out in British units because of the instrument scales. Where metric equivalents have been introduced into the text, the exact British measurements are also given. Conventional abbreviations for units of measurements are as follow:

mm, cm, m, km	millimetre, centimetre, metre, kilometre
g, kg, M.T.	gram, kilogram, metric ton
s, yr	second, year
C	centigrade (degrees)

Additional conventional abbreviations are:

a.s.l.	above sea level (elevation)
B.P.	(years) before present, "present" is conventionally referred to the year A.D. 1950
ppm	parts per million (concentration)
$\emptyset$	phi unit ($-\log_2 x$, where x is the size in mm)
μ	micron (10^{-3} mm)

Because of various standard usages, some symbols have more than one meaning: the particular use is always clear in context. The normal units of measurement are quoted for all dimensioned quantities.

A	area, m^2; cross-sectional area, m^2; area of flow, m^2
a	coefficient in the relation $w_s = aQ^b$ (after Leopold and Maddock, 1953)
B	constant in the Meyer-Peter and Müller equation, $= 0.25$
b	exponent in the relation $w_s = aQ^b$ (after Leopold and Maddock, 1953)
C_i, c_i	various constants
C	concentration (of dissolved solute); ppm; $\simeq$ gm/m^3
c	coefficient in the relation $d_* = cQ^f$ (after Leopold and Maddock, 1953)
a, b, c	principal axes of clasts; lengths of principal axes (long, intermediate, short, respectively); mm
D	particle size (based on intermediate axis length); mm
D_m	mean particle size, mm; effective diameter of live bed materials; mm
D_{90} etc.	proportional size limits than which 90 per cent of the material is finer, etc., in a grain size distribution; mm
d	differential symbol; water depth; m
d_*	measure of mean flow depth, $= A/w_s$; m

e	base of natural logarithms, = 2.718
F	clast flatness measure (section 6), = $2c/(a+b)$ F-statistic
f	function . . . ; exponent in the relation $d_* = cQ^f$ (after Leopold and Maddock, 1953)
g	acceleration of gravity, = 980 cm s^{-2}
g_s	specific rate of bedload sediment transport, by weight; gm $s^{-1}m^{-1}$
g'_s	specific rate of bedload sediment transport, by submerged weight; gm $s^{-1}m^{-1}$
H	proportion of heavy minerals
h	number of nodes in a braided channel network
i, j	indices
j	exponent in the relation $Q_s = pQ^j$ (after Leopold and Maddock, 1953)
k	intrinsic permeability; darcys; coefficient in the relation $\bar{v} = kQ^m$ (after Leopold and Maddock, 1953)
k_r	coefficient of particle roughness on a smooth bed in the Meyer-Peter and Müller equation, = $26/D^{1/6}_{90}$
k_t	coefficient of mean flow resistance in the Meyer-Peter and Müller equation, = $1/n$
L	number of links in a braided channel network
M	number of lags
m	exponent in the relation $\bar{v} = kQ^m$ (after Leopold and Maddock, 1953)
N	number of observations
N'	equivalent number of independent observations in a serially correlated sequence
n	Manning's roughness coefficient
P	number of days since the last precipitation, plus one day (section 4)
p	number of division points in a braided channel network; coefficient in the relation $Q_s = pQ^j$ (after Leopold and Maddock, 1953)
Q	discharge; m^3s^{-1}
Q_{max}, Q_{min}	maximum/minimum discharge; m^3s^{-1}
Q_{gl}, Q_{gr}	glacier/groundwater derived runoff; m^3s^{-1}
q	specific discharge, = Q/w_s; $m^3s^{-1}m^{-1}$
Q_b, q_b	bedload sediment discharge; gm s^{-1}; gm $s^{-1}m^{-1}$

Q_c, q_c	dissolved sediment discharge; gm s^{-1}; gm $s^{-1}m^{-1}$
Q_s, q_s	suspended sediment discharge; gm s^{-1}; gm $s^{-1}m^{-1}$
R	hydraulic radius (section 5); m; clast roundness measure (section 6), $= 2r/a$ (after Cailleux)
r	coefficient of correlation; smallest radius of curvature in principal axis plane of clast (section 6); mm
r_i	serial correlation coefficient after i lags
r^2	coefficient of determination
S	clast intercept sphericity (section 6), $= \sqrt[3]{bc/a^2}$ (after Krumbein); concentration of suspended solids; ppm
s	hydraulic gradient
s_s	density ratio of solid to fluid, $= \rho_s/\rho_f$
S, s	standard error
T	Temperature, heat; ° C
t	time; t-statistic
u_i	serial correlation coefficient after i lags in a residual series
v	velocity; m s^{-1}
v_*	friction velocity, $= \sqrt{\tau_o/\rho_f}$; m s^{-1}
W	load in the channel (section 5.3.2)
w	width; m
w_s	fall velocity of sediment; m s^{-1}; surface width of channel; m
w_{eff}	effective width of channel (section 4); m
z	stream elevation (section 5.3.2); m; measure of stream depth (section 5.4.2 and section 6); m
z_c	critical depth, $= \tau_o/\rho g\theta$; m
x, y, z	Cartesian coordinates
$\mathbb{C}$	electrical conductivity; μmho
$\mathbb{C}_s$	specific conductance; μmho at 20° C
$\mathbb{F}$	Froude number
$\mathbb{f}$	Darcy-Weisbach resistance coefficient
α	probability level in confidence limits or statistical significance tests
β	constant in the long-profile equations (section 5.3.2)
Γ	specific weight, $= \rho g$ (kgm $m^{-2}s^{-2}$)
γ	specific gravity of fluid

γ_s	specific gravity of sediment, taken to be 2.65
Δ, δ	increment, difference
μ	dynamic viscosity of water (kgm $m^{-1}s^{-1}$)
ν	kinematic viscosity of water, $= \mu/\rho_f$ (m^2s^{-1})
ξ	random variate
Φ	bedload function, $= q_s/wD$
Ψ	reciprocal of the entrainment function, $1/\Psi = \tau_o/\gamma(s_s - 1)D$
$1/\Psi_c$	critical (threshold) value of the entrainment function
π	3.1416
θ	energy gradient of a stream; phase angle (section 3.1.3)
ρ	density; gm cm^{-3}
ρ_f, ρ_s	fluid/grain density; gm cm^{-3}
σ	standard deviation
τ	tractive force; newtons m^{-2}
τ_c	critical tractive force; newtons m^{-2}
τ_o	shear stress at the channel bed; newtons m^{-2}

Overbar indicates a mean value.

APPENDIX B

TABLE 1

Summary Data of the Lewis River Watershed[1]

	Land	Ice	Total
Flitaway Lake drainage area (km^2)	7.2	4.3	11.5
Direct Lewis River drainage area	15.8	178.1	193.9
Total drainage area	23.0	182.4	205.4
Proportion of total on ice			88.8 %
Area of Flitaway Lake (1964) (km^2)			0.63
Highest elevation in the watershed (m)			965
Gauge elevation (m)			230
Basin relief (m)			735
Mean elevation (m a.s.l.)			738
Hypsometric integral			0.624
Maximum runoff length (along a channel) (km)			25

[1] Data pertain to the guaged watershed. The stream gauge was located about 500 m upstream from the confluence of the Lewis and Striding Rivers.

TABLE 2

Summary Data of Ekalugad Sandur Watershed

	North Valley	Middle Valley	South Valley	Total Watershed
Total area (km^2)	188.59	105.94	89.84	384.37
Area of permanent snow and ice[1] (km^2)	21.41	16.10	43.91	81.42
Proportion	0.113	0.151	0.489	0.212
Basin relief (m)	1060	865	1025	1060
Mean elevation (m)	602	582	645	600
Hypsometric integral	0.560	0.665	0.576	0.512
Length of longest channel (km)	30.0	34.7	26.3	34.7

[1] Data for the area of permanent snow and ice are from 1960 air photography. Changes in this decade are probably relatively minor.

TABLE 3

The water budget of the Lewis River watershed: 1963-1965 (All data are expressed in terms of water equivalent)

Period	1963 25 June - 23 August		1964 31 May - 14 August		1965 29 May - 13 August	
Budget component	Specific[1] (cm)	Total ($10^6 m^3$)	Specific[1] (cm)	Total[2] ($10^6 m^3$)	Specific[1] (cm)	Total[2] ($10^6 m^3$)
Runoff (Q)	81.8	168.0	29.4	60.5	48.5	99.6
Precipitation	7.4	14.3	9.1	17.6	13.6	26.4
Snow storage: May	13 (est)	25 (est)	21.1	40.9	11.8	22.8
August	-	-	-	-	-	-
Flitaway net outflow	20.0	2.3	17.8	2.1	7.8	0.9
Ice ablation	63.3	112.8	-3.5	-6.3	31.0	55.3
Sum of the elements of runoff composition (C)	75.2	154.4	26.4	54.3	51.3	105.3
Discrepancy (Q-C)		+13.6		+6.2		-5.7
Runoff as a percentage of runoff composition		109.0		111.5		94.8

1 Note that the specific budget values will not add up, as they are computed for the area to which they apply, which is not the same in all cases:

- runoff is computed over the entire watershed;
- precipitation is computed over the direct Lewis watershed: that is, the total watershed less the Flitaway drainage zone;
- snow storage is computed over the direct Lewis watershed;
- Flitaway net outflow is computed over the Flitaway drainage zone;
- ice ablation is computed over that portion of the glacier which falls in the direct Lewis watershed;
- total runoff composition is computed over the entire watershed.

2 Partial budget only: significant runoff occurred after the end of measurements.

TABLE 4

Proportional Composition of Lewis River Runoff, 1963-1965

Source	1963	1964[3]	1965[3]
Snowmelt from winter accumulation[1]	14.9%[2]	57.2	22.9
Precipitation during the runoff period	8.5	29.1	26.5
Ice ablation[1]	67.0	-	55.5
Flitaway Lake outflow	1.4	3.5	0.9
Totals	91.8	89.8	105.8
Discrepancy of the water budget over measured runoff	-8.2	-10.2	+5.8

[1] In 1964 the negative ablation indicated that a net addition occurred to ice storage. This amount has been arbitrarily deducted from the contribution of snowmelt, since the accumulation was in the form of superimposed ice formation at the base of the snowpack (cf. Müller, 1962; Anonymous, 1967).

[2] Based on estimated accumulation only (cf. Table 3).

[3] Partial budget only: significant runoff occurred after the end of measurements.

TABLE 5

Total Runoff from the Ekalugad Fiord Watersheds

River	Specific (cm)	Total ($10^6 m^3$)
North, 1967[1]	61.9	121.0
Middle, 1967[1]	54.2	58.1
South, 1967[1]	62.0	57.2
South, 1968[2]	18.6	16.7

[1] Period July 1 - September 8. Data for July 1-4 and September 8 were estimated for Middle and South Rivers.

[2] Period June 29 - August 14.

TABLE 6

Calculations of Groundwater Discharge on Ekalugad Sandur

Intrinsic permeability: (darcys)	2.6×10^2	9.9
Upper sandur in South valley: local gradient: 0.0097 width of sandur: 700 m	$0.10\ m^3 s^{-1}$	0.004
Distal end of the sandur near the sea: local gradient: 0.0066 width of sandur: 1500 m	0.14	0.006

TABLE 7

Variance Reduction of Diurnal Runoff by Harmonic Analysis

No. Harmonics	r_i^2	r_T^2	r_i^2	r_T^2	r_i^2	r_T^2	Period (hrs)
Lewis River	1963		1964		1965		
1	.950	.950	.920	.920	.937	.937	24
2	.016	.964	.068	.988	.033	.970	12
3	.014	.978	.005	.993	.013	.983	8
4	.009	.987	.002	.995	.004	.987	6
Ekalugad Fiord, 1967	North		Middle		South		
1	.968	.968	.967	.967	.827	.827	24
2	.013	.981	.009	.976	.056	.883	12
3	.007	.988	.007	.983	.031	.914	8
4	.003	.991	.004	.987	.019	.933	6

r_i^2 = additional variance reduction achieved by the i^{th} harmonic.

r_T^2 = total variance reduction achieved by the first i harmonics.

TABLE 8

Serial Correlation of Daily Runoff

Data Series	N	r_1	u_1	N'	Variance removed in seasonal trend
Lewis River 1963	64	0.925	0.131	58	91.7%
Lewis River 1964	64	0.721	0.238	44	65.2%
Lewis River 1965	64	0.845	0.327	36	77.1%
Ekalugad Rivers: 1967					
North	64	0.854	0.352	33	82.0%
Middle	64	0.869	0.401	26	81.4%
South	64	0.521	0.178	49	42.1%

r_1 = one-step serial correlation before removal of seasonal trend.

u_1 = one-step serial correlation of the residual series after removal of seasonal trend.

N' = equivalent number of independent observations in the residual series.

TABLE 9A

Summary Statistics of Summer Climate: Field Stations

Period	Mean Temperatures (°C.) Max.	Min.	Mean	Precipitation (mm)	Relative Humidity (%)	Cloud (tenths)	Wind (km/hr)
A. Lewis River 1963							
June (last 17 days)	7.9	-0.4	4.1	4		6.5	
July	9.7	2.6	6.6	57		6.5	
August (first 24 days)	10.2	3.5	7.0	13		8.0	
Season	9.4	2.2	6.1	74		7.0	
1964							
June (last 23 days)	4.0	-0.9	1.7	22	78.4	9.3	10.7
July	9.3	3.0	5.9	56	73.6	7.7	11.4
August (first 14 days)	7.6	2.0	4.9	13	69.9	6.4	16.3
Season	7.1	1.5	4.3	91	74.5	8.0	13.0
1965							
June	4.8	-1.8	1.4	8	70.3	7.1	12.8
July	8.4	3.9	6.2	110	67.4	7.2	14.7
August (first 12 days)	7.6	2.8	5.4	18	78.9	9.2	17.4
Season	6.8	1.3	4.1	136	70.8	7.4	14.2
B. Ekalugad Fiord							
1967							
June (last 28 days)	3.4	-0.5	1.4	43	83.7	9.3	17.4
July	11.4	3.8	7.6	54	80.4	5.9	11.9
August	7.5	3.5	5.3	125	86.4	9.0	15.3
September (first 10 days)	2.3	-2.1	0.1	30	83.0	8.4	18.0
Season	7.1	2.0	4.4	252	83.5	8.0	15.1
1968							
June	4.5	-0.5	2.0	25	89	8.0	13.4
July	6.3	1.4	3.6	65	92	8.4	14.4
August (first 18 days)	6.0	1.6	3.8	10	80	7.3	12.5
Season	5.5	0.7	3.0	100	88	8.0	13.6

TABLE 9B

Summary of Summer Climate: East Baffin Stations

Period	Mean Temperature (°C.) Hooper	Clyde	Dewar	Exceedance Probability (%) (Dewar)	Precipitation (mm) Hooper	Clyde	Dewar	Exceedance Probability (%) (Dewar)
1963								
June (last 17 days)	0.6	1.2	0.1		9	T	1	
July	2.8	3.3	3.5	80	4	23	28	62
August (first 24 days)	4.3	4.7	6.0		6	20	12	
Season	2.8	3.3	3.5		19	43	41	
1964								
June (last 23 days)	3.4	1.2	-1.6		T	T	19	
July	3.0	4.0	3.7	76	20	13	85	4
August (first 14 days)	0.1	4.0	1.8		28	10	98	
Season	2.5	3.1	1.5		48	23	202	
1965								
June	-0.2	0.4	-1.3	50	25	12	13	60
July	3.4	2.7	5.2	60	36	43	35	50
August (first 12 days)	1.8	1.7	2.8		5	27	21	
Season	1.7	1.6	2.1		66	82	69	
1967								
June (last 28 days)	-1.6	-0.3	-2.2		35	11	39	
July	6.2	5.2	6.0	51	5	1	53	28
August	2.2	3.2	2.8	68	42	6	95	25
September (first 10 days)	-2.1	-0.8	-4.2		26	9	13	
Season	2.0	2.4	1.7		108	27	200	
1968								
June	1.2	1.8	2.5	1	17	24	35	14
July	2.7	2.4	3.5	80	32	49	35	50
August (first 18 days)	0.9	2.6	3.7		19	19	4	
Season	1.7	2.2	3.2		68	92	74	

Data abstracted from Canada Department of Transport, Monthly Record: Meteorological Observations in Canada. 1959-1968.

TABLE 10

Weight of Material Removed from the T2 Terrace Scarp, 1967-1968

Location	Eroding Scarp Length (m)	Weight of Material Removed[1] (metric tons)			Mean Scarp Retreat (m)		
		25 June - 1 Sept. 1967	1 Sept. 67 - 25 June 1968	25 June - 17 Aug. 1968	25 June - 1 Sept. 1967	1 Sept. 67 - 25 June 1968	25 June - 17 Aug. 1968
North River	839	4,828	1,375	529	2.56	0.73	0.28
Middle River	2,395	1,239	1,420	6,925	0.23	0.26	1.29
South River	8,860	35,360	63,452	27,913	1.77	3.18	1.40
Total/Mean	12,094	41,427	66,247	35,367	1.52	2.44	1.30

[1] Specific weight of the material is assumed to be 2.25

TABLE 11

Summary Data on Major Ions in Baffin Island Streams, with Comparisons from other Waters

Source	Cations (%)				Anions (%)			Total Load ppm	pH
	Ca^{++}	Mg^{++}	Na^{+}	K^{+}	HCO_3^{-}	SO_4	Cl^{-}		
Lewis (4 samples during snowmelt period)	39.7	11.0	44.6	4.7	13.2	12.3	74.5	56.83	6.0
Lewis (4 samples during high flow)	33.1	7.5	47.0	12.4	10.5	41.7	47.8	5.33	5.7
Lewis (3 samples of Supraglacial water[1])	35.9	6.2	45.3	12.5	28.5	39.6	31.9	7.33	6.0
Ekalugad Fiord. North River (9 samples)	27.2	13.2	36.6	22.9	16.1	65.9	18.0	4.93	5.8
Middle River (10 samples)	41.0	12.5	28.3	18.2	2.1	85.2	12.7	4.55	5.2
South River (7 samples)	53.8	11.9	18.1	16.3	0.0	87.6	12.4	7.87	5.7
Baffin rainwater[2]	26.3	21.1	0.0	52.6	0.0	87.5	12.5	1.20	5.9
Coppermine River at mouth, N.W.T.[3]	63.8	28.4	4.9	3.9	93.2	5.4	1.4	42	7.4
Great Whale River at mouth, Quebec[3]	55.6	11.1	22.2	11.1	26.7	36.7	36.6	9.5	5.9
Koksoak River at Fort Chimo, Quebec[3]	58.0	20.0	20.0	2.0	77.8	16.8	5.4	17	7.1
Teton Glacier,[4] Wyoming, U.S.A.	43.2	12.6	16.3	27.9				2.52	6.4
Nigardsbreen, Norway[4]	59.0	4.0	24.0	13.0				1.58	4.1
Panarak, Jötunheim[4] Norway	63.1	4.0	12.1	20.7				3.42	4.2
Bover, Jötunheim[4] Norway	53.0	6.1	12.1	28.8				1.80	4.5
"Average" fresh water[5]	64	17	16	3	73	16	10		
Average composition of igneous rocks[6]	32.6	18.8	25.4	23.2					

[1] Probably some admixture of rainwater.

[2] Amongst anions, there is, in fact, a significant proportion of NO_3^{-}. Proportion of anions is SO_4^{--}, 79.8%; Cl^{-}, 11.4%; and NO_3^{-}, 8.8%.

[3] Data from Thomas (1964).

[4] Data from Keller and Reesman, 1963 for single samples.

[5] Data from Gorham (1961).

[6] Data from Hem (1959).

TABLE 12

Relationships Between Concentration of Dissolved Materials (C) and Discharge (Q), and Dissolved Load Rating Equations

Rivers	Relationship with Discharge (Q) (ppm) C =	r^2	Improved Relationship	r^2	Predictor Equation for Sediment Discharge (gms/sec) Qc =	r^2 (apparent)
Lewis	$36.87Q^{-0.52}$	0.919	$59.90Q^{-0.76}Q_{gr}^{0.56}$	0.974	$36.87Q^{0.48}$	0.908
Ekalugad Fiord:						
North River, with storm event of 14 July, 1967	$5.98Q^{-0.05}$	0.129				
North River, without storm event	$6.61Q^{-0.10}$	0.343	$6.54Q^{-0.05}P^{-0.01}$	0.600	$6.54Q^{0.95}P^{-0.01}$	0.990
Middle River	$7.40Q^{-0.17}$	0.353	$7.15Q^{-0.17}P^{-0.003}$	0.375	$7.40Q^{0.83}$	0.928
South River	$17.77Q^{-0.32}$	0.556	$18.15Q^{-0.30}P^{-0.005}$	0.606	$18.15Q^{0.70}P^{-0.005}$	0.874

P = number of days since the last rain, plus one day (i.e. the day of the last rain).

Q_{gr} = "ground-environment" discharge.

TABLE 13

Discharge of Dissolved Materials in Baffin Island Streams

River	Sediment Transport (metric tons)	Material Removed after Adjustment for Rain and Melt-water Contents[1]	Material Removed/ sq km of Watershed	Volume Removed (m^3)[2]	Equivalent Surface Lowering of Watershed (mm/1000 yr)
Lewis, 1963	930	666	3.24	251	1.22
Lewis, 1964	501	436	2.12	165	0.80
Lewis, 1965	647	482	2.35	182	0.89
Ekalugad Fiord:					
North River, 1967	599	381	2.02	144	0.76
Middle River, 1967	263	193	1.82	73	0.69
South River, 1967	431	363	4.03	137	1.52
Ekalugad Total, 1967	1293	937	2.44	354	0.92
South River, 1968	186	166	1.85	63	0.70

1 At Ekalugad Fiord adjustment was made to the total runoff from precipitation only, since the bulk of the runoff was derived from precipitation or recent snowfall. At Lewis River a different correction was made for the runoff calculated to be from glacier ice melt. The basic data are 1.2 ppm total dissolved solids for rainwater and 1.9 ppm for runoff from the glacier surface. Since the only rainwater sample was from Ekalugad Fiord, it is possible that the corrections for cyclic salts at Lewis River are nonrepresentative, especially since there is some possibility (cf. p. 46) of the rainwater composition at Lewis River being different from that indicated at Ekalugad Fiord.

2 Assuming a specific gravity of 2.65 in bedrock (i.e., no allowance made for voids in unconsolidated materials).

TABLE 14

Relations between Suspended Sediment Concentration (S) and Suspended Sediment Transport (Q_s) and Discharge (Q)

River	Relationship of Concentration and Discharge S =	r^2	Optimum Relation S =	r^2	Rating Equation for Sediment Discharge Computations (gms/sec) Q_s =	r^2 (apparent)
Lewis	$10.20Q^{0.82}$	0.737	same		$10.20Q^{1.82}$	0.932
Lewis Lateral	$23.87Q^{1.03}$	0.787	$38.82Q^{0.97}P^{-0.58}$	0.832		
North	$9.11Q^{0.77}$	0.400	$21.60Q^{0.76}\mathbf{C}^{-0.14}P^{-0.29}$	0.511	$9.74Q^{1.84}P^{-0.38}$	0.800
Upper North	$3.48Q^{0.95}$	0.620	$1.21Q^{1.14}P^{-0.29}\mathbf{C}^{5.14}(dQ/Q)$	0.795	$3.48Q^{1.95}$	0.872
Middle	$6.19Q^{0.76}$	0.310	$0.0038Q^{1.21}\mathbf{C}^{1.82}P^{-0.26}$*	0.457	$6.39Q^{1.89}P^{-0.38}$	0.718
Upper Middle	$1.30Q^{1.48}$	0.280	same		$1.30Q^{2.48}$	0.709
South	$3.72Q^{0.82}$	0.465	$3.58 \times 10^{-8}Q^{1.78}\mathbf{C}^{3.05}T^{-0.85}P^{-0.22}$*	0.676	$4.08Q^{1.91}P^{-0.31}$	0.765
Upper South	$0.36Q^{1.46}$	0.482	$2.32Q^{1.50}P^{-0.38}$	0.539	$0.43Q^{2.50}P^{-0.38}$	0.755

* The very small value of the coefficient in these two equations results from the large power on the conductivity term.

TABLE 15

Summary of Suspended Sediment Transport Observations

River	Sediment Transport (metric tons)	Sediment Yield metric tons/km^2 Watershed	Volume[1] Removed (m^3)	Equivalent Lowering of Watershed (mm/1000 yrs)
Lewis, 1963	47,761	232.5	18,023	87.7
1964	13,978	68.1	5,275	25.7
1965	26,067	126.9	9,837	47.9
Ekalugad Fiord:				
North, 1967	21,659			
Upper North, 1967	16,325	86.5	6,160	32.7
Middle, 1967	3,852			
Upper Middle, 1967	6,934	65.5	2,616	24.7
South, 1967	3,523			
Upper South, 1967	4,378	48.8	1,652	18.4
Total Ekalugad, 1967	27,637	71.9	10,429	27.1
South, 1968	239			
Upper South, 1968	72	0.8	27	0.3

[1] Assume s.g. = 2.65 in bedrock (i.e. no allowance made for voids in unconsolidated materials).

TABLE 16

Sediment Balance on the Active Sandur at Ekalugad Fiord:Suspended Sediment Load

River	Net Sediment Exchange (metric tons)	Gross Volume[1] (m^3)
North, 1967	-5335	-2657
Middle, 1967	3082	1535
South, 1967	855	426
Total Sandur, 1967	-1397	-696
South, 1968	-167	-83

[1]Assumes s.g. = 2.0 (i.e. packing intermediate between open and close).

TABLE 17

Critical Tractive Force Estimates from Bed Load Transport Measurements

	Mean of Large Cobbles on bed (mm)	Mean Size of Bed Load (mm)	Largest Cobble in Load (mm)	$v_*(m\ s^{-1})$	$\tau(m\ s^{-2})$	Flow competence from Shields' criterion (mm)
1	93.2	1.1	39	0.18	32.3	40
2	65.8	0.9	13	0.22	48.2	90
3	66.0	7.0	32	0.29	83.9	110
4	100.7	0.7	6.5	0.21	43.0	60
5	76.2	1.3	12	0.17	28.2	35
7	56.8	0.75	10	0.13	17.6	20
8	56.8	1.4	10	0.15	23.7	35
9	18.1	0.2	5	0.07	5.4	7
10	63.2	0.95	10	0.13	15.6	18

TABLE 18

Values of the Entrainment Function at the Threshold of Bed Motion, as Measured from the Critical Tractive Force Diagram[1]

Data	$1/\psi_c$	Remarks
Ekalugad Fiord	0.023	Marked cobbles set on top of bed.
Chitty Ho	0.023	Data in Leopold, Wolman and Miller (1964).
Flume data for gravel	0.037	Kellerhals (1967). Criterion for absolute stability. Lag gravels.
Elbow River	0.039	Absolute stability (Hollingshead, 1968).
Lane Canals	0.045	D_m is probably too large due to Lane's selective sampling system. Range is 0.029 to 0.528. Probably referred to absolute stability.
Elbow River	0.055	Shields' threshold (Hollingshead, 1968).
Indian Canals	0.064	Canals became more underloose with increasing D. At D = 60 mm, $1/\psi_c$ = 0.200.
Ekalugad Fiord	0.118	From transport measurements.

[1] Mean values or modal values are listed for scattered data.

TABLE 19

Bedload Transport Equations Based on the Meyer-Peter and Müller Formula

Section	θ $(\times 10^2)$	D_{65} $(m \times 10^3)$	D_{90} $(m \times 10^3)$	$\bar{v} - Q$ Relation	Meyer-Peter and Müller Function[1]	W_{eff} (m)	$Q_c (m^3 s^{-1})$
Lewis River	1.09	9.91	145	$0.20Q^{0.52}$	$s'(0.1152Q^{0.78}-1.318)^{3/2}$	29.7	22.8
Ekalugad Fiord:							
Lower South River	0.94	3.89	183	$0.138Q^{0.497}$	$s'(0.0647Q^{0.75}-0.5174)^{3/2}$	84	15.1
Upper South River	0.33	3.89	166	$0.276Q^{0.498}$	$s'(0.1433Q^{0.75}-0.5174)^{3/2}$	34	5.5
Middle River	0.22	3.89	233	$0.107Q^{0.795}$	$s'(0.0340Q^{1.20}-0.5174)^{3/2}$	20	9.7

[1] The value $s' = \frac{2.65}{1.65}$ converts submerged weight to dry weight.

TABLE 20

Summary of Bed Load Sediment Transport

River	Sediment (metric tons)	Sediment Yield: metric tons/km^2 of Watershed	Volume Removed[1] (m^3)	Equivalent Surface Lowering of Watershed (mm/1000 yrs)
Lewis, 1963	215,542	1,049	81,332	396.0
Lewis, 1964	48,543	236	18,318	89.2
Lewis, 1965	99,094	482	37,394	182.1
Ekalugad Fiord:				
Middle, 1967	60,347	570	22,772	215.0
South, 1967	34,487			
Upper South, 1967	100,041	1,114	37,751	420.4
South, 1968	66			
Upper South, 1968	2,577	29	972	10.8

1 Assumes s.g. = 2.65 in bedrock (i.e. no allowance made for voids in unconsolidated materials).

TABLE 21

Summary of Total Sediment Transport

River	Total Sediment Transport (metric tons)	Sediment Yield metric tons/km^2 of Watershed	Volume Removed[1] (m^3)	Equivalent Surface Lowering of Watershed (mm/1000 yrs)	Proportional Distribution of Sediment Load		
					Solution	Suspension	Bed load
Lewis, 1963	264,233	1,286	99,711	485	.004	.181	.816
Lewis, 1964	63,022	307	23,782	116	.008	.222	.770
Lewis, 1965	125,809	613	47,475	231	.005	.207	.788
Mean	151,021	735	56,989	277			
Ekalugad Fiord:							
North, 1967	Data incomplete	No bed load observations					
Middle, 1967	64,461	609	24,325	230	.004	.060	.936
Upper Middle, 1967	Data incomplete	No bed load observations					
South, 1967	38,442				.011	.092	.897
Upper South, 1967	104,851	1,168	39,566	441	.004	.042	.954
South, 1968	492				.397	.487	.135
Upper South, 1968	2,835	32	1,070	12	.066	.026	.909

[1] Assumes s.g. = 2.65 in bedrock (i.e. no allowance made for voids in unconsolidated materials)

TABLE 22

Proportional Distribution of Discharge and Sediment Transport for the highest flow events

River	Days of Record		Cumulative proportional discharge/ transport during highest flow days 1	2	3	4	5
Lewis, 1963	70	Q	.053[a]	.092[a]	.130[b]	.166[b]	.200
		Q_c	.032[a]	.059[b]	.085	.111[a]	.136[b]
		Q_s	.088[a]	.146[a]	.198[b]	.248[b]	.291
		Q_b	.099[a]	.160[a]	.217[b]	.270[b]	.317
		Q_T	.097[a]	.157[a]	.213[b]	.265[b]	.312
Lewis, 1964	64	Q	.116	.199[a]	.269[a]	.335[a]	.396
		Q_c	.047	.092[a]	.133[a]	.173[a]	.211
		Q_s	.244[a]	.364[b]	.452[a]	.530[b]	.599[b]
		Q_b	.313[a]	.479[b]	.586[a]	.686[b]	.769[b]
		Q_T	.296[a]	.451[b]	.553[a]	.648[b]	.727[b]
Lewis, 1965	72	Q	.104[a]	.177[a]	.236[a]	.281[a]	.326
		Q_c	.048[a]	.089[a]	.124[a]	.158	.190[a]
		Q_s	.248[a]	.381[a]	.457[a]	.572[a]	.563
		Q_b	.293[a]	.457[a]	.555[a]	.623[a]	.685
		Q_T	.282[a]	.440[a]	.532[a]	.597[a]	.657
Lewis, 3 seasons	206	Q	.032	.059	.081	.102	.122
		Q_c	.015	.029	.042	.054	.066
		Q_s	.074	.121	.161	.200	.232
		Q_b	.080	.139	.184	.226	.262
		Q_T	.078	.135	.179	.220	.255
North, 1967	70	Q	.065[a]	.120[a]	.172	.217[b]	.262
		Q_c	.062[a]	.114[a]	.164[b]	.208[b]	.252
		Q_s	.161[a]	.282[a]	.384[b]	.447[b]	.504[a]
		Q_b			no data		
		Q_T			not calculated		
Middle, 1967	66	Q	.059[a]	.107	.153[a]	.197[b]	.240[b]
		Q_c	.051[a]	.095	.137[a]	.178[b]	.217[b]
		Q_s	.156[a]	.257[a]	.326[b]	.382[b]	.433
		Q_b	.168[a]	.269	.368[a]	.454[b]	.533[b]
		Q_T	.167[a]	.266[a]	.364	.448[b]	.526[b]
South, 1967	66	Q	.103	.155[a]	.200[a]	.245[a]	.288[a]
		Q_c	.059	.099[a]	.135[a]	.171[a]	.207[a]
		Q_s	.407[c]	.471[c]	.530	.575[a]	.612[a]
		Q_b	.477[c]	.579[c]	.653[c]	.724[a]	.793[a]
		Q_T	.466[c]	.562[a]	.635[c]	.702[a]	.767[a]

TABLE 22 (cont.)

River	Days of Record		Cumulative proportional discharge/ transport during highest flow days 1	2	3	4	5
Upper South, 1967	66	Q_s	.576[c]	.628[c]	.665[a]	.701[a]	.729
		Q_b	.254[c]	.342[a]	.413[a]	.484[a]	.550[a]
		Q_T	.268[c]	.353[a]	.423[a]	.491[a]	.556[a]
South, 1968	47	Q	.070[a]	.113[b]	.147[a]	.179[b]	.212
		Q_c	.049[a]	.084[b]	.114[a]	.143[b]	.171
		Q_s	.181[a]	.268[b]	.320[b]	.357[a]	.393
		Q_b	1.000	----	----	----	----
		Q_T	.241[a]	.297[b]	.333[b]	.362[a]	.389
Upper South, 1968	47	Q_s	.298[a]	.406[b]	.462[b]	.499[a]	.533
		Q_b	.655[a]	.811[b]	.856[b]	.891[a]	.920
		Q_T	.648[a]	.805[b]	.853[b]	.890[a]	.921

Small case letters indicate sequences of consecutive days

TABLE 23

Summary of channel geometry data

A. Means and variances

River	$(\overline{w_s/d_*})$	σ_{w_s/d_*}	Coefficient of variation	Range of w_s/d_*	N
Lewis River- all channels	69.5	70.3	1.01	14.1-383.3	25
Lewis River- sandur surface channels only	80.6	74.4	0.92	17.8-353.3	20
Ekalugad Fiord North River	105.6	74.8	0.71	13.2-213.6	11
Middle River- all channels	153.6	99.7	0.65	25.4-406.7	15
Middle River- sandur surface channels only	119.3	79.3	0.67	25.4-217.4	4
South River	79.3	44.8	0.56	23.3-182.9	42
Small channels	43.1	18.8	0.44	15.6-67.9	10

B. Mean Values of w_s/d_* in Braiding and Non-braiding Reaches

River	Braiding	Non-braiding
Lewis	86.8	81.2 (sandur surface channels only)
Ekalugad Fiord North River	122.6	85.1
Middle	--	199.3 (sandur surface channels only)
South	69.8	84.5
Others	40.4	44.9
All	81.3	82.1

TABLE 24

Data of Hydraulic Geometry in Baffin Sandur Streams: $x = c_1 Q^{c_2}$

River	w_s	d_*	$\bar{v}$	A
A - Intercepts (c_1)				
Upper Lewis	18.01	0.12	0.46	2.17
Middle Lewis	12.15	0.21	0.39	2.56
Lower Lewis	15.13	0.30	0.22	4.55
Upper Triangle	17.00	0.18	0.33	3.03
Lower Triangle	11.77	0.28	0.30	3.33
Lewis Base	17.50	0.30	0.20	5.00
Middle 2	21.73	0.42	0.11	9.09
Middle 3	41.78	0.21	0.11	9.09
South 1	53.35	0.13	0.14	7.14
South 2	22.04	0.19	0.23	4.35
South 3	51.26	0.15	0.13	7.69
South 4	26.32	0.22	0.17	5.88
South 5	49.00	0.16	0.13	7.69
South 6	17.70	0.21	0.27	3.70
B - Exponents (c_2)				
Upper Lewis	0.18	0.40	0.42	0.58
Middle Lewis	$\overline{0.41}$	$\underline{0.10}$	0.49	0.51
Lower Lewis	0.10	0.30	$\overline{0.60}$	$\underline{0.40}$
Lewis Base	0.17	0.32	0.51	0.49
Upper Triangle	$\underline{0.09}$	$\overline{0.56}$	$\underline{0.35}$	$\overline{0.65}$
Lower Triangle	0.32	0.21	0.47	0.53
Mean	0.212	0.315	0.473	0.527
Middle 2	0.065	0.145	0.79	0.21
Middle 3	0.07	0.15	0.78	0.22
Mean	0.068	0.147	0.785	0.215
South 1	0.15	0.36	0.49	0.51
South 2	0.30	0.26	0.44	0.56
South 3	0.29	0.29	$\underline{0.42}$	$\overline{0.58}$
South 4	$\overline{0.33}$	$\underline{0.25}$	$\underline{0.42}$	$\overline{0.58}$
South 5	$\underline{0.04}$	$\overline{0.36}$	$\overline{0.60}$	$\underline{0.40}$
South 6	0.23	0.27	0.50	0.50
Mean	0.223	0.298	0.478	0.522

Over-bar and under-bar notations indicate maximum and minimum values

TABLE 25

Comparison of Exponents in Sectional Hydraulic Geometry

River	w_s	d[1]	$\bar{v}$	A	Source
Baffin sandurs (mean of Lewis and Ekalugad data, excluding Middle River)	0.22	0.31	0.48	0.52	
$\alpha = 0.05$ confidence range on Baffin sandur means	0.245-0.195	0.344-0.276	0.525-0.435	0.565-0.475	
Gauge 1: Hoffellssandur	0.17	0.39	0.44	0.66	Arnborg (1955) data (computed by present writer)
White River, Mt. Rainier (112 channels)	0.38	0.33	0.27	0.73	Fahnestock (1963)
Ephemeral streams in semi-arid United States	0.29	0.36	0.34	0.66	Leopold and Miller (1956)
U.S. Midwest	0.26	0.40	0.34	0.66	Leopold and Maddock (1953)
All U.S. (158 stations)	0.12	0.45	0.43	0.57	Leopold, Wolman and Miller (1964)
Flume channels (D_{50} = 0.0022 ft.)	0.50	0.39	0.16	0.84	Wolman and Brush (1961)
Non-cohesive river: Langbein theory	0.50	0.23	0.27	0.73	Langbein (1964)

[1] d is variously defined, but is normally A/w_s, as in this study

TABLE 26

Bed Load transport at very high flow in the measurement sections

River	Q (m^3s^{-1})	Transport per metre width (kgms/sec dry weight)	Volume transport per metre width (m^3/sec)	v (m/s)	Depth of moving material (cm)
Lewis	250	31.12	.0117	.43	2.7
Middle	50	9.20	.0035	.13	2.7
Lower South	100	3.07	.0012	.25	0.5
Upper South	100	12.87	.0049	.15	3.3

TABLE 27

Limiting Values of $\sqrt{8/ff}$ for the Hydraulic Geometry of the Sandur Sections Based on $\sqrt{8/ff} = 8.4(d_*/D)^{1/6}$

Section	Channel State[1]	Reference Discharge	Predicted Datum (from hydraulic geometry)	D_{90} Observed[2]	D_{65} of Transported Material[3]	0.5 mm[4] (medium sand)	Apparent Stability[5]
Upper Lewis	P	$100 m^3 s^{-1}$	12.81	13.14	17.04	24.23	Stable
Middle Lewis	B	100	20.39	11.00	15.07	22.23	Unstable
Lower Lewis	P	100	57.48	13.55	18.47	25.57	<u>Unstable</u>
Lewis Base	S	250	8.47	13.32	19.78	26.77	Stable
Upper Triangle	S	100	3.80	15.69	20.83	27.70	Stable
Lower Triangle	P	100	12.35	12.45	17.16	24.35	Stable
Middle 2	P	50	17.80	11.31	20.14	24.42	Neutral
Middle 3	B	50	64.77	10.93	18.25	22.87	<u>Unstable</u>
South 1	B	100	4.99	11.54	19.81	24.15	Stable
South 2	B	100	6.95	11.36	19.50	23.90	Stable
South 3	B	100	4.92	10.90	19.22	23.68	Stable
South 4	P	100	21.68	12.45	19.85	24.19	Unstable
South 5	P	100	15.63	11.98	20.44	24.67	Neutral
South 6	P	100	17.81	11.62	20.00	24.31	Neutral

1 B - indicates a braided channel
S - indicates a stable, single channel
P - indicates an apparent propensity for anastomosis, but no well defined braiding present.

2 Same particle size datum as used to construct Fig. 45, based on observed particle sizes at moderate flow levels.

3 Data derived from Table 19.

4 An extremely conservative value, almost certainly not approached in fact: equation was $\sqrt{8/ff} = 10.86(d_*/D)^{1/9}$, appropriate for $d_*/D >> 100$ (Ackers, 1958).

5 Stable if predicted datum exceeds no limiting datum;neutral if it exceeds only the observed D_{90} datum; unstable if it exceeds the D_{65} datum; <u>unstable</u> if it exceeds all data.

TABLE 28

Fitting Equations for Long Profile Trends: Baffin Sandur Streams

Stream		S_{c_1}[1]	$c_2(x) \times 10^2$	$S_{c_2} \times 10^2$	$c_3(x^2) \times 10^4$	$S_{c_3} \times 10^4$	No. Stations
Lewis River	-0.9652	0.3636	0.2139	0.0230	0.0015	0.0003	72
Triangle River	-1.4381	0.1039	0.2784	0.0025	--	--	70
Ekalugad River: Bed							
North I	-0.1612*	0.3748	0.1068	0.0167	0.0008	0.0002	68
North II	6.1689*	5.1318	0.0746*	0.0730	0.0020	0.0003	56
Middle I	-1.4002	0.4991	0.1823	0.0374	0.0002*	0.0006	40
Middle II	7.8499	1.1470	-0.03018*	0.0190	0.0011	0.0001	92
South I	-1.4731	1.2755	0.0513	0.0146	0.0005	0.00004	118
Surface							
North I	0.4417*	0.3748	0.1103	0.0167	0.0008	0.0002	68
North II (L)	7.0920*	5.1318	0.0783*	0.0730	0.0020	0.0003	56
Middle I	-0.8576*	0.4991	0.1737	0.0374	0.0003*	0.0006	40
Middle II (L)	8.7532	1.1470	0.0407	0.0190	0.0011	0.0001	92
South I	-0.9187	0.3870	0.1410	0.0183	0.0008	0.0002	64
South II	9.5378	1.2755	0.0476	0.0146	0.0006	0.00004	118

* Not significantly different from zero at $\alpha = 0.05$

[1] Standard error

TABLE 29

One-step Markov Relations for Long Profile Characteristics

River	Characteristic	r^2	Markov equation (Upstream Control)	Error[1]	u_1[3]	Range[2]
Lewis Triangle	Bed Elevation	0.480	0.723x-0.013	.604	0.136	7
North River	Bed Elevation	0.622	0.771x+0.018	.698	-0.123	10
	Surface Elevation	0.467	0.676x+0.001	.568	0.036	7
	Slope	0.551	0.742x+0.006	.020	0.132	8
	Width Deviation	0.419	0.693x+0.308	19.1	-0.047	6
	Talweg Deflection	0.558	0.748x+30.5	40.4	0.371**	8
Middle River	Bed Elevation	0.338	0.580x-0.002	.674	0.076	5
	Surface Elevation	0.539	0.739x-0.002	.508	0.082	8
	Slope	0.388	0.623x+0.005	.016	0.204**	5
	Width Deviation	0.595	0.866x+0.596	25.9	0.162**	9
	Talweg Deflection	0.681	0.825x+19.9	38.4		12
South River	Bed Elevation	0.478	0.686x-0.005	.833	0.054	7
	Surface Elevation	0.597	0.772x-0.002	.657	0.145	9
	Slope	0.305	0.552x+0.007	.020	0.228**	4
	Width Deviation	0.530	0.956x+0.689	23.7	-0.050	8
	Talweg Deflection	0.817	0.900x+9.3	25.3	0.239**	23

1 $\alpha = 0.05$ error of estimate of the dependent variable; $t_{0.05}s$, where t refers to student's t and s is the standard error of estimate.

2 $R_1^n < 0.1$

3 ** Significant at $\alpha = 0.05$;

TABLE 30

Apparent Periodicities (Wavelengths) in Long Profile Characteristics

	Lewis River	North River	Middle River	South River
A. Distances (metres)				
Short	115	105-115	115-120	115-145
Intermediate	185	160 230-275	205-230	185
Long	825	455 550	455 550	550 730
B. Proportional to Mean Width				
Mean Width (metres)		32	34.6	34/21[1]
Short		3.33-3.57	3.30-3.52	3.36-4.25/ 5.45-6.88
Intermediate		5.0 7.15-8.58	5.94-6.61	5.38/8.72
Long		14.3 17.1	13.2 15.8	16.1/21.6 26.1/35.0

[1] Mean width of South River is 34 m on the lower course, but only 21 m along the slightly entrenched upper course.

TABLE 31

Mean Values of Exponents in the Hydraulic Geometry in Riffles and Pools

Condition	w_s	d_*	$\overline{v}$	A	C_4 (Eqn. 12)
Riffle	.214	.318	.468	.532	0.902
Pool/Chute	.173	.239	.588	.412	1.585
	.225	.285	.490	.510	0.735

[1] The lower figures are computed without including the two Middle River sections.

TABLE 32

Mass Balance of Channel Bars, Ekalugad Fiord

A. Upper South River, Right Diagonal Bar. Established 6 August, 1967

Stake No.	Changes (cm) 6-11 Aug. 1967	11-17 Aug. 1967	11 Aug., 1967 25 June, 1968	25 June - 15 Aug., 1968	6 Aug., 1967 - 15 Aug., 1968
1	0.6 cm		-3.7	5.0	1.9
2	2.4	-3.2	-5.2	11.0	9.2
3	2.7		in snow		3.6
4	14.0	-15.2	-15.1	1.5	0.4
5	9.1		-0.8	-0.5	7.8
6	0.9		-3.6	0.0	-2.7
7	3.2		1.4	-0.5	5.1
8	2.4		-4.1	2.0	0.3
Mean	+4.41		-4.44	+2.64	+3.2

B. Middle River, Left Diagonal Bar. Established 3 August, 1967

Stake No.	Change (cm) 13 August, 1967 - 15 August, 1968	Stake angle degrees (from vertical)
1	-13.0 cm	60
2	1.5	45
3	-54.0	85
4	-4.0	5
5	-32.0	60
6	3.0	0
7	-29.5	60
8	-2.0	30
9	-9.0	15
10	1.0	0
11	-8.0	40
Mean	-13.3	36.4

TABLE 32 (cont.)

C. Middle River, Sedimentation Zone. Established 23 July, 1966

Stake No.[1]	Changes (cm) 23 July - 7 Aug., 1966	7 Aug., 1966 - 5 Aug., 1967	5 Aug., 1967 - 15 Aug., 1968	23 July, 1966 - 15 Aug., 1968
1	-9.1	-18.9	1.0	-29.0
2	-0.5	1.1	0.0	0.6
3	1.5[2]	7.1	1.0	9.6
4	6.6	3.4	2.0	12.0
5	5.7	-0.7	0.0	7.0
6	3.7	-15.5	-1.0	-12.8
7	10.4	-7.3	0.0	3.1
8	-3.1	0.5	-1.0	-3.6
9	3.6	-2.6	2.0	3.1
10	-11.6	-1.7	2.5	-10.8
4A	-36.2	In channel. Lost after 7 Aug., 1966		
6A	-38.1			
	+0.72	-3.46	+0.65	-2.08

1 Stake positions are mapped in Figures 56, 57 and 61.

2 Has been +9.0.

TABLE 33

Mean Exponents of Hydraulic Geometry in Braiding and Non-braiding Reaches

	N	w_s	d_*	$\bar{v}$	A	C_2 (Resistance Equation)
Braiding (B)	5	.244	.232	.524	.476	1.685
Propensity to braid (P)	7	.181	.276	.543	.457	1.160
Non-braiding (S)	2	.130	.440	.430	.570	0.545[1]

[1]The individual values of 0.04 and 1.05 are highly disparate. The high value of 1.05 for Lewis Base is the result of the very large boulders present in the reach which act, in effect, as form resistance elements.

TABLE 34

Division Points in the Lewis Sandur Channel Network

Date	Number of Division Points (p)
1958	13
1961	13
1963	22
1965	29
1967	21

TABLE 35

Mean Moments of Grain Size Distributions (Coarse Materials by Lithology) (mm measure)

Lithology	Coarse Materials (>8 mm) Type 1		Type 2		Type 3		Type 4		All sample		Fine materials (<8 mm)	
	Mean	Std. Dev.	Mean	Std. Dev.	Mean	Std. Dev.	Mean	Std. Dev.	Mean	Std. Dev.	Mean	Std. Dev.
Mean grain size (a-axis)	66.5	±41.8	67.7	±45.4	58.5	±39.5	73.9	±56.3	69.2	±47.1	1.346	±0.690
Std. deviation grain size	57.4	49.9	50.7	37.5	39.7	37.3	61.4	51.9	58.6	44.4	1.413	0.531
Skewness grain size	1.498	0.947	1.465	0.763	0.990	0.953	1.626	0.945	1.983	1.047	2.435	1.444
Kurtosis grain size	2.916	4.105	2.671	3.883	1.336	2.957	3.642	5.300	3.205	7.878	12.023	13.433
Mean largest cobble	220.4	170.4	215.8	146.0	140.9	130.7	276.6	235.8	198.5[1]	143.7		
Mean roundness	0.277	0.053	0.308	0.043	0.339	0.066	0.278	0.042	0.297	0.034		
Mean flatness	0.597	0.040	0.544	0.037	0.457	0.067	0.606	0.035	0.573	0.029		
Mean sphericity	0.725	0.026	0.691	0.024	0.642	0.046	0.732	0.023	0.711	0.019		

[1] As mean of 10 largest in sample of 96.

TABLE 36

Mean Values of Roundness and Sphericity, According to Lithology and Size Class

Lithology	(1)		(2)		(3)		(4)	
	R	S	R	S	R	S	R	S
12-16 mm	0.270	0.716	0.303	0.685	0.329	0.648	0.262	0.728
48-64	0.296	0.726	0.348	0.693	0.394	0.680	0.312	0.745
96-128	0.320	0.730	0.343	0.402	0.332	0.638	0.339	0.731
192-256	0.262	0.743	0.231	0.686	---	---	0.263	0.741

TABLE 37

Proportional Distribution of Cobble Lithology and Form Over the Ekalugad Sandur

Lithology Zingg	(1)	(2)	(3)	(4)	(5)-(7)	Total
Blade	.0161	.0451	.0226	.0249	.0001	.1088
Roller	.0507	.0628	.0165	.0729	.0008	.2037
Spheroid	.0875	.0793	.0135	.1503	.0014	.3320
Disc	.0705	.1170	.0459	.1209	.0012	.3555
Total	.2248	.3042	.0985	.3690	.0035	1.0000

TABLE 38

Summary of Trend Surface Analysis

Subject	1st order surface F^2	R^2	x-coord	y-coord	2nd order F	R^2	3rd order F	R^2
Type (1)								
Mean size	55.52**	0.439	-2.495	-3.567	11.71**	0.552	6.69**	0.626
Std. deviation size	43.07**	0.378	-2.773	-4.123	6.08	0.450	5.25	0.524
Skewness size	1.11	0.015	0.010	-0.003	0.98	0.036	0.27	0.043
Kurtosis size	0.83	0.012	0.033	-0.071	1.16	0.036	0.88	0.060
Mean roundness	35.84**	0.340	0.003	0.003	1.44	0.360	1.44	0.386
Mean sphericity	4.62*	0.061	0.001	0.001	0.76	0.076	3.01	0.152
Type (2)								
Mean size	37.04**	0.343	-2.403	-3.567	9.88**	0.459	12.43**	0.604
Std. deviation size	42.96**	0.372	-2.090	-4.123	4.97**	0.437	5.31**	0.514
Skewness size	1.24	0.017	0.0	-0.003	5.53**	0.122	0.98	0.147
Kurtosis size	1.43	0.020	0.020	-0.071	3.86*	0.095	0.94	0.120
Mean roundness	7.19**	0.092	0.001	0.003	9.24**	0.242	1.80	0.281
Mean sphericity	1.00	0.014	0.0003	0.001	1.84	0.052	1.29	0.087
Type (3)								
Mean size	29.86**	0.296	-1.871	-3.063	13.17**	0.452	7.48**	0.551
Std. deviation size	20.75**	0.226	-1.557	-2.349	10.99**	0.375	7.49**	0.488
Skewness size	0.79	0.011	-0.003	0.048	2.43	0.060	0.88	0.084
Kurtosis size	1.27	0.018	-0.0121	0.261	2.71	0.072	1.00	0.099
Mean roundness	9.04**	0.113	0.002	0.006	1.60	0.143	5.77**	0.268
Mean sphericity	4.70*	0.062	-0.0004	0.0001	0.79	0.078	2.72	0.147
Type (4)								
Mean size	36.92**	0.342	-2.970	-3.294	12.87**	0.485	11.95**	0.620
Std. deviation size	49.61**	0.411	-3.004	-2.474	10.74**	0.522	10.23**	0.633
Skewness size	3.12*	0.042	-0.013	0.042	2.93	0.099	0.77	0.119
Kurtosis size	1.96	0.027	-0.066	0.138	1.58	0.059	0.99	0.086
Mean roundness	3.02	0.041	-0.0004	0.003	7.74	0.178	2.93	0.243
Mean sphericity	13.13**	0.155	-0.001	0.001	3.74	0.218	1.71	0.256
All cobbles								
Mean size	42.72**	0.376	-2.604	-2.886	12.89**	0.512	11.39**	0.635
Std. deviation size	64.16**	0.475	-2.753	-2.780	11.57**	0.580	10.76**	0.681
Skewness size	1.92	0.026	-0.014	0.013	3.09	0.087	1.91	0.136
Kurtosis size	0.70	0.010	-0.070	-0.111	1.51	0.041	1.62	0.086
Mean roundness	8.89**	0.111	0.001	0.004	7.44	0.234	2.14	0.279
Mean sphericity	4.89**	0.064	-0.0004	0.0006	2.69*	0.116	5.03**	0.230
Fine materials								
Mean size	2.30	0.031	-0.011	-0.008	4.02**	0.109	4.18**	0.207
Std. deviation size	2.90	0.004	-0.001	-0.014	4.39**	0.090	3.23*	0.170

[1] F tests the increment of variance reduction due to each new surface; R^2 gives the total proportional variance reduction at that level.

* Significant at $\alpha = 0.05$

** Significant at $\alpha = 0.01$

TABLE 39

Analysis of Variance of the Trend Surfaces of Mean Grain Size (All Cobbles)

Surface	Sum of Squares	df	Mean Square	Residual df	Residual Mean Square	F	R^2
1 Inc.	120238	2	60119	142	1407	42.72**	0.376
Total	120238	2	60119			42.72**	0.376
2	43499	3	14500	139	1125	12.89**	
	163738	5	32748			29.23**	0.512
3	39453	4	9863	135	866	11.39**	
	203191	9	22577			26.08**	0.635
4	10595	5	2119	130	818	2.59*	
	213786	14	15270			18.68**	0.668
5	19784	6	3297	124	698	4.73**	
	233570	20	11678			16.74**	0.730
6	13454	7	1922	117	624	3.08**	
	247024	27	9149			14.66**	0.772
Residual	73038						
Total	320062						

* Significant at $\alpha = 0.05$. ** Significant at $\alpha = 0.01$.

TABLE 40

Material Deposited in Tasiujaq Cove in 1967

River	Solid volume (m^3)		Unpacked volume		Packed volume	
	Suspended	Bed Load	Suspended	Bed Load	Suspended	Bedload
North[1]	6,160	42,573	10,300	71,000	6,600	45,100
Middle	3,852	20,400	6,400	34,000	4,100	21,600
South	3,523	10,022	5,900	18,200	3,700	11,600
Total	13,535	73,895	22,600	123,200	14,400	78,300

	Increment[2] to bay bottom (m)	to delta front (m)
Packed	0.01	0.15
Unpacked	0.02	0.24

[1] Bed load estimated as 85% of total load, extrapolated from suspended load data.

[2] Note assumptions given in text.

APPENDIX C

List of tables of supporting data (not published)

Copies of all tables listed here, which constitute the data appendix of the writer's dissertation, may be obtained as part of the dissertation, from University Microfilms. Copies of the appendix are also deposited in the following places, from which it may be possible to arrange loan or reproduction of all or part of the data.

The Library, University of British Columbia, Vancouver 8, Canada.

The Library, Geological Survey of Canada, Department of Energy, Mines and Resources, 601 Booth Street, Ottawa, K1A 0E8, Ontario, Canada.

A film copy of the dissertation is also held in the National Library of Canada.

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