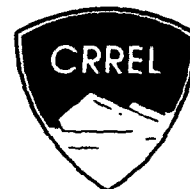


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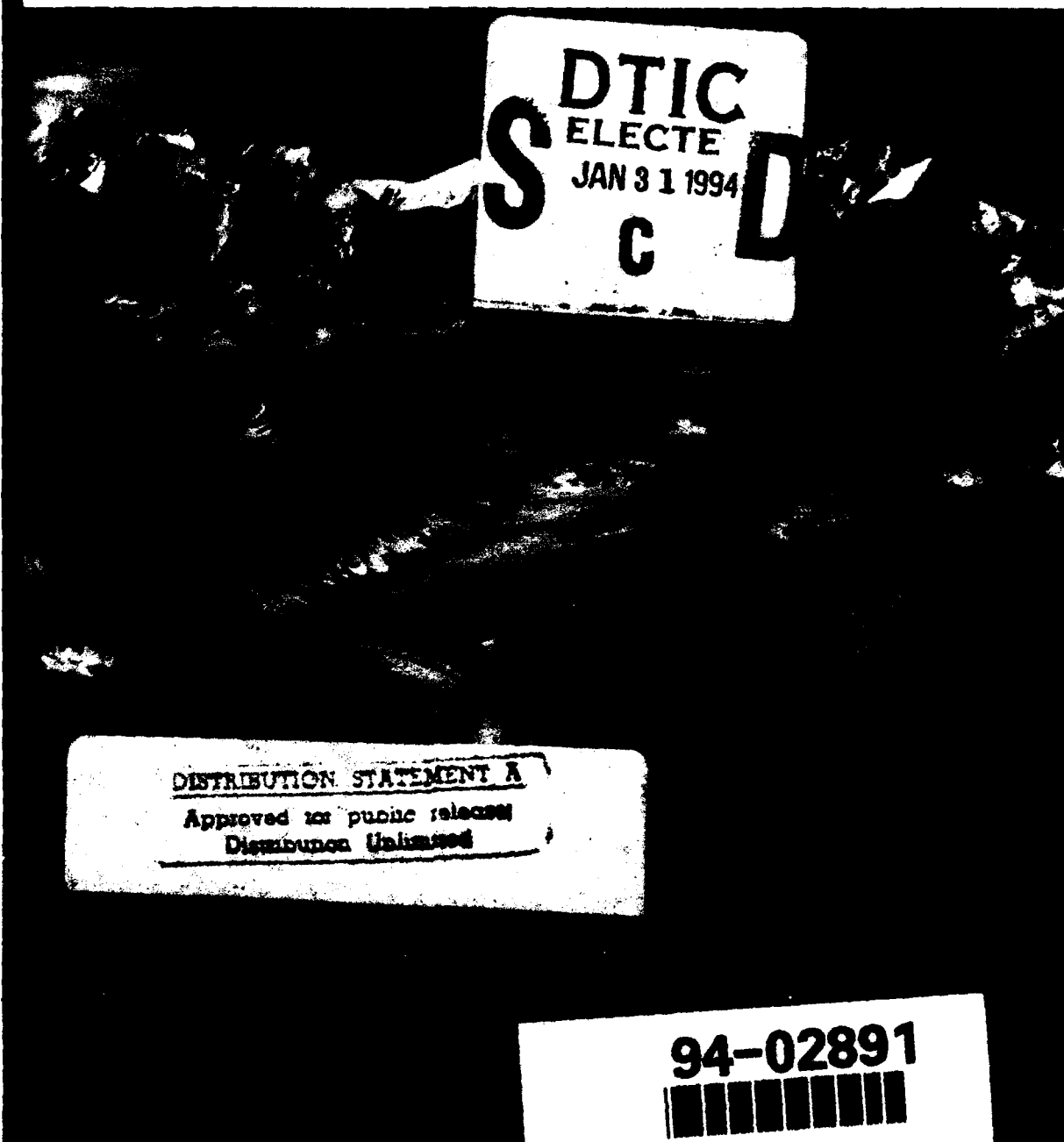


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Glaciohydrologic and Glaciohydraulic Effects on Runoff and Sediment Yield in Glacierized Basins

Daniel E. Lawson

November 1993



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Abstract

Glaciers exert significant control on runoff and sediment yield of partly glacierized basins, such basins being inherently more complex than non-glacierized basins. Glaciohydraulic and glaciohydrologic processes and factors determine the characteristics of runoff and of sediment discharge of rivers draining glacierized catchments. Significant problems develop in predicting both short- and long-term variations in water and sediment discharge because of the complicating effects of these processes and factors. Predictions are necessary for effective management of a basin or watershed's water resources. These processes and factors must therefore be incorporated to improve predictive models of runoff and develop models for sediment yield. In this monograph, the current state of knowledge on the nature of runoff and sediment yield in rivers originating in partly glacierized basins is reviewed and integrated, in particular analyzing the glaciohydrologic and glaciohydraulic processes and factors determining basin characteristics. Current statistical and physical models for predicting runoff and sediment yield in glacierized basins are reviewed and, based upon an assessment of both the state of knowledge and modeling techniques, future research or application of existing knowledge to improve predictions are recommended.

Cover: An alpine glacierized basin in the Alaska Range (photo by Professor Edward Evenson, Lehigh University).

For conversion of SI metric units to U.S./British customary units of measurement consult ASTM Standard E380, *Standard Practice for Use of the International System of Units (SI)*, published by the American Society for Testing and Materials, 1916 Race St., Philadelphia, Pa. 19103.

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**US Army Corps
of Engineers**

Cold Regions Research &
Engineering Laboratory

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PREFACE

This monograph was prepared by Dr. Daniel E. Lawson, Research Physical Scientist, Geological Sciences Branch, Research Division, U.S. Army Cold Regions Research and Engineering Laboratory. Funding support for this monograph and the field experiences incorporated into it was provided by the Office of the Chief of Engineers, CWIS 32604, *Cold Regions Water Resources Program*; DOD International Standardization Funds; the U.S. Army Corps of Engineers European Research Office (ERO); the National Science Foundation (grant to R. Powell); and the Natural Environment Research Council (U.K.) (grant to D. Collins).

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Executive Summary

Glaciers and ice sheets are critical components of the global hydrological cycle and often are the primary freshwater source in many regions of the world. Over 80% of the world's freshwater is held within glaciers and ice sheets. They are also a primary source of sediment in both freshwater and marine environments.

Glaciers and ice sheets are also important components in the global climate and can modify regional climate through various feedback mechanisms. Glaciers are sensitive to changes in climate that affect their mass and ultimately runoff and sediment discharge. Water resources will be affected by short- and long-term changes in climate, and, therefore, the role of glaciers in the hydrological cycle and, specifically, as water and sediment sources must be understood.

Glaciers occur in parts of most mountainous alpine regions of the world, with the largest concentrations in Antarctica, Greenland, the Canadian Arctic and the Gulf of Alaska region. Many major rivers in western North America head in glacierized basins, and runoff and sediment discharge from these basins exert significant control on their flow and water quality.

Predicting runoff and sediment yield from drainage basins or watersheds is an important part of efficient water resources management. Glacierized drainage basins are inherently more complex than nonglacial drainage basins, and these complexities pose serious problems for predicting runoff or sediment yield over both the short and long term. In particular, runoff predictions can affect hydropower and reservoir management, as well as downstream water use and allocations, including community water supplies, fisheries, recreation, navigation, mining and irrigation.

Glaciohydraulic and glaciohydrologic processes and factors determine the characteristics of runoff and sediment yield of rivers draining glaciers. Currently, models are incapable of accurately predicting either runoff or sediment yield from glacierized drainage basins. To improve predictive models of runoff

from partly glacierized catchments, as well as to develop a model of sediment yield for such basins and watersheds, requires incorporating these processes and related factors into their formulation.

The purposes of this monograph are, therefore, to:

1. Review the current state of knowledge on the nature of runoff and sediment yield in rivers originating in partly glacierized basins, emphasizing the glaciohydrologic processes and factors determining the characteristics of runoff, and the glaciohydraulic processes and factors determining sediment yield, fields of study in which there have been significant increases in knowledge over the last 10 years.

2. Review current statistical techniques and physical models for predicting runoff and sediment yield in glacierized drainage basins.

3. Assess the state of knowledge and modeling techniques, and recommend future research or applications of existing knowledge to improve predictions.

Glaciers, whether they occupy only a small percentage of the basin area or cover the majority of it, significantly control runoff, originating mainly as meltwater from snow and ice, precipitation and groundwater, and from surface water originating in unglaciated parts of the basin. Glaciers modify peak discharges, the variability and volume of hourly, daily and seasonal discharges, the lag between precipitation and resultant increase in runoff, and long-term trends in the annual volume of flow. Glaciers play the roles of daily and seasonal water sources and water storage media, long-term water reservoirs, and routes for meltwater and meteoric water from the glacier's surface to its discharge at the terminus.

The amount of energy available for producing meltwater from snow and ice is a critical component of daily and seasonal runoff. This energy is mainly net radiation, sensible heat, latent heat of condensation or evaporation, and heat released by cooling or freezing of rainwater falling on the snowpack and ice.

Long-term (decadal to century) trends in runoff largely reflect the effects of climate on the mass

balance of the glacier. The mass balance is the net change in glacier mass over 1 year, being the difference between total accumulation of snow and ice and the total amount lost to ablation. Positive and negative mass balances can be associated with either increases or decreases in runoff over short and long periods.

Meltwater originating on the snowpack percolates through it under the influence of gravity. The process of meltwater percolation can be described in a general sense by Darcian flow theory for an unsaturated, homogeneous porous medium. Above glacier ice, the seasonal snow rapidly becomes saturated with meltwater; within this zone, Darcy's law for saturated flow applies. Within the permanent snowpack of the accumulation (or firn) area, a thick saturated layer commonly exists on temperate glaciers and acts as an unconfined aquifer that stores meltwater and retards its movement into the glacier's drainage system.

Meltwater flowing through the seasonal and permanent snowpacks either discharges at their bases onto the glacier surface, where it flows within supraglacial channels, or enters the glacier directly through intergranular passageways or veins, and larger drainage sinks such as crevasses and moulins. Within the glacier, the idealized englacial drainage system consists of a tree-like network of small veins and capillary tubes feeding progressively larger conduits at depth. Overall, flow is perpendicular to equipotential surfaces, which are a function of water pressure gradient and potential energy. At the bed in the subglacial environment, water moves generally perpendicular to the intersection of these equipotential surfaces with the substrate.

The precise nature of the englacial drainage system remains uncertain, however. Indirect analyses indicate that within different parts of a glacier, as well as in different glaciers, drainage may vary in rate, suggesting that both direct drainage pathways and indirect, tortuous routes of flow with significant water storage may exist.

Water entering the subglacial environment (below the ice/bed interface) theoretically flows through one of three drainage systems: progressively interconnected conduits and larger tunnels (conduit-tunnel system), an anastomosing pattern of smaller conduits that link cavities in the bed (linked-cavity system), or, in the absence of conduits or channels, as a thin film between the ice and bed (distributed flow system). Each system may coexist within different parts of the bed of the same glacier. Channels at the glacier bed may be incised downwards into the substratum (N-channels), incised upwards into the ice

(R-channels) or partly in both. Water may also flow preferentially in existing bedrock features or abandoned channels.

In the established conduit-tunnel system, there is an inverse relationship between water pressure and discharge, whereas in the linked-cavity system, they vary directly. This relationship allows the linked-cavity system to maintain a given water flux capability under water pressures that are higher than in a conduit-tunnel system, and ensures a mostly stable system that can persist with seasonal changes in discharge.

In contrast, a conduit-tunnel system will evolve and grow with increases in discharge, but collapse by ice deformation as discharge decreases, pressures drop and heat flux is reduced. Runoff will reflect such changes. By peak summer discharge, a steady-state system evolves with rapid flow-through, reduced and minimal water storage, and low water pressures.

While drainage system configuration and hydrology are reasonably straightforward when the substrate consists of bedrock, beds composed of sediment complicate things. In this case, conduits or tunnels may be incised partly into sediments or fully cut into basal ice with a sediment floor, and, therefore, closure or expansion is affected by the rate of sediment deformation into them. Sedimentary substrates are also permeable and a certain small proportion of water is transported within them. Water can then flow in a distributed system.

The majority of runoff from glacierized basins occurs within several months or less, lasting a few weeks in the Arctic and 4 to 5 months in lower latitude basins. Glaciers effectively moderate annual stream flows by storing as ice some portion of the precipitation received. Annual runoff is therefore not correlated with annual precipitation as it is in nonglacierized basins. Release of water from ice storage depends on the energy input and extent of the seasonal snow cover, whereas runoff from rainfall in summer depends on the type and seasonal evolution of the englacial and subglacial drainage systems. The most variability in daily discharge, therefore, takes place in spring, as the melt season begins, and generally decreases through the summer. The seasonal discharge peaks 1 to 2 months later in glacierized basins than it does in adjacent nonglacierized basins, reflecting early season meltwater retention in the snowpack, the progressive increase in meltwater production and drainage system development. Minimum runoff, sometimes decreasing to near zero, occurs in mid- to late-winter.

Sediment transport and sediment yield in partly

glacierized basins are complex functions of the internal or glaciohydraulic processes, which erode, entrain, transport and deliver sediments into glacial discharges, and of climatic conditions, which determine the rate, magnitude and timing of meltwater production. Thermal conditions within and below a glacier affect sediment entrainment and release, while the englacial and subglacial drainage systems, their seasonal evolution and stability, and their interaction with the various glaciologic processes determine sediment transport and discharge to rivers at the terminus.

Under normal flow conditions, glacial rivers in general transport significantly more sediment in suspension and as bedload annually than do nonglacial rivers. Sediment discharge is, however, more variable daily, monthly and annually, with up to 95% of the annual load transported during the melt season, the duration of which varies with latitude and altitude.

Sediment transported by glacially fed rivers is derived from sediment in transport by the glacier itself and from sediment in transport by meltwater within the englacial-subglacial drainage system. Sediment also comes from nonglacial processes in glacier-free areas of the basin.

Ice flow and sliding determine the predominant transport paths for debris derived at the surface and bed of the glacier. Sediments deposited on the glacier surface in its accumulation area by wind, mass wasting or other processes are most commonly transported englacially along flow lines above and out of contact with the bed and emerge at the glacier surface within the ablation zone. Debris incorporated at the highest elevations is, however, carried along flow lines that may intersect the bed, where basal processes and subglacial transport could take place.

Debris transported within a relatively thin zone near the glacier's sole makes up the majority of sediment in transport within the glacier. Basal debris originates by a variety of processes, including mechanical abrasion and plucking, regelation and freeze-on.

The annual flux of sediment transported by ice can only be grossly estimated, if ice flow velocity is known and if measurements or estimates are available of supraglacial, englacial and basal debris volumes.

Sediment entrainment and transport by englacial and subglacial discharges have not been directly observed or measured in most cases. Estimates of substrate erosion rates are based strictly on theoretical assessments of bed material properties and the erosive capacity of meltwater flowing in channels or conduits.

Seasonal (and daily) variations in sediment flux

reflect seasonal variations in the extent and development of the englacial-subglacial drainage system, the type of drainage system and seasonal variations in sediment production and availability, the latter being a function of previous hydrological events that remove or deposit bed materials. As the drainage system is progressively established in spring, new sediments produced during winter are entrained and episodically removed when different areas of the bed drain and are flushed of sediment. Diurnal increases in meltwater discharge can likewise cause episodic opening and closing of conduits and daily variations in sediment discharge.

Through the melt season, discharge increases, and the drainage system, whether linked-cavity or conduit-tunnel type, becomes established and well developed. In the conduit-tunnel system, increased discharge leads to reduced water pressures and a decrease in the bed area subject to erosion, correlating with a progressive decrease in sediment discharge. Increased discharge in the linked-cavity system results in increased water pressures, with conduit and cavity growth possible, entraining basal debris and bed material. In both fully developed systems, parts of the bed can remain inaccessible to runoff and accumulate sediments. Only flooding, which can rapidly expand flow across other parts of the bed, can then produce high sediment discharges. Late in the melt season, a lowered sediment flux appears to be a function of an exhausted sediment supply and fully developed drainage system. Severely reduced winter discharges ensure sediment storage on the bed that then provides larger volumes of sediment for spring meltwater discharges.

Sediment yield of glacierized catchments is highly variable from region to region and can vary from glacier to glacier and from year to year. This variability is not related directly to discharge; increases in sediment load commonly occur out of phase with that of runoff, but may also take place without any change in discharge. Sediment discharge can also vary in the extreme, with nearly the equivalent of the annual load released in a single day. However, insufficient data exist to define the magnitude of long-term fluctuations in sediment yield.

Statistical techniques and physical (conceptual) models are currently incapable of accurate predictions of either runoff or sediment yield over the short or long term. Statistical techniques suffer from a lack of data, in particular long-term records, and cannot account for the extreme annual or seasonal variability common to runoff and sediment yield. Statistical treatments are not applicable to more than the single basin for which they were developed, and each must

be calibrated independently for other basins with basin-specific data.

Physical models of runoff are better because they consider processes and can be continually adjusted to simulate observed hydrometeorological conditions, or altered to consider existing conditions and runoff history. Although meltwater production at the glacier-snowpack surface can be reasonably modeled, physical simulations suffer from an incomplete treatment of the glaciohydrologic processes and controls on runoff. Models are basin-specific and have rarely been applied to more than one basin. In these basins, they are generally incapable of accurately predicting diurnal, peak or seasonal flows.

Physical models of sediment yield of glacierized basins have not been developed.

Future research requirements include a need for long-term data on runoff and sediment yield that are linked to both downstream hydrology and hydraulics and climate. Such data are required for testing,

evaluating and developing statistical relationships and physical models. Data from glacierized catchments in North America are particularly lacking, especially considering the vast area and volume of ice in both Canada and the U.S.

Research on the glaciohydrologic and glaciohydraulic processes and factors that control water and sediment production is required to formulate realistic and accurate physical models of runoff and sediment yield.

The relationships between climate and glacier mass balance, glaciohydrology and glaciohydraulics, and runoff and sediment yield need to be defined and methods developed that minimize the need for costly, labor-intensive measurements of annual glacier mass balance, runoff and sediment transport. Remote sensing techniques using both active and passive sensors need to be examined for regional predictions of runoff, both annually and in near-real-time.

Glaciohydrologic and Glaciohydraulic Effects on Runoff and Sediment Yield in Glacierized Basins

DANIEL E. LAWSON

Introduction

Glaciers and ice sheets are critical components of the global hydrological cycle and, as such, are often primary sources of water for many rivers, lakes and other freshwater environments (Fig. 1). They are also one of the primary sources of sediment in freshwater and marine environments.

Glaciers and ice sheets are, in addition, critical components in the global climate system, and may modify regional climate through various feedback mechanisms. Glaciers can be extremely sensitive

to climatic variations, responding through changes in their mass that ultimately affect runoff as well as sediment discharge and water quality.

With water being one of the world's most precious resources and with over 80% of the freshwater being held within glaciers and ice sheets, the response of glaciers to short- and long-term changes in climate will affect water resources globally. It is, therefore, important that we understand their role in the hydrological cycle and, more specifically as



Figure 1. Glaciers supply large amounts of fresh water to rivers and lakes in glacierized regions, exerting significant control on river hydrology and water quality. The Matanuska River largely originates from several subglacially fed streams draining the Matanuska glacier in south-central Alaska.

addressed in this monograph, their role as water and sediment sources affecting the hydrology and water quality of rivers and lakes.

Glaciers and ice sheets concentrate in Antarctica, Greenland and North America, primarily in the Canadian Arctic and Gulf of Alaska regions. Glaciers are also found, however, throughout most mountainous alpine regions of the world. In western North America, for example, glaciers are found in the states of Washington, Oregon, California, Wyoming, Montana, Idaho, Colorado, Nevada and Utah, and the Canadian provinces of Alberta and British Columbia. Within Alaska, approximately 3% of the land (74,700 km²) is covered by glacier ice.

Most Alaskan rivers, as well as many of the largest river systems in the western U.S. states and Canadian provinces, originate at glaciers and provide a reasonably sustained annual water supply. In Alaska, the estimated average runoff from glaciers is 220 km³/year, nearly 35% of the total runoff in the state and 10% of the total runoff for the U.S. (Mayo 1986). This runoff exerts a significant control on the nature of flow and water quality of all major Alaskan rivers, except the Colville River on the North Slope (e.g., Benson et al. 1986). Even the smaller glaciers in drainage basins of the western U.S., however, have marked effects on river hydrology (e.g., Krimmel and Tangborn 1974).

Because runoff and sediment yield from glacierized basins are inherently more complex than they are from nonglacierized basins, significant problems develop in predicting both short- and long-term variations in water and sediment discharge, predictions that are necessary for effective management of our water resources.

For example, hydroelectric power generation depends on runoff, requiring both long-term projections of the frequency and magnitude of peak and low flows during the planning of a new project and short-term forecasts, including daily and hourly variations in the magnitude and timing of runoff, for efficient operation. In the long term, predictions of decadal or longer trends in runoff may be decisive in determining if a new project is viable. Inaccurate forecasts may hurt a hydroelectric project and its financial status after completion if the projected trends are too high, or result in inefficient use of the water if projected volumes are too low. In general, over-forecasting results in a loss of power generation capacity because reservoir water levels will be lowered in anticipation of large inflows, whereas under-forecasting can cause inefficient water use and down-

stream problems necessitating mitigation measures.

Large-scale hydropower schemes with large reservoir storage capacities, such as that of the Columbia River system in the northwestern U.S. and southwestern Canada, are not significantly affected by usual diurnal variations in runoff, but total annual flows, the timing of peak runoff and the occurrence of unusual peak or flood flows are critical to daily operations. Smaller scale, but commonly complex, hydroelectric schemes depend particularly on accurate daily and even hourly discharge predictions for the best use of water and for power production (Young 1985).

The Grande Dixence S.A. hydroelectric project in Switzerland is a prime example of an operating system fed by glaciers with requirements for both short- and long-term runoff predictions. This hydroelectric project uses meltwaters from the alpine valleys of St. Nicolas (Matteral) and Herens, Valais, Switzerland, a region that encompasses 35 glacierized drainage basins ranging in area from 1 to 80 km². In the Grande Dixence hydroelectric scheme, a more or less continuous adjustment of water use to water yield requires daily and hourly forecasts of meltwater yield from each glacierized basin, which include determining when to pump water between multiple storage reservoirs and the storage rates throughout the entire system (Lang and Dayer 1985).

Grande Dixence also relies heavily upon accurate annual and decadal runoff predictions for each glacierized basin because the design incorporates low volume reservoirs based upon projected 18-year means in runoff (Bezing 1982). To make operations optimum, runoff, sediment yield and related data in these basins have been measured and analyzed since 1948. This data set is perhaps the best available, yet even with these data, a drop in meltwater production owing to increased storage of precipitation as ice caused an unanticipated shortfall of water for energy production from 1965 through 1979 (Bezing 1978, 1982).

The downstream uses of water for irrigation, potable supplies, navigation, fisheries, industry and recreation also depend upon accurate prediction of the seasonal and annual availability of water (Meier and Roots 1982, Power 1985). Water availability in glacierized basins is complicated because it depends upon the amount of water stored within the glaciers and the timing of its release, items which are not necessarily seasonally related. Incorrect predictions of water availability are especially critical to downstream arid or semi-

arid lowlands that receive their water from rivers originating in glacierized basins of mountainous areas that may be several hundred kilometers or more distant (Dreyer et al. 1982, Tarar 1982, Young 1985, Hewitt et al. 1989, Butz 1989). In addition to the complexities of the hydrologic system under the present climatic conditions is the probability of long-term changes in climate (e.g., Collins 1985). Thus, current water use plans and operations, as well as the design and construction of new water supply systems, must consider the effect of climate change (Gleick 1987, 1990). Climatic change will not only result in long-term warming, but also an increase in the frequency and magnitude of extreme climatic events. Decisions on water use require us to predict such changes in climate, or the effects of hydrological changes may be exacerbated by incorrect decisions (e.g., Power 1985, Dracup and Kendall 1990, Peterson and Keller 1990, Cooper 1990).

An incident on the Columbia River system provides an example of the economic impacts resulting from an incorrect runoff prediction. The Columbia River basin receives water from multiple glacierized basins in the Rocky and Cascade mountains. This water is used to power 14 hydroelectric plants along the river's main stem, while also meeting downstream water uses; thus, both daily and seasonal hydrologic forecasts are required (Tangborn 1980b, Power 1985). Forecasts following the 1976–1977 winter drought in the western U.S. predicted runoff of less than 30% of normal in Oregon, but this prediction did not account for the effects of glaciers within the upper parts of the Columbia River basin. The seasonal runoff was actually twice the predicted amount because of a sunnier and warmer than normal summer that increased meltwater runoff from the glaciers, causing an estimated \$18,000,000 loss in potential hydroelectric generation (Tangborn 1980b, Power 1985). In addition, economic losses to agriculture resulted from large investments in drilling water wells, cloud seeding and land use changes that were made to deal with the anticipated water shortages (Glantz 1982).

Flood forecasting depends largely on long-term records for predicting 100-year and other probable floods. Basic data for flood predictions in glacierized basins are generally limited, but also have several complicating factors associated with them. Unanticipated, sometimes catastrophic, floods in glacierized basins are commonly induced by high-intensity rainfall during warm periods in which the melt rates of glacier ice and snowcover are high

(e.g., Bowling and Trabant 1986, Young 1985, Rothlisberger and Lang 1987), but the flooding may lag unpredictably from delays attributable to the internal drainage system of glaciers.

Flood forecasting may be further complicated by catastrophic water releases called glacier outburst floods or jökulhlaups. Outburst floods can far exceed "normal" floods; on the Knik River in Alaska, for example, the measured peak outburst flood from a glacier-dammed lake was $22.3 \text{ (m}^3/\text{s)/km}^2$ [$304 \text{ (ft}^2/\text{s)/mi}^2$], whereas the 100-year flood is calculated at $3.9 \text{ (m}^3/\text{s)/km}^2$ [$53.7 \text{ (ft}^2/\text{s)/mi}^2$] (Parks and Madison 1984). Jökulhlaups have been documented globally, including on the conterminous U.S. (Richardson 1968, Driedger and Fountain 1989), in the European Alps (e.g., Haeblerli 1983), in Norway and Iceland (e.g., Thorinsson 1953, Bjornsson 1974), in the Himalayas and in other parts of Asia (e.g., Konovalov 1990).

As with runoff, the variability and volume of sediment in discharges are important issues in the use of water from glacierized basins. Numerous problems can result from both the amount and characteristics of the sediments suspended in the water column and from the resulting deposition of transported sediments within navigable channels, river confluences, reservoirs, lakes, harbors and inlets. The seasonal variability in sediment transport can also affect channel aggradation or degradation, bank erosion and channel migration.

In Europe and Scandinavia, where development of glacierized basins is more advanced than elsewhere, sediment problems are considered during both the design and operating phases of hydroelectric projects (Harrison 1986). Prior to design, sediment data are collected within the glacierized basins to estimate sediment delivery and short- and long-term yield. These data establish the size of settling basins, reservoir sedimentation rates, the capacity of purge valves and structural design. Sediment suspended in the water column as well as sediment transported as bed-load must be separated from the water before it enters the intakes of a power generating plant. Sand traps, sedimentation chambers or settling basins are constructed to minimize damage to waterways and mechanical equipment.

Sediment transport volumes and grain size distribution must also be considered in subglacial intake design, including construction of subglacial sedimentation chambers designed to trap bed-load material (Wold and Ostrem 1979). Large, daily volumetric variations in sediment discharge can characterize glacier sources, and can reduce

the effectiveness of such measures and cause rapid sedimentation by coarse bedload material (up to boulder size), covering the intakes and damaging equipment (e.g., Ostrem et al. 1967). Sedimentation in diversion tunnels or galleries also poses serious problems for hydroelectric projects (Wold and Ostrem 1979). Sedimentation in storage reservoirs results in loss of volume and, hence, loss of potential energy production during the project life; remediation of this problem increases costs for dredging or for mitigations resulting from the downstream effects on fisheries of purges (e.g., Harrison 1986, Bogen 1986, Wang et al. 1987). The size and mineralogy of the suspended load is analyzed for designing turbines and other mechanical equipment that will be affected by water transporting suspended sediment (Bogen 1986, 1989).

Mechanical equipment, including turbines and pumps, will wear excessively from the glacially derived suspended sediment, as do structures such as intake walls and galleries. In Switzerland at Grande Dixence, for example, pump rotors must be replaced, at a cost of \$100,000 each (1968 dollars), for every 30–40 million m³ of water pumped (Bezing and Schafer 1968). In Norway, decreased turbine efficiency because of wear by suspended sediment at three power stations resulted in a loss of energy output of about 7 to 10 GW hr/yr (Bogen 1986). The size and mineralogy of the suspended sediment affects its abrasive nature, and both need to be considered when designing turbines and other mechanical equipment (Bogen 1989). In addition, suspended sediments affect river and reservoir water quality and ecological conditions, including photosynthesis, while the coarser bedload provides material in river substrates. The environmental impacts of sediment discharge modifications must therefore be considered in project design and operations as well (e.g., Bogen 1986). Sediment data continue to be gathered and analyzed, and used to modify the project design as sediment problems emerge during operations in Europe and Scandinavia. Such items as intake grill spacing, enlarging or adding settling chambers, pressures in valves and pipes, slopes of gallery basins and others may subsequently be modified to improve operations. At Grande Dixence, for example, a floating, telescoping intake for supply reservoirs was designed after construction to re-

duce the concentration and size of sediment in meltwater before it flowed into pumps and turbines, causing excess wear and failures.*

Techniques for forecasting runoff and sediment yield are necessary components to the decision-making process in water resources management, yet they remain generally inaccurate and unable to forecast the highly variable nature of glacierized basin flow and sediment discharge. The basic reasons for inaccurate predictions of runoff have resulted primarily from our lack of understanding of the processes and factors within glaciers (or ice sheets) that determine water movement and storage, and our inability to incorporate these complex controls into predictive models. In addition, the response of glaciers to changes in climate, which determine the glacier's mass balance (ice loss or gain) and subsequently runoff, is likewise complex and insufficiently understood for full incorporation into predictive techniques. In both instances the lack of basic hydrological, climatic and glaciological data from within glacierized basins seriously compromises predictions.

As with runoff, predictions of sediment yield and the effects of glacially derived sediment require an understanding of the complex processes that determine sediment discharge within and below glaciers. Sediment production is related to glaciohydrologic processes, and thus, ultimately, to climatic controls, as well as other physical controls. Recent advances in the understanding of internal glaciohydrologic processes over the last 15 years or so warrant their inclusion in future models. Better predictive techniques will be important in the future as use and development of glacierized regions increases (e.g., Partl 1977, Benson et al. 1986, Bezing 1981, Thomsen et al. 1989, Bogen 1989).

In this monograph, the glaciohydrologic and glaciohydraulic processes that directly or indirectly control runoff and sediment discharge from glacierized basins, including their relationship to climate, are described. Existing techniques and models for predicting runoff and sediment yield from glacierized basins are summarized in relation to their treatment of these processes. Based upon this synthesis, future basic research and applications of existing technology are recommended.

* Personal communication with A. Bezing, Grande Dixence S.A., 1985.

Part 1. Runoff Processes, Factors and Characteristics

INTRODUCTION

The effect of glaciers on river hydrology is extremely important, whether it be in a small basin in which the glacier covers most of the basin area, or in a regional-size basin in which glaciers occupy a small percentage of the area. Studies have shown that even a few percent of areal cover by glacier ice modifies peak discharges, the lag between precipitation and resultant runoff, the variability in discharges annually and seasonally, and the total annual flows from the basin (Fig. 2) (e.g., Meier and Tangborn 1961, Ostrem 1973, Church and Gilbert 1975, Tvede 1982, Collins 1984, Fountain and Tangborn 1985a, Braithwaite and Olesen 1988, Hewitt et al. 1989, Young 1990).

Fountain and Tangborn (1985a) found, for example, that the presence of glaciers increased runoff volume in comparison to nonglacierized basins

of similar precipitation characteristics in the North Cascades, Washington (Fig. 3). For 20% glacier cover, annual runoff is 10% greater than in the nonglacierized case, while summer melt season flows are 50% greater. A reconnaissance study by Clarke et al. (1986a,b) concluded that for 5.2 and 5.9% glacier cover above the proposed Devil Canyon and Watana dam sites on the Susitna River, Alaska, approximately 15 and 18%, respectively, of the runoff is produced by glaciers in the upper Susitna River basin (790 km² area), or nearly three times as much water as estimates based upon area alone would suggest. In the Tanana River of interior Alaska, Anderson (1970) estimated that 47% of the summer runoff was derived from glaciers, even though only 5% of the basin is ice covered. Glacially derived runoff in this continental basin is nearly an order of magnitude greater than that in similar nonglacial basins; in maritime Alaskan

Figure 2. Mean summer season runoff (May to September) for the Massa River basin of the Aletschgletscher (67% glacier cover) and the Thur River basin (0.01% glacier cover) for 1965–1983 (after Röthlisberger and Lang 1987). Runoff varies in virtually opposite response to similar meteorological conditions.

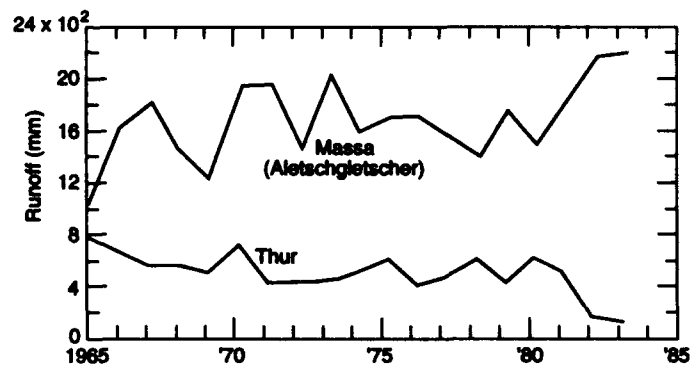
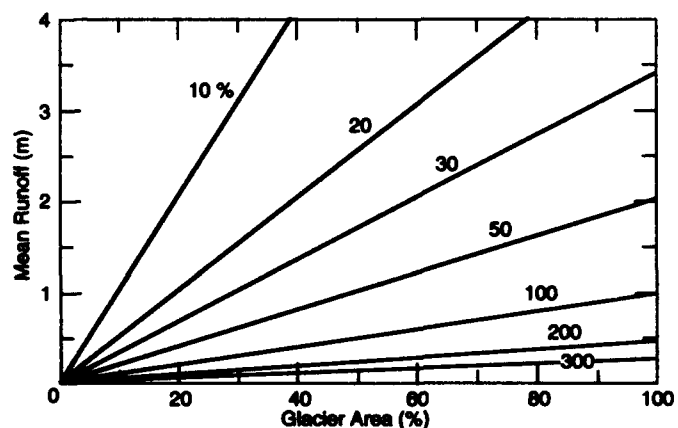


Figure 3. Percent increase in runoff from a glacierized basin with similar precipitation characteristics as the area covered by glaciers increases. Calculations assume a net mass balance of -1.0 m (after Fountain and Tangborn 1985a).



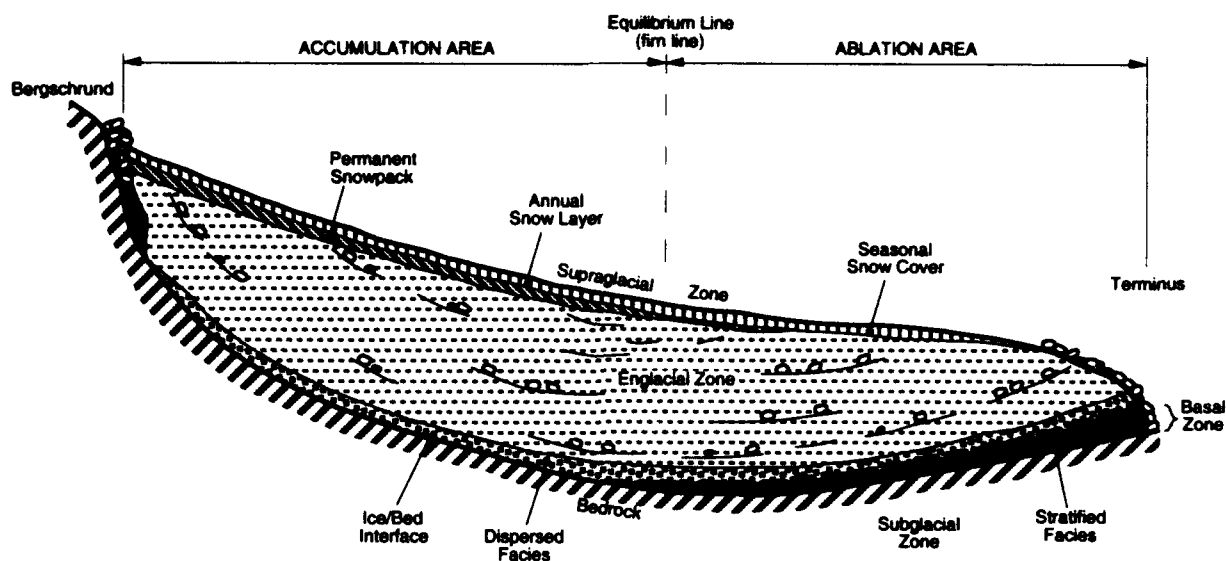


Figure 4. Idealized glacier-defining terms. Supraglacial refers strictly to the surface of the glacier, englacial to within the main body of the glacier, and basal to a thin zone just above the bed that develops in response to the interaction of the glacier and the subglacial materials and drainage system.

areas, it commonly exceeds nonglacial sources by 2 to 1 (Mayo 1984). In contrast, in the arid basin of the upper Indus River in Pakistan, an estimated 80% of the runoff is derived from 20% of its area, encompassing the snowfield and glacierized zones above 3500-m elevation (Hewitt et al. 1989).

The difference between glacierized and nonglacierized basins lies with the glacier itself, including its roles as a water source, storage site and routing medium. The production of ice and water are ultimately a function of climate through its role in determining the amount of energy available for



Figure 5. At the end of the melt season, the snowline (arrow) marks the lower limit of the accumulation area and the elevation of the equilibrium line. Meltwater produced seasonally in both the accumulation and lower ablation areas is the largest component of glacier river discharges (photo by E. Evenson of Eureka Glacier, Alaska).

meltwater production and the amount of precipitation that is added to the mass of the glacier as ice. Factors determining water movement through or storage within the seasonal and perennial snowpack and within (englacially) and below (subglacially) the glacier in turn affect the timing and volume of runoff as well (Fig. 4).

In glacierized basins, runoff is composed of snow-melt, ice-melt, precipitation and the release of water stored within the glacier and its snowpack, with surface-derived meltwater from snow and ice the principal components of summer discharges (Fig. 5) (Rasmussen and Tangborn 1976). In addition, groundwater flow from unglaciated areas surrounding the glacier may be an additional component of runoff, although this quantity is generally believed to be small. Internally derived meltwater from geothermal heat and frictional heating by ice sliding and flow are likewise considered insignificant components in runoff calculations.

Seasonal variations in the form of precipitation and in the energy supply result in a seasonal periodicity to the quantity, quality and timing of runoff. Annual runoff variations ultimately reflect both short- and long-term climatic trends, but more immediately are a function of annual meltwater production and the annual precipitation volume as snow or liquid. Because precipitation may be stored as a solid or a liquid, seasonal and annual variations in the volume of water in storage can significantly regulate runoff as well.

Long-term trends in runoff are mainly a function of the mass balance of the glacier, which is ultimately a function of long-term trends in climate. The volume of precipitation and amount stored in the glacier versus the amount of energy available to melt determines the balance between gain and loss.

ENERGY AND MASS BALANCE

Over seasons, years and decades, the amount of energy available for meltwater production during each period is a critical component of the runoff process that must be considered in predictive models for glacierized basins. The energy (Q_M) available for melting snow or ice consists of the following components (e.g., Röthlisberger and Lang 1987)

$$Q_M = Q_{NR} + Q_S + Q_L + Q_P + Q_G \quad (1)$$

where Q_{NR} = net radiation

Q_S = sensible heat

Q_L = latent heat of condensation or evaporation

Q_P = heat provided from precipitation

Q_G = heat from heat conduction through the snowpack to the ice.

The energy available Q_M is then the heat used to melt ice or snow or is that gained from refreezing of meltwater.

If all the components of the energy budget are known, the melt rate (M) (melted water equivalent of snow and ice per unit area over a unit of time) can be calculated as

$$M = Q_M / S \quad (2)$$

where S is the latent heat of melting (333.7 J g^{-1}). Of the energy components, studies have shown net radiation (Q_{NR}) to be, in general, particularly important in determining ablation rates. Q_P is negligible, except in the situation where rain water freezes in a cold snowpack, when it can be extremely important as a heat source and modifier of snowpack properties.

Net radiation can be calculated from

$$Q_{NR} = G(1 - \alpha) + \epsilon_a \sigma T_a^4 - \epsilon_e \sigma T_e^4 \quad (3)$$

where G = global (direct and scattered) radiation
 α = surface albedo (ratio of reflected to incident solar radiation)

$\epsilon_{a,e}$ = emissivity of the atmosphere (a) and of the surface (e); for new snow, a value of 0.98–0.99 is used

σ = Stefan-Boltzmann constant ($5.67 \times 10^{-8} \text{ J m}^{-2} \text{ K}^{-4} \text{ s}^{-1}$)

$T_{a,e}$ = radiation temperatures of the lower atmosphere (a) and of the snow or ice surface (e).

The last two terms represent the rate of incoming long-wave radiation and the rate of emission of long-wave radiation at the surface respectively. Short-wave net radiation [$G(1 - \alpha)$] can be estimated from records of sunshine duration or cloud observations, if latitude, local geographical factors and time of year are considered. Albedo can be roughly estimated with various empirical relationships. Long-wave net radiation can also be estimated empirically, but actual measurements of each factor are best.

Although eddy heat fluxes can be measured directly, they require highly sophisticated instrumentation and are, therefore, usually estimated using the gradient method based upon standard temperature and humidity measurements.

The sensible (Q_s) and latent (Q_L) heat fluxes are proportional to the gradients of potential air tem-

perature (θ) and specific humidity in the atmospheric boundary layer (q)

$$Q_s = c_p \rho K_s \cdot \frac{d\theta}{dz} \quad (4)$$

$$Q_L = L \rho K_L \cdot \frac{dq}{dz} \quad (5)$$

where c_p = specific heat of dry air

ρ = air density

K_s = eddy diffusivity for sensible heat

K_L = eddy diffusivity for latent heat

L = latent specific heat of evaporation

θ = potential air temperature; for gradients over a few meters, the actual air temperature can be used

q = specific humidity (grams of water vapor per gram of moist air); $q = 0.622e/p$, where e is vapor pressure and p is atmospheric pressure

z = height of the sensor above the surface.

The equations are based on the assumption of constant flux with height. K_s and K_L are determined from wind profiles, which depend on the roughness of the surface and temperature stratification. Under neutral conditions in a constant flux layer, the profiles are expected to be logarithmic. From the analogy between eddy diffusion of momentum, heat and water vapor, it can be assumed that $K_s = K_L$ for neutral stability.

On the basis of these concepts, the sensible and latent heat are estimated using either the mixing theory of Sverdrup (1946) or the gradient method of Thornthwaite and Holzman (1939).

For the mixing length theory

$$Q = L_p k^2 \cdot \frac{u_2(e_2 - e_0)(0.622/P)}{[\ln(Z_2/Z_0)]^2} \quad (6)$$

where k = von Karman constant (usually taken to be 0.4)

u_2 = wind velocity (cm s^{-1}) at level 2

e_2 = vapor pressure at level 2

e_0 = saturation vapor pressure for the surface temperature at height Z

Z_0 = roughness parameter (the height above the surface at which $u = 0$)

Z_2 = height of level 2.

The sensible heat flux is estimated by

$$Q_s = C_p \rho k^2 \frac{u_2(\theta_2 - \theta_0)}{[\ln(Z_2/Z_0)]^2} \quad (7)$$

For the gradient method

$$Q_L = L \rho k^2 \frac{(u_2 - u_1)(e_2 - e_1)(0.622/P)}{[\ln(Z_2/Z_1)]^2} \quad (8)$$

$$Q_s = c_p \rho k^2 \frac{(u_2 - u_1)(\theta_2 - \theta_1)}{[\ln(Z_2/Z_1)]^2} \quad (9)$$

Observations of wind velocities (u_2, u_1), vapor pressures (e_2, e_1) and temperatures (θ_2, θ_1) are made at two levels (Z_2, Z_1) above the surface.

The release of heat by cooling and freezing of rainwater is important when it is falling on a snowpack, but is generally considered negligible when rainwater falls on glacier ice (Röthlisberger and Lang 1987). In particular, the release of latent heat of freezing (Q_m) is important in ripening or metamorphosing the snowpack by reducing its cold content (heat necessary to raise the temperature of one unit area to 0°C) and increasing its temperature nearer to its melting point. Ambach's (1961) analyses at Hintereisferne, Austria, however, have suggested that rainfall may be important in reducing the cold content of the upper layers of glacier ice late in the melt season.

Heat from precipitation is given by

$$Q_p = C m_p (T_p - T_0) \quad (10)$$

where m_p = mass of rainwater (g cm^{-2}) falling on the ice surface at temperature $T_0 = 0^\circ\text{C}$

T_p = temperature of the raindrops at impact

C = specific heat of water ($4.1868 \text{ J g}^{-1} \text{ } ^\circ\text{C}^{-1}$).

The heat released by freezing Q_m is simply

$$Q_m = S m_p \quad (11)$$

where S is specific latent heat of freezing and m_p as above. To define its effects on the snowpack, the cold content (C) can be calculated from

$$C = c_s \int \rho T_s(z) dz \quad (12)$$

where c_s = specific heat of snow or ice ($2.1 \text{ J g}^{-1} \text{ } ^\circ\text{C}^{-1}$ at 0°C)

ρ = density of the snow pack at depth z

T_s = temperature at depth z .

Thus, if $Q_m = C$, the released latent heat of

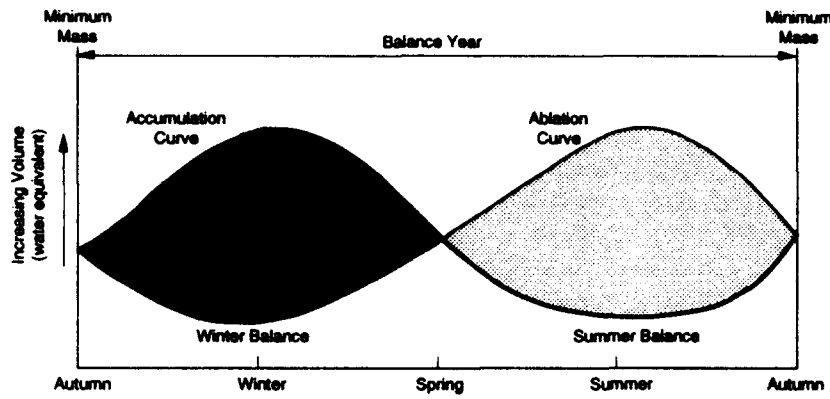


Figure 6. Mass balance illustrated by plots of accumulation and ablation through 1 balance year. The difference between the negative summer balance and positive winter balance defines the annual mass balance for the glacier (after Sugden and John 1976).

freezing from a quantity of rainwater falling on the snowpack would have the potential to warm it to the melting point temperature, a factor considered important in calculating meltwater retention during runoff.

Overall, net radiation is the largest component of total heat available for melting, while latent heat flux is important in governing daily variations in melt rate (Lang et al. 1977, Lang 1980). Whereas net radiation increases with altitude, both the sensible and latent heat tend to decrease as a result of the vertical lapse rates of air temperature and vapor pressure. These rates control the level above which melting processes are strongly reduced or cease altogether. Meltwater production in general is thus a function of net radiation, snow-ice albedo and cloud cover duration (e.g., Meier 1969).

As stated previously, long-term (decadal to century) trends in runoff largely reflect the effects of climate on the mass balance of the glaciers, rather than simply the daily and seasonal variations in energy. Mass balance is the net change in mass of the glacier over some period, generally considered over 1 year (Fig. 6) (Paterson 1981). Mass balance (b) is thus the difference between the total accumulation of snow and ice (c) and the total amount lost to ablation by evaporation and sublimation (a) or

$$b = c - a. \quad (13)$$

In a hydrological sense, the accumulation area of the glacier S_{ac} is that area in which there is a net gain or positive mass balance observed at the glacier surface (see Fig. 5). Below the equilibrium line or climatic snowline, where ablation and accumulation are in balance, ablation exceeds accumulation, resulting in a negative mass balance across the lower ablation area S_{ab} .

Direct measurements of glacier mass balance are commonly made at representative points within the accumulation and ablation areas on the same date during each mass balance period. These data are then integrated over the glacier surface area s

to define annual net mass balance B as

$$B = \int_0^{S_{ac}} bn_{ac} ds + \int_0^{S_{ab}} bn_{ab} ds \quad (14)$$

where bn_{ac} is net mass balance at point n in the accumulation area and bn_{ab} at a point in the ablation area.

Changes in mass are reflected in glacier activity, with glacier flow responses generating changes in total ice volume and in ice distribution within the basin. These changes may be reflected hydrologically in runoff over decades and longer (Fig. 7).

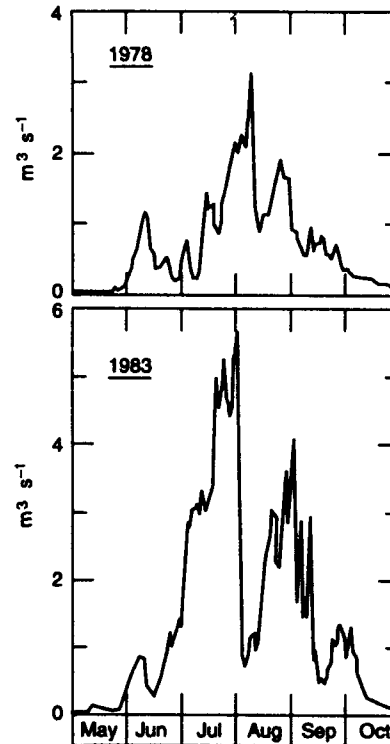


Figure 7. Mean daily flows at Vernagtbach, Austria, with 84% glacier cover during an ablation minimum period and positive mass balance year (1978), and during a maximum ablation period and negative mass balance year (1983) (after Röthlisberger and Lang 1987).

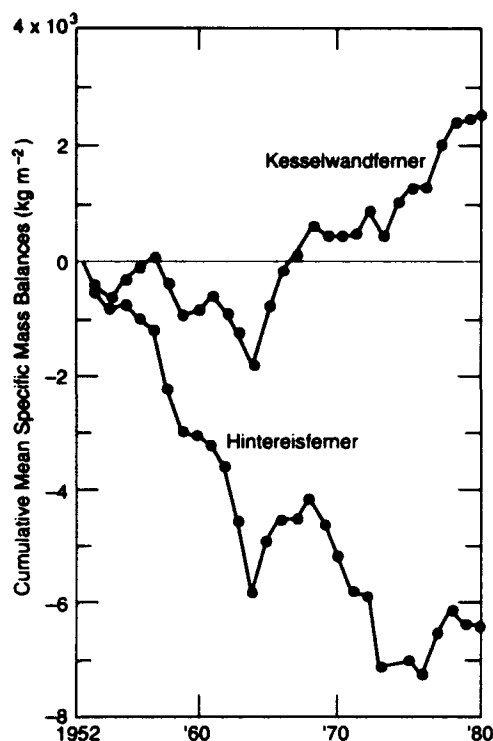


Figure 8. Cumulative mass balances of two adjacent glaciers—Hintereisferner (9.08 km²; 3050-m elevation) and Kesselwandferner (4.44 km²; 3180-m elevation)—Oetztal Alps, 1952–1980. Elevation as well as other factors contribute to the sharp contraction in mass balance (after Röthlisberger and Lang 1987).

Gradual changes in glacier mass balance may affect annual runoff volumes as well, particularly in basins with smaller glaciers that are apparently in equilibrium with the present climate and therefore respond relatively rapidly to short-term climatic changes (e.g., Kuhn et al. 1979, 1985).

Hydrological estimates of net glacier mass balance equate changes in basin water balance to changes in mass balance. While this method is also used, Tangborn et al. (1975) concluded that it did not reliably predict mass balance because water storage within the glacier is not considered in the estimate. Röthlisberger and Lang (1987), however, indicate that this method can provide a reasonable estimate of mass balance if it is applied over periods of several hydrological years, each beginning and ending during the winter low flow season when water storage is at a minimum (Fig. 8).

In the hydrological method, storage change (mass change equivalent) in water balance ΔS is

$$\Delta S = P - R - E \quad (15)$$

where P = precipitation measured across the basin
 R = runoff measured at a gauging site
 E = amount of ice, snow and water lost to evaporation and sublimation.

In order for $\Delta S = B$, the volume of liquid stored in the glacier must be negligible, a condition that does not generally exist. Further, this relationship assumes that essentially there are no annual changes in the glacier's ice volume or geometry, and thus the unlikely condition that the glacier is in equilibrium with a stable climate. In addition, measurements of precipitation and evaporation are commonly not available within the immediate basin and therefore meteorological stations away from the basin are used as surrogates to calculate each parameter, introducing further errors that are difficult to assess.

Mass balance estimates may additionally be based upon repeated photogrammetric surveys run at minimum snow cover and spanning periods of 5 or more years (Lang and Patzelt 1971, Tangborn et al. 1975). Comparative analyses suggest that these estimates provide a reasonable approximation of the mass balance. Significant improvement results, however, when a minimum number of spot measurements of accumulation and ablation are incorporated into the estimate, even when they are only taken intermittently.

Unfortunately, the effects of climate on glacier mass and energy balance and runoff are complex and poorly understood (e.g., Bowling and Trabant 1986, Röthlisberger and Lang 1987). Glacier mass balance is itself a complex product of regional and micro-climatic regimes, local and regional topography and other processes and parameters that are likewise poorly defined. The nature of the response and the time scale of that response to these controls vary from glacier to glacier and region to region (e.g., Reynaud 1988). The distribution of a glacier above and below the equilibrium or climatic snowline is an essential parameter of the climate-mass balance relationship that is reflected in the diurnal, monthly and annual variations in runoff, as will be described subsequently (Röthlisberger and Lang 1987).

Studies have shown that long-term changes in climate and mass balance are reflected in runoff, although not necessarily as a direct response to such changes. Glacier recession during periods of negative mass balances, for example, results in runoff additional to that which would have been produced if the glacier existed in a steady state; however, sustained deglaciation during sustained negative mass balances results in diminished flows. Although mass balance changes affect glacier thickness and

thus area and length, historical fluctuations in length and area over time are not directly related to mass balance changes, making it difficult to use such information directly in long-term runoff predictions (Reynaud 1988). For example, Kasser (1973) has shown that runoff was significantly reduced by 36% in the upper Rhone River basin in Europe during 1916–1968, when the glacier cover declined from 16.8 to 13.6%. This reduction was a longer-term response that was opposite to the initial increase in discharge as the general warming in climate took place at the beginning of the century. During the same period, the high Alpine Matter Vispa basin above Zermatt, Switzerland, had discharge decrease by 22% as glacier cover was reduced by 15% (Bezingue 1982). It may in part reflect changes in precipitation, with a reduced input counteracting any increase in meltwater production. In addition, Kasser (1981) analyzed the overall effects of changes in glacier cover for all of the Swiss Alps from 1876 to 1973, concluding that a loss of about 26% glacier cover resulted in an annual reduction in specific runoff of about 3.4%.

Runoff data for estimating future water yield must therefore be considered in lieu of glacier mass balance and climatic changes during the period of record (e.g., Bredthauer and Harrison 1984). Collins (1987), for example, analyzed the period of 1931–1960 for several Alpine basins in Switzerland, concluding that runoff predictions based upon simply this period of record, without considering mass balance changes that resulted from climatic changes, would over-predict future runoff. During shorter time spans, to which many records for glacierized basins are limited, similar responses can affect runoff records. Data from Aletschgletscher, Switzerland, showed record discharges for the period 1940–1949, the warmest period on record, when glaciers exhibited negative mass balances. However, during the late 1960s and 1970s, when weather was cooler, glaciers had

Table 1. Example of runoff variations ascribable to mass balance changes in the Nigardsbreen Glacier, Norway (after Ostrem and Wold 1986).

Year	Observed runoff (10^6 m^3)	Glacier mass balance (10^6 m^3)	Theoretical runoff*
Negative mass balance effects			
1969	249.3	-61.7	187.6
1970	210.9	-21.1	189.8
Mean	230.1	-41.4	188.7
Positive mass balance effects			
1964	166.9	40.2	207.1
1965	154.2	37.1	191.3
Mean	160.6	38.7	199.3

* Corrected to eliminate glacier influence.

positive mass balances, and runoff was at a minimum (Fig. 9) (Röthlisberger and Lang 1987).

A relatively rapid response of runoff to annual changes in mass balance may also characterize some glaciers, as is illustrated by data from the basin with the glacier Nigardsbreen in Norway (Ostrem and Wold 1986). During 1964 and 1965, the glacier had positive mass balances, decreasing by approximately 25% the total annual discharge from the river when compared to discharge for the glacier in a theoretical, steady-state condition (Table 1). In contrast, in 1969 and 1970, the glacier had negative mass balances that resulted in an increase of nearly 18% in runoff.

In general, the prevailing negative mass balances of glaciers in most regions of the world during the last century must be considered in interpreting runoff records. Because of the nonlinear relationship of climate to mass balance caused by various feedback mechanisms, and the nonlinear relationship between mass balance and glacier cover and distribution, runoff predictions based upon future scenarios of climatic change are difficult to make. In

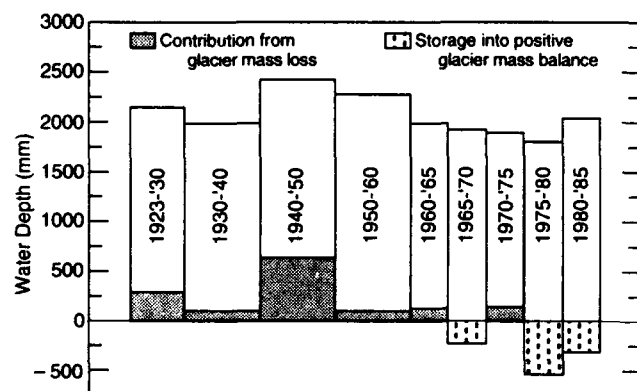


Figure 9. Mass balance and runoff in the Aletschgletscher basin for the period 1922–23 to 1984–85. Minimum runoff occurred during 1975–1980 in a period of positive mass balance and a maximum runoff occurred during 1940–1950 in a period of negative mass balances and glacier recession (after Röthlisberger and Lang 1987).

addition, glaciers located within the same climatic regime, including those located in adjacent basins, may react differently to global climatic changes because of regional variations in temperature and precipitation, as well as because of other microclimatic effects on glacier mass balance that result from global changes (e.g., Meier 1983).

Conclusions from statistical analyses of long-term hydrologic records versus specific climatic variables and glacier mass balance remain speculative (e.g., Kasser 1959, Collins 1985, 1987, Kuhn et al. 1985). Even in areas with the best runoff records, data on mass balance and climate are commonly missing or problematic. Meteorological stations are generally located at some distance from the glacierized portion of the basin and are limited in number. Furthermore, the parameters measured consistently (mainly temperature and precipitation) are not necessarily those of importance. Observations of glacier advance or retreat, which are more commonly available through historical archives, have been used to infer mass balance, but may be incorrect because, owing to a time lag in response, positive or negative mass balances may be associated with either advancing or receding glacier termini (e.g., Mayo and Trabant 1984).

GLACIOHYDROLOGIC PROCESSES AND FACTORS

Runoff variability over decades or shorter periods is linked directly to the glaciohydrologic processes and controls on meltwater movement or storage within the glacier, including the snowpack of the accumulation area. Our understanding of these complex processes and controls is still rather limited, but improving.

In many, if not most, instances, the alpine and valley glaciers in basins of interest for water resource development are those that are relatively warm and within which meltwater is relatively abundant (see cover photo). Thermally, such glaciers are classified as temperate, the ice being at or near the pressure-melting point throughout. Glaciers that are mainly at the pressure-melting point, but that are in some part below it, are called subpolar. Temperate and subpolar glaciers, which are prevalent in North America, Europe, Scandinavia and Greenland, not only have a significant volume of meltwater on and within them, but also have significant seasonal increases and decreases in discharge that result in similar variations in rivers draining these basins.

In general, water moves from the surface of these glaciers through intergranular passages, veins and conduits, and reaches the interface between the glacier's sole and its bed. Here, flow may be a thin film or be concentrated in conduits, cavities or large tunnels that feed streams draining to the glacier terminus. While snow-melt, surface and subglacial ice-melt, and precipitation form the bulk of this discharge, water from subglacial groundwater aquifers and from tributary streams draining the adjacent nonglacierized areas of the basin are also potential water sources for this drainage system.

Snowpack drainage

Runoff from snow-covered areas of the glacier consists of meltwater from the upper parts of the snowpack and rain water, both of which affect the energy balance at the snowpack surface. Meteoric water and meltwater introduced into the snowpack progressively change its physical and hydrologic properties, eventually resulting in a mature snowcover that possesses somewhat more uniform characteristics. As water moves into and percolates through the snowpack, it causes a rapid metamorphism of snow particles at depth. In this process, larger grains grow at the expense of smaller grains; the latter are gradually eliminated and the edges of larger grains are progressively rounded until a spheroidal shape emerges (Wakahama 1968, Colbeck 1973b). Material properties change as the grains grow and density increases.

Snowpacks are, however, heterogeneous, even after alteration by metamorphic processes. Properties reflect the history of snow accumulation, the character of the snow deposited (grain size, water content), the number of cycles of diurnal melting and refreezing at the surface between snowstorms, the number of freezing rain events that form ice layers of variable thickness and extent, and the activity of wind in creating dense, hardened crusts (e.g., Male 1980, Colbeck 1982). Snowpacks are therefore stratified, with vertical and lateral variability in grain size and shape, density, porosity and permeability. These properties in turn affect melt metamorphism.

The paths of flow through the snowpack can be rather tortuous, reflecting its heterogeneous nature. Important in this regard are dense, low-permeability horizons and ice layers that reduce vertical water percolation, causing local variations in saturation and rerouting of flow downslope (Fig. 10) (e.g., Colbeck and Davidson 1973, Colbeck 1973a, 1978). Delays of several hours in the

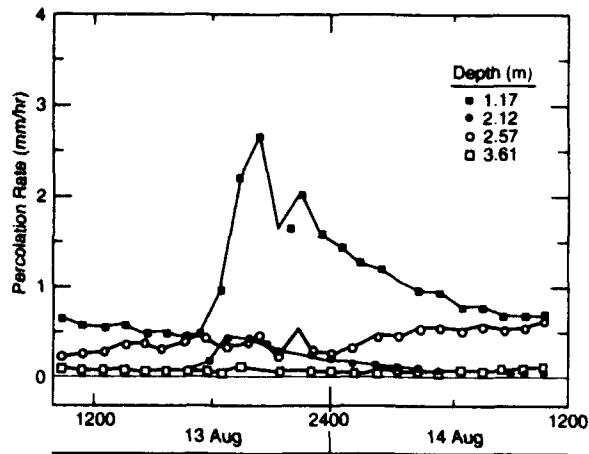


Figure 10. Percolation rates of meltwater into snow stratified with ice layers and lenses in the accumulation area of Ewigschneefeld, Grosser Aletschgletscher, Switzerland. Rates based on water volume collected in funnels set at depths indicated (after Lang et al. 1977).

diurnal pattern of meltwater runoff may result from the ponding of water above the ice layers and the resultant dispersion of meltwater moving through the snow. The introduction of large amounts of meltwater can, however, form vertical drainage channels through the ice layers, thus reducing their role in retarding vertical percolation (Gerdel 1954, Langham 1974). Such layers may still cause a net lateral flow downslope, as each successive layer is encountered (Fig. 11). Vertical channels or veins can continue to grow by melting, as heat is introduced by meltwater percolation, conduction and solar radiation (Langham 1975). Between ice layers, continued grain growth, which increases the intrinsic permeability of the snow, and the formation of vertical, more permeable coarse-grained flow "channels" within the maturing snow, which increase in size as the meltwater percolation rate increases (Gerdel 1954), also enhance drainage, even though melt metamorphism increases snowpack density.

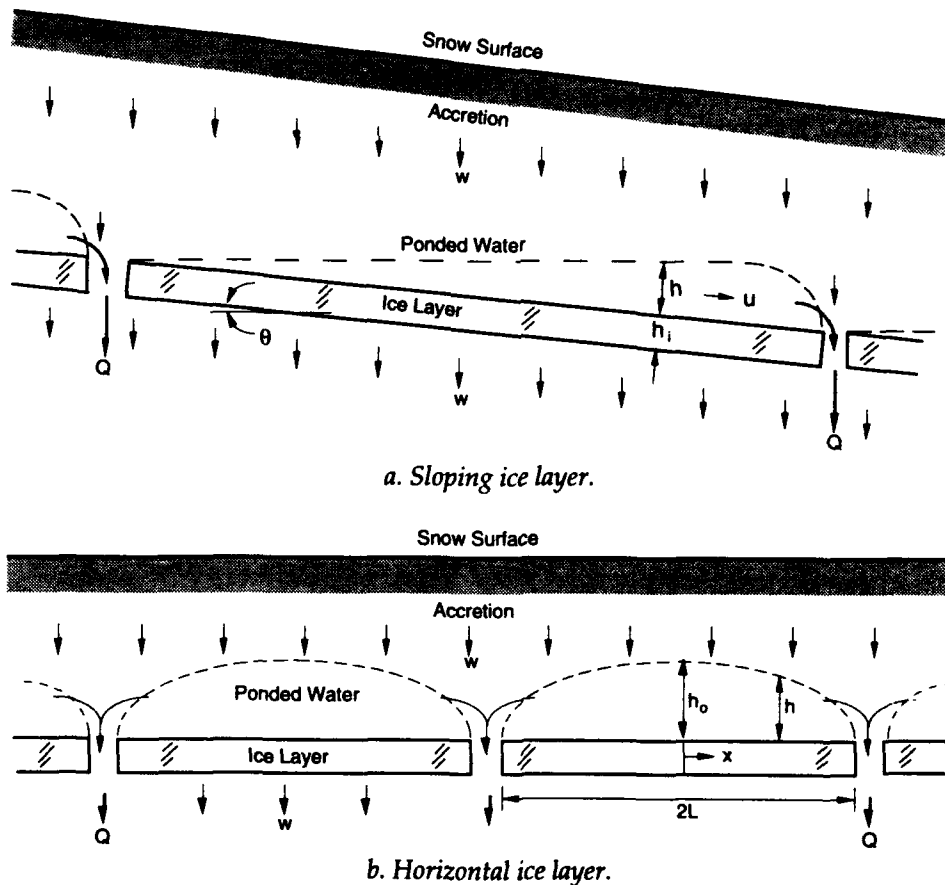


Figure 11. Effects of sloping and horizontal ice layers on vertical water movement through pores w and drains Q in a snowpack (after Colbeck 1973, 1978). Vertical drainage channels can develop that effectively increase percolation rates, but cause retardation and lateral flow downslope.

The process of water movement through the snowpack is dominated by gravity (Gerdel 1945) and can be described generally using Darcian flow theory for an unsaturated, homogeneous, porous medium (Colbeck 1971, Colbeck and Davidson 1973). According to Darcy's law, the volume of water flowing per unit area per unit time (u) is

$$u = (K_1/\mu_1)[\partial p_c/\partial z + \rho_1 g] \quad (16)$$

where K_1 = permeability
 μ_1 = water viscosity
 ρ = water density
 $\partial p_c/\partial z$ = capillary pressure gradient
 g = acceleration of gravity.

Colbeck (1974b) has shown that the pressure gradient related to capillarity can be ignored in the case of a rapidly metamorphosing snowpack and, therefore, the volume flux of water in the unsaturated snowpack is given by

$$u = \alpha K S^{*3} \quad (17)$$

where α = a constant = $\rho_1 g_1 \mu_1^{-1}$
 K = intrinsic permeability
 S^* = effective liquid water saturation or the fraction of pore volume with moving water (Colbeck and Davidson 1973).

The latter term can vary greatly, depending upon the magnitude of surface water flux and the snow's permeability (Colbeck 1977).

Water seepage through the snowpack with time is then given by

$$\frac{dz}{dt}\bigg|_u = 3\alpha^{1/3} k^{1/3} \phi_e^{-1} u^{2/3} \quad (18)$$

where $dz/dt|_u$ = rate of downward movement of the flux

ϕ_e = effective porosity [$\phi_e = \phi(1 - S_{wi})$]
 ϕ = porosity
 S_{wi} = water held by capillary forces (Colbeck and Davidson 1973).

S_{wi} is relatively small, as capillary forces are not transmitted readily within snow (Colbeck 1979). Equation 18 expresses the important relationship showing that larger values of flux move more quickly than smaller values of flux, and that at any given depth, the passing diurnal wave of meltwater is first characterized by a rapid increase to a peak value followed by a slower decline, and an extended recession that results in a continuous flow of water at depth.

Colbeck (1979) has shown that within snowpacks that are complex and stratified, the calculated surface flux for the gravity flow theory overestimates peak flows while underestimating the timing of runoff. Dunne et al. (1976) have shown that these inaccuracies result from the heterogeneous nature of the snowpack, which is not considered by this model. Improved predictions using the gravity flow model could result if field measurements of the surface and bottom fluxes of the pack are used to calculate values for $k^{-1/3} \phi_e^{-1}$, thus effectively integrating the effects of the snowpack's characteristics on the timing and magnitude of wave front propagation and water flux (Colbeck 1977).

Runoff prediction can be significantly improved by modeling the heterogeneous nature of flow in complexly stratified snowpacks, which actually have multiple flow paths, as well as distinct flow channels. Flow is also affected by internal variations in permeability and the transient ponding of water on ice layers (Colbeck 1979). These complexities can be somewhat simplified by considering simultaneous flow down a series of flow paths, with the flux of water within flow channels considered separately from a background flow or flux of water associated with propagation of the wetting front.

Colbeck (1979) modified his original gravity flow theory (Colbeck 1971) to account for the generation of flow "fingers" at crusts, for flow down existing drainage channels, and for temporary ponding on ice layers. Flow channels are generated in dry, cold snow as "fingers" that propagate from fine-grained crusts ahead of the wetting front, leaving behind dry, cold snow. After flow fingers are established by the first movement of water through a snowpack, they become preferential paths for later waves of meltwater because of the grain growth and permeability increase within them. Larger-grained flow channels and drainage channels in ice develop subsequent to flow fingers. Flow fingers, as well as channels, are thus critical to routing water through a layered snow cover and contribute a substantial amount to the overall flux of meltwater through a snowpack.

Multipath flow is considered in terms of wetting front propagation over time t (Colbeck 1978)

$$\frac{d\xi}{dt} = \alpha^{1/3} \frac{k^{1/3}}{\phi_e} (u_1^{2/3} + u_1^{1/3} u_2^{1/3} + u_2^{2/3}) \quad (19)$$

where ξ = depth of the wetting front
 α = constant as in eq 17
 ϕ_e = effective porosity

u_1 = water flux feeding the wetting front
 u_2 = water flux ahead of the wetting front
 k = intrinsic permeability.

The flux of water is a function of the intrinsic permeability and effective saturation (S^*)

$$u = \alpha k S^{*3}. \quad (20)$$

The flux of water through each finger (f) is related to the flux of water arriving at the crust layer ($\bar{u}$) as

$$f = \bar{c} \bar{u} \quad (21)$$

where $\bar{c}$ is a constant greater than 1, and the number of fingers (n) is also proportional to the flux ($n = cu$).

The rate of propagation of an advancing finger is then described by

$$v = \alpha^{1/3} \frac{k^{1/3}}{\phi_e} (\bar{c}^{2/3} \bar{u}^{2/3} + \bar{c}^{1/3} \bar{u}_2^{1/3} + u_2^{2/3}) \quad (22)$$

where u is the flux feeding the crust that generates the finger and u_2 is the flux being overtaken by the rapidly advancing finger. The effects of ice layers on storage and reducing overall infiltration rate can also be considered in a step-by-step iteration using

$$V = \frac{1}{8} \pi \phi L \sqrt[3]{I/\alpha k} \quad (23)$$

where V , the volume of ponded water, is related to the infiltration reaching the ice layer (I), the distance between drains (L), the snow porosity (ϕ), and the snow permeability (k).

To apply this model, as Colbeck (1979) illustrates, requires construction of the flow field following eq 18 and propagation of the wetting and finger fronts by eq 19 and 23 respectively. Discharge is then calculated by combining the flux of each component over its fractional area. The effectiveness of this model is clearly related to the amount of data available on the state of the snowpack, including the number, thickness and depth of crusts and ice layers, dimensions of fingers and channels and related snowpack properties. These data are generally not known in detail, however, thereby limiting practical application of the model at this time.

A simplified version, in view of the lack of data, was also proposed by Colbeck (1979). It modifies the gravity flow model to consider multiple flow paths, each of which acts independently of one

another, and the number, size and properties are based upon judgment and whatever data are available.

Total flow (u) is then the summation of the contribution of each flow path (f_i)

$$\bar{u} = \sum_{i=1}^n u_i f_i \quad (24)$$

and the average snowpack properties can be defined from

$$\frac{\phi}{k^{1/3}} = \left[\sum_{i=1}^N f_i \left(\frac{\phi_i}{k_i^{1/3}} \right)^{3/2} \right]^{2/3} \quad (25)$$

and based upon measurements of input and discharge. Flow must be routed along each path separately and based upon properties assigned to it.

Below the seasonal snow cover, density gradually increases with depth as the snow densifies into firn, predominantly as a result of regelation (Colbeck and Parssinen 1978). At increasingly greater depths, the dense snow changes from firn into glacier ice, with the most rapid transformations also taking place when meltwater is present (Shumskiy 1964). The depth at which ice transforms apparently differs from glacier to glacier and ranges from as little as 13 m to over 40 m in temperate regions where meltwater is important (e.g., Paterson 1981).

When a seasonal snow layer overlies glacier ice in the ablation area, a fully saturated horizon develops above it, especially early in the year when the uppermost part of the ice is seasonally frozen and water is kept from entering the glacier. Flow within this saturated layer can also be treated using Darcy's law as (Colbeck 1974a)

$$u = -(k/\mu_l) (\partial p_l / \partial x - \rho_l g \theta) \quad (26)$$

where k = permeability
 μ_l = water viscosity
 ρ_l = water density
 x = distance measured along the slope
 g = acceleration of gravity
 θ = slope of ice surface.

Because metamorphism takes place rapidly under saturated conditions, the grain size is larger than in the overlying unsaturated snow and the effective permeability and thus seepage rates are higher in the saturated zone. In addition, frictional heating opens and enlarges channels within the

lower-most part of the saturated layer as melt increases, thereby decreasing runoff response times later in the melt season. The flux of water (u) through this layer is therefore given by (Colbeck 1974a)

$$u = \alpha k_s \theta \phi^{-1} \int_{t_0}^{t_L} I(t) dt \quad (27)$$

where ϕ = layer porosity
 k_s = layer permeability
 I = infiltration rate
 L = length of flow
 θ = slope angle
 α = constant.

The duration of flow over a length L occurs over time ($t_L - t_0$) and is given by

$$t_L - t_0 = \frac{L\phi}{\alpha k_s \theta} \quad (28)$$

Water flow in the basal layer of the snowpack integrates the diurnal water fluxes from the unsaturated zone, with the extent of integration into a single flow front determined by the slope angle and the length of flow to the point of discharge from the snowpack.

Runoff from the permanent snowpack or firm of the accumulation area is more complex than in the ablation area. Here, not only must the seasonal snowcover undergo melt-metamorphism and drainage develop, but flow paths through the underlying, denser "aged snowpack," which is being constantly altered by compression, compaction and regelation to form glacier ice, must develop. In the first case, the gravity flow theory can be applied (e.g., Ambach et al. 1981). In the latter case, for which there are few measurements, vertical water movement and retention within the firm or permanent snowpack may be related to intergranular processes and flow in small tubes or capillaries as occurs in true glacier ice (as described later).

There often is a saturated horizon within the firm of temperate glaciers that acts as an unconfined aquifer, storing water and retarding its movement into the glacier's drainage system. The upper surface of this saturated layer or water table varies in depth from several meters to 40 m or more (e.g., Sharp 1951, Vallon et al. 1976, Schommer 1977, Oerter and Moser 1982, Östling and Hooke 1986). Its configuration roughly approximates that of the ice surface, but is progressively deeper with distance up-glacier of the equilibrium or firm line, and

is depressed around glacial drainage features such as moulins and crevasses (Schommer 1977, Lang et al. 1977). Calculations of water storage within the firm area of Cascade Glacier were made by Fountain (1989), who concluded that a 1.25-m seasonally saturated layer of water formed, representing about 12% of total meltwater storage within the glacier. This water layer fluctuated daily after the seasonal production of meltwater was sufficient to fully saturate it. Water flow in this saturated horizon can be defined according to eq 26 following Colbeck (1974a).

Glacier drainage

Meltwater flowing from the seasonal and permanent snowpacks either discharges at the firm line onto the glacier surface within supraglacial channels, or enters the glacier directly through a variety of ways. The internal drainage system is complex and knowledge of it is based to a large degree on theory, with only limited field measurements and observations.

The idealized drainage system within temperate and subpolar glaciers (Fig. 12) consists of 1) small veins (micrometer size) feeding progressively larger tubes (millimeter size) and conduits (centimeter size and larger) with increasing depth below the ice surface, and 2) larger drainage sinks, including moulins and crevasses, which feed surface meltwater directly into conduits that likewise eventually join together at depth as an arborescent (tree-like) system of passageways (e.g., Shreve 1972, Röthlisberger and Lang 1987, Hooke 1989). At the bed of the glacier, the subglacial system may consist of both a series of conduits, which may be linked together by cavities, and of one or more larger tunnels (greater than 1 m in diameter) into which the majority of the conduits feed. There probably is a thin film of water at the bed in those areas removed from conduits or tunnels.

Englacial

Nature and behavior. Both the existence and significance of intergranular flow or seepage of water through glacier ice remain the subject of debate. Nye and Frank (1973) have argued that there should be free drainage in temperate glaciers through a system of veins along three-grain boundaries (Fig. 13), and that such veins should join at four-grain intersections as nodes (Fig. 14) to form a network of capillary-size tubes (Nye 1989). Raymond and Harrison (1975) found veins in Blue Glacier, Washington, of up to 25 μ m in diameter in ice cores from depths to 60 m. They concluded that, owing to a

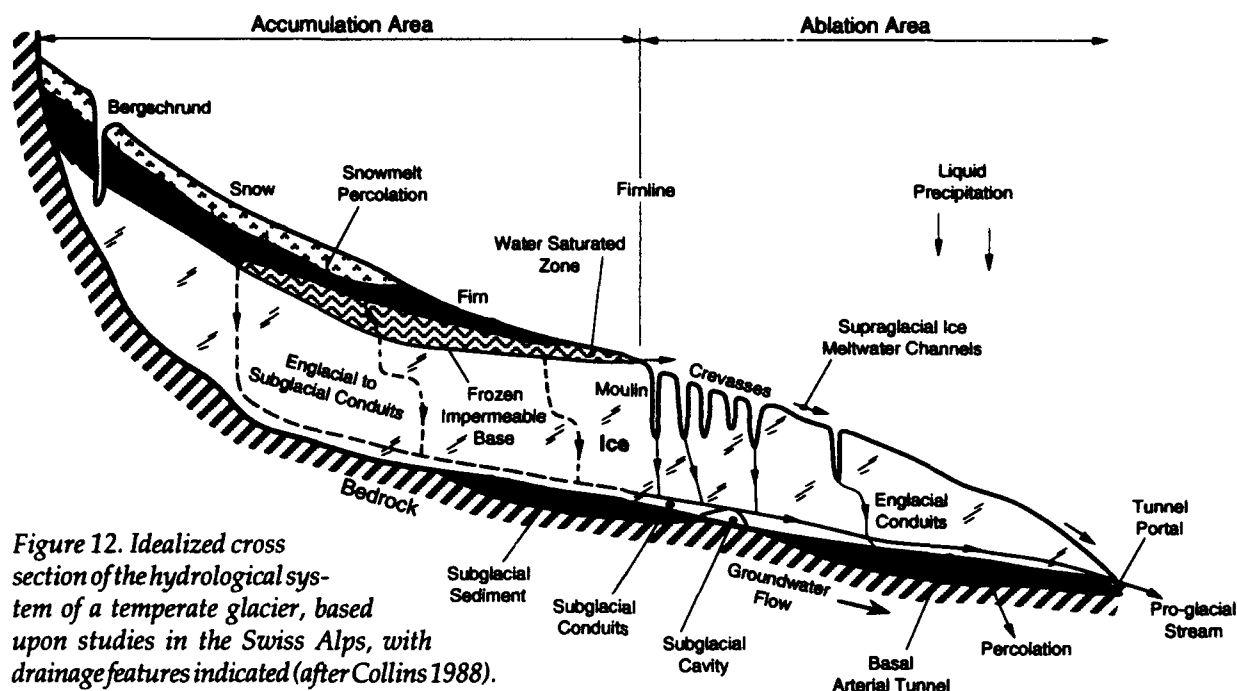


Figure 12. Idealized cross section of the hydrological system of a temperate glacier, based upon studies in the Swiss Alps, with drainage features indicated (after Collins 1988).

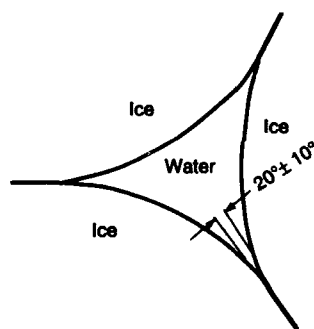


Figure 13. Geometry of water vein at a triple grain junction (after Nye and Frank 1973).

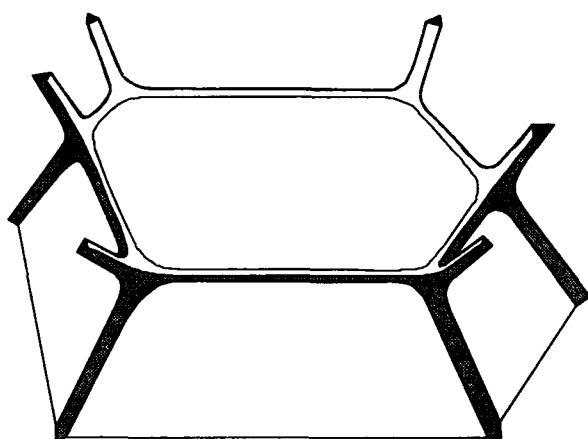


Figure 14. Veins and nodes in polycrystalline glacier ice (after Nye [1989] as modified from Smith [1948]).

variety of factors, drainage through such veins was negligible. Calculations suggested discharge of less than 10^{-1} m year⁻¹. Lliboutry (1971) has argued against the importance of intergranular flow, as processes such as deformation and recrystallization are likely to close veins, and air bubbles within the veins are likely to block free flow. In contrast, Wakahama et al. (1973) described water drainage through an intergranular vein network at rates of 10^{-7} to 10^{-5} m s⁻¹, while Berner et al. (1977) calculated rates of 7.6×10^{-8} m s⁻¹.

Raymond and Harrison (1975) also found 1- to 2-mm tubular conduits within ice cores from 9.6- and 20-m depths, the former exhibiting an upward branching network of several tubes. In contrast to intergranular veins, they concluded that a few such capillaries per square meter of the glacier's surface would be sufficient to drain the season's ablation runoff of 25 mm day⁻¹ or 9 m year⁻¹ at Blue Glacier.

Theoretical analyses by Shreve (1972) have suggested that such small veins and tubes are likely to grow with depth into the glacier, under the assumption that the release of potential energy with discharge results in more heat being generated per unit of wall area in the larger, rather than smaller, veins. Thus, they will expand more rapidly and increase in size at the expense of smaller veins to develop a tree-like network of conduits fed from the surface by intergranular seepage. Röthlisberger's (1972) analysis of conduit flow hydraulics also indicates that, on the basis of head losses in com-

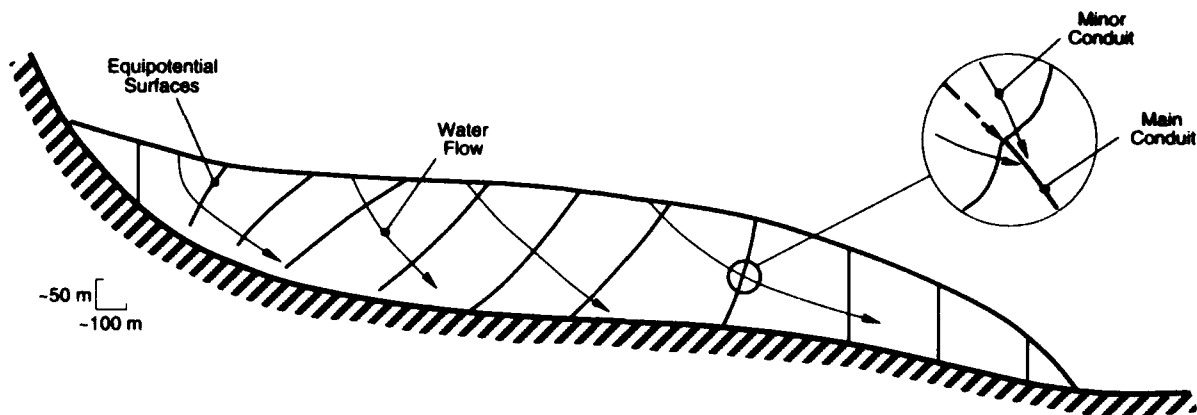


Figure 15. Longitudinal section of a glacier illustrating upglacier dipping equipotential surfaces and directions of englacial water flow (Shreve 1972). Near larger conduits, equipotential surfaces bend up-glacier, resulting in diversion of flow into smaller conduits and veins towards the larger conduits (after Hooke 1989).

peting parallel channels of different size, conduits should progressively enlarge and coalesce with depth.

Shreve's analysis examined the overall flow pattern and the trend and orientation of veins and conduits by defining a hydraulic potential (Φ) for the drainage network that is primarily a function of ice pressure (Fig. 15). Flow is expected to be normal to equipotential surfaces, defined by

$$\Phi = \Phi_0 + \rho_w g Z + \rho_s g (H - Z) + p(\dot{r}) \quad (29)$$

where Φ_0 = reference potential

ρ_w = density of water

ρ_s = density of ice

g = acceleration due to gravity

H = elevation of glacier surface

Z = elevation of a point within or at the bed of the glacier

$\dot{r}$ = rate of conduit closure by plastic flow of ice

$p(\dot{r})$ = pressure that is a function of $\dot{r}$ and is related to the difference in pressure between water in a conduit and ice adjacent to the conduit.

The second and third terms of eq 29 are the potential energy of water from its height above a datum, and the pressure in the water from the overlying ice respectively.

By differentiating eq 29 with respect to distance s , setting the result to 0 and solving

$$-(\rho_w - \rho_i) s \, dz/ds = \rho_i g \, dH/ds \quad (30)$$

or

$$dz/dx = -\rho_i / (\rho_w - \rho_i) \, dH/dx \quad (31)$$

where s is along an arbitrary direction in which Φ is constant, x is horizontal distance, and dz/dx is the angle of the equipotential surface.

We find equipotentials to slope downward in the up-glacier direction at about 11 times the slope of the ice surface (Fig. 16). The general direction of drainage perpendicular to equipotentials Φ is given as the angle δ

$$\delta = \arctan \frac{\rho_i \cos \alpha}{\rho_w - \rho_i \cos \alpha} (\tan \alpha) \quad (32)$$

where α is surface slope. Thus, water will tend to flow steeply downwards in the direction of areas with a gentle surface slope, but the flow will become more nearly parallel to the surface as the surface increases in steepness.

At the bed or in the subglacial flow system, Shreve's theory indicates that water will move perpendicular to the intersection of the equipotential surfaces with the bed. Thus, the direction of flow may not be in the direction of maximum bed slope, but may in fact be diagonally across such

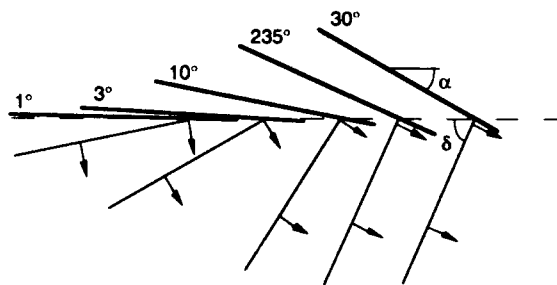


Figure 16. Inclination δ of equipotential lines (solid) with seepage gradients (arrows) indicated for various surface slopes α (after Röthlisberger and Lang 1987).

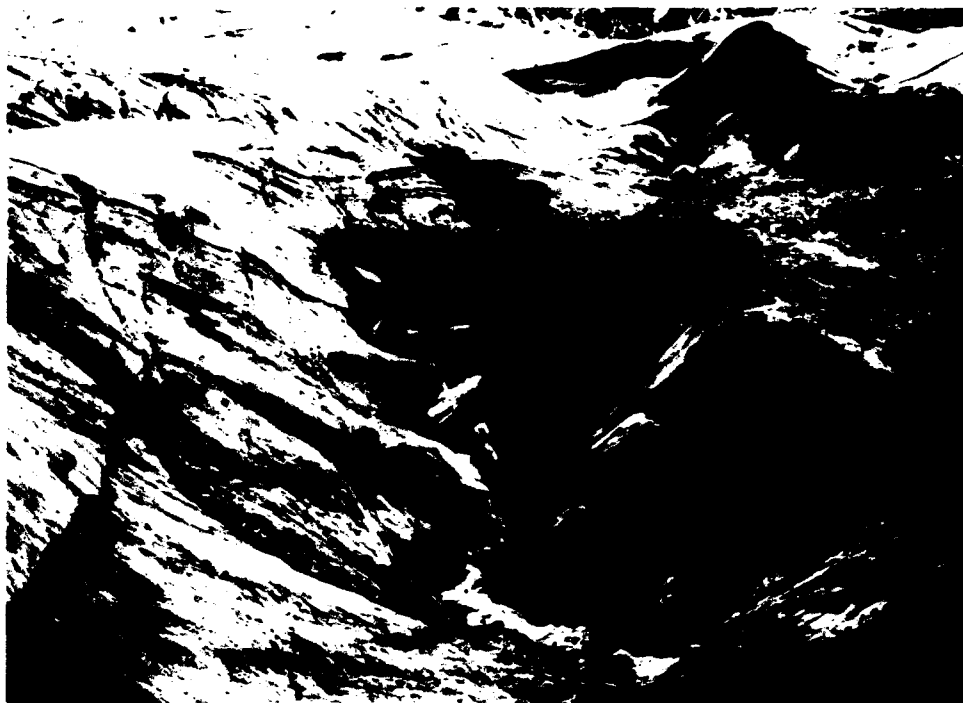


Figure 17. Moulin fed by supraglacial stream in the ablation area of the Matanuska Glacier, Alaska. Water from this moulin discharges subglacially at the glacier's terminus. The opening at the ice surface is about 15 m wide; the conduit fed by the moulin is about 2 m wide, but narrows at depth.

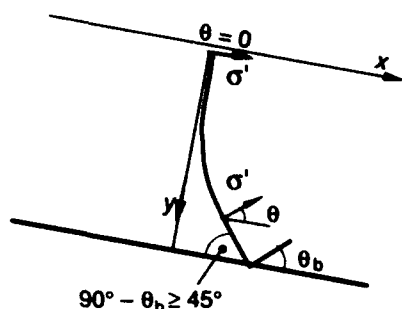
slopes. Movement of water up an up-glacier dipping bed (against glacier flow) is also possible. However, such conclusions need to be tempered because water may tend to flow in existing channels in the bed, rather than creating new ones (Röthlisberger and Lang 1987). This still depends on equipotential orientations and on localized pressure variations as the ice flows around meter-scale objects in the bed.

The actual existence of this intergranular-fed conduit flow system has not been fully verified and its importance to total volume of runoff is unknown. The volume of surface water reaching the subglacial drainage system in this mode is probably a small percentage of the total glacier runoff during the peak of the melt season. If parts of the glacier are seasonally or permanently below the melting point in a layer near their surface (e.g., Hooke et al. 1983, Röthlisberger and Lang 1987), water is prevented from seeping intergranularly. Seasonal low temperatures in the ablation area, for example, can cool ice to below the pressure-melting point and create frozen, impermeable near-surface layers.

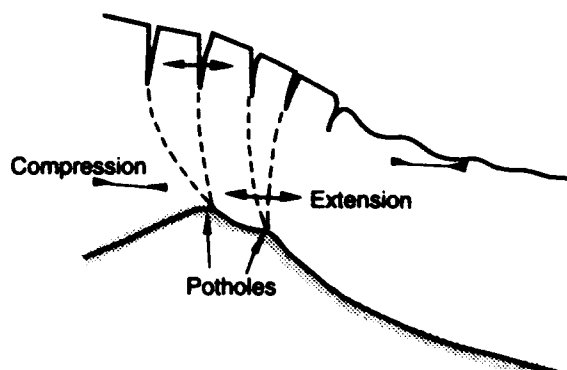
In addition to seepage through veins or capillaries, surface-derived meltwater and rain water

enter the glacier through much larger passageways created in the process of glacier flow and deformation (Fig. 17). Fracturing and crevassing are the primary means of establishing vertical passageways into the ice. Crevasses form where tensile stresses exceed the tensile strength of the ice, either through changes in the stress field along the ice flow path, or through changes in the stress field generated by subglacial water pressures. Moulins, sink holes in the glacier surface, appear to form when supraglacial streams are intersected by a newly formed fracture or crevasse, the water plunging into the base of the opening. If the crevasse has intersected an active englacial conduit, water may then freely drain through it. In other cases, energy released by descending water may rapidly enlarge any smaller passageways to accommodate this water and establish drainage in conduits to greater depths (Hooke 1984, Holmlund 1988).

Robin (1974) and Weertman (1974) have also hypothesized that a crevasse that fills with water could propagate downward to the bed simply in response to the difference in density of the water relative to that of the ice. As the excess pressure created by the pooled water increases with depth,



a. Orientation of the principal stress deviator σ' as a function of depth y in the glacier.



b. Hypothetical relationship among stress field, crevasse propagation, drainage and pothole formation.

Figure 18. Crevasse propagation and englacial drainage (after Röthlisberger and Lang 1987).

the cracks tend to extend preferentially downward but also tend to dip down-glacier and flatten as they approach the bed in response to the existing stress field within the ice (Fig. 18). Crevasse propagation would provide large drains for meltwater; potholes commonly found in bedrock of former glacier beds may be evidence of this process.

Conduits leading from the bottoms of moulin or crevasses, which may be up to 30 to 40 m deep, wander in the general direction of the source crevasse, possibly joining other conduits. Eventually, they plunge down near-vertical conduits at greater depth below the moulin (Holmlund 1988). Indirect field measurements of a moulin conduit's orientation showed it to gradually change from near-vertical to a lower angle of inclination in the down-glacier direction (Iken 1972, 1974) consistent with the equipotential concept of Shreve (1972) (Fig. 19). Calculations by Hooke (1984) on the other hand, suggest that, for veins and conduits greater than 3 to 4 mm in diameter, melting of passage walls by mechanical energy would result in faster melting on the gravitationally lowest side, increasing in rate with depth, and thus resulting in a near-vertical conduit orientation that would be enhanced by glacier flow.

The length of time over which moulin-fed conduits remain open at the glacier surface appears to vary, with a rather limited number of observations (e.g., Hooke 1989) suggesting 1 to 2 years, or possibly longer. Conduits will close through plastic flow of ice, but water input will affect the size of the opening and determine whether plastic flow results in closure. Seasonal reductions in meltwater flow could therefore result in conduits clos-

ing annually. Such closures affect the location of drainage from the surface to the glacier's bed and can alter the timing and volume of discharge. Following complete closure when water flow is lost, a new passage to the subglacial drainage system must be established to transport the annual supply of meltwater and meteoric water that is on the glacier's surface during the melt season.

The precise nature of the englacial drainage system below moulin is basically unknown. Only

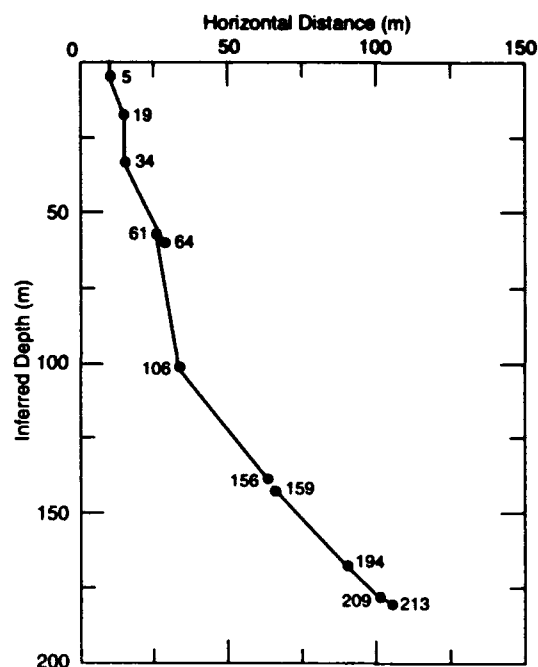


Figure 19. Indirectly measured slope of conduit draining moulin on White Glacier, Axel Heiberg Island, Canadian Arctic (after Iken 1974).

a few visual observations have been made (e.g., Carol 1947, Holmlund 1988). Dye tracing studies or meltwater chemical analyses (e.g., Behrens et al. 1971, 1982, Collins 1977) indirectly indicate the existence of relatively direct routes of conduits to terminus streams in some instances, as well as relatively slow flow along indirect and possibly tortuous routes in others (e.g., Stenborg 1967, 1973, Ambach et al. 1972, Krimme et al. 1973, Burkimsher 1983, Oerter et al. 1985, Seaberg et al. 1988, Willis et al. 1990). In some cases, the complete loss of the dye tracer suggests that meltwater is being stored internally and that the system locally lacks a direct connection between the moulins and terminus streams. The same glacier may exhibit different drainage patterns, with both rapid, direct conduit-to-terminus stream drainage and slow through-flow of meltwater discharges (e.g., Willis et al. 1990).

Whether the drainage system consists mainly of downward coalescing (or upward branching) conduits or of more discrete and separate conduits feeding the subglacial drainage system remains to be resolved. Conditions, including the nature of subglacial drainage and bed properties, and the nonsteady conditions of both water input and ice flow with time, have feedbacks that complicate simple theoretical treatments of conduit behavior and drainage system characteristics.



Figure 20. Large tunnel or portal at the margin of Maclaren glacier. Atmospheric conditions prevail in such instances (see man next to tunnel for scale) (photo by E. Evenson).

Significant questions concerning the nature of water flow within englacial conduits remain. Parameters including conduit size and shape, pressure variations within conduits, and thermal conditions around conduits affect theoretical treatments of drainage, yet there are few measurements of such parameters. Prevailing theories are therefore based upon limited field measurements and remain relatively simple.

Theoretical treatments. Water flow within conduits has generally been examined theoretically for a circular shape, with variations in hydraulic pressure being treated as a function of changing boundary conditions along the channel length using numerical integration (Röthlisberger 1972, Shreve 1972, Weertman 1972, Spring 1980, Spring and Hutter 1981, 1982, Lliboutry 1983, Hooke 1984). Each theory is based upon the concept that conduit size is determined by the rate of frictional melting of ice walls by flowing water, and the rate of conduit closure is determined by plastic deformation of the ice resulting from an overburden pressure that exceeds the water pressure within the conduit. Further, thermal conditions are considered, particularly the gain or loss of energy as water adjusts to the pressure-melting point. Glacier surface topography, conduit location within the glacier and discharge are additional factors.

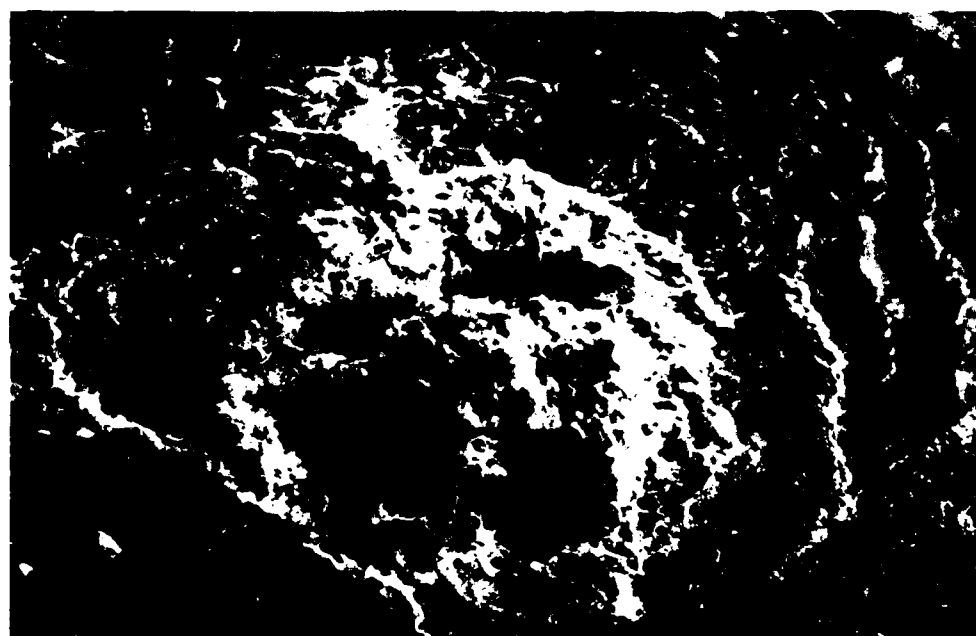
Near the surface, and possibly to depths of 100 m, atmospheric conditions may be common, depending on discharge (Hooke 1989). At greater depths, however, conduits will probably be under hydraulic pressure most of the time, and theoretical treatments typically assume this. The existence of air cavities and hydraulic pressure variations resulting from changes in channel configuration and other conditions will therefore affect the applicability of theoretical treatments (Röthlisberger 1972). Near the point of conduit discharge at the glacier terminus, atmospheric conditions are commonly assumed (Fig. 20). Such simplified conditions, however, are not always applicable and water can emerge under pressure in termini that end on land or in water (Fig. 21).

Under steady-state conditions and constant discharge Q , a conduit of unit length ds will remain at a given size (and pressure) if the effects of ice deformation to close the passageway are offset by melting of conduit walls as determined by heat transfer between the water and ice (Fig. 22). Thus, the volumetric change resulting from melting or freezing is balanced by ice flow and is expressed as

$$\dot{m}_t + \dot{m}_p = \dot{m}_c \quad (33)$$



a. Upwelling, right below center.



b. Turbulent jet.

Figure 21. Water emerging at the margin of the Matanuska Glacier, Alaska. Here, subglacially fed conduits remain under pressure.

where $\dot{m}_t$ = volumetric frictional melt rate
 $\dot{m}_p$ = volumetric rate of melting (freezing) related to changes in the pressure-melting point along the flow path
 $\dot{m}_c$ = volumetric rate of conduit closure by plastic ice flow.

For a conduit element of incremental length ds , the incremental pressure loss is equal to the change in piezometric level ($df = f_2 - f_1$) along it (measured relative to a horizontal datum), while the incremental change in water pressure is $dp = p_2 - p_1$; df is therefore the gradient of pressure loss because of friction.

The frictional melt rate $\dot{m}_t$ is then calculated using the relationship (Röthlisberger and Lang 1987, after Röthlisberger 1972)

$$\dot{m}_t = C_m^{-1} \rho_i^{-1} \rho_w g Q \left(-\frac{df}{ds} \right) ds = C_1 Q \left(-\frac{df}{ds} \right) ds \quad (34)$$

where C_m = energy of fusion
 ρ_i = density of ice
 ρ_w = density of water
 g = acceleration due to gravity
 Q = discharge
 C_1 = constant.

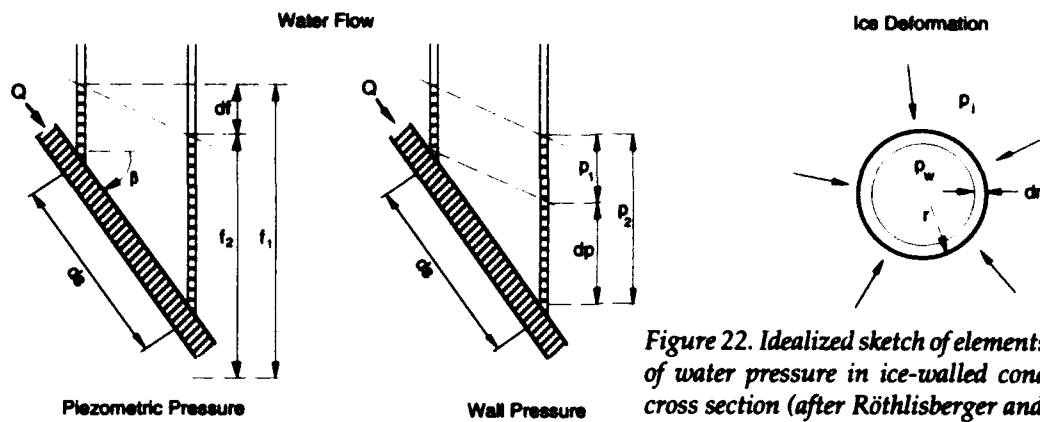


Figure 22. Idealized sketch of elements in computation of water pressure in ice-walled conduits of circular cross section (after Röthlisberger and Lang 1987).

The length of the conduit element ds is defined using standard Cartesian coordinates (Fig. 22) as

$$ds = \sqrt{dx^2 + dy^2 + dz^2} \quad (35)$$

with ds positive in the direction of flow. For a conduit sloping at angle β , water pressure change becomes

$$-dp = -df + ds \sin \beta. \quad (36)$$

β is positive. The rate of melting or freezing along the conduit wall $\dot{m}_p$ is

$$\dot{m}_p = C_t C_w C_m^{-1} \rho_w^2 \rho_i^{-1} g Q \left(\frac{dp}{ds} \right) ds = C_2 Q \left(\frac{dp}{ds} \right) ds \quad (37)$$

where C_t = change of pressure-melting point with pressure

C_w = specific heat capacity of water (Table 2)

C_2 = constant.

C_t for pure ice and water is 0.074 kPa^{-1} , while for air-saturated water and pure ice it is 0.098 kPa^{-1} .

The values for the constants C_1 and C_2 are calculated using numerical values for the physical constants of ice and water at 0°C (Table 2); for pure ice and water, $C_1 = 3.21 \times 10^{-5} \text{ m}^{-1}$ and $C_2 = -10^{-5} \text{ m}^{-1}$. For air-saturated water, $C_2 = -1.33 \times 10^{-5} \text{ m}^{-1}$. Insufficient data currently exist to determine which value of C_2 is appropriate. Röthlisberger and Lang (1987) agreed that changes in gas concentrations along the conduit length are unlikely and, on the basis of CO_2 measurements by Metcalf (1984), concluded that an undersaturated condition can be justifiably assumed.

Table 2. Physical constants of ice and water at 0°C and related properties (after Röthlisberger and Lang 1987).

Symbol	Property	Quantity	Unit
g	Acceleration due to gravity	9.807	m s^{-2}
ρ_w	Density of water	999.8	kg m^{-3}
ρ_i	Density of ice	916	kg m^{-3}
c_w	Specific heat of water	4.217×10^3	$\text{J kg}^{-1} \text{K}^{-1}$
c_m	Energy of fusion	3.34×10^5	J kg^{-1}
$C_{t,0}$	Change of pressure-melting point with pressure (pure water)	-7.4×10^{-8}	K Pa^{-1}
$C_{t,s}$	Change of pressure-melting point with pressure (air-saturated water)	-9.8×10^{-8}	K Pa^{-1}
η	Viscosity of water	1.787×10^{-3}	$\text{kg s}^{-1} \text{m}^{-1}$
A	Ice deformation coefficient used by various authors		
	Paterson (1981)	0.167	$\text{bar}^{-3} \text{a}^{-1}$
	Hooke (1984)	0.244	$\text{bar}^{-3} \text{a}^{-1}$
	Lliboutry (1983)	0.232	$\text{bar}^{-3} \text{a}^{-1}$
	Röthlisberger (1972)	1.00	$\text{bar}^{-3} \text{a}^{-1}$
n	Exponent in the power law of ice deformation	3.00	—
k	Hydraulic roughness parameter for various conduits		
	Smooth	100	$\text{m}^{1/3} \text{s}^{-1}$
	Medium	50	$\text{m}^{1/3} \text{s}^{-1}$
	Very rough (torrent)	10	$\text{m}^{1/3} \text{s}^{-1}$

Lliboutry (1983) and Hooke (1991) favor the use of larger values for air-saturated water.

Conduit closure rate assumes radial ice flow across the circular conduit cross section and $\dot{m}_c$ is defined by the relationship

$$\dot{m}_c = 2\pi r \dot{r} ds = 2\pi r^3 A n^{-n} (p_i - p_w)^n ds \quad (38)$$

where r = conduit radius

$\dot{r}$ = rate of decrease in conduit radius

A = ice flow law deformation parameter

p_i = pressure of overlying ice at the wall (expressed as $[\rho_i - \rho_w]h \cos \beta$, water column height)

p_w = water pressure in the conduit

n = exponent in the flow law of ice (Nye 1953).

The flow law of ice is assumed to follow the general form of

$$\dot{\epsilon} = (\tau/A)^n \quad (39)$$

where τ is effective shear stress and $\dot{\epsilon}$ is effective strain rate.

Discharge Q is defined in terms of either laminar flow

$$Q = \frac{\pi \rho_w g}{8\eta} \cdot r^4 \left(-\frac{df}{ds} \right) \quad (40)$$

where η is the viscosity of water, or turbulent flow as (Röthlisberger 1972)

$$Q = 2^{-2/3} \pi k r^{3/8} \left(-\frac{df}{ds} \right)^{1/2} \quad (41)$$

where k is the hydraulic roughness factor. Turbulent flow has also been represented by the Darcy-Weisbach law (Spring and Hutter 1982), in which discharge is proportional to the two-fifths power of r , rather than the three-eighths power.

Substituting for r using a generalized relationship for Q from eq 41 and 42 in eq 40, we see that

$$\dot{m}_c = C_3 Q^{-b} (p_i - p_w)^n \left(-\frac{df}{ds} \right)^m ds. \quad (42)$$

where exponents q and n are determined from eq 41 and 42 and the factor C_3 varies with conditions in the particular reach of the conduit (Röthlisberger and Lang 1987). Factors such as mode of water flow (turbulent, laminar), channel roughness and ice flow deformation will cause C_3 to vary spatially and temporally within the drainage system. Exponents b and m are less than 1.

The relationship in eq 35 can be defined in terms

of water pressure p_w or hydraulic gradient ($-df/ds$). For steady state in a sloping conduit (angle β), the relationship for dp follows eq 38, with the $\tan \beta$ equaling the conduit slope. Thus, following Röthlisberger (1972, eq 20, p. 185), who derives this expression in more detail, the governing equation for water pressures within a conduit is

$$\begin{aligned} & \left[\frac{dp_w}{ds} + \rho_w g \tan \beta \right]^{11/8} \\ & - 0.316 \left[\frac{dp_w}{ds} + \rho_w g \tan \beta \right]^{3/8} \frac{dp_w}{ds} \\ & = D k^{-3/4} (nA)^{-n} Q^{-1/4} (\cos \beta)^{-11/8} (p_i - p_w)^n \end{aligned} \quad (43)$$

where D is related to ice and material properties by $2^{3/2} \pi^{1/4} C_m \rho_i (\rho_w g)^{3/8} = 3.63 \times 10^{10} (\text{N m}^2)^{11/8} \text{m}^{-3/8}$ from Table 2.

This relationship (eq 43) can also be expressed in terms of the pressure or hydraulic gradient $-df/dx$

$$\begin{aligned} & \left(-\frac{df}{dx} \right)^{1/8} + 0.462 \rho_w g \tan \beta \left(-\frac{df}{dx} \right)^{3/8} \\ & = \frac{D}{0.684 k^{-3/4}} (nA)^{-n} Q^{-1/4} (\cos \beta)^{-11/8} (p_i - p_w)^n \end{aligned} \quad (44)$$

(Röthlisberger 1972). An alternative relationship for the steady state has also been developed by Nye (1976) in terms of the effective pressure $N = p_i - p_w$ as

$$N = \left[\left(\frac{v_i \rho_w g_s}{KL} \right) \left(\frac{\rho_w g_s}{N_T} \right)^{3/8} \right]^{1/n} Q^{1/4n}. \quad (45)$$

Here $v_i = 1/\rho_i$ = specific volume of ice

ρ_w = density of water

ρ_i = density of ice

K = viscosity constant proportional to A (ice deformation parameter)

L = latent heat

g_s = downslope acceleration

N_T = turbulent drag coefficient.

Because the hydraulic gradient is inversely related to discharge ($-df/ds \propto Q^{-b}$, where b is a constant related to the nature of flow), large discharges are associated with smaller potential gradients and hence lower water pressures, and smaller discharges with higher hydraulic gradients and

higher water pressures (Röthlisberger 1972). Thus, for example, seasonal decreases in discharge under steady-state conditions may result in higher water pressures in winter than in summer. Therefore, where adjacent conduits are constantly discharging, water will tend to drain from the smaller-diameter conduits into larger-diameter ones, in keeping with the concept of an arborescent englacial drainage network as described by Shreve (1972) and Hooke (1989).

Changes in the orientation of the conduit relative to that of the glacier surface and thickness, other factors remaining constant, can significantly alter water pressures (p_w) within them, and thus influence whether they will contract or expand. These variations along the length of flow define the hydraulic grade line (Fig. 23). Of particular consideration for runoff are a conduit's stability for different slopes and orientations relative to ice flow and thickness variations. Figure 24 illustrates changes in these parameters and the resultant configuration in the hydraulic grade and equivalent lines. In the situation where conduits descend under ice that slopes down-glacier (Fig. 24b and c), pressure flow can only occur at a sufficient depth such that ice flow compensates for ice loss from frictional heating (Röthlisberger and Lang 1987). Therefore, in glaciers that are relatively thin (less than 100 m or so), flow would generally be under atmospheric conditions in circular conduits (Hooke 1984). Hooke et al. (1990) concluded, however, that conduits that were broad and low tend to close more rapidly than circular ones and thus maintain pressures under thinner ice near the terminus.

In the situation where conduits are ascending (dipping up-glacier) at an angle much steeper than the slope of the ice surface (Fig. 24d), the hydraulic head or piezometric pressure may equal the overburden pressure, whereupon the water equivalent line of overburden pressure is the hydraulic gradient of flow. Frictional heat is balanced by the heat required to warm the ascending water to the increasingly higher pressure-melting point (Röthlisberger 1972). Because water and ice pressures are equal, ice flow will not close the conduits and new ones cannot form because of the required balance in heat. This situation characterizes conduits ascending at about 30% steeper gradient than that of the ice surface. If the bed slope is 1.5 to 2.0 times steeper than the surface slope (depending upon the degree of air saturation of the water), heat loss will exceed heat gain and supercooling of the water may cause ice growth to maintain the water at the pressure-melting point. This could poten-

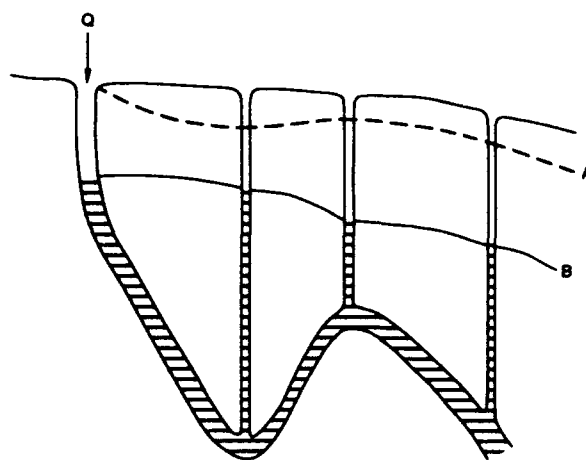


Figure 23. Hydraulic terms of Röthlisberger (1972) illustrated for idealized conduit in a glacier cross section (after Röthlisberger and Lang 1987). Water enters a moulin at ice surface (Q). Dashed line A represents the height of the water column equivalent to the overburden pressure, or water equivalent line. Solid line B represents the height to which water under pressure within the conduit would be free-standing in an open borehole, or the piezometric pressure or hydraulic head variations and the hydraulic grade line.

tially close conduits (Röthlisberger 1972). Ice growth with a constant discharge could cause water pressures to rise, exceeding ice pressures. Hooke (1991) has suggested that water pressures could actually increase to the point where water is forced out of conduits into the interface between the glacier's sole and its bed.

Under nonsteady conditions, in which discharge varies in magnitude with time (conditions that are more representative in general of water flow in glaciers and particularly during large magnitude floods), water temperatures and the rate of heat transfer along the length of flow are important factors. The extreme case of nonsteady flow is outburst floods (jökulhlaups) from ice-dammed lakes (Nye 1976, Spring 1980).

Discharge, frictional heating and the rate of heat transfer, which depends on the initial water temperature, all change during such outbursts. During the initial phase, the potential discharge is greater than the conduit can transport; frictional heating enlarges it to increase discharge. Conversely, reduction in discharge during flood recession would produce conduit contraction under steady-state theory. If heat transfer is considered, however, heat is actually lost by transport to outlet streams as discharge increases and conduit growth is thus limited. If water temperatures are initially

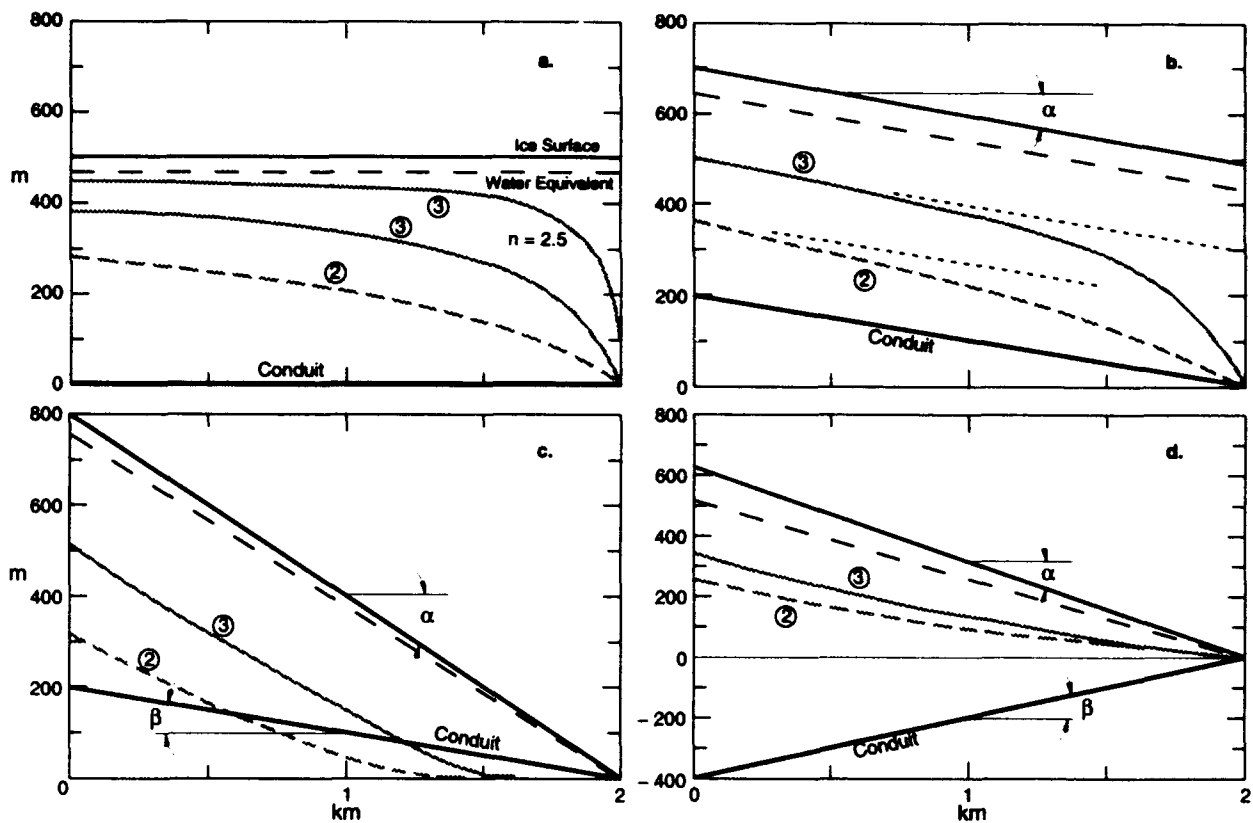


Figure 24. Orientation of conduit relative to ice flow and surface slope under different ice overburden pressures (water equivalent line) determine conduit pressures and stability (after Röthlisberger 1972). Conduit hydraulic heads or piezometric pressures (hydraulic grade lines) were modeled by Röthlisberger (1972) for constant discharge $Q = 10 \text{ m}^3/\text{s}$, channel roughness $K = 10 \text{ m}^{1/3}/\text{s}$ and $n = 3$. Calculations assumed P_w at atmospheric pres-

sure at the discharge point at the terminus and other values for physical constants given in Table 2. Curves (2) use $A = 0.167 \text{ bar}^{-3} \text{ a}^{-1}$ and (3) use $A = 1.00 \text{ bar}^{-3} \text{ a}^{-1}$. For case (a), conduit and surface are horizontal; for case (b), both are at low angles dipping down-glacier; for case (c), the surface dips down-glacier more steeply than the conduit; and for case (d), the conduit dips up-glacier, in the opposite direction of the surface slope.

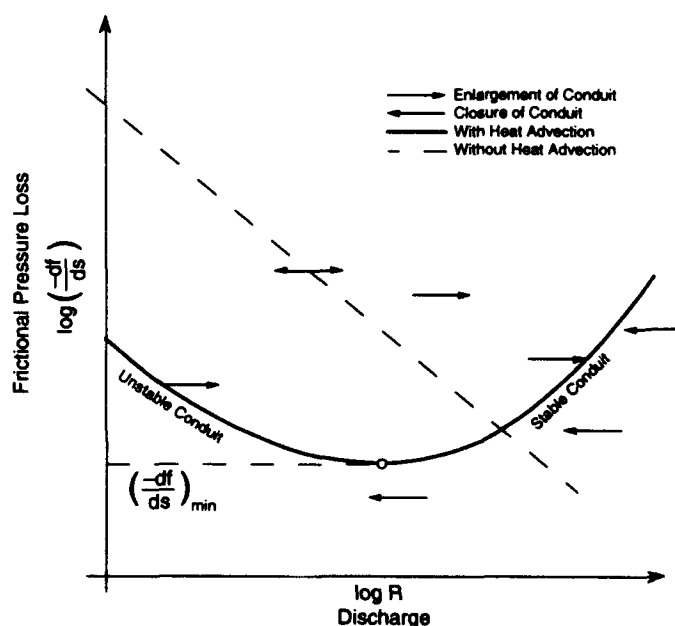


Figure 25. Heat transfer considerations in the stability of ice conduits draining a large reservoir of water on pressure gradient vs. discharge plot. Below the curve, insufficient frictional heat results in conduit closure. Above, surplus heat enlarges the conduit. Dashed line shows closure or enlargement (arrows) if heat transfer is neglected. Right of the dashed line along the solid line, conduits are stable under increasing discharge; below the solid line and left of the dashed line they are unstable (from Spring, unpublished data [after Röthlisberger and Lang 1987]).

at the pressure-melting point, conduit growth may be inhibited, high hydraulic pressures maintained and discharge limited.

Spring and Hutter (1981, 1982) have derived equations for the nonsteady-state conditions of water flow in conduits or tunnels, considering both varying discharge and the effects of heat transfer on conduit growth or shrinkage (Fig. 25). Solutions are not generally straightforward, however, as there are nine unknown parameters, which can vary with time and changes in discharge. Significant simplifying assumptions are thus required to resolve the problem.

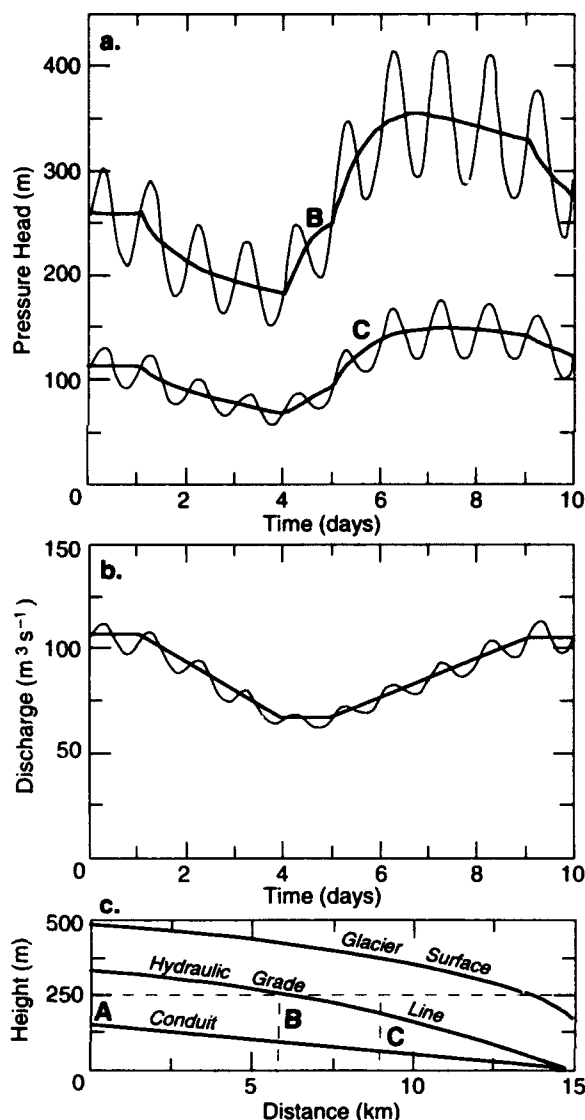


Figure 26. Computer simulation of variations in piezometric water pressures (from Röthlisberger and Lang 1987). Variations at points B and C are in (a) with discharge variations given in (b); locations of points B and C in idealized glacier configuration shown in (c).

The effects of changes in discharge with time can also be evaluated using repeated numerical integration of the steady-state theory along the length of the conduit. An example of temporal variations in discharge are harmonic fluctuations that take place diurnally (Fig. 26). For a discharge with high-frequency fluctuations, over periods ranging from a few hours to a day, the geometry of the conduit cannot respond to the fluctuations and the pressure head varies in phase with the discharge (Röthlisberger and Lang 1987). For very long periods of a year or more, conduits may adjust to a near steady state for the existing conditions and, as a result, pressure variations are out of phase with discharge variations in accordance with the simple steady-state theory. In essence, larger discharges are associated with lower hydraulic pressures and vice versa.

Subglacial

Water flow in the subglacial environment is more complex and less understood than strictly englacial water flow. As with englacial drainage, subglacial theoretical models are based upon limited field observations and measurements.

Subglacial drainage may take place in interconnected conduits or tunnels (conduit-tunnel system) (Röthlisberger 1972), in smaller conduits that link cavities in the bed (linked-cavity system) (Walder 1986, Iken and Bindenschadler 1986, Fowler 1987a, Kamb 1987) or in the absence of channels, as a thin film between the ice and bed (distributed-flow system) (Weertman 1972, Alley 1989a). Channels at the glacier bed may be incised downwards into the substratum (Nye or N-channels [Nye 1973]), or incised upwards into the ice (R-channels [Röthlisberger 1972]). Water may also flow, perhaps preferentially, in existing subglacial fractures, joints or abandoned subglacial channels in bedrock. Which forms of drainage are most important will vary in response to various controlling factors, including the composition and topography of the glacier's bed, ice thickness, ice surface slope and water supply. The proportion of water carried by each drainage system varies from one glacier to another, between different parts of the same glacier, and in both cases, with time (seasonally and annually) (Walder and Hallet 1979, Hantz and Lliboutry 1983, Iken and Bindenschadler 1986, Seaberg et al. 1988, Hooke 1989, Fountain 1991, 1992).

The prevailing direction of subglacial water flow can be assumed to be normal to the contours defined by the intersection of the Shreve equipotential surfaces with the bed (Shreve 1972). In a

glacier-scale sense, ice surface slope, ice thickness and bed topography influence channel location because of the general tendency for water movement from areas of high to low piezometric pressure. Field evidence also indicates that existing drainage features will, in effect, override other controls on direction and become preferential avenues of flow, rather than new channels being established (Röthlisberger and Lang 1987).

The hydraulics of flow within subglacial conduits incised into glacier ice can be treated theoretically using the steady-state approach of Röthlisberger (1972) to determine the pressure gradient in conduits under a constant discharge, as described earlier. In the subglacial environment, however, certain factors differ in their importance from the englacial environment and can have major effects on spatial and temporal differences in conduit or channel hydraulics. While the type of passageway clearly affects flow hydraulics, channel cross section, direction and magnitude of channel slope, and hydraulic roughness are important as well. Beds composed of unconsolidated gravel, for example, will have roughness factors 10 times or more that of an ice-encased conduit (e.g., Röthlisberger and Lang 1987).

While a circular cross section is generally held to approximate englacial conduit shape, conduits in the subglacial environment may differ considerably from this idealized configuration, depending upon related factors such as ice deformation rates, bed composition and bed roughness. Existing passageways or structural controls in bedrock may result in highly asymmetrical channel cross sections. Recent field observations also have suggested that a low, broad profile may be a more realistic representation of conduits incised upwards into or occurring within basal ice (e.g., Hooke et al. 1990). Water pressures in such conduits are higher than in those with the more commonly assumed circular or semi-circular cross section (Hooke 1989, Hooke et al. 1990). Such broad tunnels with limited height may be characteristic of adverse bed slopes where pressures are high and there is a relatively limited supply of heat available to melt conduit walls (Shreve 1985a,b).

Further, ice at or near the bed may locally undergo intense, enhanced deformation, particularly in association with small-scale changes in bed topography, and affect closure rate as well as channel cross section. Additionally, water flow being in contact with the bed and occurring under progressively thinner ice towards the terminus influences discharge, heat transfer and flow hy-

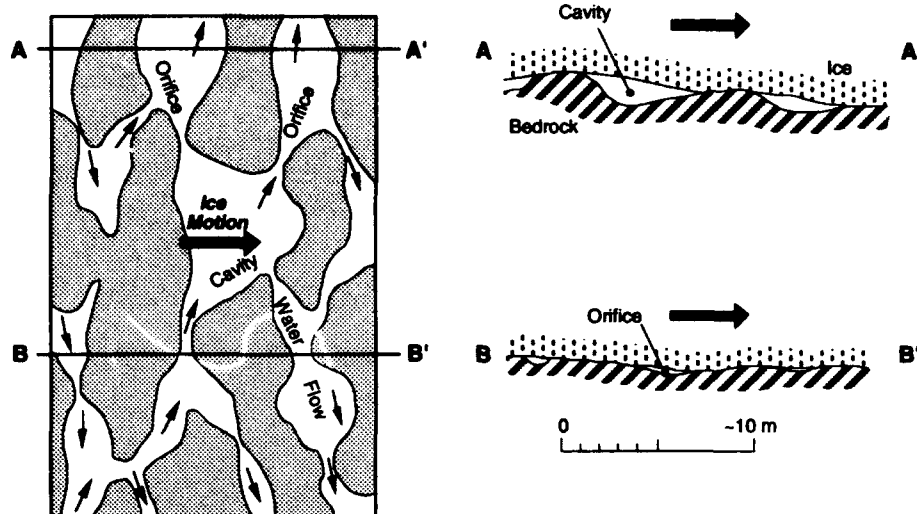
draulics (Spring and Hutter 1981, 1982). Atmospheric conditions within tunnels and conduits may also prevail under the thinner ice near glacier margins (Lliboutry 1983, Hooke 1984).

Bedrock beds

Although there are few quantitative treatments of subglacial drainage owing to a lack of data and direct observations, several physical models of the nature of flow within different drainage systems have evolved. The linked-cavity model of a subglacial drainage system developed on a bedrock substrate consists of an anastomosing series of channels linking cavities in the glacier bed, which are transected by and feed larger conduits or tunnels that trend generally toward the terminus (Kamb et al. 1985, Walder 1986, Iken and Bind-schadler 1986, Fowler 1987a). This model is based primarily upon measurements of water pressures in drill holes and moulins, dye tracer studies of water movement, analyses of the former beds of glaciers, and ice velocity measurements in conjunction with each of these.

Theoretical treatments. Theoretical analyses of the hydraulics and stability of the linked-cavity system and its interaction with the conduit-tunnel system—which basically follows Röthlisberger's (1972) model—have been developed by a number of authors (e.g., Walder 1986, Fowler 1986, 1987a, Iken and Bind-schadler 1986, Humphrey 1987, Kamb 1987). Although the analytical approaches differ, the basic principles governing the hydrology of the drainage system models are similar.

The interconnected or linked-cavity model represents drainage at the bed of a glacier as a series of cavities of different cross-sectional areas and longitudinal sizes and shapes that are linked by short segments of conduit or channel called "orifices" by Kamb (1987) (Fig. 27). Cavities are the result of the separation of ice from the bed where basal sliding is comparatively rapid, the bed is rough and basal water pressures are high (Lliboutry 1968, Kamb 1970, Fowler 1986, 1987a). Normal stresses generated on the stoss side of topographic highs are larger than those on the lee side and result in ice separation on the lee of it (Fig. 27b). Separation may be expanded by (or induced by) water under pressure that has access to the stoss side low pressure zones. Thus, any point where the normal pressure across the ice/bed interface drops below the water pressure, the ice and bed will separate as the water enters it. Normally, large cavities tend to be isolated from one another; however, when the sliding velocities are



a. Schematic of linked-cavity conduit system, in map view, on a small area of the glacier bed. Areas of ice contact with the bed are shaded, areas of ice-bed separation (cavitation) are blank. Vertical cross sections along lines AA' and BB' are shown in Figure 21b. The large arrow indicates the direction of basal sliding. Water flow direction is indicated by small arrows.

b. Vertical cross sections AA' and BB'. Cavity areas are water filled.

Figure 27. Linked-cavity drainage system (after Kamb 1987).

rapid enough and basal water pressures sufficient, connecting conduits develop and provide hydraulic connections between the cavities (Kamb 1987).

Observations of the beds of former glaciers (e.g., Walder and Hallet 1979, Hallet and Anderson 1980) suggest that most cavities are relatively small features, with heights of less than 1.0 m (but variable) and lengths of several decimeters to several or more meters (e.g., Walder 1986, Kamb 1987). The connecting channels are smaller (about 0.1 m in height), but variable in dimension and are important hydraulically in controlling water flow through the linked-cavity system (Kamb 1987). These orifices or conduits transmit the water through the subglacial system and, as such, ice-bed separation and the orifice geometry are more critical hydraulically than cavity characteristics per se.

The actual nature of the linked-cavity system will depend in large part on the configuration and dimensions of bed roughness features and their spatial distribution in relation to basal shear stress, water pressure and ice overburden pressure (Walder 1986, Kamb 1987). The existence and pattern of flow in the linked-cavity system are thus tied to the topography of the bed as in Nye chan-

nels. They differ in theory from Nye channels, however, in being associated with bedrock roughness rather than being channels eroded into bedrock. Similarly, because they are linked to cavitation processes, conduits or orifices tend to be oriented transverse to ice flow direction and their geometry is, in large part, controlled by ice sliding and ice flow rates (Walder 1986, Kamb 1987).

While the hydraulic parameters and processes considered in Röthlisberger's (1972) model for flow in conduits in ice are similar to those considered in the models of the linked-cavity system, the systems differ in the relationship of conduit geometry to the separation or cavity formation process and the roles of heat flux and water pressure. Heat transfer resulting from viscous heating by water flow results in melting of the conduit and cavity ice roofs in the linked-cavity model as it does in the conduit-tunnel system. Heat dissipation can therefore enhance cavity growth during periods of higher water flux and increased water pressure, but overall, it plays a less important role in comparison to the mechanics of glacier sliding in determining and maintaining cavity size (Walder 1986). Heat dissipation may be critical if higher water pressures result in excess ice melt and sliding

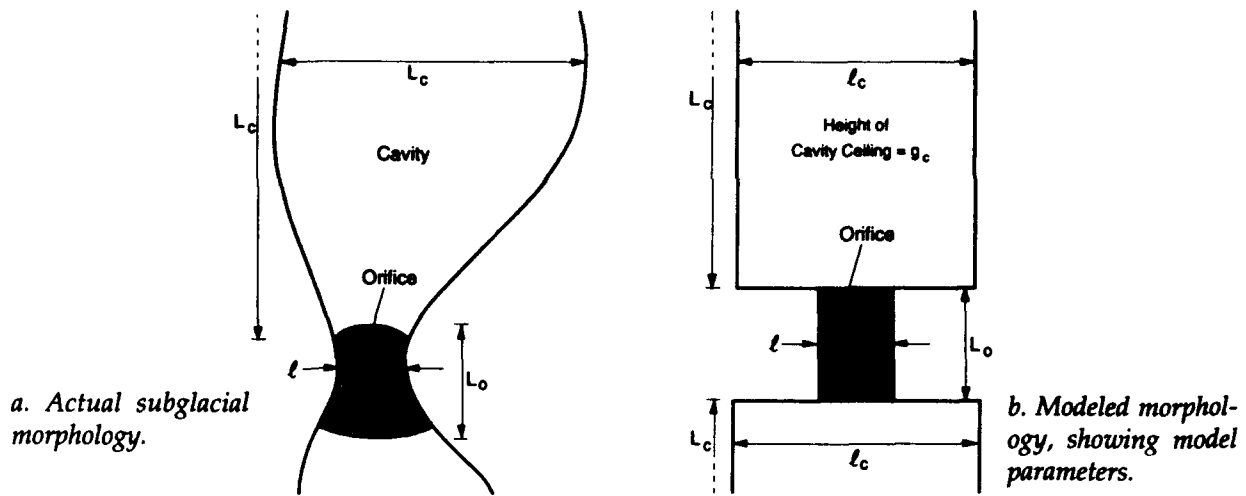


Figure 28. Map view of linked-cavity system (after Kamb 1987).

rates, and lead to an unstable evolution from the linked-cavity system to a Röthlisberger-type conduit-tunnel system (Kamb 1987).

Let us consider the Kamb (1987) linked-cavity model to illustrate the general relationships in the linked-cavity drainage system. Basic geometric relationships and dimensional parameters of cavities and conduit orifices assumed as "typical" for modeling are given in Figure 28.

We assume water flow through the linked-cavity system to be controlled primarily by the hydraulic behavior of the conduit orifices. Hence, water flow is defined in terms of their geometry and, assuming flow is turbulent, can be described by the Manning formula.

Hence

$$\bar{u}_w(x) = M^{-1} \left(\frac{g}{2} \right)^{2/3} \left(\frac{\alpha \Lambda}{\omega} \right)^{1/2} \quad (46)$$

where $\bar{u}_w$ = mean water flow velocity (averaged across the gap height g)

M = Manning roughness factor

g = local height of the orifice at a point $g(x)$

$\alpha \Lambda / \omega$ = hydraulic gradient of the orifice, where

α = surface slope of the glacier

$\Lambda = L_c / L_o$, where L_c is the length of the cavity between orifices and L_o is the length of the orifice. Λ is termed the average head gradient

ω = tortuosity of the flow path.

Water flux through the system Q_w is determined by integrating the mean flow velocity along the breadth of the orifice ℓ for successive gap heights $g(x)$ and summing the contributions of N_o independent orifices across a transverse section of the bed

$$Q_w = \frac{N_o}{2^{2/3} M} \left(\frac{\alpha \Lambda}{\omega} \right)^{1/2} \int_0^{\ell} [g(x)]^{5/3} dx. \quad (47)$$

We assume for simplicity that ice melting during heat transfer from the flowing water to the ice roofs of cavities and conduits takes place at an equal rate per unit area in both features. Viscous heating is defined as the product of the water flux and the potential gradient ($\rho_w g_r \alpha \Lambda / \omega$, where ρ_w is density of water and g_r is gravitational acceleration). Heat transfer is assumed to occur directly to the ice roof and is defined in terms of an equivalent volumetric rate of ice melt. The local heat production and meltback rate $\dot{m}$ per unit area of the orifice roof is

$$\dot{m} = \frac{(\alpha \Lambda / \omega)^{3/2}}{2^{2/3} D M} g^{5/3} \quad (48)$$

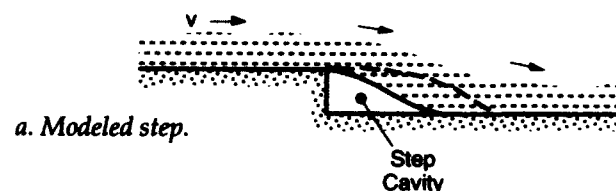
where D = constant, equal to $\rho_i H / \rho_w g_r$, with dimensions of length taken equal to glacier flow length

H = latent heat of melting

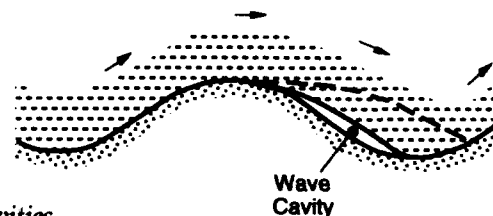
ρ_i = density of ice.

The initial and subsequent configurations of the cavities and cavity roofs, which change with ice sliding and time, thus are important to defining heat dissipation and water flux through the system. Of various idealized cavity configurations illustrated by Kamb (1987), two types of bed topographic features and their resultant separations, referred to as step cavity and wave cavity forms, were chosen for modeling (Fig. 29).

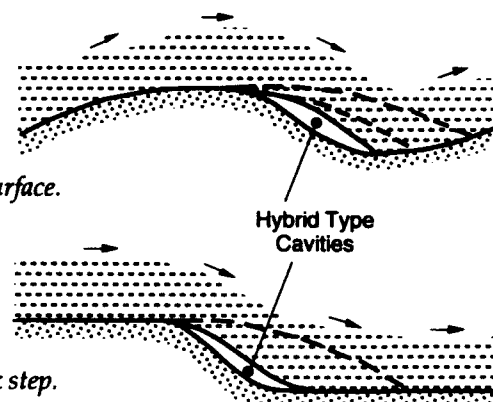
For both step and wave cavities, Kamb uses geometric parameters (cavity height $g(x)$ and length



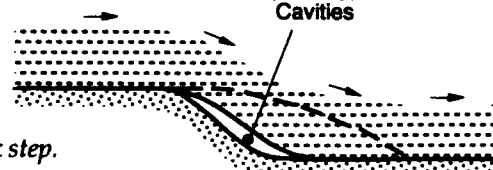
a. Modeled step.



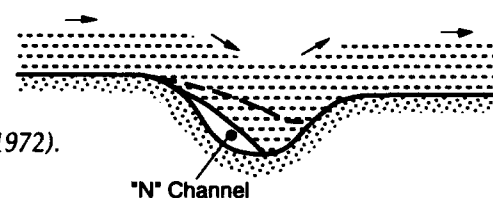
b. Wave cavities.



c. Bedrock surface.



d. Rounded bedrock step.



e. N channel of Weertman (1972).

Figure 29. Various idealized configurations for cavities in the linked-cavity model of Kamb (1987). Figures 29c and d are variations on a and b that would develop in the lee of a "faceted wave" in the bedrock surface (c) and of a rounded bedrock step (d). Separation or gap heights modeled by Kamb are ≈ 0.1 m (after Kamb 1987).

ℓ) to define cavity opening as a function of ice sliding in the down-glacier direction x and roof melting from viscous heat dissipation. Cavity expansion is opposed by a closure rate $w(x)$ resulting from ice flow into the cavity in the vertical direction z . Changes in the gap profile with time following the initial separation from the bed must therefore be considered.

According to the Kamb (1987) model—parameters for idealized step and wave cavities are shown in Figures 30 and 31—the cavity heights $g(x)$ and orifice breadths ℓ are as follows.

For step cavities

$$g(x) = h \left(\frac{1}{2} - \frac{1}{\pi} \sin^{-1} \frac{2x - \ell}{\ell} - \frac{2(2x - \ell)\sqrt{x(\ell - x)}}{\pi \ell^2} \right) \quad (49)$$

where $0 < x < \ell$ and

$$\ell = 4 \sqrt{\frac{\eta v (h + m)}{\pi \sigma}} \quad (50)$$

with h = step height

η = ice viscosity

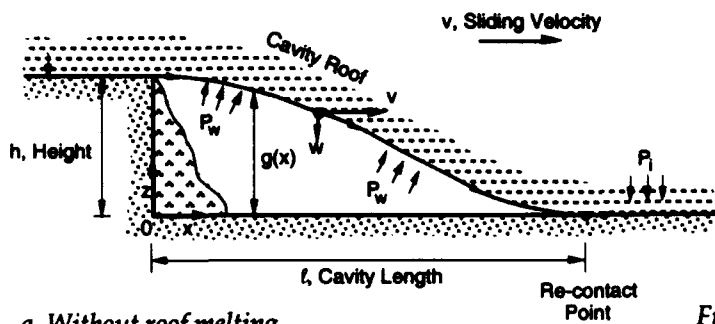
v = ice sliding velocity

$\sigma = P_I - P_w$, with P_I overburden pressure and P_w basal water pressure (i.e., the effective confining pressure)

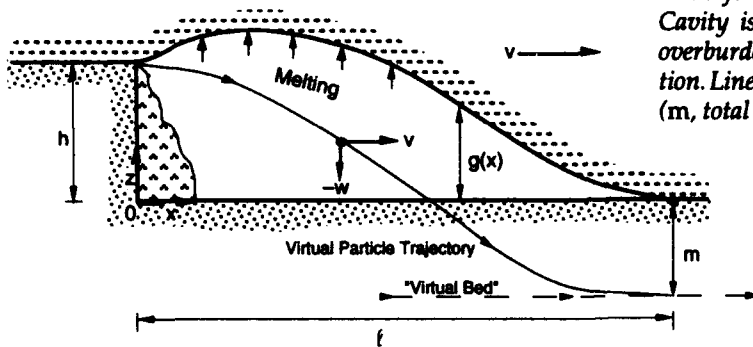
m = constant that equals 0 if there is no melting of the roof.

Closure rate by ice flow $w(x)$ is defined as

$$w(x) = v g'(x) = -\frac{\sigma}{2\eta} \sqrt{x(\ell - x)} \quad (51)$$

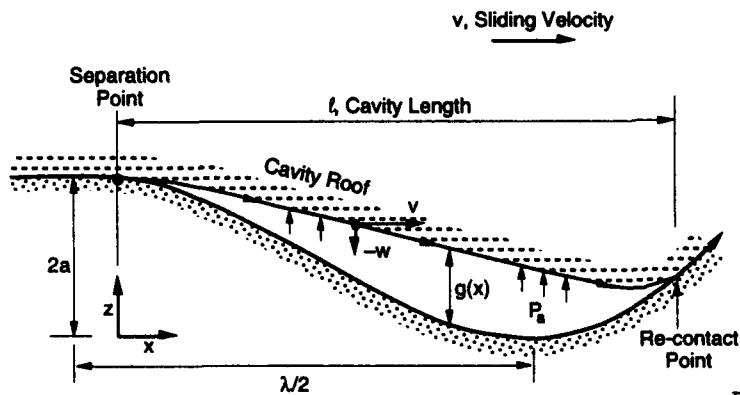


a. Without roof melting.

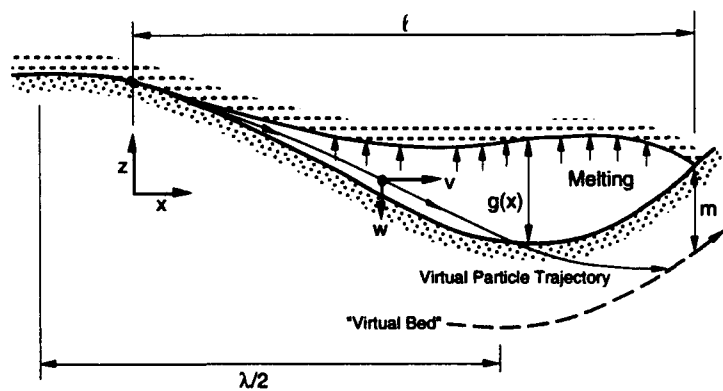


b. With roof melting.

Figure 30. Parameters of the Kamb linked-cavity model for the idealized step cavity (after Kamb 1987). Cavity is filled with water at pressure P_w and ice overburden pressure P_i ; $v(x)$ is ice motion in z direction. Lines with arrows (b) show particle path if melting (m , total meltback) were absent.



a. Without roof melting.



b. With roof melting.

Figure 31. Parameters of the Kamb linked-cavity model for the idealized wave cavity (after Kamb 1987). Lines with arrows (b) show particle path without roof melting.

For a nonlinear ice rheology, η is defined by

$$\eta = \eta_R \left(\frac{\sigma_R}{\sigma} \right)^{n-1} \quad (52)$$

where η_R is the effective viscosity at a reference shear stress level σ_R . The flow law parameter η is assumed to be 3.

For wave cavities the relationships differ in that the bed is assumed to have a quasi-sinusoidal shape, with a wavelength λ and half amplitude a . Kamb assumes that the normal stress varies as $(x - x_0)^2$ and follows a linear ice flow law

$$g(x) = \frac{2\pi^2}{3} \frac{a}{\lambda^4} x^{5/2} (\ell - x)^{3/2} \quad (53)$$

for $0 < x < \ell$ and

$$\ell = \frac{4\lambda}{\pi} \left(\frac{2}{5} \right)^{1/2} \left(1 - \frac{\sigma}{\Sigma} \right)^{1/2}, \quad \sigma \leq \Sigma \quad (54)$$

where, as before, $\sigma = P_1 - P_w$, and where

$$\Sigma = 8\pi^2 \eta a v / \lambda^2 \quad (55)$$

Σ being termed the wave cavitation parameter. This parameter is simply the effective confining pressure for cavitation; when $\sigma < \Sigma$, cavitation occurs. Wave cavities form only where water pressures P_w are sufficiently high such that $P_w > P_1 - \Sigma$. In contrast step cavities exist where P_w is low and the ice sliding rate is relatively slow. The wave cavitation parameter can be further defined in terms of a nonlinear ice flow law, such that

$$\Sigma = 2N \left(\frac{4\pi^2 e^2 a v}{\lambda^2} \right)^{1/3} \quad (56)$$

for

$$\eta = N \dot{\epsilon}^{-1/(1/n)} \quad (57)$$

where $N = \text{constant}$

$$n = 3$$

$\dot{\epsilon}$ = second invariant of the strain rate tensor.

Σ therefore varies with $v^{1/3}$, and sliding velocities are relatively fast for wave cavities to exist. The relationship for $w(x)$ follows that of step cavities with gap height $g'(x)$ (eq 53) and a nonlinear flow law relationship for cavity growth by viscous heat dissipation.

The general relationship

$$g'(x) = w(x) + \dot{m}(x) \quad (58)$$

defines the gap profile evolution beneath moving ice and accounts for progressive widening or narrowing. Substituting eq 48 and 51 for m and w , respectively, we see that the steady-state gap profile of a step cavity is

$$vg'(x) = -\frac{\sigma}{2\eta} \sqrt{x(\ell - x)} + \frac{(\alpha\Lambda/\omega)^{3/2}}{2^{2/3}DM} g^{5/3}. \quad (59)$$

Substituting the value of σ from eq 50 and making the relationship nondimensional using the variables $\gamma = g/h$, $\xi = x/\ell$ and $\mu = m/h$, we define the steady-state gap profile $\gamma(\xi)$ by the nondimensional differential equation

$$\frac{d\gamma}{d\xi} = 2 \Xi \sqrt{(1 + \mu) \gamma^{5/3}} - \frac{8}{\pi} (1 + \mu) \sqrt{\xi(1 - \xi)} \quad (60)$$

where

$$\Xi = \frac{2^{1/3}}{\pi^{1/2}} \frac{(\alpha\Lambda/\omega)^{3/2}}{DM} \left(\frac{\eta}{v\sigma} \right)^{1/2} h^{7/6}. \quad (61)$$

The dimensionless quantity Ξ is termed the melt-stability parameter and provides a measure of the importance of roof melting by viscous dissipation in the linked-cavity system.

Integration of eq 60 for values of $\xi = 0$ to 1 results in a series of curves representing changes in the gap profile under meltback for various values of μ (Fig. 32). As the roof melts, the roof line rises and its peak shifts downstream, with the gap length ℓ increasing by nearly a factor of 3 at a maximum stable value of the parameter $\Xi = 1$ in eq 61 (see also Fig. 7 on p. 9091 in Kamb [1987]).

For any given step cavity profile, water flux Q_w is defined from eq 47 and 48 as

$$Q_w = \frac{2^{4/3}}{\pi^{1/2}} \frac{N_0}{M} \left(\frac{\alpha\Lambda}{\omega} \right)^{1/2} \left(\frac{\eta v}{\sigma} \right)^{1/2} h \frac{13}{6} \Phi \quad (62)$$

where

$$\Phi = \sqrt{1 + \mu} \int_0^1 \gamma^{5/3} d\xi \quad (63)$$

is the flux factor and determined by integration of the cavity profile as a function of the melt-stability parameter Ξ .

For wave cavities, the relationship follows eq 60 and the nondimensional results are similar in form, but $w(x)$ must be defined differently because the points of cavity separation and down-glacier ice-bed contact continually changes as a function of roof melting during sliding (Fig. 31). Kamb uses

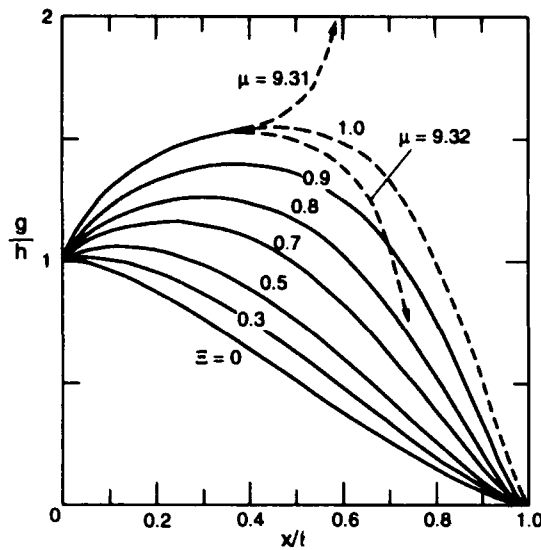


Figure 32. Steady-state configuration of the step orifice roof profile values of the melt stability parameter Ξ from 0 to 1.0. For each value of Ξ , the gap height $g(x)$ is shown in terms of g/h as a function of dimensionless longitudinal coordinate x/l , where l is the gap length. At $\Xi = 1$, the system becomes unstable (after Kamb 1987).

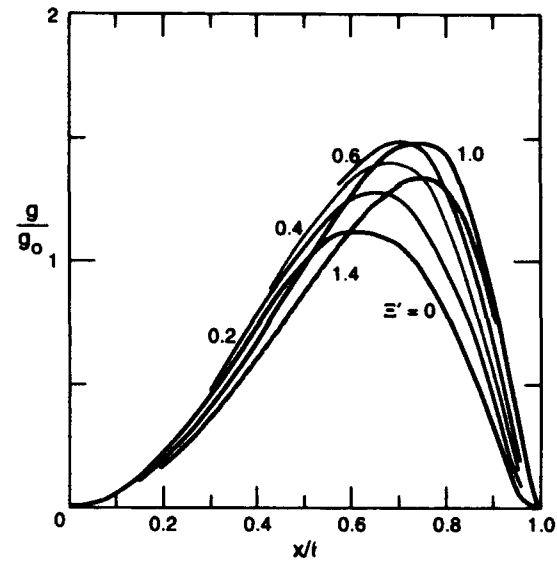


Figure 33. Steady-state configuration of the wave orifice gap for values of the melt stability parameter Ξ' from 0 to 1.4. For each value of Ξ' , the gap height $g(x)$ is shown in terms of the ratio g/g_0 , where g_0 is the height midway along the length of the gap ($x = 1/2$) in the absence of roof melting ($\Xi' = 0$). The longitudinal coordinate x is scaled by the gap length l (after Kamb 1987).

the dimensionless quantities $\xi = x/l$, and $\gamma = g/g_0$ where

$$\frac{g_0}{a} = \left(\frac{128}{75} \right) \left(1 - \frac{\sigma}{\Sigma} \right)^2 \quad (64)$$

and

$$B = \left(\frac{5}{(5 - 2v + v)^2} \right) \quad (65)$$

with v , a meltback parameter, defined by

$$v = \frac{2^4}{\pi^5} \left(\frac{\lambda}{l} \right)^4 \frac{m}{a} \quad (66)$$

to develop a differential equation for the progressive evolution of the wave gap profile. Gap profile changes are defined by

$$\frac{d\gamma}{d\xi} = \Xi (5B)^{1/2} \gamma^{5/3} - (B)^{2/3} \xi^{3/2} (1 - \xi) (8\xi + 3v - 5). \quad (67)$$

The melt-stability parameter Ξ for wave cavities is

$$\Xi' = \frac{2^7}{3^{2/3} 5^{7/3}} \frac{(\alpha \Lambda / \omega)^{3/2}}{DM} \left(\frac{\eta}{v \Sigma} \right)^{1/2} a^{7/6} (1 - \frac{\sigma}{\Sigma})^{11/6}. \quad (68)$$

Integration using $\xi = 0$ to 1, for successive values of Ξ' , defines v (eq 66) and the meltback ratio m/a . The steady-state configuration of wave cavities for various values of Ξ' is shown in Figure 33. In contrast to step cavities, melting of the roofs modifies cavity height greatly, but has little effect on cavity length. Step cavity evolution involves greater elongation of the gap, rather than increased gap height (compare Fig. 32 and 33; see also Fig. 9 on p. 9092 in Kamb [1987]). In addition, changes in configuration with time in response to ice movement result in increased cavity height in the down-ice direction, while lowering of the roof takes place up-ice of its peak height because of the down-ice shift in the cavity separation point (Kamb 1987).

Discharge Q_w through wave cavity forms follows from eq 47 and 64, with m/a defined by eq 66 and integration of eq 67 for various values of Ξ' . Kamb's relationship is

$$Q_w = \frac{2^{5/2}}{75^{5/3} \pi} \frac{N_0}{M} \left(\frac{\alpha \Lambda}{\omega} \right)^{1/2} \lambda a^{5/3} \left(1 - \frac{\sigma}{\Sigma} \right)^{23/6} \Psi \quad (69)$$

where

$$\Psi = B^{1/2} \int_0^1 \lambda^{5/3} d\xi \quad (70)$$

is a "flux factor" analogous to Φ (eq 63).

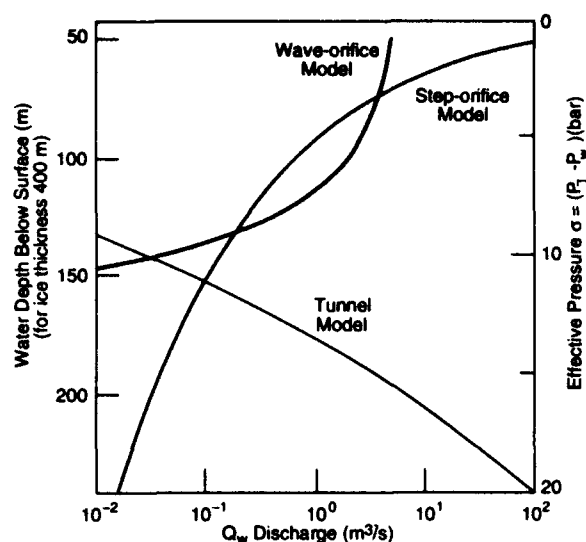


Figure 34. Comparison of the effective confining pressure vs. discharge relationship for the linked-cavity and conduit-tunnel models, assuming an ice thickness of 400 m (after Kamb 1987).

Kamb (1987) provides a quantitative evaluation of his linked-cavity model using Variegated Glacier, a surging glacier in south-central Alaska. Field measurements of various glacier parameters before, during and after a surge in Variegated Glacier are some of the most extensive available. The results of the model, using appropriate values for unmeasurable parameters (given by Kamb [1987]), compare reasonably well with the field data. The direct relationship of water pressure to discharge for both step and wave cavity profiles in the linked-cavity system contrasts with the inverse relationship of pressure and discharge in the conduit-tunnel drainage system (Fig. 34).

In addition, Kamb suggests that the linked-cavity model explains the cause of surging as illustrated by Variegated Glacier. A surging glacier is one that rapidly changes from a quiescent state to one of extremely rapid flow and terminus advance (e.g., Meier and Post 1969). Flow rates can commonly exceed tens of meters per day. Surges apparently take place periodically—over periods of about 15 to 100 years—and may last a year or more. The surge ends equally abruptly and the glacier returns to the quiescent state, during which it recedes. Only certain glaciers surge. Theories have been proposed for the cause of surging, most of which require higher than normal basal water pressures, but they remain mostly unverified (e.g., Clarke 1987). A surging glacier is important to runoff or sediment discharge predictions, as both may increase by several orders of magnitude during a surge over that of normal water and sediment flux when the glacier is not in surge.

In Kamb's model, the stability of the drainage system is determined by orifice roof melting and ice sliding and cavitation rates. Instability results when the magnitude of the viscous heat dissipation, as indicated by the melt stability parameters Ξ or Ξ' , exceeds a certain value (for a step orifice when $\Xi' > 1$, for a wave orifice when $\Xi' > 1.5$), which results in unstable growth of the cavity orifices as effective pressures decrease (Kamb 1987). Kamb suggests that the resultant perturbation manifests itself in an increase in cavity length and progressive growth into a conduit that, with time, is progressively joined along its length by other orifices also growing unstably. The result is evolution to a tunnel system (Fig. 35). As ice flow increases,

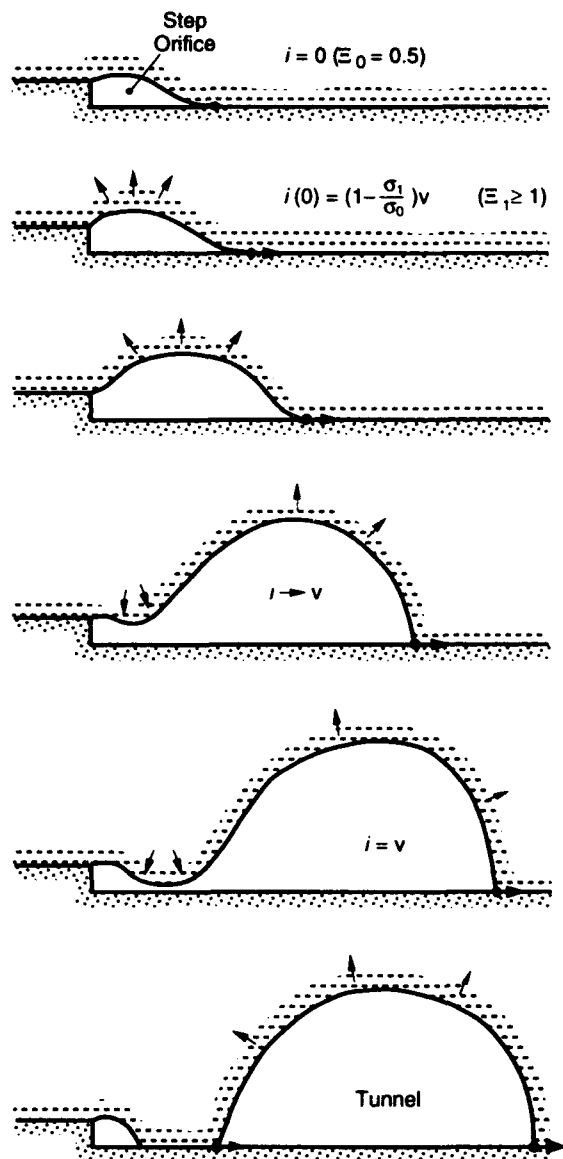


Figure 35. Transition from linked-cavity system to conduit-tunnel system as linked-cavity system undergoes unstable growth during a surge (after Kamb 1987).

the tunnel moves down-ice, Kamb speculates, to a new orientation relative to ice flow direction. Surges stop as basal water pressures drop in the newly formed conduit-tunnel system, as it is better able to handle the equivalent discharge that taxed the linked-cavity system and, thus, sliding speeds are reduced.

Surges could begin in late winter, when discharge is low and a linked-cavity drainage system covers the bed. As meltwater influx increases in spring, Kamb speculates that certain conditions, perhaps related to the increased ice thickness before surging and the subsequent increase in basal shear stress, cause the melt-stability of the orifices to remain low and hence the system does not immediately evolve into a conduit-tunnel drainage system. Basal water pressures could then rise to near those of the ice overburden pressures, reducing shear stress at the bed to near zero and initiating a surge. Unstable growth and linked-cavity system decay lead to its collapse, while initiation of drainage via a conduit-tunnel system then reduces basal water pressures, increases basal shear stresses and reduces sliding velocities below those necessary for the surge to continue.

System behavior. Several important aspects of subglacial drainage in the linked-cavity system have emerged from theoretical analyses. Differences between this and the subglacial conduit-tunnel system of Röthlisberger bear on the nature of runoff (and sediment discharge) from glaciers. As stated previously, water flow in the linked-cavity system follows a rather tortuous path in an anastomosing pattern, with potentially more than one outlet and inlet for each cavity (Fig. 27). This contrasts with a conduit network, wherein the smaller passageways are tributaries to progressively larger ones, and the direction of flow is mainly in the direction of ice flow, both of which are governed by the ice surface slope and water flow is in response to equipotential contours as defined by Shreve (1972).

Water flux or discharge is directly related to water pressures in the steady-state linked-cavity system (Walder 1986, Fowler 1987, Kamb 1987). This direct relationship results from heat being dissipated only by melting of the roofs of the cavities and conduit links, which in turn modifies their geometry and increases flow at a higher pressure. Conversely, a decrease in discharge and thus in the generation of heat by viscous dissipation reduces melting and lowers the roofs, but it reduces their cross-sectional area by a smaller factor than would apply if the cavity walls were

also composed of ice (Walder 1986). The system therefore maintains a certain water flux capability that is proportionally higher than that at a larger discharge and can adjust to a lower steady-state water pressure.

This behavior contrasts with that of the conduit-tunnel system in which water pressure varies inversely with discharge (Röthlisberger 1972). Thus, the larger tunnels with lower water pressures tend to capture flow from smaller conduits with higher pressures, resulting in a system that will tend to evolve continually by conduit capture into a single, larger tunnel of lower pressure as discharge increases (Röthlisberger 1972, Shreve 1972). The mean flow rate through a conduit-tunnel system is also higher than in linked-cavities (Walder 1986). The fully developed system under maximum discharge, therefore, consists of one or more tunnels of low tortuosity and several meters diameter (depending in part upon glacier dimensions). Tunnels trend approximately parallel to ice flow along a large proportion of the glacier's length and apparently are fed by the arborescent englacial conduit system through a series of conduits forming tributaries to the tunnels.

The direct relationship of pressure and discharge also ensures a mostly stable configuration for the interconnected cavities and conduits through the entire year. Thus, the linked-cavity system can persist even with seasonal changes in water discharge, requiring only that sliding rates be sufficient to maintain cavitation (Walder 1986, Kamb 1987, Fowler 1987). This situation is not true for the conduit-tunnel system, which will tend to evolve seasonally from multiple smaller conduits at the bed to fewer large tunnels as meltwater discharge gradually increases through the spring and summer. In the fall, however, a drop in discharge closes the tunnels and the conduit-tunnel system collapses, effectively shutting off water flow across most of the bed.

Theory suggests that the stability or instability of these two systems lies mainly in each system's capacity to adjust to changes in viscous heating. That is, if the size or water pressure of the conduit or tunnel under a fixed hydraulic gradient is increased, conduits will grow at an accelerating rate because the melting rate increases faster than the closure rate caused by plastic ice flow. The converse results in tunnel contraction and closure. There is, thus, no truly stable steady state for the conduit-tunnel system.

The linked-cavity system in a steady state under a fixed hydraulic gradient, however, can ad-

just to finite changes in size or water pressure. Water pressure changes result in increased or decreased ice roof melting and a change in size to accommodate it, while a change in size eventually results in a return to the original steady-state dimensions (Kamb 1987).

A linked-cavity system is thus inherently stable because its minimum flow capacity is determined by cavitation at the existing water pressure, and hence its existence can, in theory, be independent of melting by viscous heat dissipation (Kamb 1987). The system can, however, be unstable if viscous heating increases at a rate faster than the system can adjust for and higher pressures lead to increased basal sliding rates. Then, orifice and cavity growth can result in larger and more direct interconnection of linked-cavities across a large area of the bed, actually forming a continuous conduit or tunnel (Kamb 1987). As stated previously, this unstable evolution to a lower pressure tunnel system may stop surges.

In addition, for a given discharge, water pressures in the linked-cavity system (beyond a certain threshold discharge that depends on system parameters and cavity-conduit configuration) are higher than in a true conduit system at steady state (Kamb et al. 1985, Walder 1986). Seasonal changes in *system parameters, for example, may result in relatively rapid water pressure changes*. The existence of a linked-cavity system may explain increases in ice flow rate that have been measured in association with increases in water pressures (Iken 1981), but typically *not* in association with simultaneous increases in stream discharge (e.g., Kamb et al. 1985, Iken and Bindshadler 1986).

Seasonal variations in subglacial drainage will have important ramifications for flow routing, water storage and runoff. The more or less continuous evolution and degradation that characterize the conduit-tunnel system as meltwater production increases and decreases with the seasons results in morphological changes to the drainage system. In fall, for example, decreased meltwater production causes the system to degenerate and shrink and closes many, if not most, of the conduits and tunnels.

In spring, as meltwater production increases and water influx from the surface begins, there must be temporary water storage because a complete drainage system does not exist, and water pressures therefore increase as the system begins to reestablish itself, in some cases following remnants of the conduit-tunnel system and in other areas developing new passageways. Short-term

increases in sliding rates may result from such transient pressure increases (e.g., Iken and Bindshadler 1986). Increased discharge results in melting of conduit roofs and walls and their expansion. Pressure gradients associated with variations in water influx result in water migration from one area of the bed to another. Areas where flow becomes concentrated will have more rapid conduit growth because of higher melting rates than in other areas and eventually create larger conduits in smaller numbers. As conduits enlarge, water pressures will drop, reducing glacier sliding velocities, and eventually one or more primary tunnels will be established that transport the majority of runoff to the terminus. By peak seasonal discharge, a conduit-tunnel system is fully established, operating at water pressures that are lower than those present during its development. Flow-through of meteoric water becomes rapid, with little lag in meltwater efflux and, overall, meltwater storage is reduced.

The linked-cavity system, however, remains mostly stable as discharge rises or falls, and water pressures and sliding speeds fluctuate, and therefore at least in part will remain open through the fall and winter unless meltwater production stops. Then, connecting conduits (orifices) can degenerate, shrink and close because of ice deformation. Water storage within cavities that remain is likely. Given the limited sources of water that exist in winter, conduits probably close over parts of the bed, but not in others. A linked-cavity system can, therefore, remain at least in part operational throughout the year in areas where groundwater and meltwater from geothermal heating of the glacier sole locally produce sufficient water to fill cavities and maintain hydraulic pressures within connecting conduits.

As discharge increases in spring, concomitant water pressure increases result in the expansion of conduit orifices and cavities as meltwater reaches the bed. Locally, pressures increase as the system adjusts to the increased discharge. Basal sliding rates increase, causing a positive feedback to increase cavitation and the linking of cavities with conduits. Areas without an active linked-cavity system will become reestablished at a rate that depends on water influx, but before this happens, water retention, dispersion across the bed and slow flow-through are characteristic. Once the system is established, its configuration adjusts progressively to discharge and ice flow for the remainder of the melt season. Mainly, the cross-sectional area of conduits and cavities adjusts to

discharge variations. The nature of the linked-cavity system is also such that daily meltwater and precipitation inputs can lag in reaching terminus streams, with flow rates slower than in tunnels and the cavities generally storing meltwater.

The coexistence of the linked-cavity and conduit-tunnel systems in the subglacial environment of a single glacier has been suggested by studies of the former beds of glaciers (e.g., Walder and Hallet 1979, Walder 1986) and has been examined theoretically (e.g., Fowler 1987a, Walder 1986). Limited evidence suggests that an integrated system, where pressures within the linked-cavities and conduit-tunnels are roughly equal, may characterize many temperate glaciers, while shifts between the two systems may be associated with seasonal discharge variations and fast ice flow (e.g., Engelhardt et al. 1978, Röthlisberger and Iken 1981, Iken et al. 1983, Kamb et al. 1985, Iken and Bindshadler 1986, Walder 1986, Fowler 1987a, Kamb 1987, Seaberg et al. 1988, Hooke et al. 1989, Hooke 1989).

Theoretical analyses by Fowler (1987a,b) suggest that each system, while stable separately under steady-state conditions, will coexist in general because a glacier sliding over its bed will develop cavities, and these will become interconnected hydraulically owing to the inherent roughness of the bed. A combined drainage system is in a steady state when pressures in cavities (P_k) and in tunnels (P_T) are equal and not perturbed. A significant increase in P_T will result in water movement into the linked-cavity system, which in turn increases discharge, water pressure and cavity enlargement (Walder 1986). Under this condition ($P_k > P_T$), the combined system is again stable and the pressures will equilibrate to a steady state, with flow feeding from cavities back into conduits.

Because the cavity volume depends on sliding velocity (whereas it does not in the tunnel system), system stability is also related to it. At sufficiently high sliding velocities, the tunnel system breaks down to cavity drainage. The converse, at slow velocities, is also assumed true. As discussed previously, the tunnel system can become unstable because of perturbations in discharge and hence the conduit melting rate; either increases or decreases in discharge could cause its collapse (Kamb 1987). Similarly, linked-cavity systems can become unstable under extreme discharge rises and extreme water pressures, as in the case of flood drainage, or under the extremely slow sliding rates and minimal discharges of winter, when cavities can close or be cut off by orifice closure. In

general, the limit of stability for the conduit-tunnel part of the drainage system appears to be when discharges are low, conduits are shrinking and water pressures are decreasing (Kamb 1987, Fowler 1987a). Conversely, unstable growth of conduit orifices, owing to excess viscous heat dissipation (Kamb 1987) under rapid and extreme discharges, will make the tunnel system dominate (Walder 1986).

Sediment beds

Where the glacier lies on unconsolidated sediments, often called a deformable or soft bed, current theory on subglacial drainage is limited by few data and observations. Because of the difficulties in analyzing these "unknown" conditions, drainage on sediment beds has largely been ignored until the last 10 years or so. Currently, prevailing theory suggests that water may move mainly within a conduit-tunnel system or a distributed-flow system, but the linked-cavity system may also exist on a sediment substrate.

Nature and behavior. Theory suggests that englacial conduits will feed conduits or tunnels that are either partly incised into subglacial sediments and partly incised into the overlying ice, or fully cut into the basal ice with the channel bed being the sediment interface. Hydraulic conditions basically follow Röthlisberger's (1972) theory (Fowler 1987b), except that conduits may close as the result of sediment deformation (Boulton and Hindmarsh 1987). Some part of the drainage system may also be englacial, with conduits moving from a subglacial to an englacial position, such as in areas of intense compressional deformation associated with deep depressions within the bed (Hooke et al. 1988, Hooke 1989). In the general absence of englacial meltwater sources, water supply is sufficiently small such that only a distributed flow system develops, the water moving in a thin film between the ice and bed (Alley 1989a).

In contrast to bedrock beds, unconsolidated beds may be permeable and thus transport some proportion of the subglacial water within the substrate (e.g., Clark et al. 1984, Shoemaker 1986a). In general, the permeabilities of typical subglacial materials are assumed to be such that any influx of water from the surface will rapidly outstrip the sediments' capacity to transport water, and therefore water pressures will increase until a drainage system of some form develops.

If there are no surface sources of meltwater (such as in polar regions, but also during the winter in more temperate regions), water at the bed is

derived mainly from basal melting generated by frictional and geothermal heating, and is minimal in volume. Under such stringent water conditions, drainage will occur as groundwater within the permeable materials below the ice/bed interface, until their capacity is exceeded. Then, if hydraulic pressures equal or exceed the local ice pressure, water can accumulate between the ice and sediment (Alley 1989a). Water pressures will be relatively high and near that of the overburden pressure (within about 1×10^{-5} Pa [1 bar]) of the ice. A significant groundwater source at the bed could, however, maintain a water layer between the ice and the bed.

If the water pressure where groundwater influx is limited increases sufficiently, areas of water accumulation will thicken and expand laterally, communicating hydraulically, and result in a more uniform distribution of water pressure across the bed (Lliboutry 1987). Continued expansion and communication leads to water flow within a thin layer, which may be characterized by local thickenings that depend on variations in the effective pressure at the bed, but still not exceeding a few millimeters (Walder 1982). Flow within locally thick water "films" may act as a channel, or cycle between a film and channel, with the configuration depending on water supply, bed geometry, bed properties and sliding velocity (Alley 1989a).

It is not likely, however, that a thin film drainage system will evolve into fully channelized flow, unless input from meltwater sources significantly increases and there is sufficient supply to maintain flow continuously in the new conduits (Weertman and Birchfield 1983). Such a transition might characterize seasonal changes in the drainage system of temperate glaciers when meltwater input increases or decreases, but such a transition will probably only occur in areas removed from other englacially fed conduits (Weertman and Birchfield 1983).

Theoretical treatment. In a distributed flow system, water drainage is governed by the effective pressure associated with the roughness of the bed, which is related to the grain size of the bed material (typically of wide range, from clay to boulder size) and which determines cavitation at the ice/bed interface. This relationship is expressed as (Alley [1989a] following Nye [1969] and Kamb [1970])

$$N = \frac{\beta \tau_b}{f} \quad (71)$$

where $N = P_i - P_w$ (Weertman 1972)

- P_i = average normal ice stress on the bed
- P_w = water pressure
- τ_b = basal shear stress
- β = geometric factor related to stress on the bed roughness feature
- f = fractional area of bed occupied by the water film, where $f = 1 - s$
- s = fractional area of ice in contact with clasts in bed.

Water flow within a film of thickness d is described by Weertman (1972) and Weertman and Birchfield (1982) as

$$d = \left(\frac{12\mu q}{P_g} \right)^{1/3} \quad (72)$$

- where q = water flux
- P_g = pressure gradient
- μ = viscosity of water.

In areas removed from conduits, P_g is calculated as (Alley 1989a)

$$P_g = \rho_i g \alpha_s + (\rho_w - \rho_i) g \alpha_b \quad (73)$$

- where ρ_i = density of ice
- ρ_w = density of water
- g = acceleration of gravity
- α_s = surface slope
- α_b = bed slope.

Thus, given appropriate parameters, calculation of the expected discharge of water in a distributed flow system on an unconsolidated bed can be estimated. The relationship among basal shear stress, drainage and sliding velocity u_s is given by Alley (1989a), based upon Weertman's (1957, 1964) sliding theory, as

$$u_s = K_1 \tau_b^2 \left(1 + 10 \frac{d}{d_c} \right) \quad (74)$$

or

$$u_s \equiv K_1 \tau_b^2 d \quad (75)$$

where the approximation is valid if $d \geq O(d_c)$ (O indicates on the order of), d_c being the controlling obstacle height. The controlling obstacle height is equal to the height that results in a maximum resistance to sliding. d_c is typically on the order of 1 to 10 mm. K_1 and K_s are constants that depend upon bed roughness and related factors (Weertman 1964).

The situation where meltwater is more plentiful and surficially fed englacial conduits reach an unconsolidated bed requires that water be transmitted to outlet streams within a conduit system. Increased water pressures associated with the increase in discharge to the bed apparently result in conduit establishment, with the influx of water Q_x ($\text{m}^2 \text{s}^{-1}$) dependent upon the rate of water transport to the bed and the area of the bed that supplies water to the newly established conduit (Shoemaker 1986a).

Boulton and Hindmarsh (1987) considered drainage in conduits with sediment floors, the upper part of the conduit either being incised into the basal ice (R-channels) or fully incised into the bed (N-channels). They assume that water pressure in the conduit offsets the effects of ice pressure, which could cause both the ice and sediment to flow into the conduit. Thus, as with englacial conduits (following Röthlisberger 1972), conduit closure results from the difference between the overburden pressure P and water pressure P_w . Deformation of sediment under stress is given by (Boulton and Hindmarsh [1987], following Alley's [1989a] notation)

$$\dot{\epsilon} = K_b \frac{(\tau - \tau^*)^a}{N^b}, \quad \tau > \tau^* \quad (76)$$

and

$$\dot{\epsilon} = 0, \tau \leq \tau^*$$

where $\dot{\epsilon}$ = strain rate
 τ = shear stress
 τ^* = sediment strength
 a, b and K_b = constants.

The effective stress N is the difference between the overburden normal stress and the water pressure P_w . At the base of the ice, where the overburden normal stress on the bed is the ice normal stress P_i

$$N = P_i - P_w \quad (77)$$

Deformation results when the yield strength of the subglacial sediment τ^* is exceeded

$$\tau^* = N \tan \phi + C \quad (78)$$

where $\tan \phi$ is internal friction and C is cohesion.

Creep closes a conduit incised into sediment in response to the effective stress N . For a conduit of circular cross section of radius r , and assuming

that the shear stress τ is equal to the effective pressure N , tunnels will close (ξ_c) at a rate $\dot{r}$ following (Alley 1989a)

$$\begin{aligned} \xi_c &= K_b \frac{(N - \tau^*)^a}{N^b a^a}, \quad N > \tau^* \\ &= 0, \quad N \leq \tau^* \\ &\equiv \frac{\dot{r}}{r} \end{aligned} \quad (79)$$

For a channel incised into the bed, the closure rate ξ_c must be balanced by removal or erosion of the sediment deforming into the conduit x_e (Alley 1989a). By approximating the amount of sediment that can be eroded and thus that amount that can be transported by water flowing through the conduit for a given water flux Q_x ($\text{m}^3 \text{s}^{-1}$), estimates of the range in conditions necessary for conduit stability can be defined. Thus, the erosion rates for laminar ξ_{el} and turbulent ξ_{et} flow are

$$\xi_{el} = \frac{J_0 P_g^2 r^2 Q_x}{64 \pi \mu^2} \quad (80)$$

$$\xi_{et} = \frac{3 J_0 M^2 P_g Q_x}{4 \pi r^{2/3}} \quad (81)$$

where J_0 = constant related to sediment flux

P_g = water pressure gradient

μ = water viscosity

M = inverse of the Manning roughness coefficient.

These equations thus define the maximum sediment input from the bed adjacent to the conduit, and a steady state occurs when the closure rate $\xi_c = \xi_{er}$, the erosion rate.

Water influx Q_x depends on the rate of water supply at the bed and the area from which it can be drawn to the conduit. Alley (1989a) estimated Q_x using

$$Q_x = \frac{2 \pi K N}{\rho_w g \ln \left(\frac{2N}{r_w g \tau} \right)} \quad (82)$$

where K is the hydraulic conductivity of the till and g is the magnitude of gravitational acceleration. By limiting the area of water input from the bed

$$Q_x = 10^4 r \dot{m} \quad (83)$$

where $\dot{m}$ is the basal melt rate. This relationship assumes that a channel collects all water generated by melting in a band 10^4 times as wide as its radius, a value that Alley considered to be an upper limit.

The relationships for the effective pressures under laminar (N_l) and turbulent flow (N_t) are (Alley [1989a] following Weertman [1972])

$$N_l = \left(\frac{P_g^2 r^2}{16B\mu L} \right)^{1/3} \quad (84)$$

$$N_t = \left(\frac{MP_g^{3/2} r^{2/3}}{2LB} \right)^{1/3} \quad (85)$$

where B is a constant related to the creep hardness of ice, L is the latent heat of fusion of ice and ice is assumed to obey a power-law for creep with an exponent of 3 (Weertman 1972).

The difference between the creep-closure rate and the melt rate of an R-channel (ξ)_{net} for laminar and turbulent flows within the conduit is then defined by (Weertman 1972)

$$(\xi_t)_{\text{net}} = B N^3 - \frac{g}{16\mu L} \quad (86)$$

$$(\xi_l)_{\text{net}} = B N^3 - \frac{MP_g^{3/2} r^{2/3}}{2L} \quad (87)$$

Alley (1989a) used estimates for the various parameters in eq 82–87 to define the nature of subglacial drainage in channels of 0.001- to 10-m radii in a bed composed of diamicton (a clay- to gravel-sized mix). He concluded that R-channels would be generally unstable and close rapidly by the creep of subglacial sediment into the conduit. Only R-channels with floors of extremely high yield strengths that actually approach the yield strength of a bedrock bed can remain stable.

In the case of channels incised into the diamicton ("till channels" per Alley), stable configurations require that a low driving stress (N) exists, which is just slightly higher than the critical stress (N_c), as governed by the bed material's yield strength

$$N_c = \frac{C}{1 - \tan \phi} \quad (88)$$

This critical, but slight, difference in stresses reflects erosion of the channel beds by water flow as being limited by discharge. Given a perturbation in drainage, such as a sharp, rapid increase in discharge, "till channels" are likely to grow (Alley 1989a). Variations in discharge, perhaps as occur

seasonally or during flooding, could cause periods of expansion and contraction. In all cases, unless there is sufficient water flux, creep of the sediments will close subglacial conduits incised into sediment.

Channel formation (N-type) within bed sediments, initially in intimate contact with the ice, may begin by the buildup of pore water pressures within the sediment and the reduction of its effective shear stress to zero. This sediment would be unstable and could fail, possibly by liquefaction (Clark et al. 1984, Shoemaker 1986a, Boulton and Hindmarsh 1987). Under lower confining pressures, such as beneath the thin ice of a glacier's terminus, piping of these liquified materials may develop channels within the bed. Once developed, the draw-down of water from the sediments adjacent to them would increase the effective stress of these adjacent sediments and increase their stability. If discharge continues to increase, progressive up-glacier growth of conduits would increase the areas of the bed supplying groundwater to them (Boulton and Hindmarsh 1987). In addition, Shoemaker (1986a) suggested that subglacial channels in bed sediments may develop wherever surface meltwater flowing into moulins or crevasses reaches the bed at new locations, including each spring as englacial conduits reestablish themselves.

Kamb (1987) also briefly suggested that a linked-cavity conduit system may form over a bed consisting largely of unconsolidated rock debris. The bed must have a sufficient number of roughness features that do not move with the ice, such as bedrock protuberances or large boulders protruding from the underlying sediments. The local conditions for cavitation, and the shape of the cavities and orifices that form, are not fundamentally different from those over bedrock. Consequently, Kamb (1987) expected that a linked-cavity system within sediments below ice moving at a sufficient rate for cavitation would have a flux-versus-pressure relationship that is qualitatively similar to that of a bedrock substrate. A linked-cavity system would develop when the water flux in spring or summer exceeds that which can be transmitted by the subglacial sediment (excluding very coarse gravel) (Iken and Bindshadler 1986, Shoemaker 1986a).

It is important to reiterate that our knowledge of glacier dynamics and subglacial drainage on beds composed of sediment is extremely limited. Therefore, the role of sediment beds in water transmission, storage and temporary retention or lags in discharge over days, seasons and longer periods is basically unknown.

DISCHARGE CHARACTERISTICS

Glaciers behave as natural storage reservoirs, retaining a major portion of their winter precipitation in their accumulation areas and releasing large quantities of water annually, with strong seasonal and diurnal peaks related to solar energy input and precipitation (Meier and Tangborn 1961). The characteristics of runoff from these "storage reservoirs" reflect the effects of glaciohydrologic processes on meltwater production, storage and movement, and of climate in its roles as the source of precipitation and control on glacier mass balance.

Most meltwater runs off over only a few weeks in Arctic areas, but it may happen over several months in lower latitude alpine areas such as in the western U.S., Canada and Europe. In the upper Rhone basin of the Swiss Alps, for example, 85% of the total annual runoff occurs between 1 May and 30 September, with this total varying by up to $\pm 20\%$ (Collins 1985). In the Ferpèclé basin, which has a 65% glacier cover, 5% of the runoff occurs from October to April, while the remaining 95% occurs from May to September (Bezingé 1981). Higher elevation basins in the Swiss Alps have the majority of runoff between mid-June and September 30; above 2400 m, for example, only 1% of annual runoff is generated during winter months (Bezingé 1981).

Annual variations in runoff from glacierized basins are not related to annual variations in precipitation (e.g., Meier and Tangborn 1961, Meier 1969, Stenborg 1970, Tangborn et al. 1975, Larson 1978, Bezingé 1981, Fountain and Tangborn 1985a, Collins 1982a, 1985). In fact, annual variations may be almost independent of recent precipitation (Bowling and Trabant 1986). Kasser (1981) and Collins (1987), for example, found that runoff from ice-covered basins actually correlated negatively with precipitation, but positively with temperature. The correlation between precipitation and runoff, however, increases as the proportion of glacierized area in the basin decreases (Bezingé 1981, Chen and Ohmura 1990a). Bezingé (1981), for example, presented data from three adjacent basins showing that, in warm and dry summers, the nonglacierized basin had very low runoff compared to its annual average, whereas in the heavily glacierized basin (67% cover), the volume was well above average. In contrast, heavy precipitation during a cool summer yields the opposite pattern, with well-above average annual runoff in the nonglacial basin and well-below average runoff in the glacierized basin.

The release of water held in glacier storage,

therefore, has a moderating effect on annual stream flows (Krimmel and Tangborn 1974). Each year a portion of the precipitation received across the basin is added to storage as ice in the glacier and a quantity, which may be equal to, more, or less than the total amount of precipitation stored, may be released as runoff by melting. During warm, dry and sunny summers, release of water from storage by melting of ice compensates for the reduced contribution of runoff from precipitation, whereas more runoff is contributed by rainfall when cloudy conditions reduce melting of ice and cooler and wetter conditions prevail (Collins 1986). Both the snowpack in the accumulation area and the glacier itself retain liquid water for some period as well (e.g., Golubev 1973, Colbeck 1977, Oerter and Moser 1982, Östling and Hooke 1986, Fountain 1989). A lag in runoff results from the combined effects of storage and the complex drainage system that exists in both the snowpack and glacier (Meier and Tangborn 1961, Stenborg 1969, Tangborn et al. 1975, Lang 1968).

This moderating effect of storage, however, differs with the percentage of glacier cover in the basin, but can result from as little as 0.5% ice cover (Krimmel and Tangborn 1974). The annual variability for runoff in nonglacierized basins, expressed as the coefficient of variation (CV), is approximately the same as the CV for precipitation in the region; increased ice cover reduces the value of the CV. In the Swiss Alps, for example, the Rhone River basin (16.2% glacier cover) has a precipitation CV of 33 to 41% in July, whereas the CV for July runoff was only 5% during the same period of 1927 to 1947 (Röthlisberger and Lang 1987).

Glacier covers ranging from about 30 to 50% apparently have the minimum CV. The absolute value of the CV, however, is not consistent from region to region. Fountain and Tangborn (1985), for example, concluded that the minimum variability occurred when glacierized catchments were covered by about 36% ice (Fig. 36). Kasser (1959) observed that annual runoff in the Alps varied minimally with 30 to 40% glacier cover. Chacho (1986) reported a less consistent relationship for Alaskan catchments, with minimum variability for ice coverage ranging from 25 to 60%. More recently, Röthlisberger and Lang (1987) concluded that the minimum CV during summer in the Alps also occurred with a 30 to 60% ice cover (Fig. 37), while Chen and Ohmura (1990a) concluded that the minimum ranged from 39 to 44% cover (Fig. 38).

Collins (1985) concluded that if coverage ex-

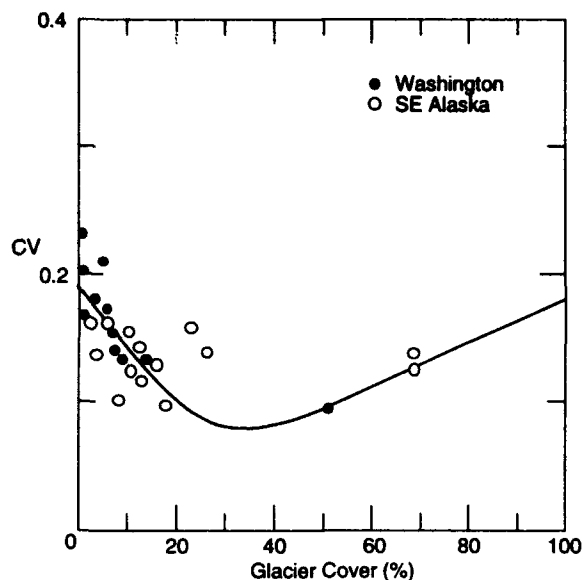


Figure 36. Coefficients of variation (CV) for glacierized basins in northwestern U.S. and southeast Alaska. The estimated minimum is about 36% glacier cover; nonglacierized basins have CV values of around 0.2, whereas the CV is 0.08 at 36% cover (after Fountain and Tangborn 1985a).

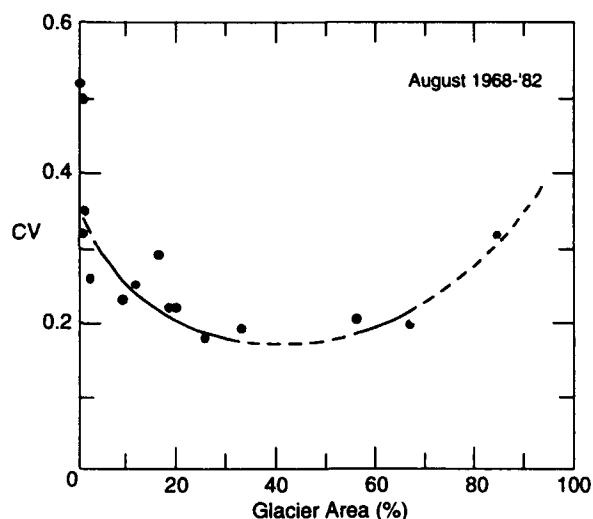


Figure 37. Coefficients of variation (CV) for runoff in August in relation to percentage glacier cover for 14 glacierized basins in the Swiss and Austrian Alps. River basins without a glacier cover have a CV of about 0.3 to 0.5, similar to precipitation for this region. Minimum values (less than 0.2) occur for a glacier cover between 30 and 60% (after R  thlisberger and Lang 1987). Dashed lines indicate apparent trend.

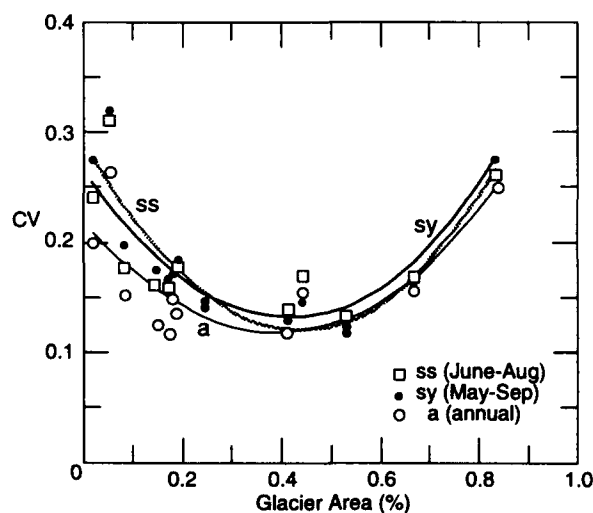


Figure 38. Coefficients of variation (CV) calculated for runoff over various periods (after Chen and Ohmura 1990a). Minimum variability is at 39 to 44% ice cover.

ceeds 50% by area, the CV remains relatively constant or increases. An increase in CV however may reflect the increased importance of glacier controls on runoff from precipitation (lags caused by routing and storage) as the unglacierized area decreases (Collins and Taylor 1990). Runoff variability is also, in general, a factor of local and regional climate, basin topography and range in altitude. Distance of the hydrologic station from the ice source, which is used for CV calculation, will affect the apparent magnitude or intensity of glaciers on runoff variability.

Additionally, monthly variations in runoff characterize glacierized basins, with a minimum CV at the height of the summer ablation season. There is a maximum CV during the early spring melt season, mainly because of the large differences in discharge that can characterize the initial melting and runoff of the previous winter's snowpack (Röthlisberger and Lang 1987).

Because water movement and storage are complex, reflecting the seasonal evolution of the snowpack's properties and the glacier's drainage network, the rate and quantity of meltwater runoff vary at the glacier's terminus through the melt season (Lang 1968, 1988; Stenborg 1969; Colbeck 1977; Collins 1977; 1982a,b, 1988; Fountain 1989). The result is that annual runoff in heavily glacierized basins more closely follows energy input than do the monthly or weekly discharges (Collins 1986).

The spatial distribution of snow cover during the year determines the proportion of basin area in which glacier ice is exposed to melting; thus, this area of meltwater production must be factored with energy input. Annual runoff may be reduced by a high accumulation of winter snowfall because it delays the rise in the transient snowline, thereby limiting the extent and duration of exposure of the underlying ice to melting (Bezing 1981, Collins 1987). If there is sufficient melt because of favorable weather, however, increased snow-melt runoff may compensate for loss of ice-melt (Krimmel and Tangborn 1974).

Similarly, snowfalls during the ablation season can rapidly and significantly reduce runoff by interrupting ice melt over periods of several days or longer (Röthlisberger and Lang 1987) by reducing glacier surface albedo (Hoinkes and Rudolph 1962, Tronov 1962). Lang (1966) estimated a July runoff reduction of 27% and annual reduction of 7% because of the effects of a heavy July snowfall on the Hintereisferner glacier. The timing of early spring snowfalls has a pronounced effect by delaying the rise in the transient snowline, maintain-

ing areas with a high albedo and lessening melt rates (Young 1990).

Although the annual variability of runoff in glacierized basins is less than that in nonglacierized basins, larger, sometimes erratic, daily and diurnal fluctuations, as well as those on a time scale of minutes to days, are common in glacial rivers.

Rapid fluctuations in discharge that can include catastrophic outburst floods result mainly from the release of water stored within and below the glacier (e.g., Collins 1979a, b, 1982b, 1988; Haeberli 1983; Iken et al. 1983; Oerter and Reinwarth 1988; Driedger and Fountain 1989). Extreme floods commonly appear when meltwater flow is amplified by heavy rainfall (Fig. 39) (Röthlisberger and Lang 1987). A combination of high summer precipitation during a maximum melt period, for example, resulted in flooding in 1984 in the Tanana River basin, Alaska (Bowling and Trabant 1986).

Extreme discharge events—termed *jökulhlups*—result from the catastrophic release of glacier-dammed lakes (Fig. 40) or internal water bodies; their timing and frequency are generally unpredictable (Fig. 41). Clague and Mathews (1973) developed an empirical function to predict peak discharge ($Q_{\max} = 75 [V_{\max}]^{0.67}$, where $V_{\max}$ is total volume drained). Clarke (1982) derived a series of equations to estimate $Q_{\max}$ using additional factors, including lake and glacier geometry, and

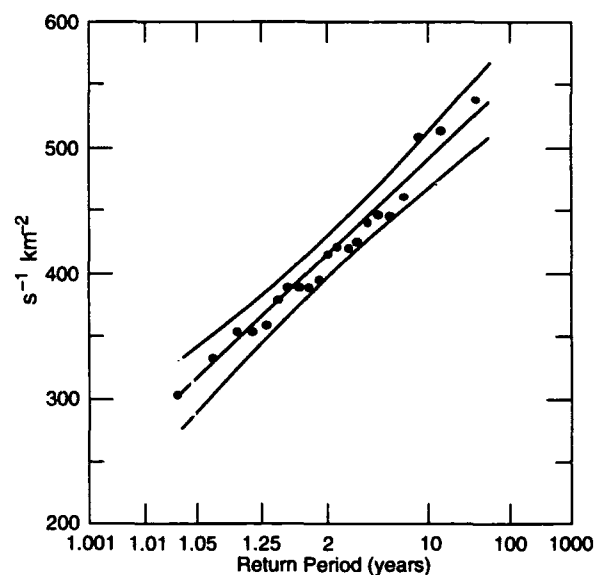


Figure 39. Maximum annual instantaneous discharges from 1965 to 1985 for river basin Massa/Blatten (195 km²) that contains the Aletschgletscher, in Switzerland. Annual maximum values were caused by meltwater flow, not precipitation alone, but may be amplified by rain storms (after Röthlisberger and Lang 1987).

Figure 40. Discharge resulting from sudden drainage of the Gornersee, an ice-dammed lake on the margin of the Gornergletscher, Switzerland. Total discharge (shaded) was approximately $6.2 \times 10^6 \text{ m}^3$ (after Bezing 1981). The bottom line shows theoretically calculated meltwater discharge without an ice-dammed lake contribution.

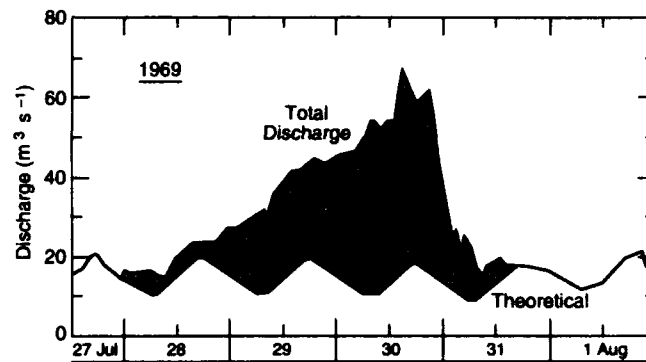
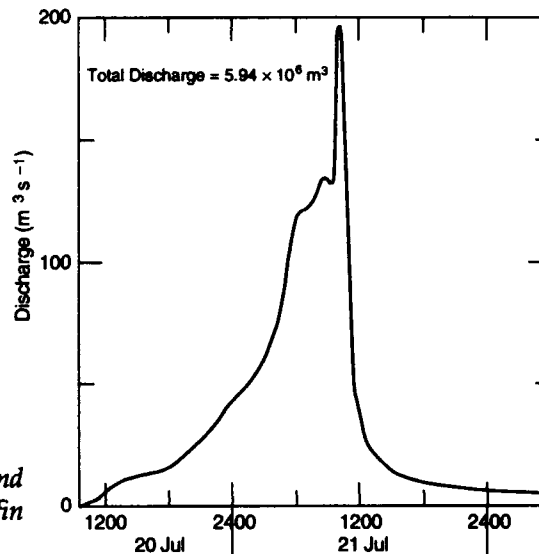


Figure 41. Jökulhlaup hydrograph for ice dam burst and lake drainage on South River, Ekalugad fjord, Baffin Island, July 1967 (after Church and Gilbert 1975).

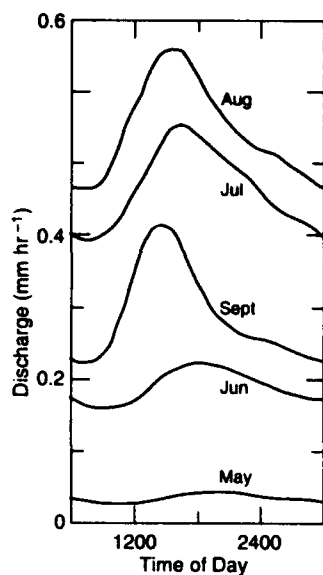


water temperature. These predictive equations are presented in a later section. Chapman (1986) also discussed a method to predict the resultant hydrograph at an Alaskan site once a jökulhlaup has begun.

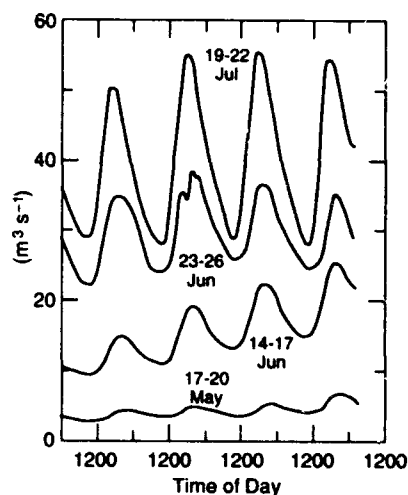
The effects of storm precipitation on runoff vary with the ablation season (Stenborg 1969, 1970, Bezing 1981). Early in the season, the snowpack is more extensive and can store rainfall, thereby reducing the runoff volume and slowing the response at the terminus. Much of this water must also flow on the ice surface in the ablation area, as the internal drainage system is not sufficiently developed to rapidly transmit water. Late in the season, when the overall potential storage of water is also reduced, rainwater can discharge directly through a well-developed drainage system of conduits. Peak flows from precipitation can happen rapidly, often within several hours. An extreme flood from rainfall took place in the Hintereisferne region, Austrian Alps, during the latter part of the ablation season, apparently in large part because the glacier's drainage system was fully developed and flow-through was rapid (Rudolph 1962).

Discharge can exhibit characteristic variations in diurnal peaks (e.g., Gudmundsson 1970, Gudmundsson and Sigbjarnarson 1972, Elliston 1973, Bezing 1981). Diurnal variations reflect primarily total energy input (e.g., Meier and Tangborn 1961). Their timing and magnitude therefore change with the time of the melt-season, with the ratio of daily maximum to minimum flows generally increasing through the ablation season (Fig. 42). The timing of diurnal peaks shifts with the season, occurring 2 to 3 hours earlier in the day near the end of the ablation season than early in it (Fig. 43). These progressive shifts reflect the early season effects of an extensive snow cover on albedo and meltwater retention, the progressive seasonal increase in altitude of the transient snowline, and the development and enlargement of the englacial and subglacial drainage systems through the melt season (Bezing 1981, Elliston 1973, Collins 1988).

The seasonal discharge from glacierized basins peaks 1 to 2 months later in the season than in adjacent nonglacierized basins (Fig. 44). This difference also reflects water storage in the snowpack and glacier, drainage system development and



a. Mean variations in runoff for 1974-1980 at Vernagtbach, Austria, with an 84% glacier cover. Progressive development of drainage system results in changes to timing of maximum flow and gradients in rising and falling limbs of the hydrograph (after Röthlisberger and Lang 1987).



b. Daily variations in discharge for Matter-Vispa river basin, Switzerland, over 4-day periods from beginning to end of melt season (after Elliston 1973).

Figure 42. Diurnal fluctuations in discharge.

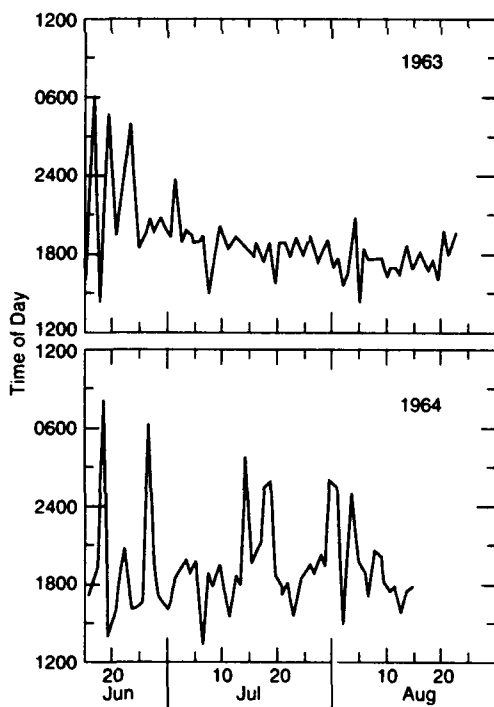


Figure 43. Seasonal shift in the diurnal peak in flow for Lewis River, Canada, as well illustrated for the year 1963. In 1964, storms were common and altered peak timing (after Church and Gilbert 1975).

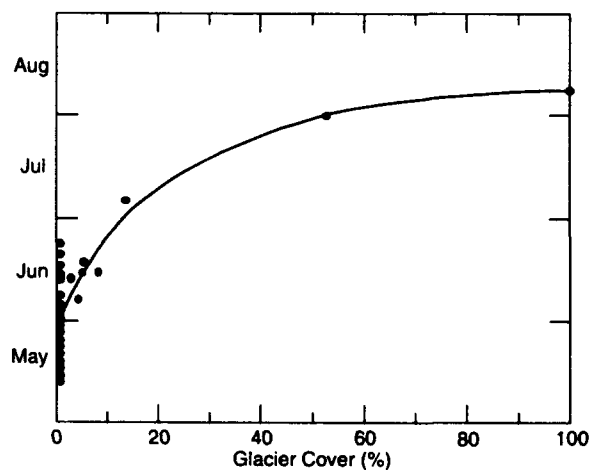


Figure 44. Timing of peak runoff as a function of percentage glacier cover for basins in the North Cascades, Washington (after Fountain and Tangborn 1985a).

Figure 45. Delay in runoff resulting from water storage within the glacier and its snowpack early in the melt season. Estimated potential runoff was computed by a recession-regression analysis of temperature and runoff patterns for late summer (after Stenborg 1970, and Church and Gilbert 1975).

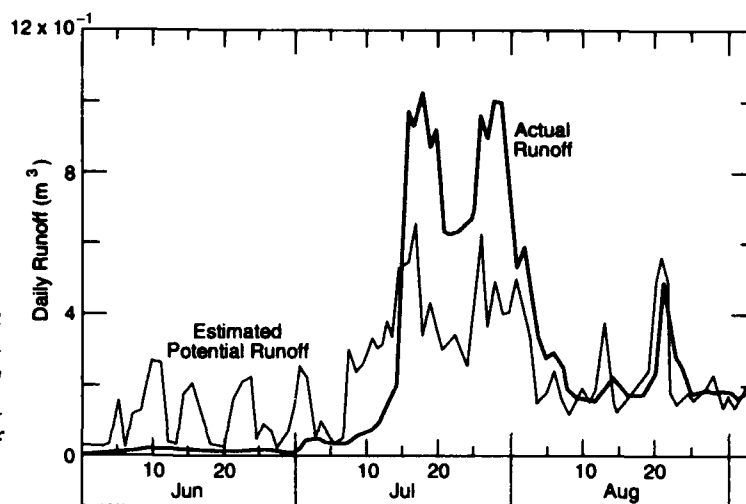
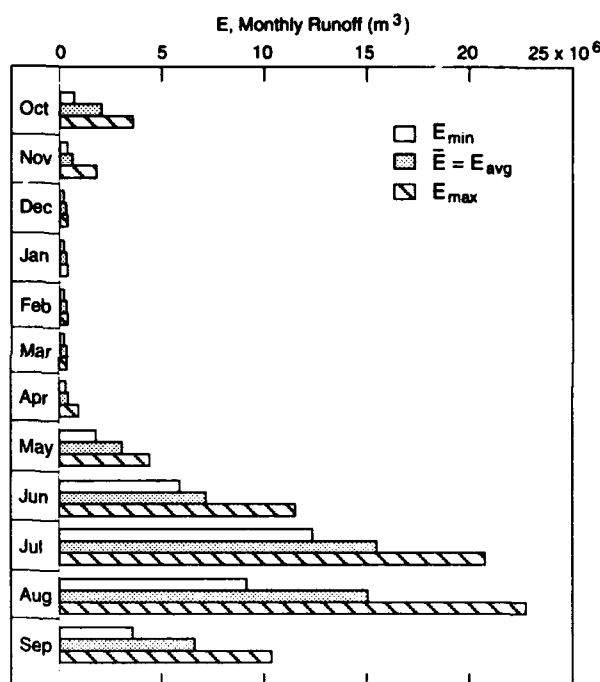


Figure 46. Runoff E from the Ferpeclé basin for the period 1966 to 1977 (after Bezing 1981). Ferpeclé basin covers 36 km², with 64% glacier cover and maximum altitude of about 3700 m. Runoff peaks annually in late July or August as is characteristic of the Swiss Alps.



energy input. Water occurs within the snowpack of temperate glaciers as an aquifer, the water table lying at a depth of typically from 5 to 40 m (e.g., Schommer 1977, Oerter and Moser 1982). The depth generally increases with distance up-glacier and varies seasonally (Lang et al. 1977, Schommer 1977, 1978), reflecting its storage role. In particular, the beginning of the runoff season is later than the beginning of the melt season because of the thermal and hydraulic retention properties of this meltwater within the snowpack (Fig. 45) (e.g., Stenborg 1970).

Within glacierized basins, maximum runoff falls later and later in the summer as the percentage of glacier cover in the basin increases. This delay results from the progressive seasonal increase in snow-melt and ice-melt in conjunction with the temporary internal storage of meltwater that takes place early in the melt season (Tangborn et al. 1975, Rasmussen and Tangborn 1976, Fountain and Tangborn 1985a). In the northwestern U.S., Fountain and Tangborn (1985a) found the maximum percentage of flow from glacier-derived discharges in late summer (July–August), when basin



Figure 47. Aufeis (frozen from water seeping from the base of the glacier) covers the stream channel next to the Matanuska Glacier margin each winter. Minimum discharge takes place in mid-March to late March and consists principally of groundwater.

albedos are lowest, snow and ice melt are greatest, and cloud cover and precipitation are minimal (Meier 1969). Peak flows are delayed by nearly a month over nearby nonglacierized basins, with a maximum runoff typically being in May in the unglacierized basins of this region.

In the Swiss Alps, summer runoff of meltwater from glacierized basins peaks in late July and early August (Fig. 46), responding to the progressive development of the internal and subglacial drainage system through June and July (e.g., Lang 1968; Elliston 1973; Collins 1977, 1988; Iken et al. 1983). Runoff from adjacent nonglacierized basins commonly peaks in June (e.g., Bezing 1981).

Minimum runoff in glacierized basins occurs during winter, with discharge in some heavily glacierized basins decreasing to near zero (Fig. 47) (e.g., Stenborg 1965). Runoff is generally characterized by an exponential decline after the ablation season ends (Röthlisberger and Lang 1987). In the Ferpeclé basin, for example, Bezing (1981) reported from 1 to 5% of total annual runoff in winter, the lower percentage characteristic of higher elevations (Fig. 48). Monthly runoff varies little from its mean value between December and March.

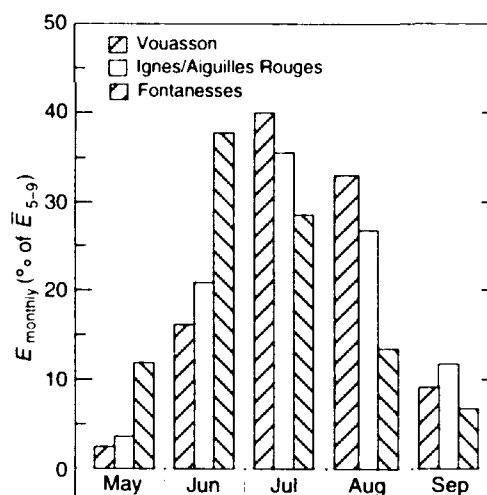


Figure 48. Percentage of monthly flow expressed as a percentage of total ablation season runoff. Average monthly flow indices for three adjacent basins with different percentages of glacier cover in the French Alps. Fontanesses—0% cover; Ignes/Aiguilles Rouges—27%; Vouasson—61% (after Bezing 1981).

Part 2. Sediment Processes, Controls and Characteristics

INTRODUCTION

Sediment transport and sediment yield in glacierized basins, particularly those with a high percentage of area covered by glacier ice, are complex functions of the internal or glaciohydraulic processes that erode, entrain, transport and deliver sediments into glacial discharges, and of climatic conditions that determine the rate, magnitude and timing of meltwater production and meteoric water input. Climate determines thermal conditions within and below the ice, conditions that affect glaciologic processes of sediment entrainment and release. Sediment discharge is therefore also linked to the nature of the englacial and subglacial drainage system and its seasonal evolution and stability.

GLACIOHYDRAULIC PROCESSES AND FACTORS

The sediment yield of a partly glacierized basin consists of the sediment in transport by the glacier, that in transport by meltwater within the englacial-subglacial drainage system and that of tributary streams feeding the main glacial river. Sediment sources include the sediments on, within and beneath the glacier. They also include new sediments created subglacially by ice and meltwater erosion of bedrock. Streams also capture surficial materials of the glacier-free areas, particularly the unstable, unvegetated sediments of the terminal moraine. Sediments of nonglacial origin are transported onto the glacier and into glacial rivers by tributary channels and by various erosion processes, including overland flow and mass wasting.

Ice entrainment and transport

The movement of sediment through the glacier and the processes of entrainment and release are

important to defining the flux of material to the terminus and glacially fed rivers. Sediments in transport within or on a glacier are referred to as debris and include supraglacial, englacial and basal debris (Fig. 49) (e.g., Dreimanis 1988). The location of sediment within the glacier is determined by the processes of incorporation, which include

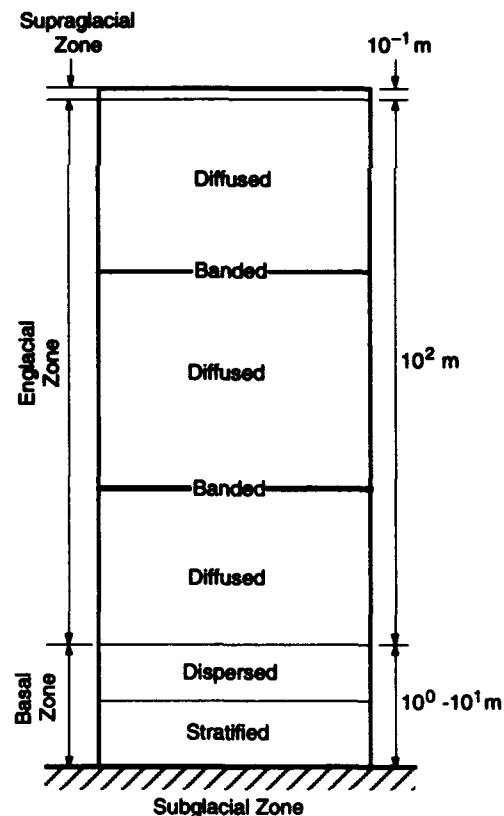


Figure 49. Ice facies and associated debris zones in a vertical ice section (after Lawson 1979a). Within the stratified facies, three subfacies are recognized—the suspended, solid and dispersed—each of which has different debris dispersal characteristics.

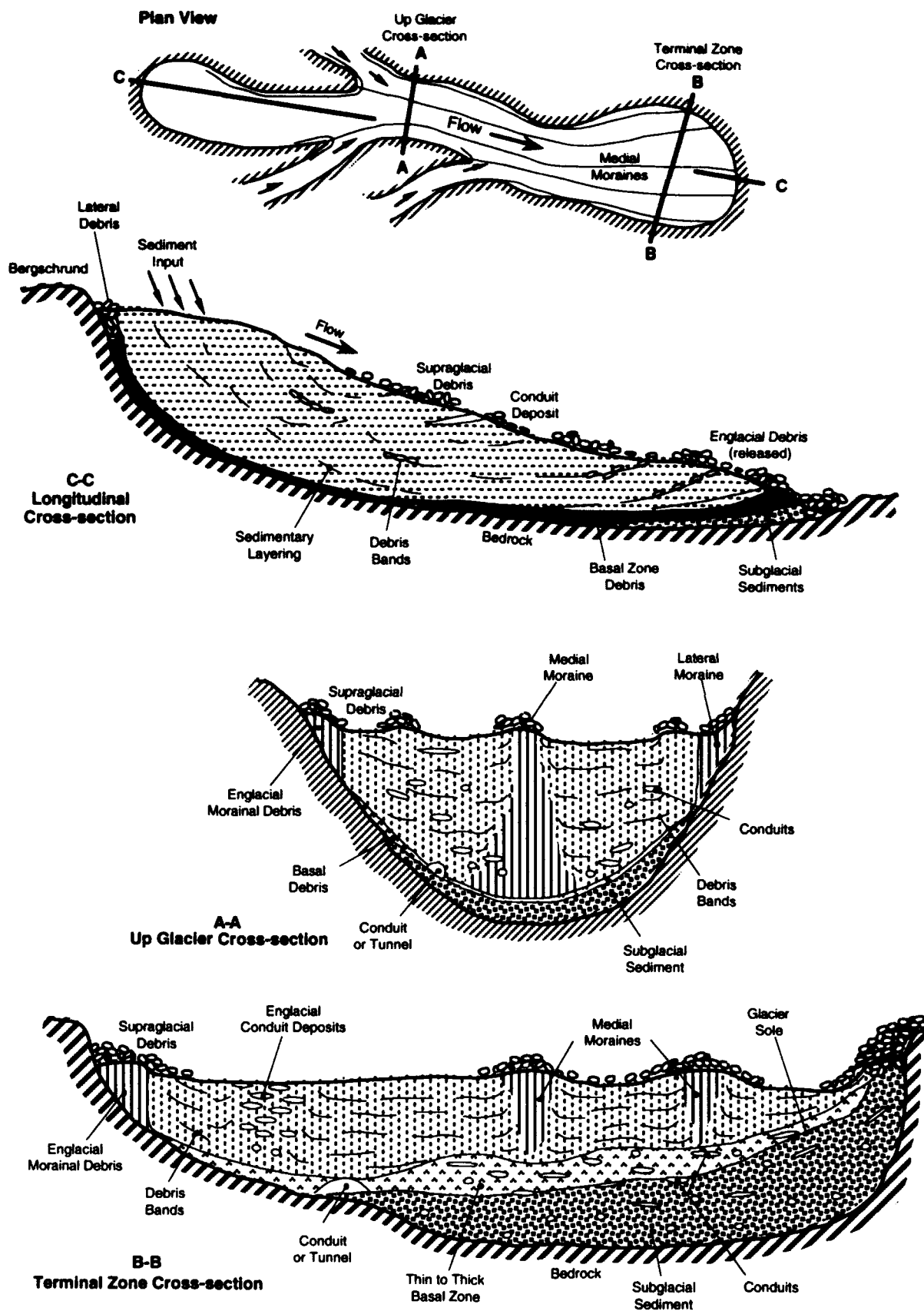


Figure 50. Debris distribution within an idealized valley glacier with multiple tributary glaciers.

snow accumulation and metamorphism, ice flow rate and mechanics, and ablation at the ice surface and the glacier's sole. Erosion processes control sediment production at the bed and the volume of new material available for ice or water entrainment and transport. Subglacial sediments, which may be transported by glacier-induced deformation (e.g., Alley 1991), form a significant source of erodible sediment.

Sediment transport by ice is a function of the location and nature of entrainment and thermal and mechanical processes associated with ice mechanics. Ice flow and sliding determine the predominant transport paths for debris derived at the surface and at the bed (Fig. 50). In particular, sediment added to the surface of the glacier's snowpack in the accumulation area will be transported englacially, while that added to the surface in the ablation zone, without the action of processes to be described subsequently, will be carried supraglacially (e.g., McCall 1960, Meier 1960, Grove 1960).

Numerous observations have shown that the majority of supraglacial debris, as well as englacial debris, is material from valley walls that is deposited on the ice by mass movements—rockfalls, avalanches, slushflows (e.g., Tarr and Martin 1914, Sharp 1949, Reid 1968). Layers of sediment deposited on snow in the accumulation area ultimately form englacial debris bands. Debris bands have a finite lateral extent and lie mostly subparallel to

the ice surface. Wind may transport fine-grained material onto the glacier surface. In the accumulation area, it becomes concentrated by melt processes as it is incorporated into the perennial snowpack, eventually forming diffuse sedimentary layers as the snow changes to ice (Meier 1960). Streams draining adjacent valley slopes may also deposit sediments on the ice surface. These materials can become incorporated englacially if deposited in the accumulation area. In the ablation zone, supraglacial streams may transport this material through moulins and conduits into englacial and subglacial locations.

Debris eroded by glacier action from the lateral confining channel walls is transported within and on the ice edge as debris septa (Sharp 1949) to form lateral moraines that include the debris that simply falls to the ice surface from the valley walls (Fig. 50). At depth, lateral moraines may blend into the basal debris. Where tributary glaciers converge, lateral moraines join to form medial moraines, which are transported as longitudinal features that parallel ice flow (Fig. 51).

Supraglacial debris is reworked and modified by weathering, especially freeze-thaw, and by rain, meltwater flow and mass movement (e.g., Sharp 1949, Eyles and Rogerson 1978). Supraglacial streams may rework and sort these sediments, transporting them into englacial conduits where deposition, conduit abandonment and ice creep



Figure 51. Medial and lateral moraines developed from material of valley walls in the Alaska Range. Medial moraines commonly separate ice from tributary valleys (photo by E. Evenson).

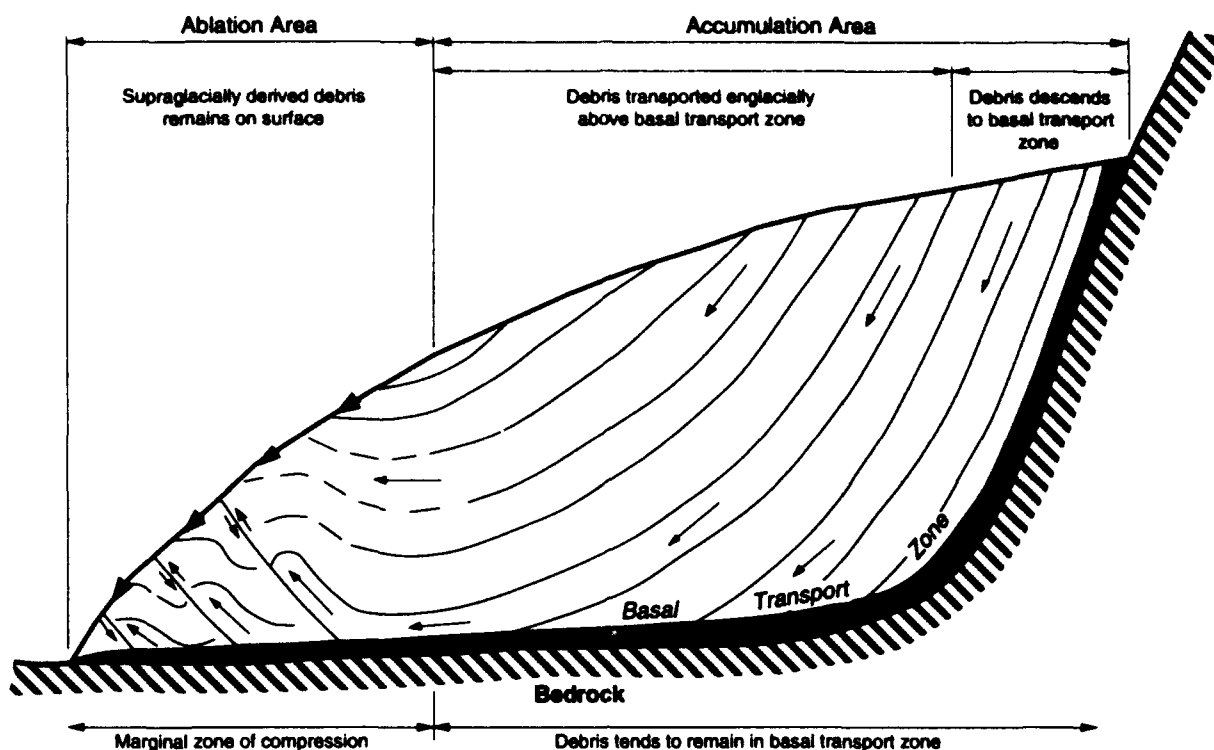


Figure 52. Idealized transport paths for debris derived supraglacially and subglacially in a valley glacier (after Boulton 1978).

may result in their remaining in an englacial position. Or they may be further transported into the subglacial drainage system, where basal processes will modify them. Supraglacial debris may also be incorporated into the ice by falling into crevasses that open during the course of ice flow and that subsequently close as the ice moves from zones of tensile to compressive stress.

Englacial debris derived from the accumulation area (essentially supraglacial debris) will be transported along flow lines (Fig. 52) and emerge at the surface of the ablation area, except if it moves to a basal position. Transport paths from higher elevations, such as the bergschründ, may intersect the bed under proper conditions, thereby resulting in modification of the debris; here, it may be incorporated into a basal or subglacial transport mode (e.g., Boulton 1978). Material incorporated from lateral valley walls at a distance from the immediate ice edge is more commonly transported above the sole of the glacier and therefore out of contact with the bed (Fig. 52).

Sediments transported within the basal zone are primarily derived from the bedrock or sediments underlying the glacier by various processes operating at the ice/bed interface. Particles within the ice just above the bed interact with subglacial sediments or bedrock while effectively in traction-

al transport as the ice slides. In the particular case of unconsolidated bed materials, they may deform continuously, with particles being physically modified by grain-to-grain interactions (e.g., MacCintock and Dreimanis 1964, Boulton et al. 1974). The effectiveness of subglacial deformation in causing modifications depends upon the materials' resistance to shearing. This resistance is a function of grain size distribution, composition, shape and related properties, and the effective stresses generated at and below the ice/bed interface, the stress level being strongly influenced by pore water pressure (e.g., Boulton et al. 1974, Boulton 1979, Clarke 1987).

Where bedrock is in contact with the base of the glacier, mechanical processes interact with meltwater to fracture and crush the rock, mainly at bed roughness features (bumps) and along existing structural features (such as joints). The strength or failure resistance of the underlying rock depends to a large degree on its composition and amount of weathering, and the presence of fractures, bedding planes, joints and other planes of weakness (e.g., Boulton 1974, Hallet 1974, 1979b, Boulton 1979, Röthlisberger and Iken 1981, Iverson 1991). These particles are plucked from the bed during and following failure. The grain size of the entrained material reflects the relative importance of

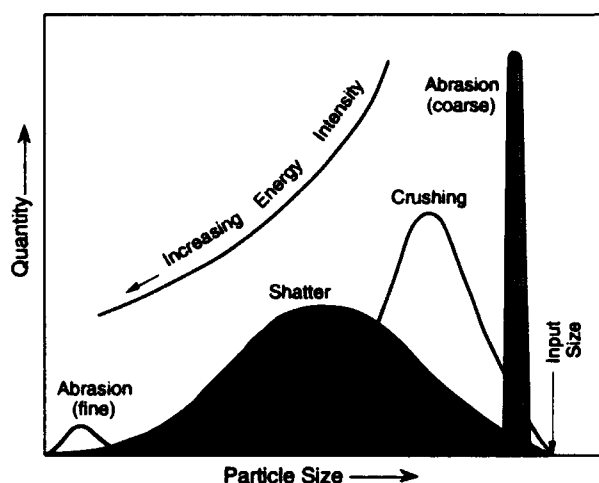


Figure 53. Characteristic grain-size distributions resulting from processes of abrasion, crushing and impact (after Drewry 1986).

each mechanical process (Fig. 53). A detailed summary of the various mechanical and chemical processes of bedrock erosion is given by Drewry (1986).

These various mechanical processes are responsible for creating "new" sediment, which is either entrained by the ice or by water within the subglacial drainage system. Rates of sediment production are difficult to estimate. Estimates from rock abrasion rates measured beneath glaciers and those from recently exposed bedrock substrates are of the same order of magnitude (e.g., Boulton 1974), but clearly can differ with the rock type, degree of weathering and presence of structural weaknesses. Sediment concentrations in meltwater discharges, which are commonly used to define a glacier's sediment yield, are also used to infer erosion rates (e.g., Thorarinsson 1939, Ostrem 1975a, Chernova 1981, Kjeldsen 1981). Erosion rates approximated by this method may be reasonable where the glacier lies directly on bedrock and transports little supraglacial or englacial debris. However, such estimates become progressively less reliable as the amount of debris in transport increases, and perhaps become meaningless when the glacier substrate is predominantly sediment. In general, however, estimates based on sediment load overestimate this rate because meltwater also carries sediment entrained by thermal and mechanical erosion of debris on or within the glacier, as well as existing sediment from the glacier substrate (e.g., Gustavson and Boothroyd 1982). Additionally, sediments can be added from adjacent glacial or glacial tributary basins by streams entering the subglacial drainage along lateral margins and at

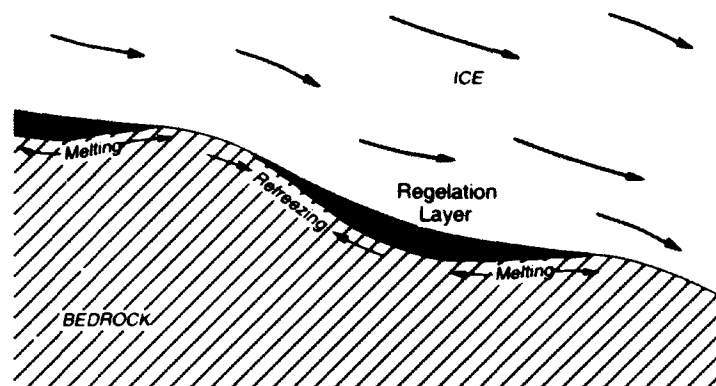
ice tributary junctures (Evenson and Clinch 1987), in some cases these sediments being the primary component in streams draining glaciers that lie directly on bedrock.

The method by which sediment and rock particles are incorporated into the glacier determines the thickness and debris content of the basal zone. Temperate glaciers (generally considered those at the pressure-melting point throughout, including at the ice/bed interface) lying on bedrock are commonly characterized by a thin (less than 1.0 m) debris-laden basal layer in which primarily rock fragments have been incorporated. These fragments are reasonably well-dispersed in small concentrations throughout the regelation ice, typically at volumes of less than 5% (Lawson 1979a, Sugden et al. 1987a, Hubbard 1991). This material is thought to be incorporated primarily by regelation in response to pressure-melting around bedrock roughness features or obstructions to sliding (Kamb and LaChapelle 1964, Weertman 1964).

Regelation is, by definition, the process of pressure-melting and refreezing at the glacier's sole in which the heat of fusion for melting is supplied by the heat of fusion released by adjacent freezing, and involves essentially no net loss of ice by melting and no net addition of ice by freezing. The basal shear stress causes pressure to be higher and the melting point to be lower on the upstream side of a bump at the bed than on the downstream side of it. Ice therefore melts on the upstream side. This water flows around the bump and refreezes on the downstream side. The heat released by refreezing is conducted back through the bump to allow further melting (Fig. 54). Because heat must be conducted through the bump, regelation is efficient only for very small bumps, typically a few millimeters or less in size. Larger bumps result in ice flow by enhanced creep (Weertman 1957, 1964). During refreezing, debris from the bed (abrasional products or sediment) may be trapped in the regelation ice, thereby being incorporated into the glacier (Kamb and LaChapelle 1964), and creating a debris-rich layer that is tens of centimeters thick.

Much thicker (several to 10 m or more) and more heavily debris-laden basal zones characterize glaciers where thermal conditions at the ice/bed interface permit the freezing of meltwater and the creation of new ice at the glacier sole (Fig. 55) (e.g., Weertman 1961; Boulton 1970, 1972; Lawson and Kulla 1978; Lawson 1979b; Strasser et al. 1992). During freeze-on, sediments or rock particles at or below the ice/bed interface can be included within the new ice. The resultant debris-rich basal zone

Figure 54. Ice formed by regelation, with ice on the down-glacier side of bed obstacles made from refreezing of water pressure-melted from the up-glacier side of the same obstacle (after Kamb and LaChapelle 1964). Maximum regelation layer thickness is controlled by obstacle size (about 10 mm), but commonly may reach 1.0 m thickness because of various mechanical stacking processes operating along the glacier flow path.



is typically stratified, with layers or lenses of high and low sediment concentration (Lawson 1979a, Sugden et al. 1987a,b). As a sediment source, limited measurements from glaciers and ice sheets suggest that this horizon commonly transports the majority of the glacier's sediment load, except in the case of an extensive supraglacial debris cover. Sediment volumes range from several percent to over 90% in individual strata, with a mean exceeding approximately 25% by volume (e.g., Gow et al. 1979, Lawson 1979a, Hubbard 1991).

Sediment incorporation by freeze-on results in characteristics that dominantly reflect the nature

of sediments at the ice/bed interface, and hence, the processes (glacigenic, hydraulic) acting within the subglacial zone (Lawson and Kulla 1978, Lawson 1979a). Their structure may be modified by subsequent ice deformation or planar shearing at the new ice/bed interface, but the actual effectiveness of these processes in altering the characteristics of entrained sediments depends upon where the sediments are incorporated and the length of flow and length of time over which they are subject to basal mechanical processes (Lawson 1988b).

Thermal conditions at the bed of glaciers are poorly characterized, either by field studies or



Figure 55. Basal ice covered by sediment released by ablation, Matanuska Glacier, Alaska. Thick sequences of debris-laden ice develop mainly by accretion of ice and sediment to the glacier's sole. Here, the exposed sequence ranges from 5 to 7 m thick, with some overthickening by localized overthrusting of the margin to 12 m thickness.

theoretical analyses, yet appear critical to sediment entrainment mechanisms. Weertman (1961) first theorized that, under nonsteady flow conditions, changes in thermal conditions at the bed may result in large-scale entrainment of blocks or rafts of material by ice accretion. Weertman assumed that under relatively thick ice, the temperature gradient in the glacier may be insufficient to conduct viscous and geothermal heat fluxes from the bed. Dissipation of this excess heat melts basal ice, producing meltwater that flows generally outward toward the terminus in accordance with the pressure gradient. As the ice thins near the terminus margin, the temperature gradient steepens, heat from the bed is dissipated and melting ceases. Freezing conditions develop where insufficient heat is generated to maintain the ice/bed interface at the pressure melting point. Weertman (1961) suggested that the location of the 0°C isotherm, the boundary where temperatures closer to the terminus are below the pressure-melting point, could change in response to various factors. A shift up-glacier would depress temperatures in areas formerly at the pressure-melting point and result in freeze-on of meltwater and the water-saturated materials just below the ice/bed interface. Large-scale amounts of bed material could be included in the basal zone by this mechanism. However, because of thermal constraints, this process is probably restricted to the thinner parts of the terminus (less than 30 m thick) where mean annual temperatures affect bed temperatures. Thus, this mechanism may account for seasonal entrainment of sediment near the ice margin.

Boulton (1972) suggested that the thermal regime at the base of a glacier can vary in response to certain physical parameters in accordance with Weertman (1961). The variations result in zones where a net loss or net addition of ice and sediment can occur at the ice/bed interface. In a general sense, zones can be identified with 1) net ice loss resulting from melting by geothermal heat and the heat of sliding and deformation, 2) no net ice loss or gain, but pressure-melting and refreezing (regelation) active, 3) net gain of ice by freeze-on of meltwater where the interface is slightly below the pressure-melting point, and 4) an interface where ice is frozen to the bed. Meltwater flowing from a zone of net melting or regelation to a zone of net freeze-on would create new ice on the glacier sole and entrain sediment. The actual mechanics of this process are, however, not yet defined. The distribution of these zones beneath a glacier can theoretically vary in response to several controlling fac-

tors, including bed roughness, topography, ice flow rates and others. Complex situations, with multiple zones alternating in response to these factors, result.

Robin (1976) theorized that cold patches that are a few tenths of a degree below the pressure-melting point and of 0.1 to 1.0 m in extent may exist, "permanently" or intermittently at the ice/bed interface. They exist in response to heat transport with pressure and temperature variations at an interface with a thin film of water. Robin proposed that water derived from melting of ice just above the bed or from external sources can freeze to the glacier's sole in areas of either reduced stresses or reduced basal water pressures. Heat conduction by meltwater from within deforming basal ice toward the ice/bed interface (a "heat pump" mechanism) will temporarily cool the basal ice and thereby produce cold patches. Changes in water pressures in concert with stress variations and heat flux around bedrock roughness features could likewise produce cold patches in high stress regions. A slight but relatively rapid reduction in pressure could then temporarily result in water at temperatures below the pressure-melting point (freezing) and permit ice growth on the glacier sole. Lliboutry (1987) similarly theorized that freeze-on over areas of several decimeters can occur on high parts of a rough bed.

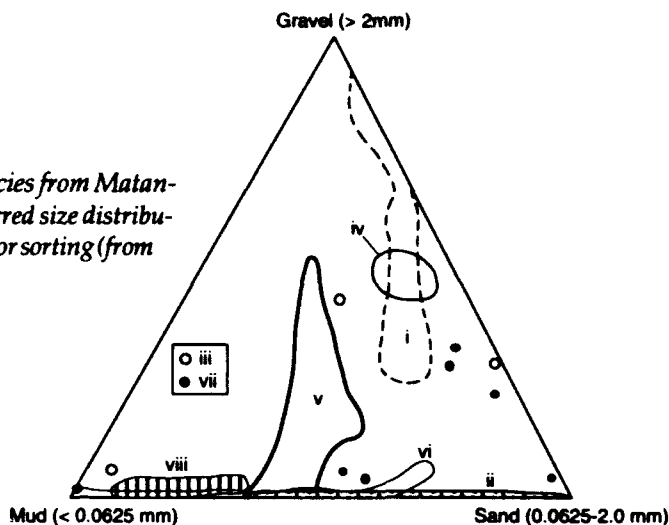
More recently, researchers have hypothesized that ice accretion and sediment entrainment may result from supercooling of water within the englacial and subglacial drainage systems in response to changes in pressure (Strasser et al. 1992, Alley et al., in prep., Lawson et al., in prep.). Supercooling in response to a rapid drop in pressure results in ice nucleation within the water and ice crystal growth on cavity and conduit surfaces. Frazil ice, as the nucleating crystals are called, continues to grow and evolve within the supercooled water, forming permeable masses or flocs that attach to ice surfaces where sediments in water passing through them are trapped within the interstices. Eventually, continued ice growth forms a solid, debris-laden mass. Similarly, epitaxial ice growth on cavity walls and ceilings can both trap sediments and close cavities. Sediments in contact with the supercooled water may also be frozen to it. This new idea remains unproven, but under study.

Debris characteristics

Because of the different sources of debris, its passive or active mode of transport, and any sub-

Figure 56. Grain size characteristics of various ice facies from Matanuska Glacier, Alaska. The solid subfacies has no preferred size distribution, with extreme textural diversity and poor to very poor sorting (from Lawson 1979a).

- i - supraglacial debris
- ii - diffused facies
- iii - banded facies
- iv - dispersed facies
- v - composite of stratified facies
- vi - discontinuous subfacies
- vii - solid subfacies
- viii - suspended subfacies



sequent modifications by glacial processes, the characteristics of debris from the englacial and supraglacial zones differ from that within the basal zone and subglacial environment (e.g., Boulton 1970, Lawson 1979a,b). The primary differences are in grain size, shape and distribution. Supraglacial and englacial debris are very coarse, poorly sorted, with few particles finer than sand-size, and extremely angular (Fig. 56) (e.g., Tarr and Martin 1914, Boulton 1978, Lawson 1979a,b). Supraglacial

meltwater flow can also remove the sand size fraction of supraglacial debris during glacier transport, coarsening it, while processes involving freeze-thaw can break down materials into angular particles (Fig. 57) (Sharp 1949).

In contrast, the sediments within the basal zone are much less angular and exhibit a wide range of grain sizes (from clay to boulder size), with the actual grain size distribution related to the characteristics and composition of the glacier bed and the



Figure 57. Supraglacial debris. Meltwater erosion and topographic inversion caused by ablation rework supraglacial debris, sometimes moving it to englacial and basal positions through crevasses and moulins.

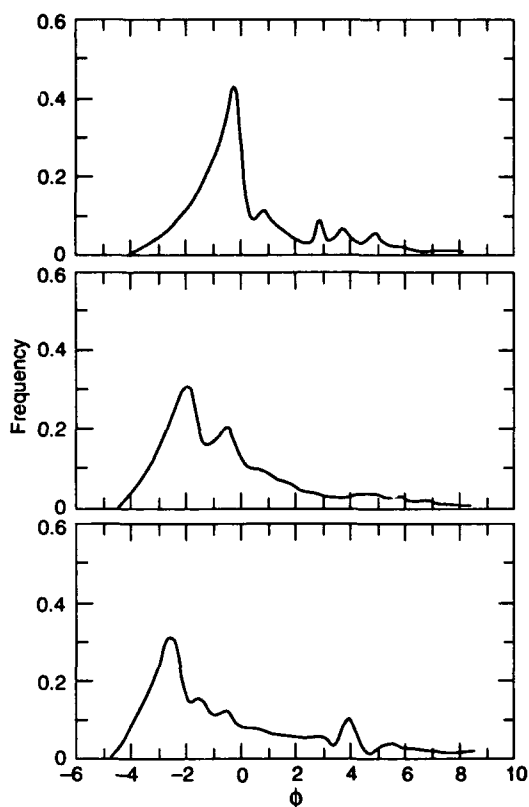


Figure 58. Typical grain size distribution curves for debris from the dispersed facies of the basal zone (from Lawson 1979a); this facies considered to be derived primarily from bedrock by the regelation process.

mechanisms by which they are incorporated (Lawson 1979a,b). In addition, the amount of reworking and meltwater transport of fine-grained silt versus debris production is important in the final grain size distribution. For example, debris in basal ice of a temperate glacier lying on granitic bedrock is typically coarse, containing sand to boulder size particles, but little debris of silt or clay size (Fig. 58).

Debris in basal ice derived from beds of unconsolidated sediment is more complex in character (Fig. 59). Grain size distribution ranges from clay to boulder size, exhibiting a poorly sorted, bimodal or even polymodal frequency distribution (Fig. 60). The actual size range, however, varies with that of the source material and the process of incorporation and therefore may vary laterally and vertically within the basal zone. Freeze-on processes produce the most variable debris characteristics because they can incorporate sediment without regard to size. Silt from meltwater is commonly important and may dominate the distribution. Within the basal zone, the debris, individual lenses and layers surrounded by clear ice can be well sorted and of limited size range where freeze-on incorporates material with such properties (Fig. 61). Gravel-size clasts (granules to boulders) are less angular than supraglacial and englacial debris, often having rounded edges and spheroidal to prolate shapes, similar to those of water-worn stream gravels (e.g., Lawson 1979a,b; Boulton 1978).

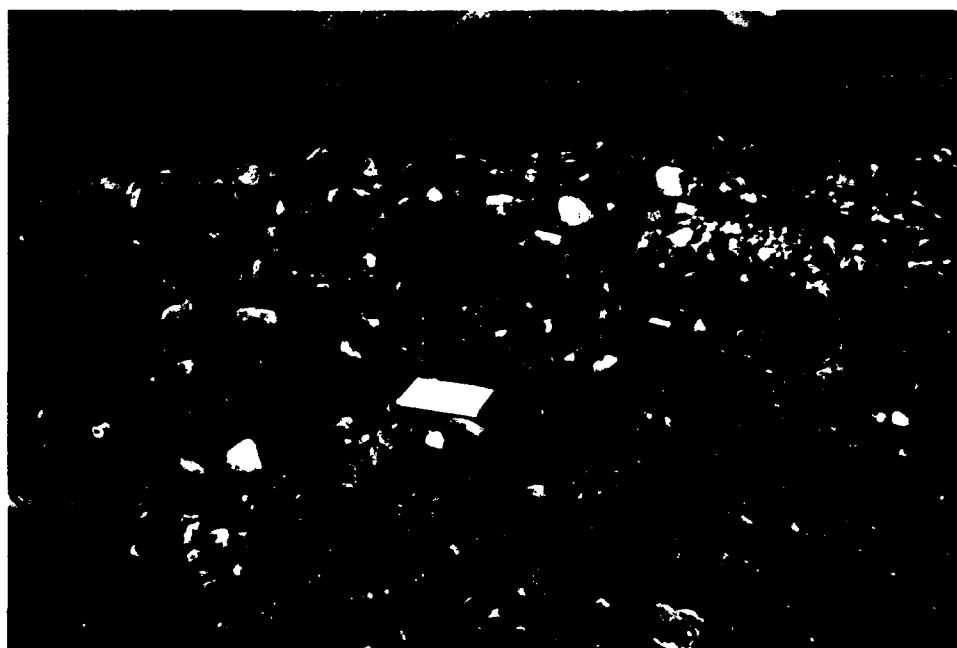


Figure 59. Basal debris. Particles released by ablation of basal zone ice. Fine-grained components are being removed by meltwater flow down the sloping ice surface. Mostly subround to round shapes characterize subglacially derived gravel-size materials. Notebook for scale.

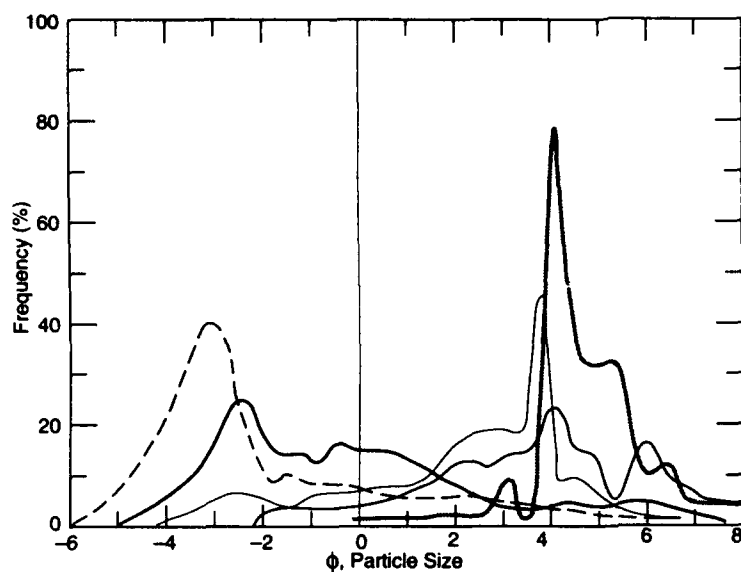


Figure 60. Typical grain size distribution curves from sediment within the stratified facies of the basal zone of Matanuska Glacier, Alaska (from Lawson 1979a). This facies is mainly derived from net freeze-on of sediment by ice accretion to the glacier sole.



a. Clear ice separates and surrounds sediment, which may occur in layers or be dispersed or suspended within it. Sediment ranges from clay to pebble size (scale in centimeters).

b. Well-stratified basal zone ice. Sediments are mainly layers of silt and sand, with occasional pebbles (scale in centimeters).



Figure 61. Examples of debris in basal zone formed by accretion processes.

The characteristics of the glacier's substrate are thus important in determining basal debris characteristics and sediment transport by ice, as well as meltwater. Substrate characteristics are as varied as glaciers themselves, and will reflect the geologic history prior to glaciation and the subsequent effects of the glacier while above it. Glaciers may lie directly on bedrock stripped clean of sediment by previous glaciations, while others develop upon sediments deposited by any number of nonglacial, glacial or proglacial processes. Perhaps more commonly beneath larger glaciers and ice sheets with lengthy flow paths, materials at the ice/bed interface may be in part bedrock and in part sediment. The characteristics of the debris and ice of the basal zone of the Matanuska Glacier in south-central Alaska, for example, suggest that the bed has an up-glacier section of mainly bedrock and a down-glacier section of mainly unconsolidated sediment (Lawson 1986, 1988b, Lawson et al. 1991). Because of the effects of bed topography, ice flow may result in localized accumulations of subglacially eroded sediment on a bed that otherwise is mostly bedrock (Hooke 1991). With time, the interaction of the glacier with the bedrock or sediment and the erosion or deposition of these materials by subglacial meltwater will determine the overall characteristics of the bed.

Because of the inaccessibility of the subglacial environment, the nature of the ice/bed interface under active ice is poorly understood. Studies of ancient and recently exposed bedrock from beneath glaciers, some with thin, discontinuous mantles of rock particles and fragments, have improved our understanding of subglacial processes

(e.g., Hallet 1974, Walder and Hallet 1979, Hallet and Anderson 1980, Sharp et al. 1989, Sharpe and Shaw 1989). Glacier interactions with beds of sediment are not as well defined by such studies because the beds are reworked by various subglacial processes during and immediately following deglaciation (e.g., Church and Ryder 1972; Boothroyd and Ashley 1975; Lawson 1979a, 1982, 1988a; Sharp 1982, Dreimanis 1988).

Theoretical treatments and observations of the effects of ice flow indicate that subglacial sediments are deformed (e.g., Engelhardt et al. 1978, Boulton and Hindmarsh 1987, Clarke 1987). Shear stresses generated by glacier flow can mechanically abrade and comminute the coarser sedimentary particles (e.g., Boulton et al. 1974, Sharp 1982, Boulton 1979) and, under proper conditions, will result in transport and a net flux of sediment to the glacier terminus (e.g., Boulton et al. 1974; Alley et al. 1987a,b; Boulton and Hindmarsh 1987; Clarke 1987; Eschelmeyer and Wang 1987; Alley 1989b, 1991). Depths of deformation may exceed 6 to 7 m and depend on shear stress and effective pressures in the bed.

Theory indicates that a glacier resting on unconsolidated sediment can move by discrete sliding between the ice and bed, ploughing clasts through the uppermost materials (Brown et al. 1987), by pervasive deformation, or by shearing across discrete planes at various depths within the bed materials (Fig. 62). The type of drainage system (distributed, conduit-tunnel or linked-cavity) affects the magnitude of the sliding rate and depth of deformation (Alley 1989a,b). Textural properties of the bed material, particularly the number

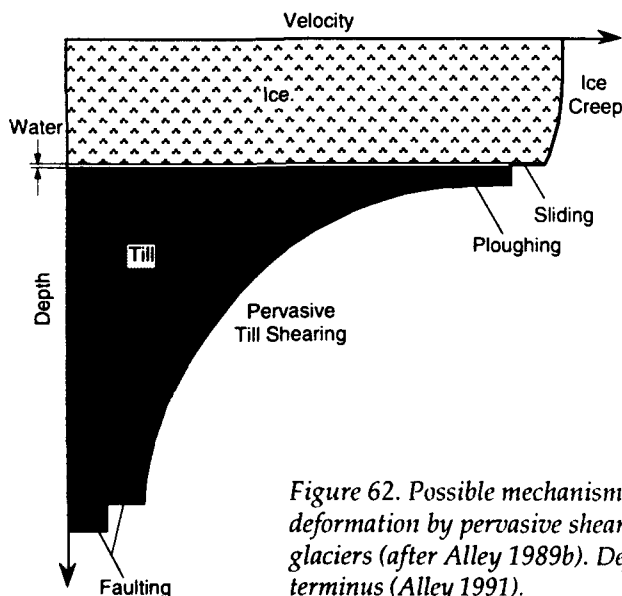


Figure 62. Possible mechanisms of ice motion on a bed of unconsolidated sediment, with deformation by pervasive shearing being the principal velocity component in wet-based glaciers (after Alley 1989b). Deformation can result in a substantial flux of sediment to the terminus (Alley 1991).

and size range of clasts, also affect basal sliding rates and the importance of ploughing of clasts at the interface. In the case of water drainage in a distributed-flow system, pervasive deformation of the underlying sediments may account for up to 60 to 100% of the total velocity of the glacier's motion (Alley 1989b).

Debris flux

The transport or flux of debris to the terminus of a glacier determines the long-term availability of sediment for entrainment by ice marginal and basin tributary streams. Sediment flux can be estimated on the basis of several parameters; however, in each case, this estimate is limited by the complex spatial and temporal variability in glacial debris, including that in a subglacially deforming layer. The total sediment budget consists of the flux of debris in transport and "new" sediment, produced by glacier erosion, that is entrained and transported in the subglacial drainage system.

Alley (1991) has suggested that sediment transport by bed deformation can be estimated by assuming that the depth of deformation is directly related to the basal shear stress, the sediment's yield strength and the effective pressure as it varies with depth. Velocity is determined by an inverse relationship between the change in basal shear stress and yield strength with depth as a function of the effective pressure at depth. In simplified form, the sediment flux by pervasive deformation SQ_D is

$$SQ_D = \alpha h_b u_b \quad (89)$$

where h_b = depth of deforming layer

u_b = velocity at top of the deforming layer, approximately equal to u_i , the ice velocity

α = constant dependent upon sediment layer thickness and water pressure, ranging from about 0.1 to 0.5.

The consequence of sediment reaching the glacier terminus is determined by the relationship between the sediment motion and ice sliding velocity, assuming they are equal, and the velocity of the glacier at its terminus (u_a). If $u_a \geq \alpha u_b$, the deforming layer must erode at the terminus to maintain a constant deforming thickness h_b during the advance. Without erosion, the deforming layer will progressively spread over a longer and longer distance and become thinner and thinner. If $u_a = \alpha u_b$, then there is neither erosion nor deposition and, if $u_a < \alpha u_b$, deposition occurs (Alley 1991).

These relationships clearly affect estimates of the probable long-term sediment flux and potential sediment discharge to glacier outlet streams.

Sediment flux resulting from the transport of debris within the basal ice zone can only be grossly estimated at present. Several key parameters required for sediment flux estimates, including sediment volume and its spatial and temporal distribution, are poorly defined by measurements. The principal assumptions for flux calculations are 1) that sediment volumes are constant spatially within the basal zone and will not vary with time; 2) that the general distribution of debris-laden basal ice within the glacier is constant; 3) that a limit to the extent and thickness of the debris-laden basal ice must be based upon a presumed thermal regime at the bed; and 4) that ice velocity and thermal regime are constant over time.

The flux of basal debris in ice per square meter width of glacier (SQ_B) can then be estimated by (Drewry 1986)

$$SQ_B = u_b h_b c_b \quad (90)$$

where h_b = thickness of basal layer

u_b = basal sliding velocity

c_b = debris concentration in the ice.

Because englacial debris is widely dispersed and highly variable in concentration, it is difficult to estimate englacial transport volumes and debris flux. In most cases, englacial debris flux (SQ_E) is a small percentage of the total sediment flux and some constant value might be assigned for long-term estimates. It might be better to estimate volumes of englacial debris from volumes of material exposed supraglacially by ablation below the equilibrium line and, as with basal debris, to make assumptions on debris dispersal and its variability with time.

Supraglacial debris transport (surficial deposits, medial and lateral moraines) can be more accurately estimated if its extent and thickness can be measured. Debris concentration at depth below the moraines must be estimated as an englacial component and is usually presumed constant with depth. Thus, supraglacial and morainal band sediment flux SQ_s is

$$SQ_s = u_s h_s w_s + u h_m w_m c_m \quad (91)$$

where u_s = surface ice velocity

h_s = thickness of supraglacial debris

w_s = width of debris cover

c_m = debris concentration in moraine septa

u = ice flow or creep velocity
 h_m = thickness of debris septa taken equal to ice thickness
 w_m = width of debris septa.

Ice flow velocity u is commonly estimated from measurements of surface velocity based upon the flow law for ice.

Thus, the total annual sediment flux for glacier or ice transport can be represented as the sum of each debris flux component, integrated with respect to the duration or length of flow

$$SQ_T = \int_0^t (SQ_D + SQ_E + SQ_S + SQ_B) dt. \quad (92)$$

For a glacier with a substrate that is mostly bedrock, SQ_D is assumed equal to 0.

Water entrainment and transport

The fluvial entrainment of sediment by water moving through the englacial and subglacial drainage system, as well as its role in subglacial erosion of the bed (e.g., Sharpe and Shaw 1989), are poorly characterized by either direct measurements or theoretical hypotheses of the thermal and mechanical processes of meltwater erosion.

Sediment concentrations in supraglacial streams and the upper reaches of englacial streams range from 0 to in excess of 60 g/L, but typically range from 50 to 400 mg/L (e.g., Dowdeswell 1982, Hammer and Smith 1983, Fenn 1983, Gurnell 1987). This value is strongly influenced by the extent and textural nature of supraglacial and englacial debris, with streams draining supraglacial moraines having the highest suspended sediment concentrations.

Debris in englacial and basal ice can be released by melting of conduit and tunnel walls by water in response to viscous heat dissipation during periods of increased discharge (Röthlisberger 1972). Similarly, expansion of a linked-cavity system can release basal debris from orifice and cavity roofs, while in a distributed flow system, pressure-melting of the glacier sole can release debris. Rates depend on the concentrations of debris within the ice and the quantity of heat available. The volume of debris released by heat dissipation in conduits and cavities has not been measured.

Similarly, there are few estimates based upon direct measurements of, or theory concerning, the rate of mechanical erosion of bedrock (abrasion, cavitation) and the volume of this material that is

entrained by water flow at the ice/bed interface. Vivian (1970) estimated bedrock erosion rates of about 20 mm/yr based upon analyses of pothole formation below glaciers, and correlated these rates with experiments on abrasion of quartzite. Rates of abrasion by sediment-laden meltwater depend to a large degree on the grain size, mineralogy and shape of the particles in transport, the total amount of suspended sediment, flow velocity and the composition of the bed (Bezing and Schafer 1968, Bogen 1989).

Bogen (1989), based upon the work of Bouvet (1958), suggested that the abrasive capacity of sediment-laden meltwaters (s) might be represented as

$$s = Q_s r q d \quad (93)$$

where Q_s = annual suspended sediment load

r = grain shape, values ranging from 1 to 5 for well-rounded to angular particles

q = proportion of quartz particles to all other mineral particles in the 0.063- to 0.125-mm size range

d = proportion of sand-size particles (0.063- to 0.5-mm diameter) to total sediment weight.

Bogen (1989) suggested that this relationship could define the magnitude of turbine wear in hydroelectric power plants caused by annual glacial discharges. Various other theories on the processes of meltwater abrasion and cavitation are detailed by Drewry (1986).

The largest potential source of material for meltwater entrainment is unconsolidated materials in contact with the ice. Theoretical estimates of the erosive potential of conduits in unconsolidated sediment may be based upon traditional river flow theory (e.g., Allen 1985), assuming a configuration for the subglacial passageway (R- or N-conduits and tunnels) and appropriate water and ice pressures.

Alley (1989a) estimates sediment flux through a subglacial conduit embedded within sediment by assuming that it must equal the influx of sediment by creep into the conduit. Water flow can generally be assumed to be turbulent within conduits (as well as within the "orifice" or connecting conduit of the linked-cavity system [Kamb 1987]) and its velocity can be defined using the Manning formula. Thus, mean velocity U_m is

$$U_m = M^{-1} P_g^{1/2} r^{2/3} \quad (94)$$

where M = Manning roughness coefficient
 r = channel radius
 P_g = local hydraulic pressure gradient.

The erosion rate ξ_e is then given by (Alley 1989a)

$$\xi_e = \frac{1}{2\pi r^2} \frac{\partial J_s}{\partial x} = \dot{r}/r \quad (95)$$

where x is distance measured along the channel, $\dot{r}$ is the rate of change in r and J_s is sediment flux in the channel as given by

$$J_s = J_0 \pi r^2 U_m^3 \quad (96)$$

where J_0 is a constant.

Alley (1989a) modifies this equation by substituting the value U_m from eq 94 into eq 96, differentiating with respect to x , and substituting the x derivative of J_s into eq 95. Based upon the relationship that

$$Q = \pi r^2 U_m \quad (97)$$

where Q is water flux, the erosion rate for turbulent flow ξ_{et} in terms of the water and sediment flux is

$$\xi_{et} = \frac{3J_0 M^2 P_g Q_x}{4\pi r^{2/3}} \quad (98)$$

A similar expression for laminar flow is (Alley 1989a)

$$\xi_{el} = \frac{J_0 P_g^2 r^2 Q_x}{64\pi \mu^2} \quad (99)$$

where μ is the viscosity of water. As stated in an earlier section, the creep closure rate ξ_c , which the erosion rate must equal or exceed for the conduit to remain open, is given by

$$\xi_c = K_b \frac{(N - \tau^*)^a}{N^b a^a}, \quad N > \tau^* \quad (100)$$

$$\xi_c = 0, \quad N \leq \tau^*$$

and

$$\xi_c =$$

where N = effective stress
 τ = shear stress

τ^* = sediment strength
 a, b, K_b = constants.

The quantity of sediment eroded and transported by conduit and tunnel flow can be limited by the capacity of the meltwater discharge. This limiting condition will affect not only the sediment flux but also limit the conduit and tunnel dimensions. Theoretical estimates of sediment flux by meltwater erosion of an unconsolidated bed for various subglacial drainage types and configurations have not yet been made.

SEDIMENT FLUX RELATIONSHIPS

If we assume that the amount of sediment in transport is mainly a function of the availability of sediment at the ice/bed interface, including the debris-laden basal ice, and of the action of meltwater within the englacial-subglacial drainage network at the interfacial zone, qualitative relationships between sediment transport-sediment yield and subglacial glaciohydraulic processes can be deduced from data obtained by monitoring the waters emerging at the glacier terminus (Collins 1979a,b). Such an assumption is required because of our lack of understanding of subglacial sediment transport processes. One of the longest multiyear records of sediment discharge in relation to water discharge and various water quality and meteorological parameters is that measured by Collins over the last 15 years at the Gornergletscher and Findlingletscher near Zermatt, Switzerland, two of the glacierized basins feeding the Grande Dixence S.A. hydroelectric scheme. In addition, various hydrological and glaciological studies have been supported by Grande Dixence over the last 40 years at these and other glacierized basins in this region.

On the basis of these long-term records, Collins (1986, 1988, 1989) has concluded that the seasonal patterns in sediment flux are related directly to the seasonal development of the subglacial drainage network, and further, that short-term variations in sediment discharge are mainly a function of sediment availability for subglacial hydraulic processes. The latter relationship is also a function of the previous history of hydrological events that have removed (or deposited) bed sediments. Subglacial hydrological events may include discharge variations as the englacial-subglacial drainage system evolves each spring, sudden drainage of ice-dammed lakes, temporary blockage of subglacial passageways by ice flow, relocation or reestablish-

ment of subglacial conduits, and flooding caused by rain. Feedbacks exist between the factors or process, but sediment discharge may also vary irregularly.

The type of subglacial drainage network will determine both the area of the bed over which water flow can withdraw sediment and the timing of active sediment withdrawal. For example, in the case of a glacier in which subglacial drainage consists of both a linked-cavity system and a conduit-tunnel system developed on bedrock, with only a thin or discontinuous sediment layer, seasonal variations in the extent and operability of each system affect sediment discharge at the terminus (e.g., Iken et al. 1983, Iken and Bindshadler 1986). Meltwater reaching the bed in spring progressively reestablishes and expands the linked-cavity system, resulting in increased pore water pressures, increased ice flow rates and increased sediment discharge.

Sediment transport may also vary rapidly as different parts of the bed are connected to the drainage system and sediment is progressively, but episodically, evacuated from new areas. In addition, early in the melt season, diurnal variations in meltwater input may result in episodic opening and closing of conduits connecting cavities (e.g., Iken and Bindshadler 1986). Subsequently, discharge via the conduit-tunnel system becomes increasingly more important, relative to the linked-cavity system, as meltwater input increases into the summer season. Overall, water pressures decrease and the actual area of the bed from which sediment can be eroded decreases. Sediment discharge progressively drops through the summer, with the frequency of large variations that often characterize early summer runoff decreasing. Then, high sediment discharge results mainly from floods (e.g., Collins 1989).

The lower sediment flux that characterizes many glacierized basins late in the melt season apparently reflects the presence of a fully developed drainage system on a bed that is mostly rock, and an exhaustion in the supply of sediment at the bed/bed interface (e.g., Ostrem et al. 1967, Fenn 1983, Collins 1979b, 1988). The overall seasonal trend in sediment supply reflects sediment being produced by glacier erosion during the winter period of low or absent runoff and remaining at the ice/bed interface until spring, when it is readily available for erosion and transport by meltwater runoff early in the ablation season.

As the melt season progresses, sediment is less readily available and meltwater less likely to con-

tact it. Because meltwater erosion is more "efficient" than ice erosion, abrasional products may be depleted during the melt season, and, thus, the volume of potentially erodible material adjacent to conduits and tunnels and across major parts of the bed may be exhausted (Collins 1988, 1989). Only through rapid, large volumetric increases in discharge, such as caused by precipitation or catastrophic floods, will the drainage system expand beyond primary conduits and tunnels, and thereby reach previously inaccessible parts of the bed with erodible sediment (e.g., Röthlisberger 1980, Collins 1986). Theoretically, this expansion is temporary and any conduits formed in either the linked-cavity or conduit-tunnel system would be unstable as the discharge decreases following passage of the flood; the drainage system would then return to its previously stable arrangement (Walder 1986).

The nature of the bed is an integral factor in the sediment discharge-drainage system relationship. When the substrate is bedrock, it is a primary but exhaustible source of glacier erosion by-products. The bed acts as a secondary sediment source through its role as a storage medium. When the substrate is sediment it's an erodible and deformable, more or less constant, primary sediment source.

In the situation where a glacier lies on a bed of sediment, supply is apparently less important than drainage system evolution, meltwater supply or the factors determining meltwater erosion rate. The grain size distribution, consolidation history, piezometric pressure and other geotechnical properties determine the deformability and erodibility of sediment beds (Clarke 1987). The nature of the drainage system and the water pressure-water discharge relationship, however, determine the amount of channel or conduit erosion that can take place and the area of the bed from which sediment can actually be withdrawn (e.g., Alley 1989a). The existence of subglacial cavities will be related to the texture and geotechnical properties of the substrate material and control whether a linked-cavity or conduit-tunnel system prevails (e.g., Fowler 1987a, Kamb 1987).

On bedrock substrates, similar relationships exist but other factors and potential feedback mechanisms include 1) bedrock erodibility and thus the potential rate of sediment production, 2) bed roughness, which will affect drainage system type and the amount of potential storage of sediment, 3) seasonal evolution of the drainage system, which may happen more rapidly on bedrock than on sediments owing to existing topographic controls

on drainage, and 4) sediment supply, which may be exhausted rapidly. Where glaciers lie on a highly resistant bedrock substrate with a well-established conduit-tunnel system, linked-cavity system, or N-channels following bedrock structures, areas of the bed may remain inaccessible to runoff for long periods and thus the abrasional products may accumulate. Only through events that flood most if not all of the bed would such areas be evacuated of sediment (Fowler 1987a). Then the drainage temporarily shifts from a distributed film to conduit flow capable of sediment erosion (Alley 1989a) and produces sudden sediment discharges. Glacier surging may represent special conditions of high water pressures that result in large, inaccessible areas of bedrock suddenly being evacuated of sediment (e.g., Humphrey 1986, 1987, Humphrey et al. 1986).

The annual yield of sediment therefore varies with bed composition, glacier dynamics, climate and the nature of the drainage system. On substrates of bedrock, its composition and roughness are directly linked to sliding, cavitation and erosion by ice, and in a less direct way to the drainage system types and their extent across the bed. Annual variations will reflect longer-term storage of sediment in areas not drained repeatedly by meltwater flows. In a similar manner, the climatically controlled annual variations in precipitation and glacier mass balance and the frequency of extremes in precipitation (both high and low) will determine the area of the bed across which a drainage system can exist stably and erode sediment.

On substrates of sediment, supply is not as significant an issue as are the climatically controlled volumes of meltwater and precipitation, and the drainage system morphology and its annual evolution.

For both types of substrates, our lack of knowl-

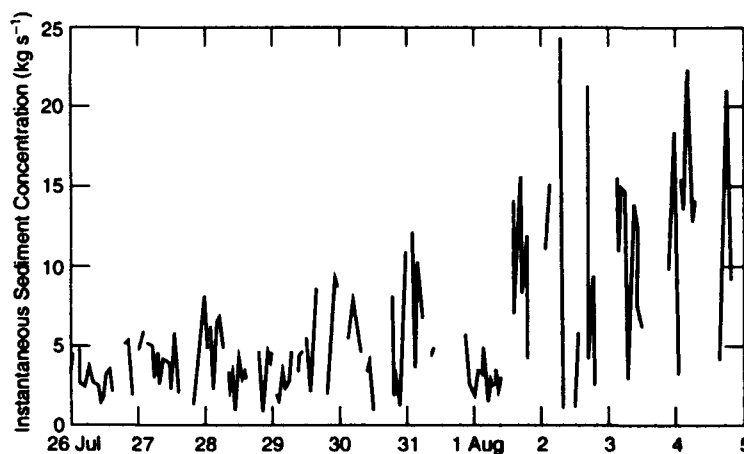
edge concerning glaciohydraulic processes of sediment production and transport and the glaciohydrologic processes of meltwater production, drainage and storage hamper attempts to define quantitative relationships for sediment flux and sediment yield.

SEDIMENT DISCHARGE CHARACTERISTICS

In general, rivers draining glacierized areas transport significantly more solid matter in suspension and bedload annually than nonglacial rivers under their normal flow conditions, with the suspended load increasingly important at distances downstream of the glacial source (e.g., Ostrem et al. 1973, Guymon 1974, Church and Gilbert 1975, Bogen 1986, Gurnell 1987). Guymon (1974), for example, concluded that glacial rivers in Alaska have suspended sediment concentrations of 10 to 20 times greater than those of nonglacial streams. Sediment delivery rates of glaciers and glacial streams are extremely variable, however, from glacier to glacier and from year to year (e.g., Ostrem 1975a, Guymon 1974).

The sediment load is transported within a limited part of the year, with up to 95% transported during the melt season, the duration and timing of which vary with latitude and altitude. In south-central Alaska, for example, sediment is discharged between late May and early October, but between mid-June and late August in northern Alaska. Sediment transport during the winter is severely reduced owing to the seasonal controls on the glaciohydraulic processes supplying sediment to glacial discharges. Few data sets are sufficient, however, to determine if more detailed trends exist and to identify the factors controlling sediment yield.

Figure 63. Diurnal ranges in suspended sediment transport, Gornera stream, Gornergletscher, Switzerland (after Collins 1979a). Fluctuations of over 100% occurred within 1- to 2-hour periods without corresponding large increases in water discharge.



Sediment discharge in rivers draining highly glacierized basins can be extremely variable over hours, days or seasons (Rainwater and Guy 1961; Vivian 1970; Vivian and Zumstein 1973; Ostrem 1975a; Collins 1979a; Bogen 1980; Beecroft 1981; Hagen et al. 1983; Gurnell and Fenn 1984b, 1985; Bogen 1989). Volumes can change rapidly within a period of several hours or less (e.g., Ostrem et al. 1967, Collins 1979a, Tomasson et al. 1980, Beecroft 1983, Humphrey et al. 1986), while a large proportion of the total annual yield may be released over 1 to 2 days (Fig. 63, Collins 1979a). Ostrem et al. (1967), for example, measured a release of nearly 60% of the total annual load in a single 24-hour period. Flash floods can transport large quantities of sediment (Fig. 64), and annual yields are highest in years with multiple floods (Bogen 1989). Thus, depending upon the dates of measurement, sediment rating curves developed from daily totals will differ (Fig. 65 and 66, Ostrem 1975a,b). Similar variabilities can characterize the differences in bed-load, suspended load and sediment yield between adjacent glacierized basins (Fig. 67) (e.g., Gurnell et al. 1988).

Sediment discharge from surging glaciers can also be extreme, varying by an order of magnitude greater than annual sediment concentrations before the surge, yet occur with only a small increase in water discharge (Humphrey 1986, Humphrey et al. 1986). Using the surge of Variegated Glacier (Humphrey 1986) as an analogy, Clarke et al.

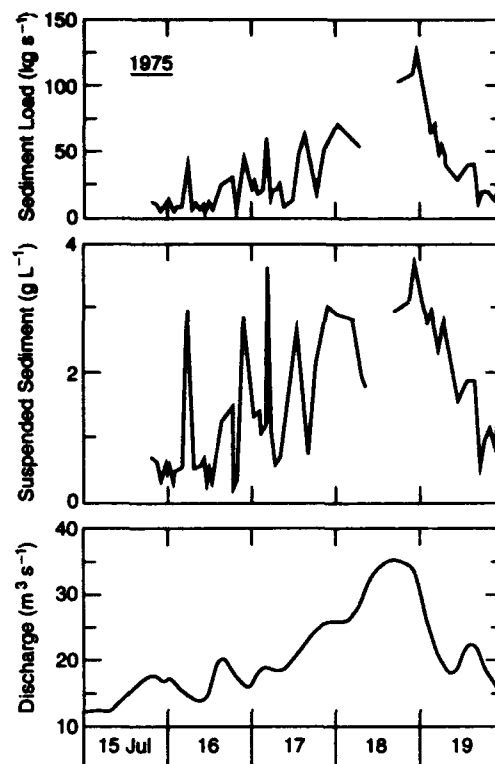


Figure 64. Suspended sediment discharge during the draining of the ice-dammed lake Gornersee through the Gornergletscher, Switzerland, 15–19 July 1975. Rapid, short-term increases in suspended sediment transport characterize the rising limb of the flood hydrograph, while total sediment load peaked prior to peak flood discharge (after Collins 1988).

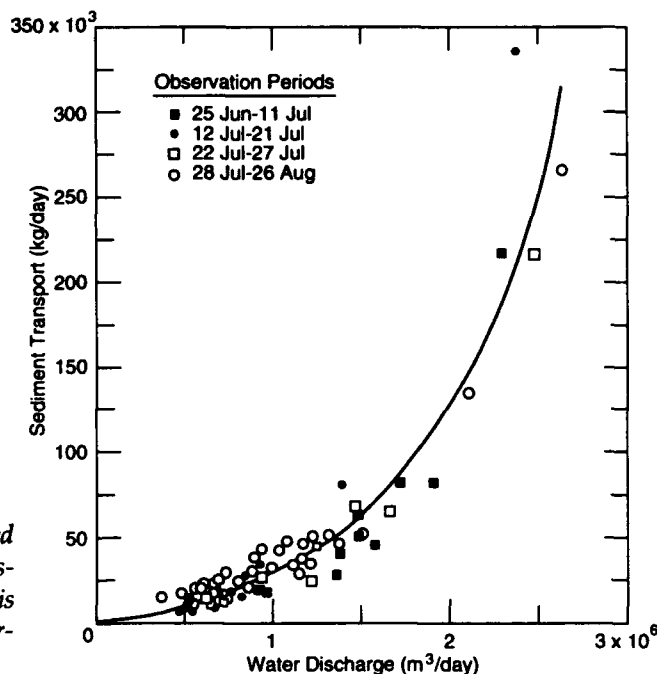


Figure 65. Sediment transport rating curve obtained for a small glacier in northern Norway (Trollbergdalsbreen), based upon daily totals. The observation period is divided into four subperiods, each of them given by different symbols (after Ostrem 1975a).

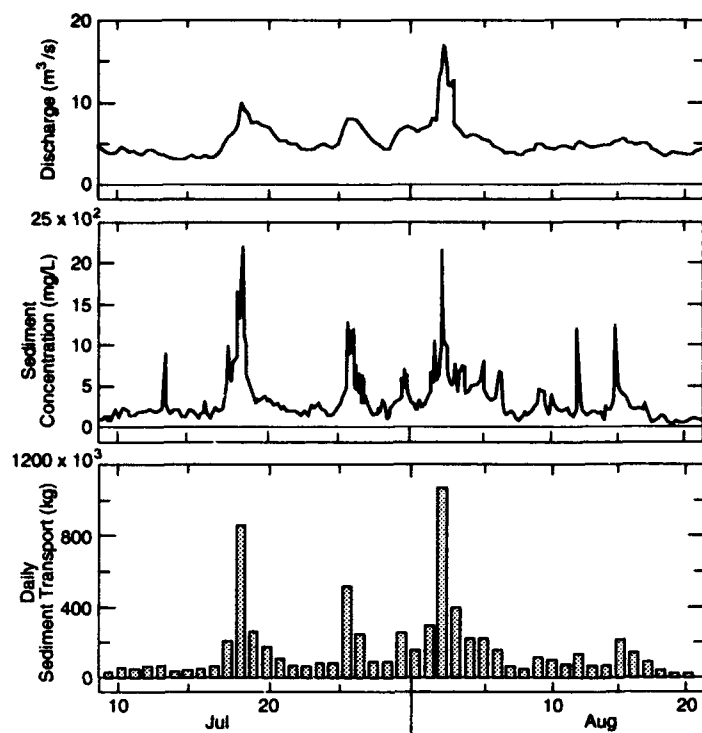


Figure 66. Discharge, suspended sediment concentration and daily suspended sediment transport data for Erdalsbreen glacier, Norway (1967). Hourly and daily variations characterize sediment concentrations (after Ostrem 1975a).

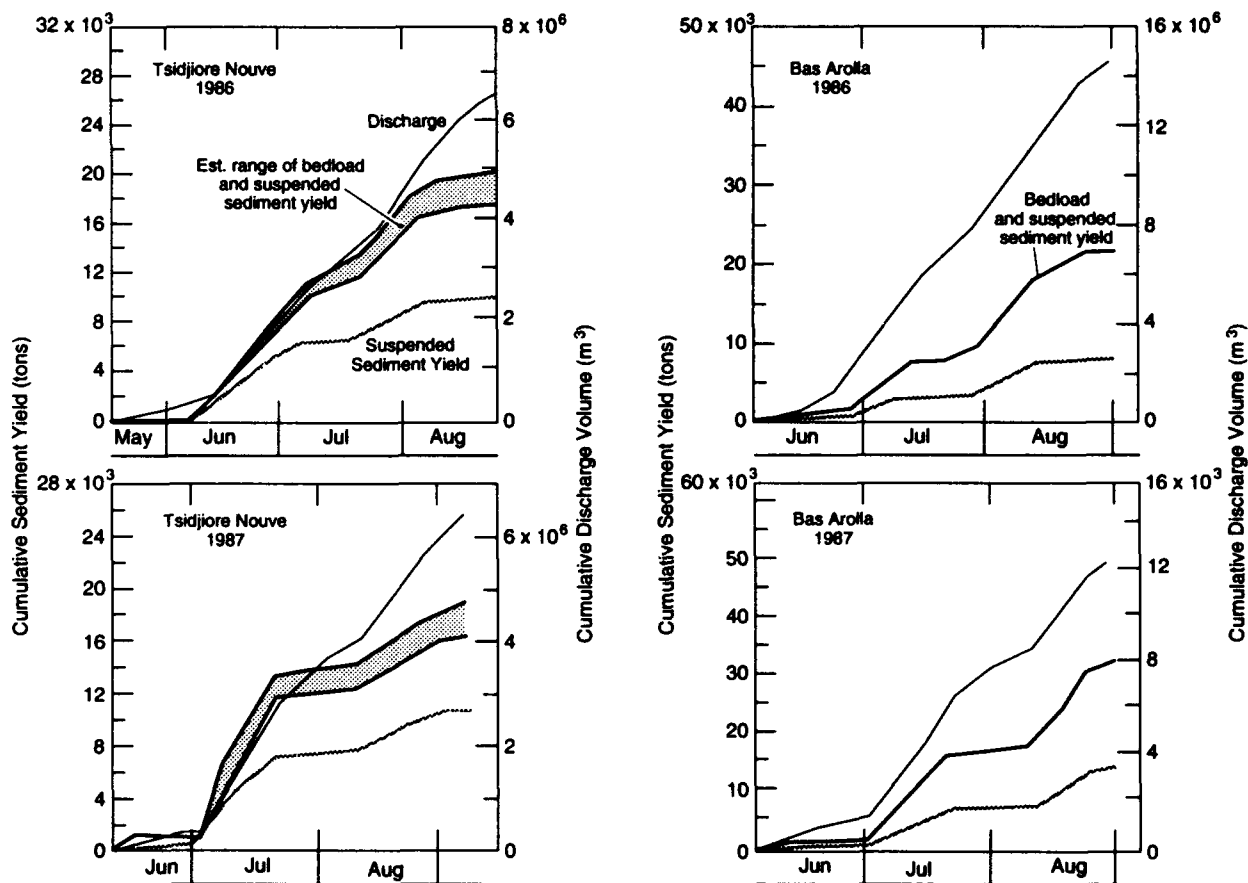


Figure 67. Accumulated weekly discharge, suspended sediment load and bedload yield from the adjacent Tsidjore Nouve and Bas Arolla basins 1986 and 1987 (after Gurnell et al. 1988).

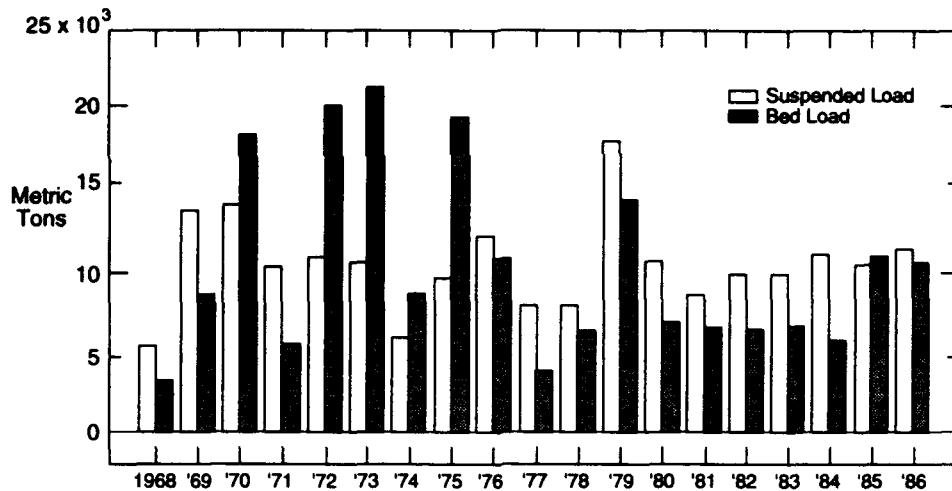


Figure 68. Comparison of annual bedload and suspended load, 1968–86, in the stream draining Nigardsbreen, Norway (after Bogen 1989).

(1986a) predicted that a surge in Susitna Glacier could discharge roughly 30 times the estimated annual sediment load into the proposed Watana Reservoir on the Susitna River, Alaska (R&M1982). Such events and the inherent variability of sediment discharge are strong evidence that aperiodic measurements are probably not representative of a basin's sediment yield over the short or long term.

The inherent variability in sediment discharge from glacierized basins reflects the characteristics of sediment in streams flowing from the ice. At the glacier margin, transport measurements indicate that the proportion of suspended load to bedload varies with the nature of the sediment supply (volume, location) and the internal drainage system (Fig. 68). Ostrem et al. (1973), for example, found that suspended sediment composed from 50 to nearly 80% of the total sediment load at the snouts

of five Norwegian glaciers. Church (1972) reported that suspended load was about 20 to 80% of total load in the Lewis River, Canada, while at Nigardsbreen, Norway, Bogen (1989) found the percentage of bedload to vary between about 30 and 60% of the total load (Fig. 69).

Bedload may only travel several kilometers or less downstream in a year and, at distances further downstream in the basin, it may be indistinguishable from bedload derived from nonglacial sources. The proportion of total load composed of bedload therefore also decreases downstream, with suspended load exceeding bedload by 3 to 6 times within 10 km or so of the glacier, but by up to 100 times at distances of 100 to 200 km from it (Harrison 1986). The latter relationship is illustrated by the Tanana River, at Fairbanks, Alaska, where the suspended load to bedload ratio is 99:1.

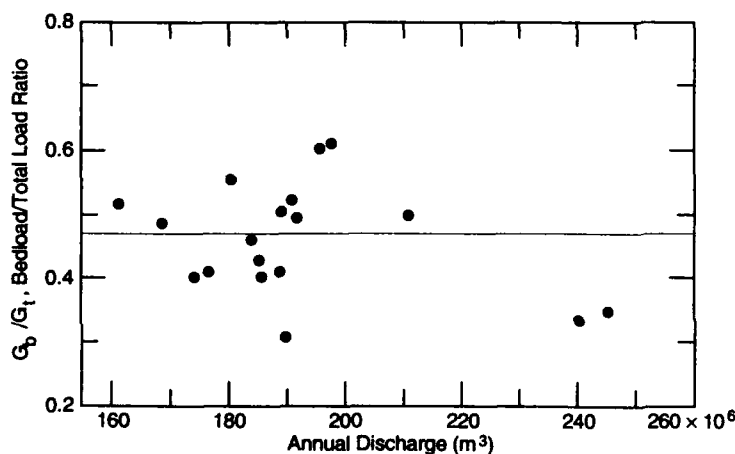


Figure 69. Ratio of bedload to total load in relation to discharge for the river draining Nigardsbreen, Norway (after Bogen 1989).

Quantitative estimates of bedload volume have been attempted by a number of researchers, including Ostrem (1975a), who trapped all bedload material behind a steel fence on the river draining from Nigardsbreen Glacier in southwestern Norway. Over a 3-week period, approximately 400 metric tons of bed material was captured by this net. Bedload in multiple glacierized basins was estimated by Bezingé (1981) to range in volume from 80 to 450 m³/km² per year from gravel traps at the Grande Dixence hydroelectric project. More recently Bezingé et al. (1989) used Grande Dixence sediment trap data to estimate that bedload had an annual mean of 12 to 14 × 10³ metric tons over the period 1977–1987, composing between 25 to 46% of the total load, depending upon the analysis used. When data from multiple basins within one or more regions are compared, bedload volumes vary widely without a consistent pattern (e.g., Church 1972, Laufer and Sommer 1982, Hammer and Smith 1983, Bezingé 1981, Bezingé et al. 1989, Bogen 1986, 1989).

Variations in suspended sediment load at the glacier outlet also characterize glacially fed rivers further downstream, reflecting the sediments' fine-grained nature and ease of transport. Suspended sediment concentrations in waters discharging directly from glaciers typically range from 0.5 to 5

g/L (Fig. 70), with peak or maximum values reported to range from 3.0 to 10 g/L, and over 20 g/L in extreme sediment discharges associated with early season floods or drainage of ice-dammed lakes (e.g., Vivian 1970, Vivian and Zumstein 1973, Ostrem 1975a, Bezingé 1978, Collins 1978, Chernova 1981, Gustavson and Boothroyd 1982, Gurnell 1982, Smith et al. 1982, Hagen et al. 1983, Richards 1984, Parks and Madison 1984, Cowan et al. 1988). Record values of 36 g/L on the Drance River and 150 g/L on the Isère were measured by Parde (1952) as reported by Bezingé (1978). Downstream, away from the ice sources, mean suspended sediment concentrations are typically less than 1.5 g/L (Gurnell 1987), but may reach 3.0 g/L in large Alaskan rivers (e.g., Guymon 1974, Coffin and Ashton 1986). During winter, minimal amounts of suspended sediment are in transport, with concentrations typically ranging from 0 to 0.01 g/L (Parks and Madison 1984).

Given the symbiotic nature of water and sediment in transport, studies have analyzed the possibility of a direct relationship between suspended sediment load and discharge. No simple correlation apparently exists between them, either within a basin (Fig. 71) or in adjacent basins (Fig. 72) (e.g., Ostrem 1975a, Collins 1978, 1988, Gurnell 1987). Maximum suspended sediment discharge can com-

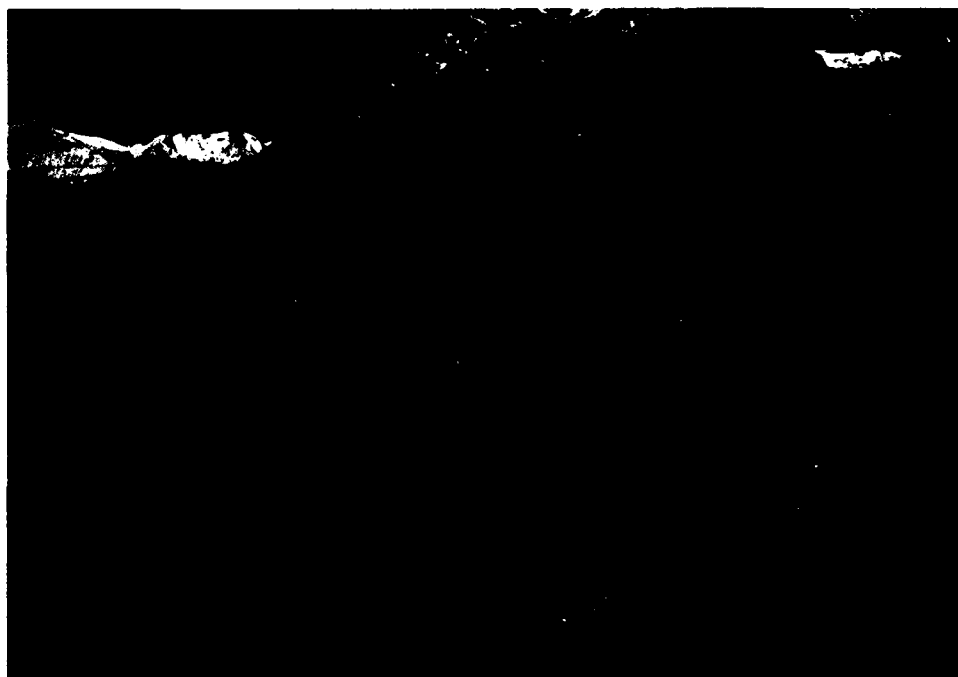


Figure 70. Sediment-laden meltwater in turbulent discharge from Matanuska Glacier, Alaska, in late May 1993. Sediment concentration averaged 2.5 g/L.

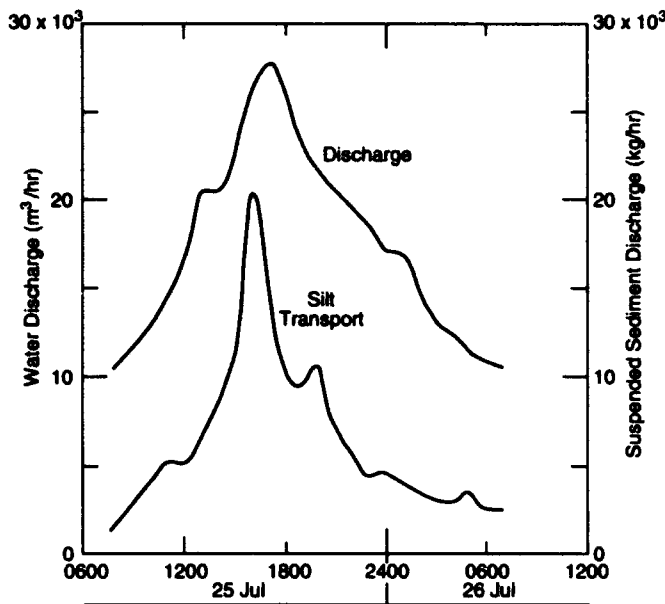


Figure 71. Rapid, ten-fold increase in sediment discharge during a two-fold increase in water discharge at Storsteinsfjell, Norway (after Ostrem 1975a). Peak sediment discharge occurred prior to peak water discharge.

monly take place before maximum runoff, either daily or annually. Over the year, for example, suspended sediment discharge is generally higher early in the melt season and decreases through the season until it is proportionally lower in volume relative to water discharge in the late summer (Fig. 73). This seasonal trend appears to be related to the availability of sediment, which is potentially greater after the low-flow winter when sediments accumulate below the glacier.

Over shorter periods this relationship can differ for the rising and falling limbs of the discharge hydrograph (Fig. 74) (Ostrem 1975, Collins 1979a, Church and Gilbert 1975). Plots of sequential sediment discharge versus water discharge measurements with this relationship can display a hysteresis or loop to the curve (Fig. 75). A clockwise hysteresis means that there are greater sediment concentrations on the rising limb of the diurnal hydrograph than at equivalent discharges during the falling stage. Ostrem (1975a) found hysteresis effects several times through the summer melt season for periods lasting several days or more in conjunction with abnormally high discharge events (Fig. 75 and 76). Clockwise hysteresis loops that included strong involutions were also found to recur daily by Collins (1979a) during extended periods with high rates of ice ablation and snowmelt (Fig. 77). Involutions reflect irregular variations in sediment concentration, which Collins (1979a) interpreted as changes in sediment avail-

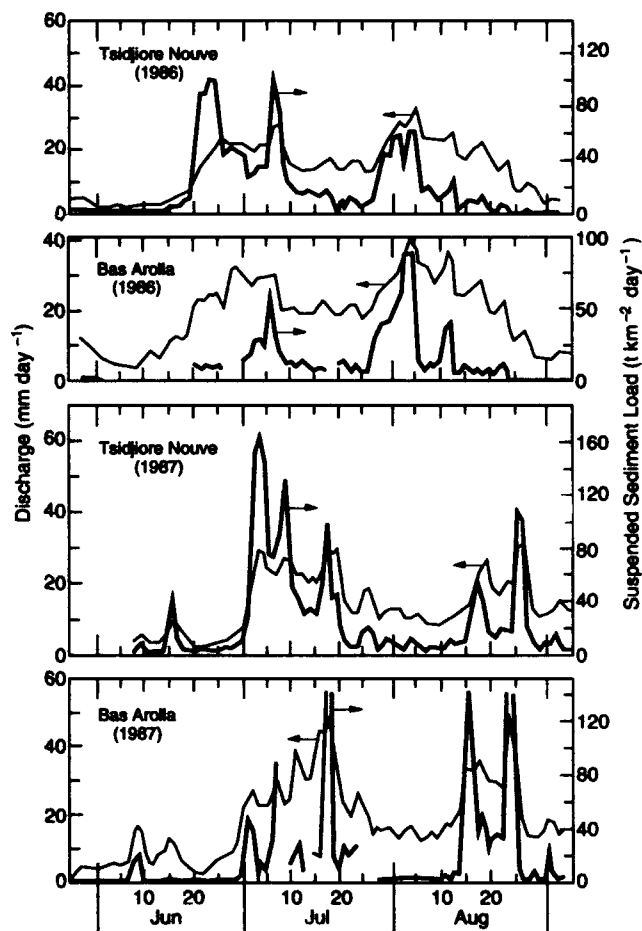


Figure 72. Variations in daily suspended sediment yield and discharge through the melt season for two adjacent glacierized basins in the Swiss Alps (after Gurnell et al. 1988).

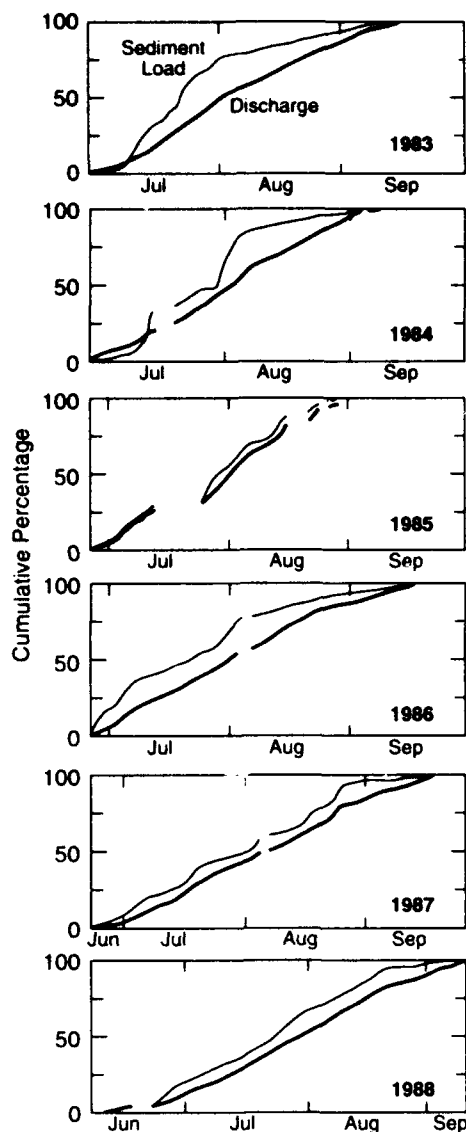


Figure 73. Variability in the cumulative sediment load and water discharge for the years 1983–1988 at Gornera Stream, Gornergletscher, Switzerland (after Collins 1990). Large volumes of sediment may be released without increases of similar magnitude in water discharge.

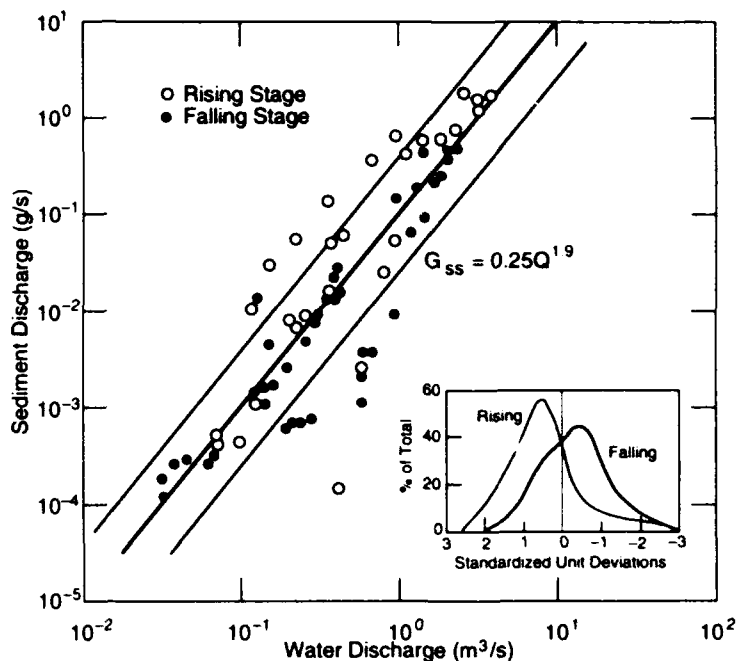


Figure 74. Suspended sediment vs. discharge relationship for Decade River, Baffin Island. Rising and falling stages clearly differ in terms of sediment discharge (after Church and Gilbert 1975).

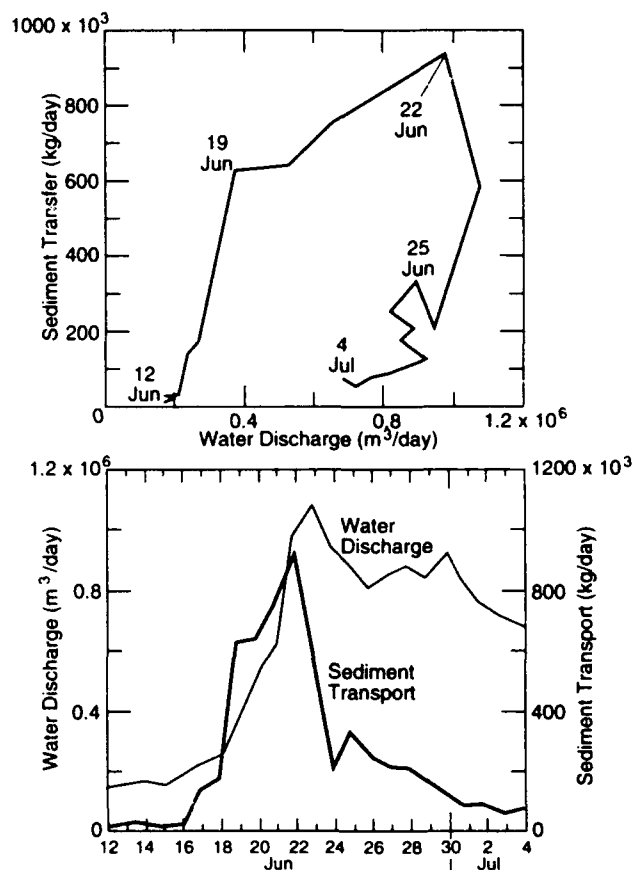


Figure 75. Relationship of daily suspended sediment transport to discharge during a 1969 flood at Erdalsbreen, Norway (after Ostrem 1975a). Clockwise hysteresis loop is characteristic of a peak in sediment discharge that culminates before water discharge.

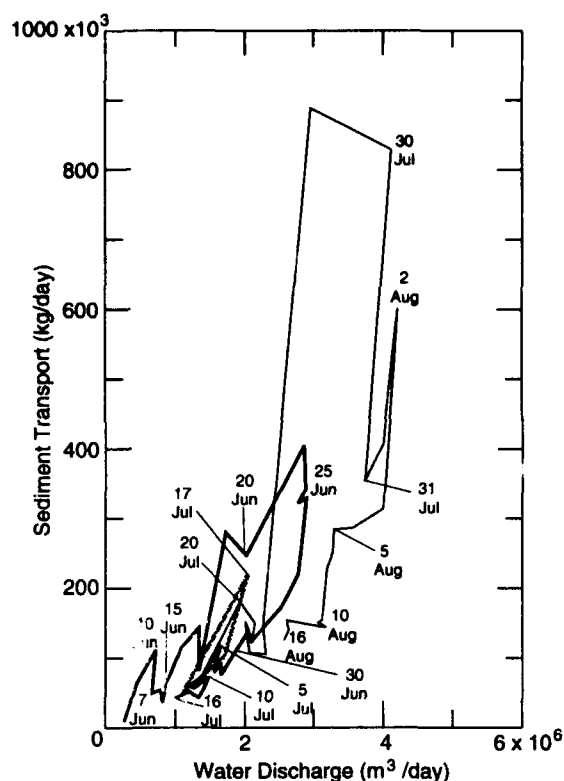


Figure 76. Hysteresis loops characterizing periods of discharge over several days during the 1969 melt season at Nigardsbreen, Norway. Each loop is associated with a period of high water discharge (after Ostrem 1975a).

ability within or at the bed of the glacier. Sediment supply problems may also be responsible for hysteresis effects over longer periods following major meltwater discharge events (Gurnell 1987). Potentially, discharge vs. suspended sediment relationships could be developed on the basis of different parts of such hydrographs; however, Ostrem (1975) and Felli (1989) concluded that such an analysis would only be valid during that year and could not be applied to out-years, even in the same basin.

Sporadic peaks in sediment discharge that are unrelated to flow are believed to reflect discrete inputs of sediment into the drainage system. These sudden but short and irregular increases in suspended sediment concentrations are common in streams at the glacier margin (e.g., Rudolph 1962, Mathews 1964a, Ostrem 1975a, Collins 1979a, Bogen 1980, Gurnell 1982, Gurnell and Warburton 1990, Johnson 1991b). The erratic nature of these sediment bursts may result from subglacial channel migrations that capture a new supply of sediment (Collins 1979a, 1988) or from glacier motion that also establishes new sediment sources for englacial or subglacial flows (e.g., Humphrey et al.

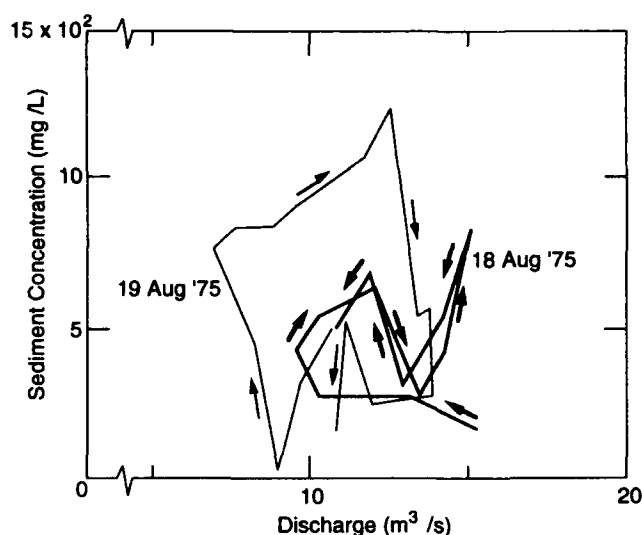


Figure 77. Diurnal hysteresis loops with involutions for suspended sediment concentration as a function of discharge plotted sequentially with time for the Gornera stream, Gornergletscher, Switzerland (after Collins 1979a).

1986). Proglacial processes tapping different sediment sources, such as unvegetated moraines (e.g., Bogen 1980, Hammer and Smith 1983, Richards 1984), may result in sediment bursts further downstream. Sediment input may also result from reworking of sediment on moraines or moraines with ice cores (Fig. 78). Melting of buried ice induces instability, erosion and transport of sediments by various mass movement processes and meltwater streams into the main river draining the glacier (e.g., Lawson 1988). As with meltwater, sediment is effectively stored within and at the glacier bed because not all of the bed is hydrologically interconnected, nor are englacial and subglacial discharges in the smaller arterial conduits necessarily capable of entraining all available materials (e.g., Collins 1988, Bogen 1989).

Currently, limited data indicate that total sediment load or annual sediment yield reflects shorter duration variability in sediment discharge; quantities of 10^3 to 10^8 metric tons/year have been measured in basins of various size and percentage of glacier cover (Gurnell 1987). Annual sediment yields summarized by Church and Gilbert (1975)

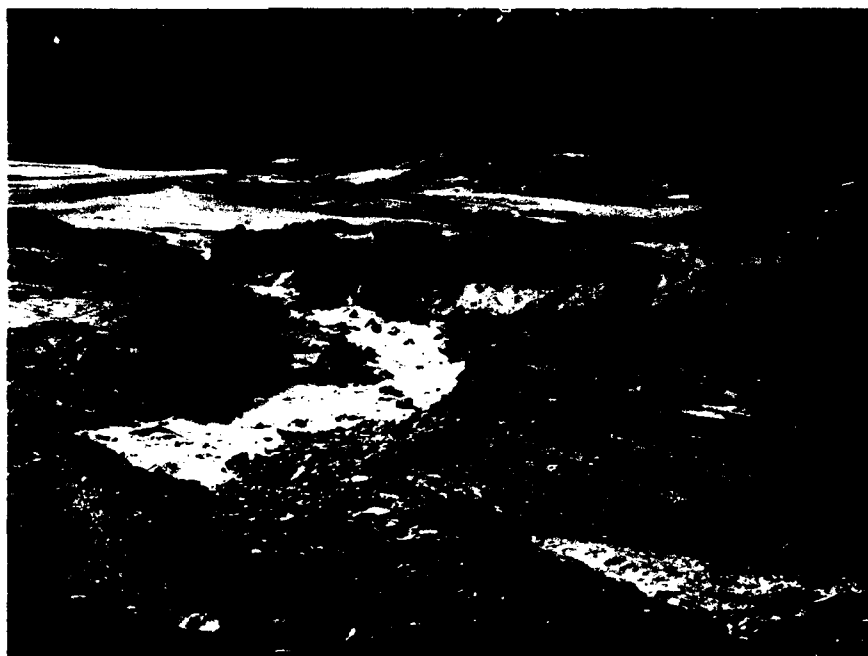
ranged from 140 metric tons/km² per year to over 1900 metric tons/km² per year, with yields from recent volcanic terrain as high as about 4000 to 8000 metric tons/km² per year. Ostrem (1975a) presented 5 years of data from streams draining five Norwegian glaciers, where the average sediment yields ranged from 414 to 1545 metric tons/km² per year, but varied by ± 25 to over $\pm 40\%$ (Ostrem 1975a,b). Sediment data from large regional catchments with multiple glacierized basins across Alaska had sediment yields ranging from about 2 to 60 metric tons/km² per day (Parks and Madison 1985).

Bogen (1989) also reported annual sediment loads ranging from as little as 3 to 10,400 metric tons/year at 14 glacier-fed intakes for the Breheimen-stryn power plant in Norway. Data from Nigardsbreen, Norway, for the period 1968 to 1986 indicate large variations in total load from year to year (10,000 to 40,000 metric tons/year) (Bogen 1989). Mean annual sediment yield of the glaciers across Norway ranged from a highly variable 100 to 1300 metric tons/km² per year (Bogen 1989). In contrast, Bezinge (1978) found a much higher maximum sediment yield of about 6000 metric tons/km² per year at Gornergletscher, Switzerland, a

glacier of similar dimension to those in Norway. Guymon (1974) analyzed records for the Knik Glacier basin, which contains a much larger glacier and higher percentage ice cover than either Gornergletscher or Nigardsbreen; annual sediment yield here is over 20,000 metric tons/km² per year.

While the basic factors affecting sediment yield have been identified, their role in determining annual trends have not been quantified and studies suggest that relationships are not straightforward (e.g., Ostrem 1975a). Sediment yield is affected by such factors as 1) basin geology and thus the erodibility of materials beneath the glacier; 2) climate and its effects on mass balance and runoff; 3) glacier dynamics and thermal regime and their effects on erosion and sediment entrainment by ice; 4) the percentage of glacier cover and its state of change; and 5) the type and extent of the subglacial drainage system in relation to the location and quantity of sediment available for transport. Currently, data on sediment discharge are representative of only a select few of the Earth's glacierized basins (e.g., Gurnell 1987), and thus they are insufficient to assess the magnitude of long-term fluctuations in annual sediment yield in either basins or regions (Church and Gilbert 1975, Harrison 1986).

Figure 78. Ice-cored moraine next to the active margin of Matanuska Glacier, Alaska. Ice core is exposed in approximately the center of the photograph, above the stream. Re-working moves previously deposited sediments, which originated from basal zone debris, into streams feeding the main river (upper left) and can cause sporadic increases in sediment concentration downstream (see people for scale).



Part 3. Models of Runoff and Sediment Yield

INTRODUCTION

Two types of models are used to predict runoff or sediment yield from basins or watersheds: statistical or physical (conceptual). Statistical models use hydrological and meteorological data to derive mathematical relationships for various parameters of interest. Such models are typically used to forecast maximum or recurrent floods for long-term planning or design. Through quantitative statistical analyses, relationships are derived that attempt to link hydrological parameters to various basin characteristics (e.g., elevation, climatic regime, basin area). However, predictions are difficult, if not impossible, to apply regionally to ungauged basins based simply upon similar, comparative data; they generally require additional data measured in the basin of interest. Statistical models are limited in application because they are not tied directly to the processes determining basin hydrology and, further, they rely heavily on long-term data records that often do not exist for glacierized catchments. Several basins within Norway and Switzerland are general exceptions (e.g., Ostrem and Wold 1986, Röthlisberger and Lang 1987).

Physical or conceptual models are formulated to simulate the actual processes determining basin hydrology and hydraulics. Physical models of the hydrological system, therefore, require knowledge of the mechanisms and controls on basin runoff and sediment yield, which in many instances are poorly understood. They, too, require data records for evaluating parameter relationships. Certain physical models (e.g., Lang [1980] discussed below) use statistical relationships between various basin parameters to develop or simulate physical process relationships, particularly when processes are complex and not sufficiently understood to model directly. Physical models can provide de-

tailed short-term predictions (weekly, daily or even hourly) if appropriate near-real-time data are available. Such models are designed with the intent that they can be applied to ungauged basins. Problems with physical models typically come from limited knowledge of the processes and controls on basin-watershed hydrology and hydraulics; this limitation is particularly true for glacierized basins.

In this part, models specific to partly glacierized basins are briefly reviewed. Other models of runoff or sediment yield that are not specific to glacierized conditions may be applicable with appropriate modifications for the glacial effect, but they will not be considered in detail in this monograph. An example of a nonglacierized physical model for basin runoff that is widely used for predicting snowmelt runoff is SSARR (USACE 1972, Speers et al. 1979, Cassell and Pangburn 1991).

STATISTICAL RELATIONSHIPS

The traditional statistical approach of using empirical equations derived from multiple regression techniques has been applied to partly glacierized basins ranging in area from several to 1000 km² or larger in a variety of locations globally. In the examples described below (and others not treated here), the statistical approach requires data records of 5 or more years, but preferably longer-term, multi-year data sets of 30 years or more. The models are in general basin-specific; they require calibration for each basin under consideration. The high variability of runoff and sediment yield in glacierized catchments makes precise short-term as well as long-term statistically generated forecasts difficult.

An inherent weakness in statistical treatments is the assumption that the variables (discharge, sediment concentration, etc.) evaluated empirical-

ly have a linear or log-linear relationship; this assumption is not generally warranted for mountainous glacierized basins (e.g., Tarar 1982, Meier 1983, Ferguson 1985, Fenn 1989). Further, relationships among the various independent and dependent variables are incompletely understood, so that errors in regression estimates cannot be accurately assessed (Parks and Madison 1984).

The general form of the multiple linear-regression equations used in statistical relationships is

$$y = a + b_1x_1 + b_2x_2 + \dots + b_nx_n \quad (101)$$

where y = hydrologic characteristic

a = regression constant

b = regression coefficients

x = independent basin variables

n = number of basin characteristics.

To account for the generally nonlinear behavior of flow characteristics relative to the independent variables, variables are transformed to logarithms (e.g., Benson and Carter 1973). The general form of the log-linear relationship is

$$\begin{aligned} \log y = \log a + (b_1 \log x_1) \\ + (b_2 \log x_2) + \dots + (b_n \log x_n) \end{aligned} \quad (102)$$

or the equivalent exponential expression

$$y = ax_1^{b_1} x_2^{b_2} \dots x_n^{b_n}. \quad (103)$$

Stepwise regression analyses are commonly performed one by one to determine the effects of independent variables and to develop the best relationships.

Accuracy or reliability is commonly defined by a statistical function such as the correlation coefficient (r), which expresses the degree of correlation between two (or more) variables, and its square, which is the coefficient of determination (r^2), indicating the goodness of fit of the data to the equation ($r^2 = 1.0$ would be a perfect fit). A standard error function defines the variation of actual data from the predicted values. The standard error, for example, is an estimated limit within which about two-thirds of the actual hydrologic characteristic is expected to fall. The coefficient of variation CV , which is equal to the standard deviation of a data set divided by the arithmetic mean, is often used for comparison of daily, monthly or annual variations in runoff.

Predicting runoff

Parks and Madison (1984) analyzed hydrologic data from major rivers draining both glacierized and nonglacierized basins in six hydrologic regions that cover the entire state of Alaska. In these multiple linear regression analyses, they examined the relationship of stream flow to several independent variables in the categories of 1) physical and climatic characteristics of the basins, 2) channel width, 3) mean annual flow and 4) 2-year peak flow. Physical and climatic characteristics of each basin are those routinely analyzed by the USGS and include drainage area, channel slope, channel length, mean elevation, water storage, forest cover, glacier cover, mean annual precipitation, mean annual snowfall, mean minimum January temperature and intensity of precipitation. Hydrologic factors include instantaneous flood peak magnitude, mean annual discharge, low-flow flood volume and flood surface flow width.

Parks and Madison (1984) found that only two basin characteristics—drainage area and precipitation—gave usable estimates of stream flow on the basis of the statistical significance of the log-linear relationships for mean annual flow, peak flow, flood volume or low flow (Fig. 79). Considering the percentage of glacier cover did not significantly improve estimates, either regionally or state wide. While reliability of the equations varies from region to region, regional estimates using regression equations did not significantly differ in error from the state-wide estimates. In addition, they found that peak flow estimates had no apparent relationship to percentage of glacier cover.

Several problems exist with the generalities of the Parks and Madison (1984) study, which are especially applicable for comparing glacierized and nonglacierized basins. Data are particularly problematic; the paucity of data for Alaskan rivers, the large regions and river basins under consideration, and non-representative data sets reflecting large regional differences in stream type and basin character clearly affect the significance of the results. In addition, the locations of gauging stations are far from the glacierized parts of the basins and therefore the apparent effects of glaciers on runoff and sediment yield are misrepresented (e.g., Borland 1961, Guymon 1974).

Parks and Madison concluded that the lack of data, their bias in time and their areal distribution compromise the reliability of estimates of stream flow, so that use of the derived relationships may be highly questionable. For example, data sets of individual factors ranged in number of samples

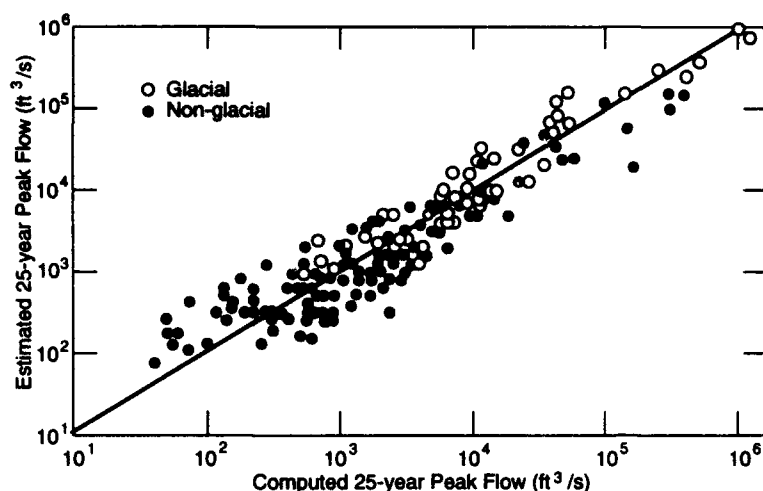


Figure 79. Example of regression analyses of runoff by Parks and Madison (1984). Actual (computed) 25-year peak flow versus that estimated by the basin characteristics equation for glacial and nonglacial streams (after Parks and Madison 1984).

from as few as 9 to as many as 200, but were mainly 30 to 70. A standard error of prediction could not be computed. There is also considerable scatter to the data, even when they are plotted in log-log format, and, in fact, the goodness of fit suggested by the r^2 values (range 0.75 to 0.99) is very likely an artifact of the limited number of hydrologic and climatic data. Currently, there are no active Alaskan stations that record stream flow or water quality daily (Parks and Madison 1984), data that are required for improving the reliability of these statistical estimates. The regression relationships for stream flow parameters presented by Parks and Madison (1984) should therefore only be considered a rough, first approximation of runoff for stations far removed from the glacierized parts of a basin.

Lamke (1978) had previously evaluated Alaskan data using only two hydrological regions (maritime and interior) and a multiple regression analysis of flood peak magnitude and flood frequency at various recurrence intervals. He examined the same set of basin characteristics as used by Parks and Madison (1984) for 260 sites. Lamke applied a log-Pearson distribution that suggested that peak flows could be related to mean annual precipitation, drainage area and water storage (lakes and ponds) in the maritime region and, in addition, to the mean January temperature and forest cover in the interior region.

Lamke's flood frequency analyses suffer from the same basic problems of reliability and apparent accuracy that plague Parks and Madison's (1984) analysis. The effect of glacier cover was not analyzed specifically, although streams affected by glacier outburst floods were eliminated from the analysis. The relatively high standard errors, particularly for floods with a recurrence interval

of 25 years or longer, is in large part ascribable to the lack of longer-term data sets. A 100-year flood recurrence interval, for example, could not be defined for either region owing to the lack of data.

Ashton (1984) considered the effects of the short Alaskan data records on estimated flood magnitudes. Various recurrence intervals were analyzed using subsets of data extracted from 30-year records at nine gauging stations. He applied a log-normal distribution to compute flood magnitudes at recurrence intervals of 1.25 to 100 years. As expected, Ashton found that 10-year records provide estimates of 100-year flood magnitudes that were closer to the 30-year record than 5-year records, but with a highly variable range of 8 to 50% deviation. A 22-year record is considered the shortest necessary for reliably estimating 100-year floods. It should be noted that Lamke's (1978) analysis had to use data from 54 stations with less than 10 years of record.

Chapman (1982) analyzed daily flow records from 57 streams in Alaska to develop mean daily flow hydrographs and relationships indicating when particular flows will be exceeded 10 and 90% of the time. A log-normal distribution was assumed for the daily data. Glacier-fed streams were treated the same as nonglacierized streams. The results are presented as representative mean hydrographs with 10 and 90% recurrence intervals. While glacierized and nonglacierized basin streams may show significant differences, such as multiple secondary peaks or much higher discharges, Chapman did not analyze their origins and the hydrographs appear difficult to use for predicting runoff.

Mayo (1986) analyzed annual runoff from glacierized basins in relation to the apparent Equilibrium Line Altitude (ELA) (mass balance = 0). His

analysis is based upon mass balance data, ELA measurements and runoff records from three basins in different regions of Alaska. This analysis indicated an inverse relationship between annual runoff rate (meters/year) and the ELA. Mayo developed a graph of this relationship to define the glacier and nonglacier runoff components, assuming that it was generally valid for Alaskan glacierized basins. From these values, average annual runoff for glacier-covered and open areas is defined.

Runoff R is calculated from

$$R = \frac{N S_n + G S_g}{S_n + S_g} \quad (104)$$

where N = nonglacier alpine runoff rate

S_n = nonglacier alpine area

G = glacier runoff rate

S_g = glacier area.

Mayo (1986) applied this relationship to the gauged, glacierized Knik River basin. Predicted average annual runoff rate (2.0 m/yr) was in remarkable agreement with measured rates for a 24-year record and suggested its may be applicable to other partly glacierized Alaskan basins. Mayo, however, did not state whether outburst flooding, which was common in this basin during the years of record, was considered. The ELA method was then applied to the ungauged Bering Glacier basin, where the estimate of annual glacier runoff rate (3.8 m/yr) accounted for an estimated 79% of total runoff.

As Mayo (1986) stated, the relationship of annual ELA to annual runoff has not been demonstrated, although he estimated a 0.2-m/yr accuracy for this technique. He suggested that it be restricted to glacierized basins within the 60 to 70°N latitude band because its application north and south of this band would over- and under-estimate runoff there respectively. An advantage of this method, if the relationship holds, is that the ELA and percentage glacier cover in the basin are parameters that can be estimated from topographic maps and remotely sensed images.

Bjerklie and Carlson (1986) evaluated multiple regression techniques for defining the daily melt component of runoff from glacierized basins, using about 20 years of data from the Phelan Creek basin containing Gulkana Glacier, Alaska (61% of total area). Regression analyses of daily discharge and temperature for June, July and August of 1977 indicated that the best parameter relationship ex-

isted with the log of the previous day's stream flow and the average temperature, similar to the findings of Lang and Dayer (1985).

Using the 1977 data relationships, Bjerklie and Carlson compared model results to data records for 1969 to 1977, finding that the regression relationship estimated annual melt volumes with an accuracy similar to other methods (e.g., Mayo and Trabant 1984). However, the timing and magnitude of meltwater peaks were commonly over-predicted, and the effect of rainfall could not be defined by this relationship. Sensitivity evaluations also indicated that the indexing of melt with temperature is basin-specific. Daily stream flow predictions, based upon the log of the previous day's flow, were considered reasonable estimates for daily forecasting.

Collins (1987) analyzed discharge records from rivers draining five heavily glacierized Alpine basins (about 36 to 68% by area) in the Rhone River area of the Swiss Alps to assess the long-term relationship of runoff to climate in the hope that accurate runoff predictions could be made without glacier mass balance data. Mass balance data are costly, difficult to obtain and generally not available. Ultimately, such a relationship might allow prediction of the effects of future climatic change on long-term variations in runoff.

Multiple regression analyses of climatic and hydrological records spanning the period of 1922 to 1983 were made for five basins ranging in area from about 39 to 778 km², with maximum elevations ranging from 3634 to 4634 m. Meteorological data were obtained from three stations, but primarily from two, one within the largest catchment at Zermatt and the other at Sion within the lower Rhone River valley. Hydrological records were from gauging stations within the lower reaches of each basin. Hydrological and climatic analyses were then compared to define temporal and spatial patterns within and between basins.

Runoff records show that annual flows behave synchronously during maximum and minimum discharge years, with a high degree of correlation between basins. Runoff varied from the mean by -25 to 30%, with up to 50% deviation in the basin with the largest glacier cover (Zermatt basin). Mean annual air temperatures from May-September also show considerable, yet parallel, year-to-year variations at each station.

Correlation coefficients for discharge and hydrometeorological variables showed that, while precipitation and discharge were only weakly correlated, mean air temperature (May-September)

is strongly related to discharge, especially in the more heavily glacierized basins. The goodness-of-fit of multiple regression equations for predicting annual runoff trends was highest using mean annual air temperature and was improved by the addition of winter precipitation totals. Collins (1987) concluded that this relationship provides a reasonable estimate of potential runoff changes in heavily glacierized basins during changes in climate, but that to improve predictions, it will require conceptually based models that consider the effects of radiation and dimensions of the ablation area.

It is important to note that Collins' (1987) analyses indicated that a 1°C shift in the 10-year mean annual temperature resulted in a significant change in the 10-year mean runoff. Cooling by 1°C in the 1960–1970 period reduced flow by about 25% and resulted in glacier expansion. Warmer periods from the early 1940s to the early 1950s, which were associated with a reduction in glacier cover, produced markedly higher than normal annual runoff. Clearly, the potential effects of climate on glacier mass must be considered when statistically evaluating periods of record.

Predicting outburst floods

Predicting extreme floods for glacierized catchments is limited by relevant data. Clague and Mathews (1973), however, developed a purely empirical relationship based upon analysis of selected case histories for estimating maximum or peak discharge during glacial outburst floods (jökulhlaups). This relationship, which produces remarkably reasonable estimates for a variety of situations, simply equates peak discharge ($Q_{\max}$ [m^3/s]) to lake volume (V_o [m^3]) at the start of flooding as

$$Q_{\max} = 75 (V_o/10^6)^{0.67}. \quad (105)$$

While not a strictly statistical analysis, Clarke (1982) improved predictions of $Q_{\max}$ by considering other variables that affect discharge (such as reservoir geometry). He also added a physical treatment of the problem by examining thermal effects using the theoretical model of Nye (1976) for jökulhlaups at Grimsvötn, Iceland. Clarke's analysis considered the role of thermal energy in tunnel enlargement by melting or its closure by ice creep following the beginning of the flood. Simplifying his model is necessary for all field applications as it requires simulation of discharge for various values of the Manning roughness coefficient

along the outlet path, and various data such as lake temperature and ice flow law parameters. Because the location and general timing of floods from ice-dammed lakes may be known, several of the physical parameters can be measured in advance when accurate predictions of $Q_{\max}$ are required.

For lake temperatures near 0°C, tunnel enlargement by melting is minimal and the predicted peak discharge $Q_{\max}$ is estimated by

$$Q_{\max} = \frac{V_o [\rho_w g z_w (t/t_o)]^{11/6}}{N^{1/2} (\rho_i L')^{4/3}} \quad (106)$$

where $N = (4\pi)^{2/3} \rho_w g n'^2$ (a constant defined by Nye for a circular tunnel)

$K_o = 2B3^{(n+1)/2}/n^n$ (also a constant related to flow law as defined by Nye) are assumed and

z_w = lake surface elevation above tunnel outlet

V_o = lake volume

n = flow-law exponent, generally 3

B = flow-law constant

n' = Manning roughness coefficient

L' = effective latent heat of fusion for ice

l_o = length of drainage tunnel

ρ_i = density of ice (900 kg m^{-3})

ρ_w = density of water (1000 kg m^{-3})

g = acceleration of gravity

s = distance along drainage tunnel.

Where the initial lake water temperatures are significantly higher than 0°C and thermal energy is dominant in tunnel melting, the relationship becomes

$$Q_{\max} = 0.424 \left[\frac{k_w (\theta_{\text{LAKE}} - \theta_i) V_o}{\rho_i L'} \right]^{4/5} \left[\frac{[\rho_w g z_w (t/t_o)]^{21/50}}{N} \right] \left(\frac{2\rho_w}{\pi^{1/2} h} \right)^{16/25} \quad (107)$$

where θ_{LAKE} = lake temperature

θ_i = ice temperature

η = viscosity of water.

Clarke (1982) illustrates application of these relationships for an estimated $Q_{\max}$ using an outburst flood case history.

Chapman (1986) has suggested a potentially simplified forecast procedure for outburst flood

hydrographs once flooding has begun. Although the recurrence interval of such floods may be known, the precise date and time when they will begin cannot be predicted. Therefore, Chapman (1986) presents a method to predict the hydrograph for the remainder of the flood to permit downstream predictions of its severity. Analyses of data from 14 glacier-dammed lake outbursts on the Snow River suggested that the cross-sectional area of the lake outlet is proportional to the amount of water that flows through it by

$$Q = CA (2gh)^{1/2} \quad (108)$$

where Q = discharge (m^3/s)
 C = coefficient of discharge
 A = cross-sectional area of the orifice (m^2)
 g = 9.8 m/s^2
 h = head (m) (i.e., lake depth).

Although the actual tunnel cross-sectional area cannot be measured during the event, it will be established early in the break-out. Then it can be estimated from past records of lake level changes, which are correlated with a current aerial survey of the outburst lake level and with discharge data from a downstream gauging station. Iterative calculations then define the best fit of the present and past data to forecast the flood hydrograph.

Predicting sediment discharge

The relationship of sediment yield to discharge has been evaluated statistically for a number of glacierized catchments. Suspended sediment rating curves have been developed by Borland (1961), Mathews (1964a), Church (1972), Guymon (1974), Ostrem (1975a), Bezing (1978), Collins (1979a), Bogen (1980, 1989), Parks and Madison (1984), and Fenn et al. (1985), among others. In general, variations in sediment concentration and water discharge did not correlate well, reflecting the lack of an apparently simple relationship between instantaneous suspended sediment concentration and discharge, and between total sediment yield and total discharge (e.g., Church and Gilbert 1975, Ostrem 1975a, Collins 1979a, Fenn et al. 1985). Examples follow.

Collins (1979a) analyzed suspended sediment concentration and discharge in the Gornera stream draining Gornergletscher near Zermatt, Switzerland. Suspended sediment concentration to discharge relationships in the log-linear form of $S_s = aQ^b$, where S_s is suspended sediment concentration, Q is discharge, and a and b are best fit estimated parameters, were evaluated for hourly data

from 1974. The range in a and b values and poor degrees of fit given by low r^2 values indicated that the relationship of sediment concentration to discharge varies between the rising and falling limbs of the diurnal hydrograph. Suspended sediment in transport also fluctuated by as much as 100% in only 1- or 2-hour periods. Considerable inaccuracy therefore resulted from sediment load calculations based upon the assumed log-linear relationship (Collins 1979a).

Richards (1984) used a multivariate regression analysis to develop short-term sediment rating curves for the Storbregrova stream draining the Storbreen glacier in Jötunheimen, Norway. As Collins (1979a) found, there were higher sediment concentrations on the rising limbs of the discharge hydrograph. Improved explanation of variance in the relationship between sediment concentration and discharge could be obtained if changes in sediment concentration with discharge were analyzed by separate, multivariate rating curves that depend on the rate of change of runoff and the storage of sediment within the immediate proglacial area. As with other attempts at defining sediment-discharge relationships, these multiple regression analyses had low degrees of fit (r^2) and thus questionable predictive capability.

Parks and Madison (1984) also evaluated suspended sediment relationships for the regionalized Alaskan "basins" that were described previously in the section on statistical runoff analyses. Suspended sediment data exist for over 850 sites in Alaska, but most measurements were taken intermittently and most sites had fewer than 20 samples for periods of record of 1 to 13 years between 1948 and 1984. Of these sites, they picked 60 for suspended sediment analyses on the basis of record length (greater than 5 years), coincident discharge data, and absence of adverse human disturbance in the basin. Basin characteristics were those available from USGS files; 31 of the streams were in glacierized basins.

Parks and Madison (1984) estimated mean suspended sediment discharge from the instantaneous sediment transport and daily flow records following the relationship

$$Q_s = 0.0027 Q_w C \quad (109)$$

where Q_s = instantaneous sediment discharge
 Q_w = instantaneous water discharge
 C = instantaneous suspended sediment concentration.

Daily means and longer-term mean suspended

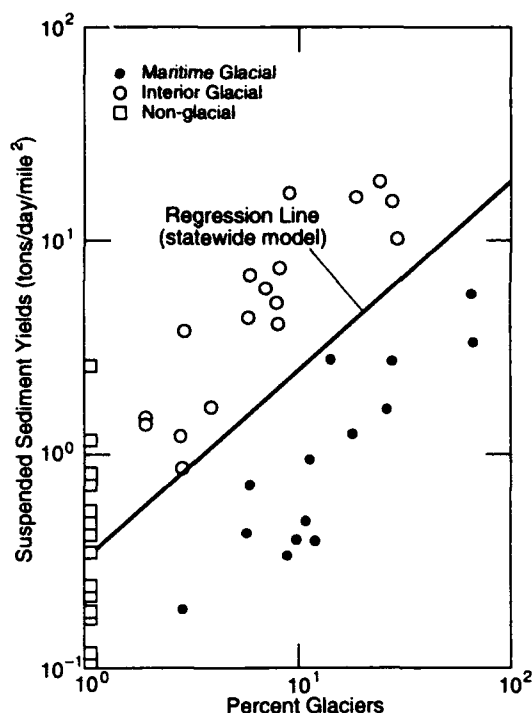


Figure 80. Relation of percentage of glaciers in the drainage to suspended sediment yield (after Parks and Madison 1984).

sediment discharges were then calculated using daily surface flow data for periods of record having no ice cover.

Parks and Madison developed stepwise multiple regression relationships with percentage of glacier cover in the basin and drainage area as independent variables. These statewide regression equations, however, had very low correlation coefficients and high standard errors. Log-log plots of residuals for sediment load, which showed the best relationship to percentage glacier cover and drainage area, were randomly scattered about the regression line, except that, in the case of glacial runoff, residuals could be grouped for interior and maritime areas (Fig. 80). They found a similar grouping for sediment yield using glacier cover as the independent variable. Daily sediment records were inadequate, however, to define annual mean yield of suspended sediment to compare to quantities predicted by the equations and only mean values could be estimated.

Regression equations for the rivers fed by glaciers provided estimates of mean suspended sediment values that had standard errors of approximately ± 50 to 100% and r^2 values of about 0.6 for sediment concentration and 0.9 for sediment load. Nonglacial streams usually had lower suspended sediment yields but greater variability in sedi-

ment-discharge relationships, something that is reflected in the larger standard errors (about ± 64 to 180%). Larger errors were also found for maritime glacial stream relationships than for interior glacial streams; these errors may be related to differences in the influence of ice-free parts of their respective basins. Unit runoff from glacierized basins in the maritime region is generally slightly higher than, but of the same order of magnitude as, that from glacierized basins in the interior. Because of the much higher maritime precipitation, however, the unit runoff from nonglacial areas may be as much as 10 times greater for maritime streams than for interior streams.

As with their runoff evaluation, the equations estimating suspended sediment yield suffer from limited data, including virtually no daily records of significant length, a limited number of gauging sites within extremely large basins that are biased both in area and time, and an incomplete understanding of the relationships among independent and dependent variables (Parks and Madison 1984). In the latter case, the basin characteristics chosen as variables are not necessarily those that are critical to suspended sediment discharge in glacierized basins, nor are they necessarily of similar importance across the large basins that they examined.

Deficiencies in the database are critical to the reliability of Parks and Madison's relationships. In particular the accuracy of the instantaneous sediment transport curve, which forms the basis of the relationships, depends on the assumption that data are reasonably well distributed over the range of expected discharges and normally distributed about the mean. Peak flow measurements of sediment concentration are few, with some data sets lacking any such observations. Predicted quantities will thus tend to be higher than actual transport.

Gurnell (1987) found similar problems in analyzing the suspended sediment yield of 43 glacierized catchments from across the world, using both single- and multiple-variable, log-transformed regression models. The independent variables that she examined included discharge volume, total basin area, glacierized area, and distance of the gauging site from the glacier source. The area of glacier cover was positively related to suspended sediment yield because of its role as a sediment source. In addition, the distance of the gauging site from the glacier source affects the apparent sediment yield of the basin, as had been previously suggested by Borland (1961) and Guymon (1974).

None of the relationships, however, was statistically significant and the regression models could not explain a large proportion of the variance in suspended sediment yield.

Fenn et al. (1985) and Gurnell and Fenn (1984a,b) examined the relationship of suspended sediment yield to discharge using multiyear as well as daily and hourly observations of the Tsidjiore Nouve glacierized basin in Switzerland. Their evaluation of regression analyses and their alternative approach to such analyses illustrate many of the problems associated with developing and using sediment rating curves for glacierized basins.

Researchers run into three major problems when estimating a suspended sediment rating curve from sample data and then applying that curve to calculate suspended sediment transport and sediment load from discharge data. First, sediment rating curves derived from different periods vary and are not valid except for the periods for which they are established (Ostrem 1975a). Ostrem (1975a) concluded that the relationship between discharge and sediment yield is, in fact, changing more or less continuously throughout the runoff season and from year to year. Second, detailed examination of the residuals of all sediment rating curves estimated by simple linear regression analysis for the proglacial stream of the glacier in Tsidjiore Nouve has shown that they are strongly serially autocorrelated, and that this autocorrelation biases the rating curve and any estimates of suspended sediment concentration derived from it (Fenn et al. 1985). Third, because the rating curve is estimated from log-transformed data, estimates of concentrations and yields of suspended sediment will, on average, be too low (Walling 1977). Thus, it is impossible to develop an unbiased simple regression model relating suspended sediment concentration to discharge for the Tsidjiore Nouve glacierized basin (Fenn et al. 1985).

To improve the relationship derived for estimating suspended sediment values, correlations between variables and the recent past behavior in discharge and sediment concentration must be considered. Specifically, suspended sediment concentration is affected not only by the present discharge of the proglacial stream, but also by the recent discharge history of the stream and the degree to which past discharges have exhausted the supply of sediment in the glacier (Gurnell 1987). Gurnell and Fenn (1984a,b) suggested that suspended sediment rating curves be based on a time-series analysis of the data. They applied a transfer function relating hourly changes in sus-

pended sediment concentration to hourly changes in discharge, assuming a 1-hour lag. Their analyses suggested that 1- to 2-day periods of record for discharge and suspended sediment data improved predictions.

Gurnell and Fenn (1984b) also attempted to separate the discharge series into glacial flow components before considering the relationship of suspended sediment concentration to discharge. This separation, in effect, indirectly considered process relationships in the glacier. A simple two-reservoir mixing model (previously used by Collins [1979b] for analyzing water quality components in meltwater) was applied to separate the total discharge into subglacial and englacial components. Their analyses suggested that peaks of suspended sediment concentration matched peaks of discharge in the englacial and subglacial discharge components, and suggested that a multivariate transfer function employing the two flow components as independent variables may improve predictions of suspended sediment concentration.

Fenn (1989) evaluated the quality of estimates from suspended sediment rating equations and transfer functions by examining the stability of rating relationships derived from 6 years of data from the Glacier de Tsidjiore Nouve basin. In general, standard rating curves based upon a linear regression analysis were applicable only to the time from which they were developed, but multi-year models improved estimates. Single year models had errors ranging from 34 to 278% when their predicted quantities were compared to data from subsequent years. The mean absolute errors ranged from 35 to 81% (mean of 52%) whereas the multi-year model had a mean error of 38%. Fenn (1989) calculated additional models, one using a least squares regression corrected for the autocorrelation associated with flow and discharge history that improved error in estimates to 15%. A simple transfer function using 25-day data sets and relating hourly changes in suspended sediment concentration to hourly discharge variations, assuming a 1-hour lag, improved predictions to only a 5% absolute error. This analysis suggested that sediment rating curves for this basin could be applied to times beyond the calibration period using transfer functions.

Although the methods of Gurnell and Fenn (1984a,b), Fenn et al. (1985) and Fenn (1989) for developing suspended sediment rating curves are more complex, each technique still incorrectly assumes a linear relationship between the independent and dependent variables and cannot yet cope

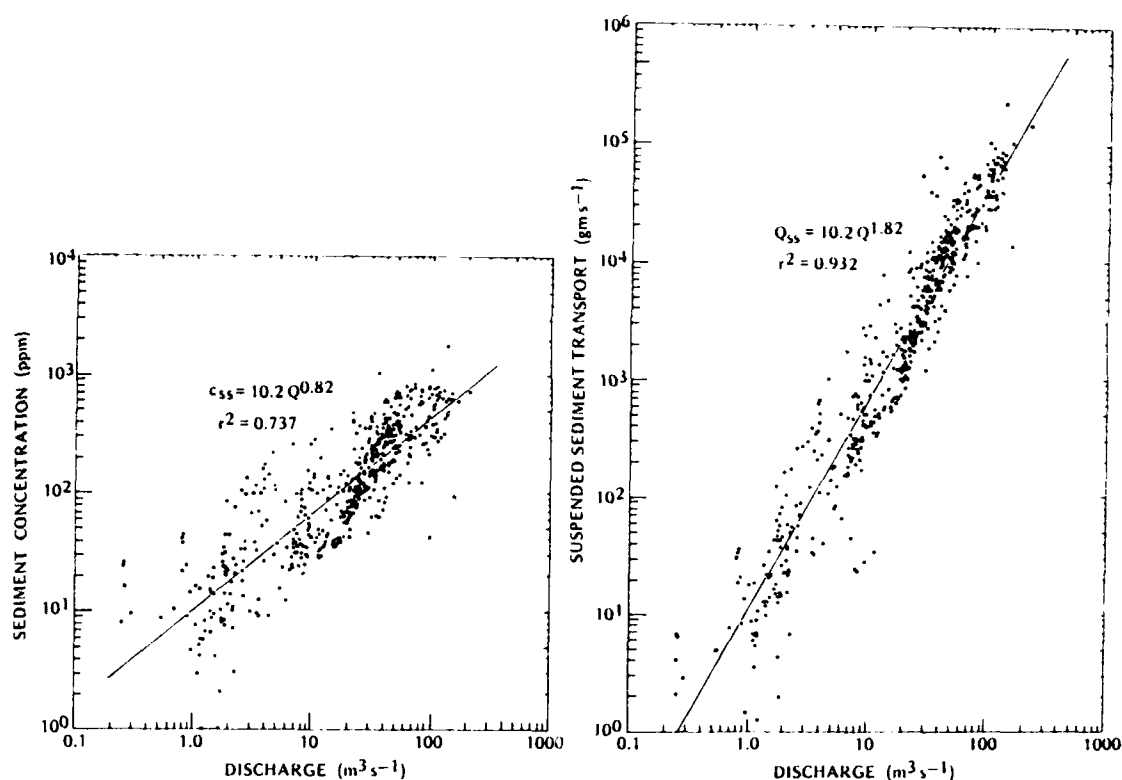


Figure 81. Suspended sediment vs. discharge relationships for Lewis River, Baffin Island. Note that the rating relationship between sediment discharge and water discharge, although useful for computing total sediment transport, is statistically spurious, since it involves a correlation between c_{ss} , Q_{ss} and Q . Standard error of estimate is the same as that derived from the left-hand plot, despite the apparently improved relationship (from Church and Gilbert 1975, used with permission of the Society for Sedimentary Geology).

with the problems of sediment availability and exhaustion (Gurnell 1987). Regression models for glacierized catchments, therefore, remain useful only in estimating sediment yield within about an order of magnitude or more (e.g., Guymon 1974, Church and Gilbert 1975). The standard error in correlation for regression analyses is usually high and the results statistically spurious (Fig. 81) (Benson 1965, Church and Gilbert 1975, Gurnell 1987). Where there are sufficient data, a simple transfer function for sediment rating curves can account for the effects of previous discharge and sediment transport history on subsequent changes in suspended sediment supply and load and can therefore improve application of sediment rating curves.

PHYSICAL MODELS

The number of physical models for predicting runoff or sediment yield from glacierized basins is limited. Most conceptual predictive techniques have been developed for specific applications and no attempt has been made to apply them to other regions.

Predicting sediment discharge

The open literature contains no physically based models of sediment yield nor any conceptual models that combine predictions of runoff with those of sediment production. The absence of a physical sediment model is a clear statement of the present lack of knowledge of sediment transport, storage and production within glacierized basins, and an indication of a need for further research.

Predicting runoff

Predictive models of runoff generally calculate discharge in two steps. One part attempts to calculate meltwater production, while the other calculates the effects of the drainage process within the glacier. Meltwater production is best simulated in terms of both the physical process and the estimated melt rate, primarily because of the quantitative data available from field studies (Fountain and Tangborn 1985b). In contrast, theories on internal drainage, which controls flow rates, storage and discharge to glacial rivers, are poorly documented by field studies and difficult to simulate through time and space.

Fountain and Tangborn (1985b) recently reviewed contemporary predictive models for runoff; the discussion that follows includes their summary and several of the more detailed case histories in Young (1985).

Conceptual model examples

Anderson (1973, 1976) model, Alaskan basins. Anderson's model is based upon snow accumulation and ablation, and uses daily precipitation and temperature measurements. This model has been used to predict runoff from Alaskan basins for periods ranging from 6 to 24 hours, but it does not consider glacier effects directly. Nibler (cited in Fountain and Tangborn 1985b) reported that the model has been modified to consider glaciers by increasing the area of snow cover to equal that of the glacierized area. Runoff is determined by treating the glacier as an extremely thick snowpack that does not melt during the summer.

The model as a whole considers snowpack melting using a linear temperature index model related to the difference in air temperature and a base temperature at which there is no melting during a dry period. Thus

$$M = M_F (T_a - T_b) \quad (110)$$

where M = snow-melt

M_F = melt factor

T_a = air temperature

T_b = base temperature at which there is no melting.

An energy balance formulation based upon air temperatures is used during rainy periods. Thus

$$Q_T = Q_n + Q_e + Q_h + Q_p \quad (111)$$

where Q_T = total energy exchanged at the snow surface

Q_n = net radiative heat transfer

Q_e = latent heat transfer

Q_h = sensible heat transfer

Q_p = conductive heat transfer by rainwater.

Air temperature from the meteorological station is empirically corrected for elevation differences in the basin to estimate melt rate. Based upon an empirical depletion curve for each basin, snowpack water equivalence is estimated and runoff is predicted when the liquid storage capacity of the snowpack is exceeded. In the glacier-modified

version, the empirical depletion curve includes the area of glacier cover. This modification presents reasonable estimates of seasonal runoff, but incorrectly predicts the daily timing and volumetric changes in runoff.

Baker et al. (1982) model, Venagtferner, Austria. Baker et al. use an energy balance model of Escher-Vetter (1980) to calculate glacier meltwater production based upon solar radiation measurements at a gauging station in the basin of Venagtferner, Austria. This gauging station also provides hourly stream flow data and meteorological data that include precipitation, air temperature, wind speed, wind direction and humidity. Glacier albedo is estimated from daily photographs of the glacier that are taken automatically from a position at the gauging station. These data are incorporated into a digital terrain model of the glacier surface, for which a radiation distribution model calculates solar radiation and albedo.

Water storage and drainage are estimated using a linear reservoir method, where runoff $R(t)$ is proportional to the volume of water stored $V(t)$

$$V(t) = K R(t) \quad (112)$$

and K is a constant. Changes in stored volume are simply considered equal to the inflow minus the outflow

$$\frac{dV}{dt} = I(t) - R(t). \quad (113)$$

Combining eq 110 and 111 yields

$$K \frac{dR}{dt} = I(t) - R(t) \quad (114)$$

or

$$R(t) = \int_0^t \frac{I(t)}{K} \exp^{-(t-1)/K} dt + R_0 \exp^{-t/K}. \quad (115)$$

where R_0 is the runoff at the start of the simulation.

An independent linear reservoir is created for each of three zones on the glacier (exposed ice, firn and snow), and runoff is estimated by combining flow from each. The constants for each reservoir are calculated by assuming the condition of inflow equal to outflow. K is then considered to be the residence time of water within each reservoir. Runoff within each zone responds differently to changes in energy input, the residence times increasing (eq 113) from the firn to the ice zone.

The model is applied to the Venagtferner glacierized basin once snow cover is restricted to only the upper glacierized elevations. Hourly predictions can then be generated. Generally, the model underestimates runoff in July and early August, while overestimating it in late August and September. Errors are apparently related to the constant storage coefficients (which should actually change through the season with the developing drainage system of the glacier), the constant temperatures used in calculating snow cover energy balance, the lack of consideration of precipitation or runoff from nonglacierized areas of the basin, and, possibly, a generally incorrect calculation of snowmelt rate (Fountain and Tangborn 1985b).

Gottlieb (1980) model, Peyto Glacier, Canada. The Gottlieb model separates the basin into ice-free and ice-covered areas, with different calculations for 1) runoff from snowmelt and glacier melt, based primarily upon an energy balance relationship, and 2) water storage and runoff, which are calculated with a linear reservoir model. The basin itself is divided into altitude bands of about 200 m elevation each, and precipitation and temperature are estimated for each band using a linear relationship for elevation change within the basin. Base-line data are taken from a meteorological station near the glacier terminus.

Within the energy balance calculations, snowmelt is determined using a modified degree-day method where the difference between air temperature and snow surface temperature defines the energy available for snowmelt. These values are further refined by considering that the solar radiation input and surface albedo vary seasonally; this effect is accounted for by a similarly varying constant. The water equivalent of the snowpack is used to define its areal extent as in the Anderson (1973) model. Water storage capacity of the snowpack is assigned a constant value. Runoff is then calculated as a sum of overland flow, interflow and baseflow components using a model by Nielson and Hanson (1973). Precipitation runoff in areas without a snowcover is also calculated.

For glacier runoff, meltwater production is calculated by a degree-day method but with an albedo value appropriate to exposed glacier ice. As with the Baker et al. (1982) model, the meltwater is routed through the glacier as a linear reservoir; water storage or water retention is assumed constant for several days at a time.

This model was first applied to the Peyto Glacier basin in the Canadian Rockies and then to

Nordbo Glacier in southern Greenland. In both applications, the first being based on 8 years of data and the second on 4 years of data, peak flows are underestimated, whereas the timing of runoff peaks precedes the actual event early in the season and lags it later in the season (Gottlieb 1980, Fountain and Tangborn 1985b). These inaccuracies appear to result from use of the modified degree-day method for snowmelt and glacier ice melt, and from the constant linear reservoir model for water storage within the glacier.

Lang (1980) model, Aletschgletscher, Switzerland. The Lang model is designed for detailed estimates of runoff over short periods of several hours to several days. Hourly data for air temperature, precipitation and incoming short-wave radiation, as well as runoff from each glacier within the basin, are required for calculations.

A multiple linear regression relationship combines melt production with runoff

$$R_t = b_0 + b_1 R_{t-1} + b_2 R_{t-2} + b_3 R_{t-3} + b_4 T_t + b_5 (T_{t-1} + T_{t-2}) + b_6 S_t + b_7 (S_{t-1} + S_{t-2}) + b_8 (PT)_t + b_9 (PT)_{t-1} + b_{10} (PT)_{t-2} \quad (116)$$

where R = discharge at time t

b_i = coefficients

$t-j$ = denotes observations at previous time step j

T = air temperature

S = shortwave radiation

P = precipitation.

Runoff forecast from eq 116 is based upon melt rates using forecasted values of air temperature, precipitation and incoming solar radiation. The runoff values for previous times are included to simulate water storage or retention within the basin as a whole, without considering glacier storage separately. The equation is further calibrated by considering the changes in ablation rate. Three intervals through the melt season are defined by changes in snow cover, albedo and probable changes in the drainage system within each glacier.

The basis of Lang's relationship is a statistical evaluation of long-term records of various basin parameters. This evaluation indicated that air temperature correlates with ablation zone melt rates, precipitation correlates inversely with incoming solar radiation, which is in turn related to duration of cloud cover, and vapor pressure correlates with longwave radiation. Because of difficulties in ob-

taining accurate data, the latter term is assigned a constant value.

Model forecasts tend to underestimate short-term (hourly–diurnal) peak flows, but predict total seasonal flow fairly accurately. The model also requires forecasts of meteorological variables, which can be unreliable, particularly where complex topography creates local variations in weather. Additionally, it is difficult to define when the model must be modified for seasonal changes in albedo.

Lundquist (1982) model, Nigardsbreen, Norway. Runoff predictions are based upon melt rates for snow and ice calculated using a modified degree-day method that considers variations in albedo and total radiation. The basin is divided into two areas that are defined by the approximate elevation of the glacier's equilibrium line. Meteorological stations in each area provide air temperature, relative humidity, precipitation and wind speed data for estimating melt rate. Snowpack water equivalent is estimated from occasional field measurements.

Meltwater drainage and runoff are calculated separately for the upper and lower elevation zones, with flow-through simulated by linear reservoirs run in parallel with each other. Storage terms differ within each, the value being chosen according to water retention in the snowpack or glacier.

In general, predicted peak flows preceded actual peak flows and were overestimated. The lack of distinction between runoff from glacier-covered and ice-free areas and the source of river discharge data from a site about 2 km from the glacier below a proglacial lake are probably the primary sources of error in the estimates.

Power and Young (1979) model, Peyto Glacier, Canada. Runoff estimates are made using the University of British Columbia (UBC) Watershed Model (Quick and Pipes 1977), with the basin subdivided into 11 area–elevation bands. The area of glacier cover in each band is specified, with the glacier treated as a thick snowpack. Snow-melt is estimated as a linear function of three mean air temperatures: 1) mean minimum air temperature, which indicates latent heat exchange at the snow surface from condensation and evaporation; 2) a mean maximum air temperature, which relates to convective heat transfer; and 3) a mean daily air temperature range, which is a measure of net radiative energy.

For glacier cover, ice-melt is calculated as

$$M_B = M_R(T_{\max} + X_C + T_{\min}) \quad (117)$$

where M_B = band-melt (mm)

M_R = point-melt factor (mm/°C)

$T_{\max}$ = elevation band maximum air temperature (°C)

X_C = radiant energy factor

$T_{\min}$ = elevation band minimum air temperature (°C)

and

$$X_C = \frac{T_{\min} + T_d/X}{D_W + T_d/X} C_m \quad (118)$$

where X_C = radiant energy factor

T_d = daily air temperature range ($T_{\max} - T_{\min}$) in the band

D_W = reference dew point controlling energy partition between melt and sublimation

C_m = limiting value.

According to Power (1985), the radiation component of melt is considered the most significant factor, and therefore melt depends on the maximum rather than the mean daily temperature. Melt caused by condensation is controlled by the vapor pressure of the air compared with the vapor pressure of the snowpack. Minimum air temperature approximates the dew point. The rate of condensation is determined by multiplying the dew point by the energy partition multiplier X_C , which was based on the minimum air temperature.

Runoff peaks are not well predicted. This inaccuracy apparently results from the neglect of water storage and drainage processes within the glacier, as well as the progressive changes that take place in the drainage system through the melt season. Ice-melt as predicted by these relationships, however, agrees with a regression analysis of Peyto Glacier data by Young (1980).

Tangborn (1984) model, basins in British Columbia, Canada. This model has been applied to various basins in British Columbia, Canada. It is calibrated with data from a nearby nonglaciated basin, then modified for the effects of glaciers using an algorithm that accounts for accumulation and ablation on the glacier surface and predicts water storage and flow delays caused by the glacier's drainage system.

Snow-melt and ice-melt are calculated using an exponential relationship for mean daily temperature, daily temperature range (used to approximate cloud cover) and albedo, including seasonal variations in angle of solar incidence. Each basin is

subdivided into altitude bands, with a maximum of five bands within which air temperature and precipitation are calculated using lapse rate corrections for meteorological data from a nearby station.

Internal glacier drainage is treated as a linear reservoir but with restricted inflow. Two periods are defined annually, the early period corresponding to inflow to glacier storage (1 November–15 July). During the remainder of the year, water is released from glacier storage and flows out as stream discharge.

An exponential relationship defines maximum inflow G_{in} as

$$G_{in} = \exp^{-C_1 K} \quad (119)$$

and maximum outflow G_{out} as

$$G_{out} = 1 - \exp^{-C_2 K} \quad (120)$$

where C_1 and C_2 are calibration constants and k is an internal storage constant.

In general, Tangborn's relationships overestimate runoff early in the season and underestimate it late in the season. These variations may result from considering glacier storage as fixed periods of inflow or outflow.

Tangborn (1986) and Clarke et al. (1986a) also applied this model to the Susitna River basin, Alaska. Here, daily data from a low-altitude weather station at Talkeetna, Alaska (105 m), and a higher-altitude station at Gulkana, Alaska (479 m), were used for calculating water storage and runoff from two equal-area altitude zones above and below the 870-m elevation. For the period of record from 1950 to 1976, the runoff forecast correlation coefficient was 0.76, with a mean standard error of about 10%. The model generally underestimated peak flows and flood runoffs associated with precipitation. Tangborn (1986) concluded that improved simulation of snowpack–glacier storage and other aspects of winter hydrology would improve the model. Tangborn (1991) discussed application of this model to the Columbia River basin, but did not specifically address glacier storage and runoff.

Lang and Dayer (1985) model, Z'Mutt Glacier, Switzerland. This model is an extension of the Lang (1980) model for forecasting hourly and daily runoff from the Z'Mutt Glacier basin for the Grand Dixence S.A. hydroelectric project. Data required for the model include discharge, air temperature, vapor pressure, precipitation and incoming solar

radiation measured at a gauging station near the Z'Mutt glacier terminus. This catchment, as well as the other 33 glacierized basins within the area of the Grande Dixence hydroelectric scheme, have extremely complex topography; this was also considered in developing the model.

The model is described as a "conceptual regression model" by Lang and Dayer (1985). Multiple regression equations relate various hydrological and meteorological variables to discharge in the form

$$Q_t = a + b_1 \times x_i + \psi \quad (121)$$

where Q_t = discharge at time t

a = constant

b_i = regression coefficients

x_i = variables

ψ = residual.

Regression coefficients are defined using a least squares method.

Seasonal variations in snow cover, snow depths, albedo, physical properties of snow and ice, and drainage system development are accounted for by dividing each melt season into three intervals corresponding to the beginning, main part and end of the ablation season. Criteria for defining each interval include air temperature, snowcover extent and the nature of the daily hydrograph. Regression equations were developed for each of the three intervals and residual values defined. Additional modifications to these relationships consider the effects of hourly versus daily parameter variations, solid versus liquid precipitation, and glacier albedo. The linear regression relationship follows eq 116. Thus, it relies upon daily meteorological forecasts of temperature, net radiation and precipitation and past daily values for discharge, which is similar to Lang's (1980) model.

The model approximates hourly runoff reasonably well for the first 12 hours. Accuracy of the meteorological forecasts to a large extent determines its accuracy within this period as well as beyond it. Lang and Dayer (1985) indicated that they were recalibrating the model using new data. Improvements suggested by them included continuously varying regression coefficients, rather than three constant periods, adding a vapor pressure term as a surrogate of latent heat flux, and increasing data collection within the basin to include measuring precipitation at more sites (two to three minimum), conducting detailed snow surveys and measuring humidity.

Power (1985), various models, Columbia River basin, Canada. Daily and seasonal forecasting is required for operations by British Columbia Hydro within the Canadian portion of the Columbia River basin. Daily time scale predictions use FLOCAST, the UBC model or SSARR. Seasonal runoff forecasts are based upon VOLCAST as well as the UBC and SSARR models.

SSARR, which uses a temperature index method to predict snowmelt runoff, was set up by the Corps of Engineers for the Columbia basin to account for runoff from nonglacial watersheds (Speers et al. 1979, Cassell and Pangburn 1991). The model, however, does not accurately represent snowmelt above the treeline in alpine areas, where solar radiation determines melt rates. SSARR also does not distinguish between snow and ice, and therefore cannot simulate directly the drainage within glacier-covered areas. An artificially thick snowpack, as used by the Nibler-modified model of Anderson (1973), described previously, is used to calculate runoff from glacierized parts of the basins, but it does not account for glacier meltwater production variabilities caused by increased or decreased albedo nor internal glacier storage and drainage.

FLOCAST is designed to forecast inflows to reservoirs and is used daily in the Columbia River basin. Snow-melt is calculated for up to 18 elevation bands by a simple degree-day formula incorporating a variable melt rate factor that is lowest in winter and peaks in June. Lapse rates are determined by maximum daily temperatures and the elevation of the freezing level.

In FLOCAST, meltwater production by glaciers is determined by assuming that 1) the snowline elevation rise is slower on the glacier than on other basin areas, and 2) the snow-melt rate on the glacier is slower than adjacent to it. Glacier snow-melt is calculated initially at one-half the rate off of the glacier, and, as the snowline rises, the glacier-melt rate is defined as a fixed value. Power (1985) compared the FLOCAST and UBC models to empirical data on snow- and ice-melt on Peyto Glacier (Young 1980), concluding that FLOCAST underestimated daily, peak and seasonal runoff, suggesting that the ice melt rates are too low. The model does not consider the effects of glacier drainage and storage, which introduce errors as well.

The UBC model uses 5-day forecasts of precipitation and of maximum and minimum temperature within 11 area-elevation bands in the watershed or basin. Snow-melt and ice-melt are both estimated using a linear relationship following eq

117 and 118. Glacier-melt depends on maximum rather than mean daily temperature, as does snow-melt. Changes in snowline elevation modify glacier- or snow-melt rates. As with FLOCAST, internal drainage and storage in glacier-covered areas are not considered.

VOLCAST is a series of statistical regression equations for forecasting monthly seasonal runoff between January and August for several areas within the Columbia basin. Parameters to define runoff volume include 1) winter precipitation (November through March, inclusive), combined with selected snow course water equivalents, 2) spring and summer precipitation (April through September, inclusive), 3) antecedent precipitation (in the previous September and October), 4) evaporation index, using February to September mean monthly maximum temperatures, and 5) a glacier-melt index, using April to September mean monthly maximum temperatures, that is related directly to glacier runoff. Regression coefficients and constants are determined using multiple parameter historical data that are updated each year.

In addition, the physical UBC model is used for seasonal forecasting at four dams from January to August. Historical sequences of maximum and minimum temperatures and precipitation from 1958 to the present are input to the model, 1 year at a time, along with the current basin conditions. The resulting set of runoff sequences is then statistically analyzed to predict maximum and minimum flow by the day, month and season.

The SSARR model was also tested for seasonal forecasts in sub-basins within the Columbia basin using median values of temperature and precipitation (Power 1985). Volumetric seasonal forecasts with SSARR were comparable to those generated by detailed multiple regression forecasts. As Power (1985) states, applying SSARR or another physically based model would be advantageous over the statistical approach because accuracy can be continually improved by making adjustments to reflect existing hydrometeorological conditions.

Braithwaite (1980, 1984) models, Greenland ice sheet. Braithwaite and Thomsen (1984a,b) applied the mass balance (MB1) and runoff (RO1) models of Braithwaite (1980, 1984) to simulate runoff from the Greenland ice sheet for evaluating hydroelectric potential. The MB1 model calculates mass balance and specific runoff based upon extrapolation of monthly air temperature and precipitation data from a nearby meteorological station as a function of elevation within the basin. Precipitation is assumed constant across the basin, but is divided into rainfall and snowfall based upon the

probability of below-freezing temperatures in each month. Temperatures are extrapolated by considering coastal versus inland locations, vertical temperature lapse rate, and the cooling effects between ice-covered and ice-free areas. Values for the parameters of each effect are based upon glacier-climate studies (Braithwaite 1984) or trial-and-error evaluations.

Monthly ablation rates for ice and snow at each elevation are assumed to be a function of monthly mean temperature, following Braithwaite (1985). Runoff simulation considers the potential for melting or refreezing of meltwater and rainfall as functions of elevation.

Specific annual runoff, based upon the sum of annual ice and snow ablation and of rainfall, is integrated over the glacierized basin area by the RO1 model to calculate glacier runoff volume. Runoff from areas that are glacier-free is calculated from annual precipitation at a nearby meteorological station.

Comparisons of the predicted runoff volumes to runoff data for 1980 to 1985 suggest that the model underestimates the volume of peak flows, overestimates low flow volume and predicts the timing of peak flows after the fact. Braithwaite and Thomsen (1984b) suggest that these differences may be lessened by including process models of the glacier-climate relationship.

Comparisons

The physical models described previously have not been compared directly by analyzing similar data sets for one or more basins or for periods covering several years or more. Fountain and Tangborn (1985b) compared six models for a single year's data and for the results of the particular basin analyzed by the respective authors (Fig. 82). The value of this comparison is limited by the fact that the dates of each year of record differ and that additional years of record, which are clearly needed to better define a model's accuracy, are not

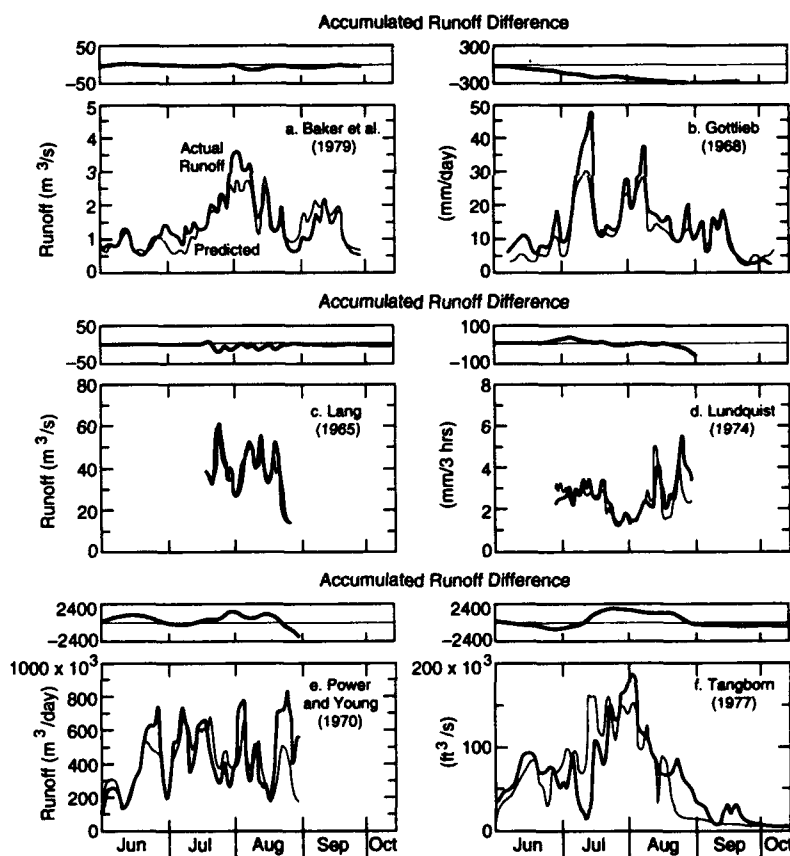


Figure 82. Comparisons of six conceptual runoff models by each respective author as analyzed by Fountain and Tangborn (1985b). Light lines are predicted runoff; heavy lines actual runoff. Data years are given in parentheses below authors' names in upper right corner of each graph (after Fountain and Tangborn 1985b).

generally available. Only the Gottlieb (1980) model has actually been applied to two basins of distinctly different climate and region using several or more years of data. The relatively similar errors of the Gottlieb model for simulated peak and seasonal runoff suggest that it may be equally applicable to different climatic regions (Fountain and Tangborn 1985b).

Fountain and Tangborn's (1985b) numerical and graphical comparison of the observed and predicted values (Fig. 82, Table 3) suggest that none of the physical models can accurately predict the timing and volume of peak flows nor the total seasonal volume of flow, with the exception of Lang's (1980) model. Fountain and Tangborn's statistical evaluation shows a significant variation in the coefficient of determination among the models. Volumetric differences are generally similar (except for Lang 1980); error calculations suggest rather large differences in model performance for each period of record. These evaluations and a

general evaluation of the other models presented suggest that each is basin-specific. Unless calibrated to a given basin using measured data, the reliability of the physical models is highly questionable.

Fountain and Tangborn (1985b) concluded that a major deficiency in each of the six models that they compared is the rudimentary treatment of meltwater within the glacier environment. This conclusion holds for the other models described here as well. Improved formulation of glacier meltwater production, drainage and storage are required to improve the accuracy of models for glacierized basin runoff. Although considerable progress has been made, Colbeck's (1977) assessment that runoff can only be accurately predicted after the snowpack has undergone melt metamorphism and the glacier's drainage system has fully developed appears still generally true today. A general physical model for predicting runoff over short or long periods has not yet been developed.

Table 3. Comparison of six physical models using data of Figure 82 (after Fountain and Tangborn 1985b).

Reference	Data year	No. of data pairs	Mean of observed runoff	Mean of calculated runoff	Coeff. of determination*	Seasonal volumetric difference†	Standard error**	Relative error††	Absolute error***
Baker et al. (1982)	1979	121	1.38 m ³ /s	1.32 m ³ /s	0.80	0.05	0.25	-0.05	0.18
Gottlieb (1980)	1968	131	13.0 mm/day	11.0 mm/day	0.82	0.15	0.31	-0.15	0.22
Lang (1980)	1965	40	36.5 m ³ /s	36.5 m ³ /s	0.68	0.00	0.17	0.00	0.13
Lundquist (1982)	1974	497	2.57 mm/3 hr	2.41 mm/3 hr	0.55	0.06	0.20	-0.06	0.14
Power and Young (1979)	1970	92	425 × 103 m ³ /day	400 × 103 m ³ /day	0.54	0.06	0.31	-0.06	0.23
Tangborn (pers. comm.)	1976	137	61.8 ft ³ /s	59.2 ft ³ /s	0.54	0.04	0.50	-0.04	0.33

* Coefficient of determination: $\frac{\sum (r_o - \bar{r}_o)^2 - \sum (r_c - r_o)^2}{\sum (r_o - \bar{r}_o)^2}$

† Seasonal volumetric difference: $\frac{[\sum (r_o - r_c)]}{\sum (r_o)}$

** Standard error: $\frac{[\frac{\sum (r_c - r_o)^2}{n}]^{1/2}}{n}$

†† Relative error: $\frac{\sum (r_c - r_o)}{n \bar{r}_o}$

*** Absolute error: $\frac{\sum |r_c - r_o|}{n \bar{r}_o}$

where r_c = calculated runoff

r_o = observed runoff

$\bar{r}$ = mean runoff

n = total number of observation.

Conclusions and Recommendations

Glaciers occupying as little as a few percent of the area of a basin exert significant control on runoff and sediment yield. Glaciers modify peak discharges, determine the variability and volume of hourly, daily and seasonal discharges, create a lag between precipitation and the resultant increase in runoff, and influence long-term trends in the annual volume of flow. Runoff variations result from the glacier's roles as a daily and seasonal water source and water storage medium, long-term water reservoir, and in routing meltwater and meteoric water through the englacial and subglacial drainage systems to the terminus.

Glaciers produce the majority of their runoff within several months or less, from a few weeks in the Arctic and 4 to 5 months in lower latitude basins. Daily variability is greatest in spring as the melt season begins and generally decreases through the summer. Runoff is at a minimum, sometimes decreasing to near zero, in mid- to late-winter.

The amount of energy available for producing meltwater from snow and ice is a critical component of daily and seasonal runoff. This energy consists mainly of net radiation, sensible heat, latent heat of condensation or evaporation and heat released by cooling or freezing of rainwater falling on the snowpack and ice. Long-term trends in runoff largely reflect the effects of climate on the mass balance of the glaciers.

Glaciohydrologic processes and factors control meltwater and meteoric water movement and storage within the permanent and seasonal snowpack and within and below the glacier itself. Most glaciers in drainage basins with the potential for water resources development, or that are in current use, are those at or near the pressure-melting point. These glaciers have a relative abundance of meltwater, the volume of which varies seasonally.

Above glacier ice, meltwater fully saturates the

snow and Darcy's law for a saturated medium, modified for inhomogeneities, applies. Metamorphic changes under saturated conditions take place more rapidly than they do under unsaturated conditions, and permeability and seepage rates are greater. Water flow in a saturated basal layer of the snowpack effectively integrates the diurnal water fluxes. Within the firm or permanent snowpack of the accumulation area, a thick saturated layer commonly acts as an unconfined aquifer that stores meltwater and retards its movement into the glacier's drainage system.

Meltwater flowing from the seasonal and permanent snowpacks either discharges at their bases onto the glacier surface, where the meltwater creates supraglacial channels, or enters the glacier directly through intergranular passageways or veins, and through larger drainage sinks such as crevasses and moulins. Within the glacier the englacial drainage system is idealized to consist of a tree-like network of small veins and capillary tubes feeding progressively larger conduits at depth. Surface water entering the glacier in crevasses and moulins moves through englacial conduits that apparently join progressively larger conduits at depth.

The water reaching the subglacial environment theoretically enters one of three drainage systems: 1) branching, progressively interconnected conduits and tunnels similar to the englacial system (conduit-tunnel system), 2) an anastomosing pattern of smaller conduits or orifices that link cavities in the bed (linked-cavity system), or 3) in the absence of conduits or channels where there is no englacial discharge at the bed, a thin, hydraulically connected film between the ice and bed (distributed-flow system). Water may also flow preferentially within existing bedrock fractures or abandoned channels. Each system can coexist within

different parts of the same glacier, with the proportion of discharge varying over time as well as distance.

The linked-cavity system maintains a given water flux capability under pressures that are higher than in a conduit-tunnel system, and this ensures that it is mostly stable and can persist with seasonal changes in discharge. These changes are accommodated by changes in the cross-sectional area of conduits and cavities. Diurnal lags in meltwater or precipitation input can result from conduits being reduced or closed over parts of the bed during winter before the full system is reestablished in the spring.

In contrast, a conduit-tunnel system will tend to evolve and grow with increases in discharge, but as discharge decreases, the system may disintegrate because of ice deformation as heat flux is reduced. The system therefore evolves and degrades seasonally and runoff reflects such changes, with closure during winter resulting in temporary water storage and, subsequently, significant lags in discharge as the system redevelops each spring. By peak summer discharge, a steady-state system, with rapid flow-through, reduced storage and low water pressures, is reestablished.

While drainage system configuration and hydrology are reasonably straightforward when the substrate consists of bedrock, beds composed of sediment result in additional complicating factors. In this case, conduits or tunnels may be incised partly into sediments or cut upward into basal ice with a sediment bed. Therefore, closure or expansion is affected by the rate of sediment deformation into the tunnels.

Further, sedimentary substrates are permeable and a certain small proportion of water is transported within them. If there is no drainage system, but there is melting, hydraulic pressures must increase sufficiently to equal or exceed the local ice overburden pressures; then, water can accumulate between the ice and sediment, thickening to a few millimeters or less, and becoming distributed across the bed.

Sediment transport and sediment yield are complex functions of the internal or glaciohydraulic processes that erode, entrain, transport and deliver sediments into glacial discharges, and of climatic conditions that determine the rate, magnitude and timing of meltwater production. Thermal conditions within and below a glacier affect sediment entrainment and release, while the englacial and subglacial drainage systems, their seasonal evolution and stability, and their interaction with the

various glaciologic processes, determine sediment transport and discharge to rivers at the terminus.

Glacial rivers in general annually transport significantly more sediment in suspension and as bed load than nonglacial rivers under normal flow conditions. Sediment discharge of glacierized basins is, however, more variable daily, monthly and annually, with up to 95% of the annual load transported during the melt season, which varies in length with latitude and altitude.

The sediment yield of glacierized catchments is highly variable from region to region and can vary from glacier to glacier and from year to year at a glacier. This variability is not related directly to discharge; sediment load commonly increases out of phase with runoff, but it may also increase without any change. Sediment discharge can also vary in the extreme, with nearly the equivalent of the annual load released in a single day.

Insufficient data exist to define the magnitude of long-term fluctuations in sediment yield. Basin geology, climate and its control on glacier mass balance and thermal regime, and the type and extent of the subglacial drainage system in relation to available sediment control the annual sediment discharge.

The sediment transported by glacially fed rivers comes from sediment in transport by the glacier itself and from sediment in transport by meltwater within the englacial-subglacial drainage systems. Sediment is also derived from nonglacial processes in glacier-free areas of the basin and may include sediments in transport by tributary rivers and various erosional processes, including overland flow and mass wasting. Unvegetated, recently deposited sediments in the glacier's terminus region are highly unstable and can be rapidly eroded into rivers.

Debris transported within a relatively thin zone near the glacier's sole makes up the majority of sediment in transport within the glacier. Basal debris originates by a variety of processes, including mechanical abrasion and plucking, regelation and freeze-on. Abrasion and related processes, such as freeze-thaw, create sediment from bedrock substrates, and these materials are incorporated into ice by plucking and regelation mechanisms. Freeze-on, which occurs under certain thermal conditions at the ice/bed interface, can create a thick, debris-rich zone from subglacial sediment and rock particles. Debris transported basally is modified actively by interactions with the glacier substrate and internally with other basal debris during deformation and flow. In the instance of

unconsolidated sediments composing the substrate, these sediments deform in concert with ice flow, not only resulting in particle interactions, modifications and comminution, but also a net transport of sediment to the terminus.

Debris is characterized by properties derived from its source and its passive or active transport mode. It is mostly coarse, angular rock fragments when of supraglacial or englacial origin, but finer-grained, more rounded particles and sediments when basally derived. Basal materials in particular are modified by subglacial mechanisms and can exhibit extremely variable grain size distributions if originating from subglacial sediments.

The annual flux of sediment transported by ice can only be grossly estimated, if ice flow velocity is known and measurements or estimates of supraglacial, englacial and basal debris volumes are available.

Variations in sediment flux by water reflect the seasonal variations in the extent and development of the englacial-subglacial drainage systems, the type of drainage, and seasonal variations in sediment production and availability, the latter being a function of a previous hydrologic events that remove or deposit bed materials.

Generally, as the drainage system is progressively reestablished each spring, new sediments produced during winter are entrained and episodically removed as different areas of the bed drain and are flushed of sediments. Diurnal meltwater discharges during this time can also cause episodic opening and closing of conduits and daily sediment discharge variations. In contrast, late in the melt season, a lowered sediment flux appears to be a function of an exhausted sediment supply and fully developed drainage system. Severely reduced winter discharges ensure sediment storage on the bed that then provides larger volumes of sediment for spring meltwater discharges.

Existing statistical and physical models are limited in their ability to accurately predict short- and long-term variations in runoff and sediment yield of partly glacierized basins. In particular, statistical relationships used for forecasting lack sufficient records and are limited in application to the specific basin and specific period for which they were developed.

Physical models of runoff are preferred because they consider processes and can be continually adjusted to simulate observed hydrometeorological conditions, or altered to consider existing conditions and runoff history. Although meltwater production at the glacier-snowpack surface can be

reasonably modeled, physical simulations suffer from an incomplete treatment of the glaciohydrologic processes and controls on runoff. As a consequence, existing models rely heavily on calculations of meltwater production and either ignore glaciological controls on discharge or typically consider water routing, storage and drainage through the glacier using a linear reservoir relationship. Models are considered basin-specific and have rarely been applied to more than one basin. In these basins, they are generally incapable of accurately predicting diurnal, peak or seasonal flows.

There are no physical models of sediment yield that are specific to glacierized basins and their development requires improved understanding of the processes of production, transport and storage of sediment and the interaction of these processes with the englacial and subglacial drainage system.

Data records, including most European sites in the Alps, are limited in length and number. The lack of measurements and analyses for glacierized catchments in the U.S. and Canada is particularly noticeable, especially given the large area and volume of glaciers in both countries.

Long-term data on runoff and sediment yield at glaciers, in conjunction with overall basin hydrology, hydraulics and climate, are needed. Such data are required for testing, evaluating and developing statistical relationships and physical models, including those expanded for use in multiple basins or watersheds. Detailed data collections and model developments that exist at operating hydroelectric projects in Switzerland, Norway and France should be evaluated for applications to North American basins. While these regions can differ significantly in terms of climate and terrain, the European and Scandinavian data and quantitative relationships can help focus the modeling. Analytical results cannot yet be applied directly, however, to North American glacierized basins until additional baseline data are collected and analyses are conducted.

Research on the glaciohydrologic and glaciohydraulic processes and factors that control water and sediment production is required to formulate realistic and accurate physical models of runoff and sediment yield. Process analyses must be coordinated with simultaneous measurements of climatic variables and standard hydrological data at downstream gauging stations. These latter data are those commonly collected and may exist for sites external to the glacierized basins of interest. Analytical comparisons may then allow extrapolation of some longer-term data to the basin of interest.

The relationships between climate and glacier

mass balance, glaciohydrology and glaciohydraulics, and runoff and sediment yield need to be defined and methods developed that minimize the need for costly, labor-intensive measurements of annual mass balance, runoff and sediment transport. Remote sensing techniques using both active and passive airborne and satellite multifrequency sensors need to be developed for regional predictions of runoff, both annually and in near-real-time.

Future climatic changes will have a major effect on regional and global water resources. Climate change effects on glacierized basin runoff, resulting from changes in the magnitude and frequency of precipitation and shifts in the mean and extremes of temperature, will affect glaciers through changes in mass balance. Positive and negative feedbacks may result. The links among climate, mass balance and runoff are poorly understood and require basic research.

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13. ABSTRACT (Maximum 200 words) Glaciers exert significant control on runoff and sediment yield of partly glacierized basins, such basins being inherently more complex than non-glacierized basins. Glaciohydraulic and glaciohydrologic processes and factors determine the characteristics of runoff and of sediment discharge of rivers draining glacierized catchments. Significant problems develop in predicting both short- and long-term variations in water and sediment discharge because of the complicating effects of these processes and factors. Predictions are necessary for effective management of a basin or watershed's water resources. These processes and factors must therefore be incorporated to improve predictive models of runoff and develop models for sediment yield. In this monograph, the current state of knowledge on the nature of runoff and sediment yield in rivers originating in partly glacierized basins is reviewed and integrated, in particular analyzing the glaciohydrologic and glaciohydraulic processes and factors determining basin characteristics. Current statistical and physical models for predicting runoff and sediment yield in glacierized basins are reviewed and, based upon an assessment of both the state of knowledge and modeling techniques, future research or application of existing knowledge to improve predictions are recommended.					
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