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The interdependence of the thermal and hydrologic processes of an Arctic watershed and their response to climatic change

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THE INTERDEPENDENCE OF THE THERMAL AND HYDROLOGIC PROCESSES OF AN ARCTIC WATERSHED AND THEIR RESPONSE TO CLIMATIC CHANGE

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THE INTERDEPENDENCE OF THE THERMAL AND HYDROLOGIC
PROCESSES OF AN ARCTIC WATERSHED AND
THEIR RESPONSE TO CLIMATIC CHANGE

A DISSERTATION

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by

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ABSTRACT

The heat and mass transfer processes which comprise the thermal and hydrologic regimes were monitored continuously from March 1985 until September 1989 in a small watershed on the North Slope of Alaska. Through these intense measurements, a better understanding of the physical processes which determine the character of an arctic watershed have been developed. The state of the hydrologic regime is a product of the thermal regime. The hydrologic and thermal regimes interact to such an extent that neither can be fully understood without considering the other. The consequences of a manmade or environmentally induced alteration in the thermal regime can have dramatic and perhaps dire effects on the hydrologic regime and vice versa.

The implications of global warming reach beyond warmer air temperatures, milder winters and longer summers. The potential effects of climatic warming on the hydrologic regime of an arctic watershed were explored with respect to physical changes in the active layer and the resultant changes in the components of the annual water balance and the nature of the hydrologic cycle. With the advent of climatic warming, the annual depth of thaw in the permafrost will increase, affecting the amount of soil moisture storage, the depth to the water

table, even the shape of the runoff hydrograph. The gradual thawing of the active layer was simulated using TDHC, a finite element heat conduction model which incorporated phase change. The results of four possible scenarios of climatic warming were input into HBV, a hydrologic model to elucidate the effects on the hydrologic regime. The results indicate an earlier, but less intense spring melt event, greater evaporation, greater soil moisture storage, and a potential for severe moisture stress on current vegetation types in early summer unless the precipitation pattern changes.

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I) INTRODUCTION

The Arctic has gained a certain amount of prominence in recent years due to oil production and expanded oil exploration. The sensitivity of arctic tundra to disturbance has prompted national concern for thoughtful management and protection. The effect of the Arctic on atmospheric circulation is of primary importance in understanding global weather processes. The advent of climatic warming is expected to become apparent earliest and display the greatest total change in the Arctic. The response of tundra, as a source or sink of carbon, will greatly influence the timing and significance of climatic warming. Yet, with all these topics of national and international concern, the Arctic remains largely a mystery in terms of understanding the basic physical processes which control every aspect of the arctic biome.

The physical and biological processes of any environment depend upon the heat and mass transfers occurring within that system. An understanding of the response of the arctic biome to a disturbance is cradled in a firm understanding of the character of the hydrologic and thermal regimes and their interactions. To develop a better understanding of the arctic environment I began an intensive investigation cf the hydrologic, meteorologic, and thermal processes occurring in Imnavait watershed. This small headwater watershed is located in the foothills of the Philip Smith Mountains on the Arctic North Slope of Alaska.

The objective of this research was to define how the thermal and hydrologic moisture regimes change throughout the year and determine how the system will be impacted by climatic warming. My central hypothesis is that one may deduce the response of the hydrologic and thermal processes to perturbations in the system given information on the other relevant processes. The course selected involved collecting detailed and continuous measurements on the major energy and mass transfers occurring in the watershed. Since most biologic and physical processes occur within the narrow confines of the active layer, I concentrated my efforts and measurements there. I began by monitoring the processes of radiative, convective, conductive and evaporative energy exchanges, precipitation accumulation and redistribution, freezing and thawing of the active layer and moisture movement. From these measurements I was able to identify many unique features such as intense snow redistribution, snow damming, skewed precipitation patterns, and remarkable variability in almost all physical properties. All of these features are inter-related and interact to form the composite picture of the physical component of an arctic biome. Each of these individual processes interact to such an extent that it is impossible to adequately predict the response of any single process without studying the entire system.

To further enhance my understanding of the integrated processes I utilized physical models to separately describe individual mechanisms of energy or mass transfers. Through the use of a finite element heat transfer program, I was able to determine the relative importance of conductive and convective heat transfer within the soil. A hydrologic model improved my estimates of evapotranspiration and the effect of snow damming. Use of both of these models allowed me to speculate on the response of the system to several scenarios of climatic warming.

This thesis represents a collection of papers on research completed in Imnavait watershed. The intrinsic theme of each chapter is the important physical processes which shape the character of this arctic watershed. However, each paper will discuss different aspects of the interaction of the thermal and hydrologic regimes and the effect of a thermal disturbance. Each paper was written to be published independently of the others, consequently there is some repetition between papers. I apologize for that. Although each paper had one or more co-authors, I am solely responsible for the contents of this thesis. The data collection and analysis for this project was truly a massive endeavor and reflects the efforts of several individuals named in the acknowledgments. My contributions to this research include collection of, or supervision of collection and reduction of all hydrologic, meteorologic and soil physical data. I was personally responsible for all hydrologic and thermal modeling analyses. On papers which I appear as first author, I wrote the first draft, on all others I contributed to the first draft and edited subsequent drafts to their final form.

Chapters I - III include this introduction, the literature review and the site description. Chapter IV, entitled "Hydrologic and Thermal Properties of the Active Layer in the Alaskan Arctic," has been accepted for publication in the Journal Cold Regions Science and Technology and my co-authors were Douglas L. Kane, Robert E. Gieck and Kaye R. Everett. Chapter V, entitled "Hydrology of Imnavait Creek, an Arctic Watershed," has been published by Holarctic Ecology and was co-authored with Douglas L. Kane, Carl S. Benson and Kaye R. Everett. Chapter VI, entitled "Permafrost Hydrology of a Small Arctic Watershed," was published in the proceedings of the Fifth International Conference on Permafrost and was co-authored by Douglas L. Kane. Chapter VII, entitled "Snow Hydrology of a Headwater Arctic Basin," has been submitted to the Water Resources Research Journal and was co-authored by Douglas L. Kane, Carl S. Benson and Glen E. Liston. Chapter VIII, entitled "Modeling the Snowmelt Hydrology of a Headwater Arctic Basin," was a companion paper to Chapter VII and was co-authored by Douglas L. Kane and was also submitted to the Water Resources Research Journal. Chapter IX, entitled "Thermal Response of the Active Layer in a Permafrost Environment to Climatic Warming," has been accepted for publication in the Journal Cold Regions Science and Technology and was co-authored by Douglas L. Kane and John P. Zarling. Chapter X entitled "Thermal and Hydrologic Response in an Arctic Watershed," was written by myself and was prepared for publication as a chapter in a book on arctic ecosystems in cooperation with other researchers conducting research in Imnavait watershed under the auspices of the Department of Energy. Chapter XI is a discussion of the results and limitations of the thesis as a whole.

II) LITERATURE REVIEW

A) The Arctic Foothills

The geologic history and climate of a region determines its topography, soils, vegetation, drainage, and in turn influences its thermal and hydrologic regimes. In order to fully understand the present environment, it is necessary to first examine its physiography. The Arctic Foothills is the area north of the Brooks Range and south of the Arctic Coastal Plain. Most streams and rivers emanating from the north side of the continental divide in the Brooks Range cross the Foothills in a northerly direction. The major rivers flowing from the Brooks Range often flow from morainal lakes. The Foothills are not currently glaciated, although there is considerable evidence of ice age glaciation throughout the region. During the Pleistocene Epoch, when much of North America was covered by continental glaciers, much of the Arctic remained ice-free (Selkregg, 1975). However, the Brooks Range and southern Foothills were completely glaciated. Glacial sediments were deposited throughout mountains and hills and alluvium was carried into the Foothills and Coastal Plain.

The Brooks Range and southern Foothills experienced intense folding, faulting and localized metamorphism in the late Mesozoic and Cenozoic periods (Selkregg, 1975) while the northern Foothills were only gently folded. In the Brooks Range, most of the rocks are Paleozoic limestones, sandstones, shales, and their metamorphosed forms. Rocks in the southern Foothills were not affected by metamorphism and are Mesozoic sandstones, conglomerate, limestones and chert under irregular ridges and hills separated by low areas of shale. The northern Foothills developed even crested hills underlain by sedimentary rock.

Permafrost has been defined as rock or soil, with or without including moisture or organic matter, that has remained continuously cooler than 0°C for 2 or more years (Mueller, 1943). The Foothills are underlain by continuous permafrost and features often associated with permafrost such as ice-wedge polygons, stone stripes and nets, solifluction lobes and sheets are quite common. The northern Foothills contain large amounts of ground ice, while in the southern Foothills, ground ice is less common (Lawson, 1983). The periglacial processes and geomorphologic features of this area have been extensively described in the literature (Brown and Kreig, 1983; Price, 1972; Carter et al., 1987). The upper and lower boundaries of permafrost are affected by the surface topography and the presence of lakes, streams and rivers (Price, 1972). The maximum depth of permafrost measured on the North Slope ranges between

200 and 600 m (Osterkamp and Payne, 1981) and the age of the permafrost is between tens of thousands and 2.4 million years (Carter et al., 1986).

Permafrost is the primary factor affecting geomorphic processes and land use in the Arctic (Carter et al., 1987). Ice can cement materials together, giving stability to land that may otherwise be subject to mass wasting and slopewash. When ice-rich permafrost is thawed however, much sediment can be released from hillsides and river banks. Differential settling can occur under roads and buildings as accumulations of ice melt. Ice-rich permafrost near the surface can prevent infiltration of rain and melt water, severely limiting water storage and accelerating runoff response.

Analysis of subsurface soil temperatures (Lachenbruch and Marshall, 1986; Lachenbruch et al., 1988) indicate that the region has experienced a warming trend within the last 100 years. Convective heat transfer is negligible in permafrost, so through heat conduction analysis of the top 100 m of soil, it was possible to demonstrate a widespread increase of 2° to 4°C within the last 20 to 100 years. Air temperatures measured at Point Barrow indicate a cooling trend in summer temperatures from the 1920's to the 1940's and slight warming from 1940's to late 1950's (Conover, 1960).

However the winter temperatures show a marked rise from 1920's to 1939 and then a rapid decrease to 1956.

B) The Hydrologic Regime of the Arctic Foothills

The hydrologic regime in the Arctic Foothills conforms to principles developed to describe temperate zone hydrology, but it is also affected by several other unique characteristics such as glaciers, icings (aufeis), and permafrost. However, unlike the hydrology of the temperate regions, little is known about the hydrologic regime of the Arctic Foothills. Most research on the Alaskan North Slope, hydrologic and otherwise, has been completed along the northern coast (Brown et al., 1968; Carlson et al., 1974; Carter et al., 1987; Dingman et al., 1980). Some hydrologic work has been completed in the northwestern region of the Arctic Foothills regarding rainfall runoff relations and groundwater capacity (Wilimovsky and Wolfe, 1966). A few studies have looked at the runoff regime of the large rivers which traverse the North Slope (Kane and Carlson, 1973; Scott, 1978). Extensive hydrologic research has been completed in the Canadian Arctic and Subarctic (Steer and Woo, 1983; Roulet and Woo, 1988; Roulet and Woo, 1986a and b; Marsh and Woo, 1979; Landals and Gill, 1972; McCann and Cogley, 1972; Lewkowicz and French, 1982; Ohmura, 1982a and b; Woo et al., 1981) but most of this work was done in boreal forest or on High Arctic Islands and does not easily compare with the Imnavait watershed. An underlying unity of all this research is the influence of permafrost, a great seasonal asymmetry of incident radiation, comparable climate, and the stark remoteness. An excellent review of permafrost hydrology was presented by Woo (1986).

A discussion of the water balance is a good way to begin an analysis of an arctic watershed. An analysis of each of the components and their interactions will demonstrate the complexity of the system and allow comparisons between watersheds in the Arctic and elsewhere. An understanding of the partitioning of the components is necessary to develop an understanding of the processes related to arctic hydrology. The water balance approach is quite common (Woo, 1983; Roulet and Woo, 1986a; Marsh and Woo, 1979; Landals and Gill, 1972; Woo et al., 1983; Motoyama et al., 1986a) and the expression below is from Dingman (1973b).

$$P - (E + R) = dW + dS + dL + dB + dG \qquad (1)$$

where

P = Precipitation (snow, rain, hoar frost)
E = Evapotranspiration

R = Runoff (including soil moisture and melting of ground ice)
W = Groundwater storage (below the permafrost table)
S = Soil moisture (above the permafrost table, including ice)
L = Storage in Lakes
B = Storage in biomass
G = Storage in Glaciers
d = Implies the net change over time period considered

Due to conservation of mass, this equation must balance. Often times it is possible to cancel some terms due to their inapplicability, little change over the time period considered, or a return to the level at the beginning of the time period. Ice-rich permafrost has a very low hydraulic conductivity (Kane and Stein, 1983b) and thus the groundwater system below the permafrost table does not interact in the hydrologic cycle. There were no lakes or glaciers located in the study area, so their characteristics will not be discussed.

1) Precipitation

Precipitation can occur as rain or snow, but the occurrence of snow is probable on any day throughout the year. Precipitation in the Arctic is notably low, with the highest amounts of precipitation normally occurring during late July, August and September (Kane and Hinzman, 1988). Estimates on the percentage of yearly precipitation which occurs as snow ranges between 30 and 80% (Haugen, 1980; Benson, 1982). The difficulty in measuring snowfall precipitation on the Arctic Slope has been noted by many authors (Benson, 1982; Black, 1954; Clagett, 1988; Hall et al., 1986; Woo and Marsh, 1978) however the importance of snow in the arctic hydrologic regime is clear (Carlson et al., 1974; Kuusisto, 1984a; Woo, 1982a). Snow accumulates, sublimates, and is redistributed by strong Arctic winds for about nine months each year resulting in well defined drifts and a very uneven distribution (Benson, 1969; Liston, 1986: Woo, 1982a). Depth of the snowpack can vary by several orders of magnitude within a few meters spatially and density can commonly vary by a factor of 3 within the snowpack profile (Liston, 1986).

2) Evapotranspiration

Evapotranspiration is a difficult parameter to measure in any environment, particularly so in the Arctic. A Soviet analysis of the arctic water balance concludes that there is almost no evaporation during the cold season (Briazgin and Korotkevich, 1975). Sublimation of wind blown snow can be a significant loss (Schmidt, 1986), however, this is debated in the literature (Heron and Woo, 1978). Theoretical and experimental work in Finland and Sweden (Kuusisto, 1984a; Lemmelä and Kuusisto, 1974; Bengtsson, 1980; Lemmelä, 1972) has shown that evaporation from the snowpack is very small during the winter, increasing during the spring melt.

Evaporation during the summer is a significant component of the water balance (Kane and Hinzman, 1988; Rouse and Mills, 1977; Ohmura, 1982b), though some studies neglect to include it in analysis (Onesti and Walti, 1983). Evaporation can be computed from the remainder term of the analysis of the energy budget (Weller and Holmgren, 1974), from the remainder term in a water balance (Kane and Hinzman, 1988); it can be estimated directly from meteorologic measurements such as Thornthwaite's equation (Kane and Carlson, 1973; Patric and Black, 1968), the Priestley and Taylor (1972) equation, or the Bowen ratio method (Roulet and Woo, 1986a; Ohmura, 1982a); it can be measured directly using lysimeters (Marsh et al., 1981); or it can be estimated by comparison to pan evaporation (Haugen et al., 1976).

3) Runoff

The runoff process is perhaps the most studied component of the water balance. It certainly has many implications relating to engineering (Seagel and Parish, 1974) and ecology (Harper, 1981; Craig, et al., 1984). The mechanisms of runoff have been subdivided into many categories including hillslope hydrology and channel runoff, including both snowmelt runoff and rainfall runoff.

a) Hillslope Hydrology

Surface runoff from hillslopes is often described as either overland (or sheet flow) or rill (inter-tussock) flow, the latter being more common in hummocky terrain (Woo and Steer, 1982). The presence of permafrost prevents infiltration and maintains a perched water table within the active layer. Consequently, surface flow is more likely due to the supra-permafrost water table rising above the ground level rather than precipitation or melt rates exceeding infiltration rates into the surficial organic soil (Dingman, 1973a). Woo and Steer (1982) have determined that the active layer must be saturated and depression storage must be satisfied before surface runoff can begin. Water tracks (Kane et al., 1990) appear to be a unique feature of permafrost hydrology. These small drainage channels are nearly parallel to each other and to the slope. The effect of these water tracks is to rapidly convey water from the slope to the stream giving rise to a very rapid response to rainfall or snowmelt events.

The character of the slope hydrology is influenced by the depth of thaw of the active layer (Woo and Steer, 1982; Woo and Steer, 1983; Wright, 1983). Since the frozen soils are impervious, the input from rainfall or snowmelt is stored in the thawed zone or runs off in overland or rill flow. Initially, the thaw rate of the active layer is rapid, gradually slowing through the thaw season (Wright, 1983), approximating an empirical function of the square root of time (Woo, 1976). Maximum storage capacity is reached when the maximum depth of thaw reaches the permafrost table (Woo and Steer, 1982).

During the spring melt, while an ice-rich layer is near the surface, the runoff response time to melt is fast (Woo and Steer, 1982). As the depth of the active layer increases, storage of moisture in peaty topsoil and mosses increases and storm recessions become more lengthy (Anderson, 1974; Anderson and Neuman, 1984; Wright, 1983). It has been noted in places where a porous organic layer overlies a fine-grained mineral soil, water would flow in the organic layer on the surface of the mineral layer regardless of the depth of thaw (Douglas, 1961) probably due to the usually saturated conditions of active layer.

b) Channel Hydrology

Although warm springs which penetrate the permafrost do exist on the North Slope, these are relatively rare and normally several hundred meters of ice-rich permafrost will isolate surficial runoff from subsurface groundwater. For this reason, arctic hydrology differs significantly from temperate or even subarctic hydrology (Dingman, 1975; Church, 1974). Extreme winter temperatures also amplify the break-up and freeze-up processes when compared to more southerly locations by preventing mid-winter thaw events and accelerating the early winter freeze (Woo, 1986).

The hydrologic regime of a non-glacierized arctic basin, with little possibility of groundwater input is determined primarily by the melting of snow and ice. Such basins have been defined as arctic nivial regimes by Church (1974). The spring snowmelt is normally the dominant hydrologic event of the year, yielding most of the annual runoff and usually the peak flows (Ambler, 1974; Anderson, 1974; Carlson et al., 1974; Woo, 1982a; Woo, 1983; Lewkowicz and French, 1982; Roulet and Woo, 1986a).

The initiation of streamflow can be delayed several days after hillslope runoff begins due to the presence of deep snowpacks in the valley bottoms (Woo and Sauriol, 1981). Unlike subarctic rivers and some arctic rivers which are fed throughout the winter by springs, the amount of ice in high arctic streambeds is minimal (Woo, 1982a; Woo, 1983). This is due to the absence of winter runoff and basal flow (Woo, 1986). However, the presence of ice in stream bottoms will increase initial streamflow velocities and can increase sediment transport (Woo et al., 1982).

Channel flow usually ceases or substantially decreases within a few days after the spring melt event generally remaining low except in response to significant rain events (Ambler, 1974; Anderson, 1974; Carlson et al., 1974; Woo, 1982a; Woo, 1983; Lewkowicz and French, 1982; Roulet and Woo, 1986b).

4) Soil Moisture

The level of moisture in the soil is dependent upon the thermal and the precipitation regimes and strongly affects the rates of evapotranspiration (Braley, 1980), runoff (Dingman, 1973a; Kane and Stein, 1983c) and active layer thaw (Drew et al., 1958). The antecedent moisture condition has been shown to be an important factor in modeling snowmelt and rainfall runoff (Kane and Stein, 1983a and b; Myrabo, 1986). The amount of moisture in the soil will be affected by many factors including weather, vegetation, topography, soil type (Anderson, 1988) and depth to permafrost (Drew, 1957); and in many basins, the natural variation is great.

Moisture will migrate from the ice-rich soil into the snowpack in response to a strong temperature gradient throughout the winter (Benson and Trabant, 1972; Woo, 1982b; Santeford, 1978; Smith and Burn, 1987). This migration will open some of the soil pores allowing limited infiltration of spring snowmelt water. It has been shown that snowmelt infiltration into permafrost is less significant than infiltration into seasonally frozen soils (Kane et al., 1978). The infiltrating water quickly refreezes when it contacts the colder soils effectively preventing further infiltration (Woo et al., 1982; Woo and Heron, 1981).

C) The Thermal Regime of the Arctic Foothills

The arctic hydrologic regime is closely tied to the thermal regime (Williams, 1974; Woo and Steer, 1986; Wright, 1983), energy and water fluxes are strongly linked (Woo, 1986), thus even a partial understanding of one must entail a study of the other. Energy transfers in the Arctic of course obey the laws of physics, but the magnitude of each component of the surface energy balance yields a study of extremes. Again, very little research has been completed investigating the thermal regime of the Arctic Foothills. Most of the data available have been collected along the Arctic Coastal Plain of Alaska or in the Canadian Archipelago.

Most analyses of the thermal regime of the Arctic are subdivided by season due to the extreme annual variation in radiation, air temperature, and the consequential effects of the snowpack (i.e. insulation, albedo, roughness). A common categorical description of each season according to the thermal regime is pre-melt, snowmelt, post-melt, mid-summer, freeze-up, and winter (Weller and Holmgren, 1974; Rydén, 1976).

The surface energy balance represents a conceptually simple method to examine the processes which control the thermal regime. The surface energy balance has been used frequently for calculating rates of snowmelt (Moore, 1983; Motoyama, 1986), ground thawing (Pavlov, 1981) or freezing (Pavlov, 1975), and for effects on climate (Ohmura, 1981). The following formulation is adapted from Price et al. (1976), Lunardini (1981) and Moore (1983). Like the water balance, the surface energy balance is the sum of its components, each of which varies in its difficulty of measurement and associated error. The surface is defined as the interface between the atmosphere and the earth, its vegetative cover or the snow surface when applicable.

The energy transfer at the surface is the sum of the radiant, convective, conductive and latent energy exchanges. Due to the principle of conservation of energy, the balance at the surface can be written as

$$Q_{rad} + Q_{conv} + Q_e + Q_f + Q_{cond} + Q_{bio} + Q_{precip} = 0 \quad (2)$$

where Q_{rad} is the net energy transfer due to radiation, Q_{conv} is the convective component at the surface, Q_e is the latent heat due to sublimation, evaporation or condensation, Q_f is the energy transformed in melting or freezing, Q_{cond} is the energy conducted through the ground or snow layer, Q_{bio} is the energy consumed in biochemical processes and Q_{precip} is the energy transfer in precipitation. This equation considers the surface a contact layer with no mass, therefore no energy is stored. Heat flow out of the surface is a loss and therefore negative, heat flow in is positive. Evaporation requires energy and is negative, condensation adds energy to the layer and is positive. Freezing releases latent heat and is positive, thawing requires energy and is negative.

1) Radiation

Gavrilova (1963) has stated the radiation is the most important component of the arctic heat balance, however the significance of radiation is greatly dependent upon the time of year. Radiation becomes the dominant heat transfer process between snow ablation and snow accumulation (Weller and Holmgren, 1974). One mechanism for melting at the base of the snowpack is thought to be due to shortwave radiation penetrating through the snow

(Kuusisto, 1986). Emittance of longwave radiation from snowfree areas is thought to accelerate the melt process (Roland, 1984).

The net radiation is the sum of the effects of incoming and outgoing shortwave and longwave radiation. Incoming shortwave radiation, S_{in} includes direct, $Q_{s\ dir}$ and diffuse $Q_{s\ dif}$, portions of which are reflected, S_{out}, depending upon the albedo, α. Incoming longwave, L_{in} is dependent upon sky temperature and cloud conditions. A small portion of incoming longwave is also reflected. Longwave radiation is also emitted, L_{emit}, from the surface as a property of the surface temperature and longwave emissivity. It is quite difficult to separate reflected and emitted longwave radiation and normally emitted is much greater than reflected so they are often considered as simply outgoing longwave, L_{out}. The net radiation can be measured directly with a net radiometer or it can be calculated from measurements of the components. Each particular measurement requires a specifically designed instrument. The relation between net radiation and its components is

$$Q_{rad} = S_{in} - S_{out} + L_{in} - L_{out} \tag{3}$$

Net radiation may also be calculated from fewer measurements:

$$Q_{rad} = (Q_{s\,dir} + Q_{s\,dif})(1 - \alpha) + L_{in} - L_{out} \qquad (4)$$

Albedo, α, is the ratio of reflected shortwave to incoming shortwave.

$$\alpha \equiv \frac{S_{out}}{S_{in}} \qquad (5)$$

The surface will emit longwave radiation as a gray body according to the following function

$$L_{emit} = \sigma\, \epsilon_s\, T_s^4 \qquad (6)$$

T_s is the absolute surface temperature, σ is the Stefan Boltzman constant and ϵ_s is the emissivity of the surface.

The longwave radiation emitted by the atmosphere, L_{sky}, towards the earth can be described by the function

$$L_{sky} = \epsilon_a\, \sigma\, T_a^4 \qquad (7)$$

where ϵ_a is the effective emissivity of the atmosphere and is itself a function of the atmospheric vapor pressure, e_o, and the absolute air temperature, T_a.

2) Convection of Sensible Heat

Perhaps the most difficult component to measure is the contribution of convective or turbulent transfer of sensible heat. This is the transfer of energy via the movement of a fluid. Convection is thought to be the primary mechanism for growth of the hoar frost crystals at the base of the snowpack (Benson, 1969). The energy exchange from the wind to the surface is a significant heat transfer mechanism and must be considered (Lunardini, 1982).

Calculations of convective heat transfer are dependent upon the atmospheric stability or the relationship between the adiabatic lapse rate and the environmental lapse rate, Γ. This can be determined from measurements of the air temperature profile, dT/dz, where z is the height above the surface and T is temperature. The near surface atmosphere is in a state of dynamic equilibrium and can be classified in three conditions, neutral, stable or unstable (Figure 2-1). In the stable case, any disturbance from the given state will be resisted and the system will tend to remain at that state. In neutral equilibrium, a disturbance will not be resisted, but the system will remain in equilibrium with the state although it may traverse a series of states before returning to equilibrium. An unstable equilibrium is such that any disturbance will be reinforced and the system will change to a new state. Following the formulation of Lunardini (1981), the equation governing sensible heat exchange in neutral conditions is:

$$Q_{conv} = \rho_a \, C_{pa} \, D_h \, (T_a - T_s) \qquad (8)$$

where ρ_a is the density of air, C_{pa} is the specific heat of air at a constant pressure, D_h is the convective heat exchange coefficient.

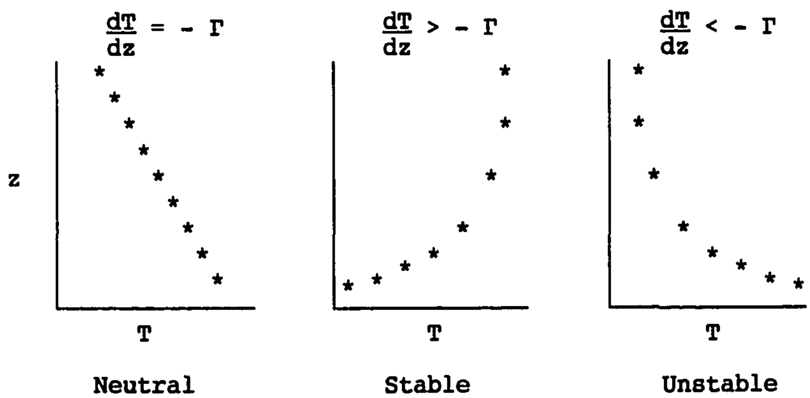

Figure 2-1. Atmospheric Stability.

3) Vaporization and Condensation

The equation governing latent heat exchange in neutral conditions for evaporation or sublimation is

$$Q_e = L_v \, \rho_a \, D_w \, \frac{0.622}{p} \, (e_a - e_s) \qquad (9)$$

where L_v is the latent heat of vaporization, D_w is the exchange coefficient for moisture, e_a is the vapor pressure of the air, e_s is the vapor pressure at the surface and p is the atmospheric pressure. The exchange coefficients for heat and moisture are related and may be derived by analogy with the transfer of momentum

$$D_h = D_w = \frac{k^2\ u_z}{[\ln(z/z_o)]^2} \qquad (10)$$

where k is von Karman's constant, 0.4, z_o is the surface roughness, z is the height at which the windspeed was measured. In stratified conditions, stable or unstable, a correction factor, such as the Richardson number, R_i, should be used to adjust Equation 10. The Richardson number is the ratio of the buoyant force to the eddy shear forces. Under stable conditions, air near the ground is cooled, has a greater density and thus will maintain its position if disturbed by turbulence. The stable exchange coefficient, $(D_h)_s$, can be found by

$$(D_h)_s = \frac{D_h}{(1 + 10\ R_i)} \qquad (11)$$

The Richardson number is derived as follows

$$R_i = \frac{g\ (dT/dz + \Gamma)}{T_{abs}\ (du/dz)^2} \qquad (12)$$

where g is the acceleration of gravity, dT/dz is the temperature gradient, T_{abs} is the temperature of the air layer in Kelvin, and du/dz is the windspeed gradient.

To determine the exchange coefficient for unstable conditions, $(D_h)_u$ we can use the following formula:

$$(D_h)_u = D_h\ (1 - 10\ R_i) \tag{13}$$

The roughness length can be determined from the wind profile.

$$\frac{u_1}{u_2} = \frac{\ln(z_1/z_o)}{\ln(z_2/z_o)} \tag{14}$$

Latent heat is also a consideration in the freeze/thaw of the active layer and the melting of snow. The amount of energy released/absorbed in the process of phase change, Q_f, can be determined through

$$Q_f = \int_0^h L_f\ W\ \rho\ dh \tag{15}$$

where W is the water content in the snow or soil, h is the thickness of the snow or active layer, and ρ is the density of snow or water.

4) Conduction

The significance of conduction varies depending upon time of year and point of view. The mechanism of heat transfer through the soil profile is primarily by conduction (Nixon, 1975) and, along with the latent heat, is the key consideration in determining rate and depth of thaw (Lunardini, 1988) in ice-rich frozen soils. Conduction is also a mechanism of heat transfer on the molecular level between the surface and the atmosphere, however, since completely calm days with stable lapse conditions are quite rare in the Arctic, this mechanism is not very important. Conduction is a very important process in basal melt of the snowpack in the more temperate regions such as Finland (Kuusisto, 1986) or Japan (Motoyama, et al., 1986b). During snowmelt in the Arctic, the near surface of the active layer is isothermal (Weller and Holmgren, 1974) and does not contribute to basal snowmelt.

Probably the easiest component to quantify is conductive heat transfer; its components are easily measured with the greatest degree of accuracy. Assuming no phase change, the one dimensional conductive heat transfer can be described by

$$\frac{\delta T}{\delta t} = \frac{K}{c} \frac{\delta^2 T}{\delta z^2} \tag{16}$$

where K is the thermal conductivity of the material (snow, ice, water, soil, or any combination) and c is the volumetric heat capacity. In problems involving a thawed layer in contact with a frozen layer, it is necessary to use two equations of the type of Equation 16 to describe the system.

$$\frac{\delta T_t}{\delta t} = \frac{K_t}{c_t} \frac{\delta^2 T_t}{\delta z^2} \tag{17}$$

$$\frac{\delta T_f}{\delta t} = \frac{K_f}{c_f} \frac{\delta^2 T_f}{\delta z^2} \tag{18}$$

where T_t, K_t, c_t represent the thawed temperature, thermal conductivity and heat capacity while T_f, K_f, c_f represent the same components of the frozen system. Equations 17 and 18 are coupled via the boundary conditions at the interface.

$$T_t(x,t) = T_f(x,t) = T_m \tag{19}$$

$$K_f \frac{\delta T_f}{\delta x} - K_t \frac{\delta T_t}{\delta x} = \rho L_f \frac{dx}{dt} \tag{20}$$

T_m = phase change temperature
at x = phase change interface
t = some time greater than zero

To correctly calculate the heat transfer into the soil, it is necessary to obtain measurements or accurate estimates of the heat capacity (c) and the thermal conductivity in both frozen (K_f) and thawed states (K_t), and the amount of latent heat of fusion (L_f) required for phase change. The soil profile on the Arctic Slope often consists of stratified layers of differing thermal properties (McGaw et al., 1978; Hinzman et al., 1990) which will impact the soil heat flow.

D) Interaction of the Hydrologic and Thermal Regimes

The response of the hydrologic regime to the thermal regime and vice versa is a complicated, dynamic and self-regulating mechanism which manifests itself in the processes of soil moisture migration (Fukuda, 1982; Lundin, 1989), evapotranspiration, sublimation, and the formation or degradation of permafrost. The latter process is the crucial determinant in the geomorphologic response to disturbance (Lawson, 1986). The energy required for phase change (ice to water and water to vapor) dissipates greater than half of the annual energy balance (Rydén, 1978; Weller and Holmgren, 1974) and dominates the energy balance during break-up and freeze-up (Wendler, 1967; Ohmura, 1981; Eaton and Wendler, 1982). Phase transition of water is the most important factor in cryogenic processes (Brown and Grave, 1979). The amount of moisture in the soil profile will affect the depth and rate of freezing; this effect is accentuated if the water is moving (Price, 1983). The soil moisture content is one of the important factors determining the temperature field in the active layer (Pavlov, 1975) as the thermophysical properties are determined primarily by density and moisture content

(Nixon and McRoberts, 1973).

With the advent of the computer age, modeling has become a common approach to explain, define, understand or predict the behavior of almost any physical or social system. The application of computer models to aid in analysis of systems in the Arctic has been extremely useful for several reasons. Computer modeling allows one to cope with the lack of extensive data sets through numerical analysis and extrapolation of existing data. It also allows an analyst to cope with data sets too large to reasonably handle without machine processing. Analysis of the interactions of varying processes is often too complex without computer assistance. Dynamic systems which change only perceptibly over large time scales can be analyzed conceptually through models to deduce mechanisms of growth or change. The effects of possible changes in a system can be investigated before such changes become reality.

As previously discussed, it is impossible to study the thermal regime without considering the hydrologic regime and vice versa, however, it is reasonable to discuss previous modeling experience based on the primary goals of the research.

1) Modeling the Thermal Regime

Extensive efforts have gone into predicting the thermal regime of the Arctic and Subarctic in order to prevent possible aggradation or degradation of permafrost, prevent frost heave or predict snowmelt runoff (Johnson, 1981). Applications include determining the effects of roads, buildings, airports and the snow they catch on the active layer. Thermal models also aid in design of structures to determine amount and placement of insulation or to estimate the number and size of thermopilings required for support or to maintain permafrost temperature.

a) Modeling the Soil Thermal Regime

Initial attempts of simulating the thermal regime of the soil profile were empirical and analytical equations, however due to the strong non-linearity of phase change, few complete solutions exist (Lunardini, 1981). For a homogeneous system with a uniform initial temperature, the modified Berggren equation is a useful one-dimensional approximation (Zarling et al., 1989). The depth of the 0°C isotherm is given by

$$X = \text{Lambda} \left[\frac{2\ K\ I}{L_f} \right]^{\frac{1}{2}} \qquad (21)$$

where K is the thermal conductivity, I is the freezing or thawing index, L_f is the volumetric latent heat and Lambda is a coefficient.

The mathematical difficulties that arise with a moving phase change boundary often necessitate the use of numerical methods, especially involving cases with complicated geometries. Based on the concept of conservation of energy, researchers were able to couple the surface energy balance and the subsurface thermal regime (Goodwin and Outcalt, 1973; Outcalt et al., 1975; Smith, 1975). Effects of tundra vegetation on soil temperature and vice versa have also been incorporated into thermal models (Ng and Miller, 1973).

Much of the work related to modeling soil temperatures has evolved in response to the problems of frost heave in frozen ground (Smith, 1985). Extensive experimental and theoretical work has been completed on the problems of coupled heat and moisture flow and mechanisms of heat flow in soil (Sheppard et al., 1981; Williams and Wood, 1985). Advances in this field have led to development of many numerical formulations of heat transfer (Goering and Zarling, 1985; Lundin, 1989).

b) Modeling Snowmelt

Investigation of snowmelt is of prime importance for most hydrologic studies in northern regions. Several techniques have been developed for estimating snowmelt, rates of snowmelt, or snowmelt runoff; these are usually based on physics, statistics, or index methods. The only physically correct method of snowmelt calculation is through a complete energy balance of the snowpack (U.S. Corps of Engineers, 1956). This technique leads to a series of complicated equations with extensive data requirements collected with expensive equipment which is usually not available to most practicing hydrologists. The processes of heat transfer within and to the snowpack are highly variable both spatially and temporally and are extremely difficult to measure. The complexity of this approach has led to development of many simplified snowmelt equations (Bengtsson, 1976; Bengtsson, 1982). These other approaches have been summarized by Kuusisto (1984b). He separated them into four divisions:

> 1. Equations involving the measurement or estimation of the radiation balance plus measurements of humidity, wind and temperature variables.
>
> 2. Equations involving humidity, wind and temperature variables. The radiation balance is taken into account by combinations of these

variables; e.g. the shortwave balance may be assumed to be proportional to the diurnal range of air temperature.

3. Equations involving only wind and temperature variables.

4. Equations involving only temperature variables and their combinations.

One may expect that performance of the method may be equated with its complexity, but that is not necessarily true. Kuusisto (1984b) compared these different techniques of snowmelt estimation and found that in most applications, methods involving only mean temperature were often adequate. Bengtsson (1976) compared the rate of snowmelt using the energy balance method and a degree-day method. Both methods produced nearly identical results. Braun (1985) compared a temperature index method, a temperature and wind index method, a combination method, an extended combination method which also considered water vapor pressure, and the energy balance approach for calculating snowmelt in several basins in Switzerland. He found during advection melt conditions, the energy balance methods yielded the best results. During radiation melt conditions, index methods were adequate. The combination method was best for operational use. Ishikawa et al. (1986) compared measured snowmelt to calculated snowmelt on an hourly and daily basis and found that simple formulae were not adequate for smaller time increments but energy balance methods could predict snowmelt for short or long-term periods. Vehviläinen and Kuusisto (1984) analyzed snowmelt data from a highly variable terrain in Finland and compared performance of snowmelt models with one to five parameters. They found melt factor, threshold temperature, and liquid water holding capacity were essential components, however little improvement developed with higher degree formulations. Dunne et al. (1976) found that snow depth and melt rate were the most important factors involved in snowmelt calculations if one used the energy balance approach. They also found lag time to be a significant problem.

The mechanisms of energy exchange within and to the snow surface are well understood, however accurate measurements of these transfer processes at a point are difficult, and extrapolation to large areas introduces several more uncertainties. Determination of snowmelt from an energy balance approach is particularly difficult over a large watershed (Male and Granger, 1981). It is difficult to extrapolate measurements from a point source to an area, this is particularly true for turbulent exchange processes (sensible heat and evaporation-condensation). It is also uncommon to have adequate data (wind velocity, air temperature, humidity) to apply an energy balance over a large areas.

The energy fluxes which control rates of snowmelt are the sensible and latent heat exchange, the radiation exchange, the ground heat flux, and the heat transfer due to precipitation. The radiative heat flux is normally considered to be the dominant mechanism of heat transfer, however the turbulent heat transfer associated with sensible and latent heat transfer is always very important (Wendler, 1967; Male and Granger, 1981; Eaton and Wendler, 1982). The ground heat transfer can be quite significant in certain regions of the world such as Japan (Motoyama et al., 1986b) or in the Pacific Northwest, but is usually much less in regions where the soil remains frozen during snowmelt (Wendler, 1967). The significance of precipitation also varies greatly regionally, having essentially no impact in northern Alaska due to very low precipitation during the spring melt period (Wendler, 1967; Weller and Holmgren, 1974), and often being the dominant term in areas such as the Pacific Northwest United States. Price and Dunne (1976) present an excellent discussion of application of the energy balance method to calculating snowmelt on large plots.

There are dozens of models available which include some routine for calculating snowmelt; among these are HBV (Bergström, 1976), SHE (Abbott et al., 1986a and b), TANK (Sugawara et al., 1984), and SWM IV (Crawford and

Linsley, 1966). In most operationally used models, snowmelt is calculated through an index method, however some models have been developed in which calculations are much more complex and perhaps even elegant. Motoyama (1986) used an energy balance method in combination with the TANK model to calculate snowmelt runoff. The SHE model uses the energy budget approach to calculate snowmelt and then uses Richard's equation to model the flow of mass within the snowpack (Morris, 1983; Abbott et al., 1986a and b). Anderson (1976) developed an energy and mass balance model for calculations of snowmelt at a point. His finite difference model is probably one of the more theoretically complete approaches.

2) Modeling the Hydrologic Regime

There is great economic incentive to create and use accurate hydrologic models almost everywhere on Earth. Management of water resources is becoming more important as more communities vie for a limited water supply, as damage from floods increase due to more development in floodplains (Hitch, 1984) and as hydro-generated electricity becomes more desirable (Whitehead, 1983). Knowledge of snowmelt runoff is necessary to operationally manage reservoir systems.

a) Snowmelt Runoff Models

In many parts of the world, input from snowmelt represents a large portion of the total yearly precipitation, especially in far northern or mountainous regions. The thermodynamics of snow and snowmelt have been studied for decades but incorporation of these processes into hydrologic models remains highly empirical. Runoff from snowmelt represents the final step in a process which includes accumulation, sublimation, evaporation, snowmelt, infiltration and storage in the snowpack or soil. Most of the important processes of snowmelt runoff are addressed in Colbeck and Ray (1978) or in IAHS (1972). Tesche (1986) states that "Development of a simulation model of snowmelt-runoff requires (1) identifying the relevant physical processes and (2) determining the most appropriate mathematical representation for all or a subset of these phenomena." There are two primary approaches to satisfying these criteria. A statistical or systems model bases its predictions upon the assumption that a cause-effect relationship exists in empirical data using indices and correlations. These models can be called operational, applied, empirical, black-box, or parametric models. This approach is convenient because it directly links measurable data (input) with the desired results (output). A distinct disadvantage is statistical models perform poorly during extreme events. The second approach is a deterministic or physical model which is based upon the mathematical forms of the conservation of mass, momentum or energy. These models are also called basic, pure, causal, dynamic or theoretical approaches. Deterministic models are inherently more complicated and require much more data to operate (Tesche, 1986).

Snowmelt runoff is so important in certain applications, specialized models have been designed with this sole purpose. Many models have been developed in different physiographic regions, but presently no model has been accepted as the universal best program (World Meteorologic Organization, 1986). This is because each model is somewhat different in the approach or method of calculation. A system developed for a prairie environment may not function well in a mountainous region. Typically, models applied in areas where the snowpack is large and uniform and topography is steep will perform better. In areas such as the prairie or tundra where the snowpack is thin and variable, depression storage is large and the components of the runoff process all represent a significant proportion of the snowpack, large errors can occur (Gray et al., 1986).

The Martinec-Rango Snowmelt Runoff Model is designed to model streamflow in mountainous basins where snowmelt is a major runoff component (Rango and Martinec, 1979). The model performed well on deep, long lasting snowpack (Miller, 1986). Granger et al., (1984) developed a snowmelt model for a prairie environment. They found accurate simulation of infiltration was a vital component to runoff prediction. Sand and Kane (1986) found that considering the changing rate of infiltration with thawing soil improved the performance of the HBV model in simulating spring snowmelt runoff and early summer rainfall. Gray et al., (1986) developed a physically based approach to describe snowmelt infiltration into frozen soils.

Dozens of other models exist. Many of these are summarized by Renard et al. (1982), Tesche (1986) and Singh (1988 b). Ten snowmelt models were directly compared against each other by the WMO (1986) on six test basins with ten years of data. The cooperating modellers found it impossible to rate the models in order of performance. The also could not relate model performance to model complexity. Bergström (1986) discusses the optimum complexity, the variabilities in natural watersheds, the problem of losses and how these factors affect modeling of snowmelt runoff.

b) Rainfall Runoff Models

Hydrologic models have proliferated in the last 15 years for several reasons. Among them are almost universal access to relatively inexpensive machine processing, refinement of the algorithms which can be used to describe the runoff processes and an increase in the need for more accurate forecasts of rainfall runoff. The increase in the knowledge of runoff processes has led to proven success in modeling efforts which in turn encourage more expanded attempts at modeling hydrologic systems.

As in snowmelt modeling, there are two primary approaches to modeling rainfall runoff, the statistical or systems approach and the deterministic or physical approach. A physical approach to modeling rainfall runoff would require specification of the system input, system structure, physical laws and the initial and boundary conditions (Singh, 1988a). The system input requires the space-time distribution of rainfall. The system structure implies information on the watershed physiographic characteristics. Physical laws must include equations for continuity, momentum, and friction. Initial and boundary conditions would include information on the antecedent moisture condition. Precipitation, evapotranspiration, interception, infiltration, overland and open channel flow, subsurface flow and surface storage are all processes which must be considered in one or all of the above (Dingman, 1984). A physical model must be independent of its intended application and it must incorporate the known physical system and the whole modeled process (Singh, 1988a).

In a systems approach, one is concerned with the input-output relationship not with the nature of the system itself (Singh, 1988a). Thus one may overcome the difficulties encountered with the complexity of input, the system structure, and the governing physics. In a systems approach to modeling rainfall runoff, one may use a spatially lumped continuity equation and a storage-discharge relation (Klemes, 1974). The watershed is represented as a series of storage elements or reservoirs. The parameters are determined from historical records and the resulting output is the hydrograph of direct runoff. These parameters are usually site specific and the model must be adjusted accordingly for each new application.

Most hydrologic systems are complex and cannot be understood in complete detail, therefore the system is replaced with a model of similar, but simpler structure. Thus the purpose of a model is to simulate the operation of a system that is exceptionally complex and predict the effect of changes on this operation (Dooge, 1972). Theoretical models are those which most nearly represent the real world system (Singh, 1988a). A theoretical model has a logical structure similar to the real world, but has been simplified somewhat and consequently is still an approximation of reality. Darcey's law could be considered a theoretical model (Dingman, 1984). An empirical model is not based upon physical laws, it is merely a representation of the data. If the conditions change, it has no predictive capability. It should not be used outside the range of the calibration data. The unit hydrograph is an empirical model (Heerdegen, 1974). A conceptual model is something in between the previous two. Generally it considers the physical laws, but in a simplified form. It provides results efficiently for some problems using parameters which have some physical significance and can therefore be estimated by concurrent observations of input and output. HBV is a conceptual model (Bergström, 1976).

III) SITE DESCRIPTION

Imnavait watershed lies in the Foothills between the Kuparuk and Toolik Rivers on the North Slope of Alaska (Figure 3-1). The topography in the watershed consists of low rolling piedmont hills with a wavelength of 1 to 2 km and amplitudes of 25 to 75 m and an average elevation of about 900 m (Figure 3-2). The bedrock is composed of shale, sandstone, conglomerate, limestone, and chert of Cretaceous, Triassic and Mississippian ages (Douglas, 1961). This area was glaciated during the Pleistocene.

Through Carbon-14 dating, Walker et al. (1989) have determined the age of these soils to be at least 11,500 years ± 140. The soils are mostly silty colluvium and residual material of glacial origin and are classified as Histic Pergelic Cryaquepts (Rieger et al., 1979). There is a thick layer of organic matter on the surface consisting of partially decomposed mosses, sedges and other associated plants. The soil profile experiences frost churning which in effect mixes pieces of the organic mat downward so a layer of organics is sometimes found on the surface of the permafrost table. The surficial organic layer is quite porous and saturates and drains quickly. However the underlying mineral soil

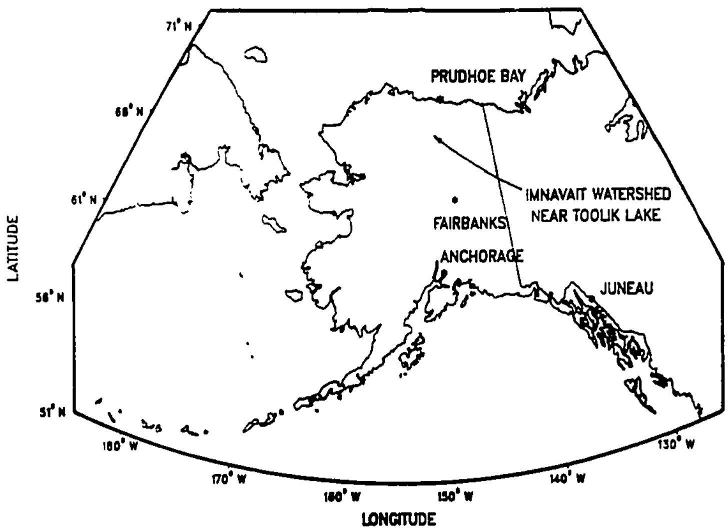

Figure 3-1. Map showing site location of Imnavait watershed.

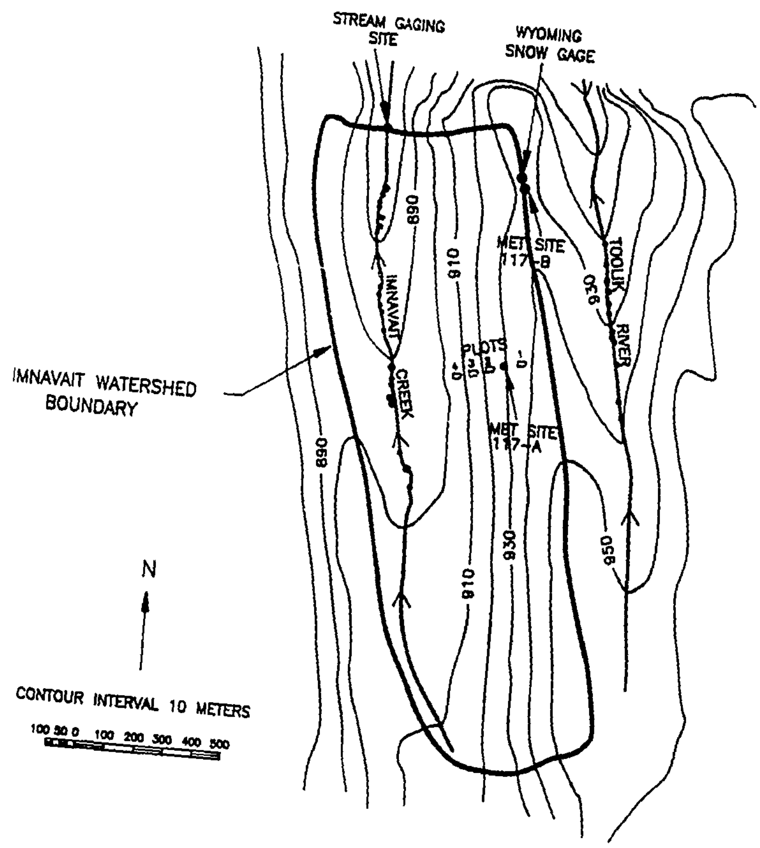

Figure 3-2. Topographic map detailing location of plots, watershed boundary and meteorologic sites.

is usually saturated with water, depleting the soil of oxygen during the thaw period. Price (1972) says frost action is the dominant factor in denudation, with wind action and running water being secondary, however Haugen (1980) says wind is a major environmental factor. Such soils are the most common on the North Slope (Rieger, 1983) and are similar in many respects to soils described by Brown (1967), Drew (1957), Douglas (1961), Kellogg and Nygard (1951), Tedrow and Brown (1962), Ugolini and Tedrow (1963), Ugolini et al. (1963), and Tedrow (1977).

Although many of the landscape features often associated with permafrost such as ice mounds, polygons, or ice wedges were not conspicuous in the watershed, the effects of frost action were still evident in the presence of translocated organics on the surface of the permafrost table and frost boils. Cryoturbation is believed to be an extremely slow process (Drew, 1957) and is apparent in a only a small area of the watershed. Ice-wedge polygons are located outside the watershed in poorly drained depressions and in drained lake basins (Brown and Berg, 1980). Extensive ground ice has been located in end and ground moraines 10 km north of the watershed (Brown and Berg, 1980). Massive ice can be present in large amounts while displaying little surface expression.

The vegetation is mostly water tolerant plants such as tussock sedges and mosses, but they are accompanied by lichens and shrubs such as willows, alder and dwarf birch. Tussock sedges are common on the ridge tops and slopes but do not occur in the lowest, wettest areas. More complete descriptions of tundra vegetation have been published (Walker et al., 1980; Jorgenson, 1986; Brown and Berg, 1980; Walker, 1985). The soils and vegetation of Imnavait watershed were especially well described by Walker et al. (1989). The geochemistry of the precipitation, water tracks, and stream of this watershed are described in detail by Everett and Ostendorf (1988).

The watershed is underlain by continuous permafrost, the presence of which affects the landscape patterns, microclimatology, hydrologic processes, and thermal regime. The climate has been described as both humid (Price, 1972) and semi-arid (Conover, 1960), but such classifications have little meaning because the nearness of permafrost to the surface and the relatively low rates of evapotranspiration keeps the moisture content of the soils generally high. The climate is most simply described, and universally agreed upon, as cold. Imnavait watershed has a continental climate, characterized by long, cold winters and short, cool summers. During the winter, the climate of the arctic tundra is affected primarily by radiative heat loss and atmospheric circulation (Ohmura, 1981). Net radiation becomes positive near the end of April and abruptly increases during snowmelt. Prior to the onset of spring melt, which normally occurs near summer solstice, the surface is characterized by a homogeneous high albedo. As the melt progresses, due to the uneven distribution of snow, the albedo of the surface also develops an extreme variability (Liston, 1986). Between the period of snowmelt and snow accumulation, the surface maintains its lowest annual albedo which results in the maximum annual energy exchange at the surface (Ohmura, 1981). The initial snow accumulation in the autumn is usually close to equinox and, since solar radiation is considerably smaller, the arrival of the new snow cover does not cause the dramatic changes in the surface energy and water balances as during spring snow ablation. (Ohmura, 1981). A detailed discussion of the climate at the interface of the tundra and treeline is presented in Rouse (1984 a, b, and c.).

Yearly precipitation is approximately 30 cm, about half of which occurs as snow (Brown and Berg, 1980; Hartman and Johnson, 1978). The average high temperature in July is about 13°C, and average low temperature in

January is about -27°C (Hartman and Johnson, 1978). During the course of the study, the minimum recorded air temperature was -56°C, the maximum was 34°C, and the average temperature was -7.4°C.

IV) HYDROLOGIC AND THERMAL PROPERTIES OF THE ACTIVE LAYER IN THE ALASKAN ARCTIC

Hinzman, L.D., D.L. Kane, R.E. Gieck and K.R. Everett. 1990. Hydrologic and thermal properties of the active layer in the Alaskan Arctic. Accepted for Publication in Cold Regions Science and Technology.

A) Abstract

Almost all biological activity in far north regions takes place within a shallow zone above the permafrost called the active layer. The active layer is the surficial layer of the soil system which thaws every summer. In Imnavait watershed, a small headwater watershed north of the Brooks Range on the North Slope of Alaska, the active layer is an extremely variable multilayered system consisting of a mat of mosses and sedges on about 10 cm of organic soil over a silt. The layer of organic soil tends to mollify thermal and hydrologic fluctuations below it. The thermal conductivity of the surface organic layer at average moisture contents is about 1/3 that of the silt and thus functions as a layer of insulation for the permafrost. Before spring melt or after a period of low precipitation, the organic mat is desiccated and will absorb 1.5 cm of water before downslope runoff occurs. The hydraulic conductivity of the organic soils is 10 to 1000 times greater than the silt, thus during a large

rainfall event, downslope movement will primarily occur in the organic layer. The subsurface mineral soil tends to remain saturated with little annual variation and shows little response to precipitation events. In comparison, the moisture content of surficial organic soil fluctuates between 10% and 90% by volume.

B) Introduction

The biological and physical processes which characterize the arctic ecosystem are all affected by the thermal and hydrologic boundary conditions of the active layer. The tundra biome is in a dynamic equilibrium to which any disturbance can upset the delicate balance and cause significant change. The thermal and hydrologic regimes are so intricately inter-related to each other and to the character and structure of the active layer that any modification to one parameter will ultimately affect the entire system. The hydrologic and thermal processes and properties of the active layer must be understood and rationally controlled to prevent unnecessary damage during development.

The objective of this paper is to describe the thermal and hydrologic regime of the active layer and to provide useful information to the researchers and developers who need to understand the dynamics of the arctic ecosystem.

Knowledge of these properties is vital for hydrologic modeling, understanding nutrient transport, explaining plant and animal species distribution and proliferation, or determining the response to various disturbances such as climatic warming. The vast majority of the North Slope, including this watershed, is classified as wetlands and understanding and preserving wetlands has long been a priority to those individuals concerned with protecting migratory waterfowl. This paper describes the important properties associated with heat and water movement and annual fluctuations associated with the thermal and moisture regime.

C) Site Description

This field study was initiated in August 1984 in the foothills of the Philip Smith Mountains on the North Slope of Alaska. The study area was located in the Imnavait Creek watershed (latitude 68°37′ N, longitude 149°17′ W, elevation ≈ 900 m), between the headwaters of the Toolik and Kuparuk rivers approximately 250 km south of the Arctic Ocean (Figure 3-1). The topography of this area consisted of low rolling piedmont hills with a wavelength of 1 to 2 km and amplitudes of 25 to 75 m. The bedrock was composed of shale, sandstone, conglomerate, limestone, and chert of Cretaceous, Triassic and Mississippian ages (Douglas, 1961). This area was glaciated during the Pleistocene.

The soils, which could best be described as Histic Pergelic Cryaquepts, were shallow and quite variable, consisting of about 10 cm of live and dead organic material over 5 to 10 cm of partially decomposed organic matter mixed with silt which overlaid a glacial till. Below the peat was a mottled, dark gray layer with occasional evidence of frost-churned organic matter. A representative soil profile description is included (Table 4-1) to demonstrate the substantial variability in physical characteristics which exists in relatively small vertical distances.

Through Carbon-14 dating, the age of these soils has been established to be at least 11,500 years ± 140 (Walker et al., 1989). The soils were mostly silty colluvium and residual material of glacial origin. There was a thick layer of organic matter on the surface consisting of partially decomposed mosses, sedges and other associated plants. The soil profile experiences frost churning which in effect mixes pieces of the organic mat downward so a layer of organics was sometimes found on the surface of the permafrost table. The surficial organic layer was quite porous and drains when saturated. In contrast, the underlying mineral soil is usually saturated with water, depleting the soil of oxygen during the thaw period. Ice lenses up to approximately 1 cm in thickness formed in the soil during the winter.

Table 4-1. Description of a representative soil profile.

Depth (cm)	Description of profile K.4 #1, near Plot 4.
0-3	Living Polytricum sp. (moss); vertical orientation; boundary clear, smooth.
3-8 Oi 1	Loose spongy mat of partially decomposed moss and roots of Vaccinium vitis-idaea, Ledum palustre, Cassiope tetragona, Eriophorum vaginatum. Boundary abrupt, smooth, principal avenue of water movement into pit.
8-16 A1 16-20 B1	Very dark greyish-brown (10 YR 2/1) decomposed Carex roots and stems. Boundary abrupt, smooth. Dark brown (10 YR 3/4) clay loam; prominent, fine, dark reddish-brown (10 YR 5/4-5/6) mottles, 2% pebbles <1 cm diameter; few fine roots; weakly thixotropic. Boundary abrupt, smooth.
20-40 C	Very dark, greyish-brown (10 YR 3/2) fine, sandy loam; few weak dark reddish-brown mottles (10 YR 5/4); few pebbles, ice lens partings are common, moderately thixotropic. Boundary, permafrost.

The area was underlain by continuous permafrost, the presence of which affects the landscape patterns, microclimatology, hydrologic processes, and thermal regime. The maximum depth of thaw over the period of study was approximately 60 cm in the small water tracks which drain the hillslopes, 40 cm near the top of the slope and 25 cm near the valley bottom. The lower boundary of the permafrost was estimated to be between 250 and 300 m (Osterkamp and Payne, 1981). Although many of the landscape features often associated with permafrost such as ice mounds, polygons, or ice wedges were not conspicuous in the watershed, the effects of frost action were still evident in the presence of translocated organics on the surface of the permafrost table and frost boils.

The vegetation was mostly water tolerant plants such as tussock sedges and mosses, but they were accompanied by lichens and shrubs such as willows, alder and dwarf birch. Tussock sedges were common on the ridge tops and slopes but did not occur in the lowest, wettest areas. More complete descriptions of tundra vegetation have been published (Brown and Berg, 1980; Walker et al., 1989).

D) Methods

The experimental approach used in this study was to perform measurements such that a water balance could be carried out for the watershed. For example, the runoff was collected at the outlet of the basin, snow surveys were made over the entire basin, and soil moisture contents were measured from the top to the toe of the slope. An extensive meteorological station was situated on the ridge near the outlet of the basin and another was located about one-half the distance up the watershed on the west facing slope (Figure 3-2).

Four plots were constructed on the 10% west facing slope (Figure 3-2). These plots were installed along a diagonal to the slope and evenly spaced from near the top of the ridge to near the valley bottom to enable the evaluation of various hydrologic processes. Each plot, measuring 89 m^2, was bounded vertically from the tops of the tussocks to the maximum depth of thaw with heavy ($\approx$1 mm) plastic to isolate it from the hillslope. At the lower end of each plot, a collection gutter was installed to intercept overland and subsurface flow. The water flowed through routing gutters to a holding tank instrumented with a water level recorder. The water levels in the tanks were continuously monitored and periodically emptied. The rates and volumes of runoff from spring melt and some summer storms were measured. The hydrology of the whole basin was also monitored through measurements of snowpack water content, precipitation, streamflow runoff and pan evaporation (Kane and Hinzman, 1988).

The snow and soil temperatures and unfrozen soil moisture were monitored at each plot. A Tektronix time domain reflectometer (TDR) was used to measure the volume of unfrozen soil moisture in the profile above permafrost. This technique has proven to be effective for this application and accurate for moisture determination within 2% (Patterson and Smith, 1980; Stein and Kane, 1983). Eight pairs of 30 cm long stainless steel probes were installed horizontally in the organic and mineral soils at depths between 5 and 40 cm adjacent to each plot. Piezoelectric crystal thermistors were installed in 5 cm increments alongside the TDR probes to complement the soil moisture data. Three thermistors were also mounted above the ground surface in 10 cm increments from the surface to measure snow temperatures. Soil heat flux plates were installed at the interface of the organic and mineral soils and at 20 cm depth in the mineral soil near plot 3. The meteorological stations included air temperature, relative humidity, wind speed and direction, and the radiation components of incoming and reflected shortwave, incoming and emitted total, and net absorbed. Radiation components were only measured between April and early September. Precipitation was measured with a Wyoming snow gage and with two tipping bucket rain gages located at 3 different elevations in the watershed.

The moisture contents were measured using TDR about every four days during the summer and during site visits in the winter. All other instrumentation were monitored by two Campbell Scientific 21X data loggers. Soil temperatures and heat fluxes were measured every hour and averaged daily, except for the summer months of 1986 and 1987 and throughout 1988 when the heat flux plates and some thermistors were sampled every 30 minutes and averaged every hour. All other sensors were measured every minute and averaged hourly.

Soil samples were occasionally collected throughout the soil profile to determine total moisture content. Large samples of the soil profile were collected near the study site for analysis in the laboratory. Soil properties such as hydraulic and thermal conductivity, soil moisture characteristic curves and bulk density were measured in the laboratory. Saturated hydraulic conductivity was determined in a manner similar to that described in Kane and Taylor (1978). In order to develop the characteristic curves, the moisture contents associated with the higher extraction pressures were developed utilizing a pressure plate apparatus. The values measured at lower pressures were determined using Tempe cells. The mineral soil samples were oven dried at 105°C for 24 hours to determine moisture content. Organic samples were dried to constant weight in a microwave oven on low power.

The soils' unsaturated hydraulic conductivity was determined through measurement of saturated hydraulic conductivity and the characteristic curves using the method of Green and Corey (1971). The relationship of unsaturated hydraulic conductivity and moisture content (percent by volume) is as follows;

$$K(\theta)_i = \frac{K_s}{K_{sc}} \cdot \frac{30\Gamma^2}{\rho g \mu} \cdot \frac{\epsilon^p}{n^2} \sum_{j=1}^{m} [(2j + 1 - 2i) h_j^{-2}] \qquad (1)$$

where

$K(\theta)_i$ = the calculated conductivity for a specified water content or extraction pressure (cm/min)
θ = the water content, percent by volume (cm^3/cm^3)
i = denotes the last water content class on the wet end, e.g., i = 1 identifies the pore class corresponding to the saturated water content, i = m identifies the pore class corresponding to the lowest water content for which conductivity is calculated.
K_s/K_{sc} = the matching factor (measured saturated conductivity/calculated saturated conductivity) when $K_s/K(\theta)_i = 1$
ρ = the density of water (g/cm^3)
μ = the viscosity of water ($g/cm\ sec^{-1}$)
g = the gravitational constant (cm/sec^2)
Γ = the surface tension of water (dynes/cm)
p = a parameter that accounts for interaction of pore classes
ϵ = the porosity (cm^3/cm^3), defined in various ways depending on the method of calculation.
n = the total number of pore classes between $\theta = 0$ and θ_s, the saturated water content, $n \geq m$
h_j = the pressure for a given class of water filled pores (cm of water)

The effects of temperature and moisture upon soil thermal conductivity were determined in the laboratory utilizing a Dynatech model TCFGM guarded hot plate. Each sample was relatively undisturbed, measuring approximately 15 cm in diameter and 5 cm thick. A controlling heat flux was introduced across each sample to maintain a temperature gradient of 2 to 3°C and the resulting thermal conductivity was calculated. This guarded hot plate operates by putting a known amount of energy through two nearly identical samples. A guard heater prevents thermal energy from leaving the warm (heated) side of the samples. The temperatures are set on each of the cool sides and an energy input is determined to give the desired thermal gradient across the samples. The effective thermal conductivity within the samples is calculated using Equation 2 and represents an average of the two samples.

$$K = (Q/A) * \frac{1}{(dT_1/d_1) + (dT_2/d_2)} \tag{2}$$

where

K = effective thermal conductivity (W/m°C)
Q = main heater power input (W)
= voltage * amperage * factory calibration
dT = temperature difference across the sample (°C)
d = full sample thickness (m)
A = the surface area of the main heater (m^2)
= 0.00835 m^2
subscripts 1 and 2 denote sample number.

E) Results and Discussion

The hydrologic response to the thermal regime and vice versa is a complicated, dynamic and self-regulating mechanism which manifests itself in the processes of soil moisture migration, evapotranspiration, sublimation, and the formation or degradation of permafrost. The latter process is the crucial determinant in the geomorphologic response to disturbance. Phase transition of water is the most important factor in cryogenic processes (Brown and Grave, 1979). The energy required for phase change (ice to water, water to vapor and vice versa) dissipates greater than half of the annual energy balance (Weller and Holmgren, 1974) and dominates the energy balance during break-up and freeze-up. The amount of moisture in the soil profile will affect the depth and rate of freezing (Drew et al., 1958); this effect is accentuated if the water is moving (Price, 1983). The soil moisture content is one of the most important factors determining the temperature field in the active layer (Pavlov, 1975) as the thermophysical properties are determined primarily by density and moisture content (Nixon and McRoberts, 1973).

The hydrologic and thermal regimes represent the integration of the surface water and energy balances. Although the surface hydrology and surface energy balance differed somewhat from year to year, the thermal and hydrologic regimes of the active layer remained much the same (Tables 4-2 and 4-3, Figures 4-1 and 4-2). Only two years of data are presented graphically because the subsurface thermal and hydrologic regimes were very consistent during the four years of the study.

Table 4-2. Average monthly air temperatures (°C) from Imnavait watershed.

	JAN	FEB	MAR	APR	MAY	JUN*	JUL*	AUG*	SEP*	OCT*	NOV	DEC	MEAN
1985					-0.6	5.2	9.3	7.4	-1.7	-15.2	-7.3	-12.1	X
1986	-20.4	-15.8	-19.4	-16.0	-5.4	9.0	12.1	5.6	0.8	-10.9	-16.9	-16.8	-7.8
1987	-20.9	-18.9	-16.3	-14.0	0.1	8.5	11.3	6.7	-3.0	-7.8	-21.9	-18.1	-5.1
1988	-13.8	-16.8	-18.4	-11.8	-1.3	8.3	10.6	5.5	-1.1	-14.4	-23.8	-15.3	-7.7
1989	-26.3	-8.1	-16.7	-7.9									X
MEAN	-20.4	-14.9	-17.7	-12.4	-1.8	7.8	10.8	6.3	-1.3	-12.1	-17.5	-15.6	-7.4+
sd	5.1	4.7	1.5	3.5	2.5	1.7	1.2	0.9	1.6	3.5	7.4	2.6	11.2+

* - data from Kelley (1985).
\+ - mean and standard deviation of all data.
X - not calculated due to incomplete data.

Table 4-3. Average monthly soil surface temperatures beneath live vegetation (°C) near Plot 3.

	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	NOV	DEC	MEAN
1985				-10.3	-5.8	-0.2	2.8	5.2	1.0	-1.8	-6.4	-8.8	X
1986	-9.4	-10.1	-13.1	-12.4	-6.8	4.7	7.9	3.2	1.9	-0.8	-2.6	-5.0	-3.5
1987	-7.2	-8.1	-9.2	-8.4	-2.6	5.9	7.3	3.7	0.0	-0.6	-2.6	-5.4	-2.3
1988	-7.1	-9.4	-9.4	-9.4	-1.5	6.3	8.4	4.6	0.3	-0.8	-3.8	-7.0	-2.4
1989	-8.4	-8.2	-8.9										X
MEAN	-8.0	-9.0	-10.2	-10.1	-4.2	4.2	6.6	4.2	0.8	-1.0	-3.9	-6.6	-3.1+
sd	1.1	1.0	2.0	1.7	2.5	3.0	2.6	0.9	0.8	0.5	1.8	1.7	6.0+

\+ - mean and standard deviation of all data.
X - not calculated due to incomplete data.

1) Hydrologic Properties and Processes

The layered system of soil horizons regulates moisture movement into and through the active layer. Considerable redistribution of moisture in the active layer occurred during the winter. The strong temperature gradients which developed during the winter induced moisture migration upward, out of the surface soil, causing a desiccation of the organic layer (Loch and Kay, 1978; Santeford, 1978; Woo, 1982b). Measurements in the laboratory revealed that the organic layer required about 1.5 cm of water for rewetting in the spring (Kane et al., 1989). Soil samples were collected periodically and analyzed for total moisture content. Ice lenses up to 1 cm in thickness were frequently found in the frozen mineral soil samples. Therefore, the total moisture content often had considerable spatial variation within a few centimeters.

The unfrozen soil moisture was measured periodically throughout the year using TDR (Figures 4-1 and 4-2). Although the soil temperatures decreased below -10°C during the winter throughout the active layer, the unfrozen soil moisture remained about 8% by volume in the fine-grained mineral soil and 6% or less in the organic soil. This is consistent with results of other researchers (Anderson and Morgenstern, 1973). The unfrozen soil moisture remained fairly constant during the winter but increased abruptly when meltwater entered the surface organic layer or when the water changed phase from solid to liquid during thawing in the mineral layer. In addition to measuring unfrozen soil moisture content, the TDR technique is useful to indicate when

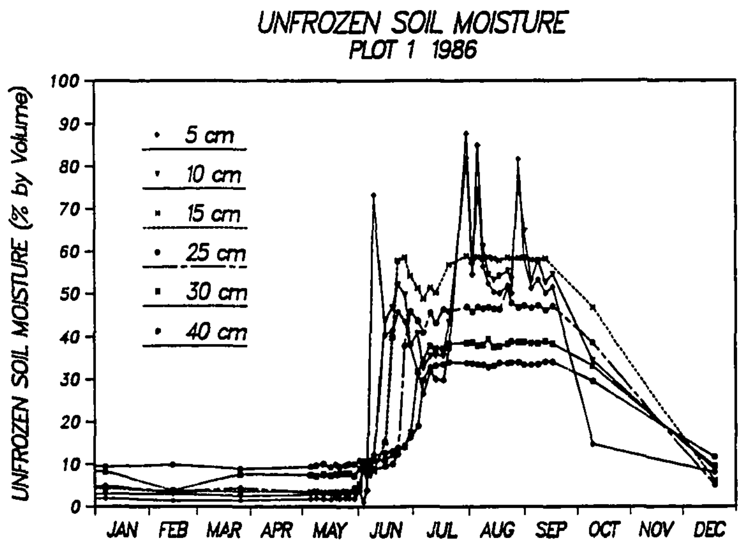

Figure 4-1. Variation in soil unfrozen water content for several depths in 1986.

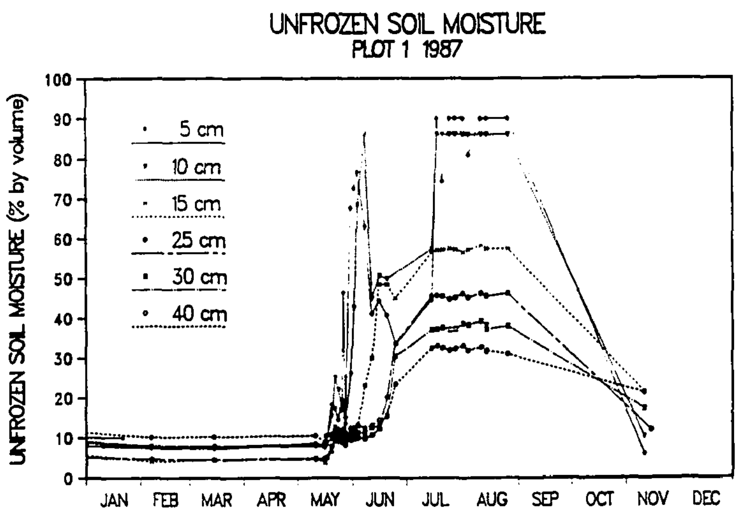

Figure 4-2. Variation in soil unfrozen water content for several depths in 1987.

the soil has thawed and when the meltwater has infiltrated the organic soil (Baker, et al., 1982; Stein and Kane, 1983).

At the time of the spring melt, most pores in the mineral soil are filled with ice. Initially the meltwater refreezes when it contacts the colder soil releasing latent heat and warming the subsoil. This in effect seals any open pores and greatly restricts potential infiltration (Woo and Heron, 1981; Kane and Stein, 1983 a and b), thus meltwater movement vertically into the frozen mineral soil is negligible. Figures 4-1 and 4-2 show a marked increase in unfrozen moisture for the organic soils near the beginning of spring melt. However, the unfrozen moisture content just 10 cm into the mineral soil did not greatly increase until after snowmelt runoff was complete. This demonstrates that the snowmelt runoff occurred above the mineral soil in the upper portion of the organic layer.

During the period of the study, yearly precipitation averaged approximately 34 cm, about one-third of which occurred as snow. Several significant precipitation events occurred in July, August and September of each year, however the mineral soils were not noticeably impacted by these events. The organic soils were immediately responsive to rain events, saturating and draining quickly (Figures 4-1 and 4-2). The mineral soils being much less influenced by individual rain and snowmelt events, have relatively stable moisture contents throughout the summer. The reason the organic layer responded so quickly to the rain events is due to the porous nature and high hydraulic conductivity associated with this horizon. In similar soils, Drew (1957) noticed that much of the horizontal flow of water occurred at the interface of the organic and mineral soils. This was also evident in the characterization of the soil profiles (Table 4-1). The unfrozen hydraulic conductivities of the organic soils ranged from 3 to 20 times greater than that of the mineral soil (Table 4-4). The reason the mineral soil remained saturated and thus unresponsive to surface events was due in part to the presence of ice-rich permafrost near the surface.

Table 4-4. Summary of the physical properties of soil samples taken from Imnavait watershed.

Horizon Type	Depth (cm)	Hydraulic Conduct. (cm/s) (10^{-3})	Range (cm/s) (10^{-3})	Bulk Density (g/cm^3)	Range (g/cm^3)	Porosity	No. of Samples
organic	0-5	19.4	16-22	0.15	0.13-0.16	0.90	4
organic	5-10	10.4	1.1-12	0.10	0.16-0.18	0.86	3
org/min	10-15	3.76	0.4-3.4	1.39	1.38-1.40	0.70	3
mineral	15-20	0.87	0.08-2.6	1.53	1.43-1.57	0.55*	6
mineral	20-25	1.42	0.6-3.7	1.33	1.30-1.42	0.54*	3
mineral	25-40	0.94	0.13-3.3	1.40	1.36-1.43	0.46*	3

* Values determined from TDR when soils were saturated.

The large differences in bulk density, porosity, and hydraulic conductivity of the mineral and organic soils (Table 4-4) also had a great effect on the process of snowmelt runoff. In the four years of this study, the snowmelt process was a very brief event, lasting about 10 days or less (Kane and Hinzman, 1988). Snowmelt runoff was always completed before the active layer had thawed below 10 cm. The ice-rich mineral soil prevented significant infiltration, so all subsurface drainage occurred only in the top 10 cm. This affects not only the rates of runoff and lag time, but also the source of nutrients transported with spring meltwater. During periods of intense melt, inter-tussock or rill flow was common particularly in areas where the soil was still frozen very near the surface. However the primary mechanism of downslope water movement was near surface flow through highly porous organic mat.

The meltwater stored in the organic layer is needed for plant growth which begins soon after spring melt is initiated. Precipitation is usually quite low in June and evapotranspiration can significantly lower the moisture content of the surface soils. After a period of low precipitation, the organic soils can absorb a significant amount of water before downslope runoff can occur. The water content of the soil profile will be at the annual minimum after the early summer dry period. An examination of the range of total soil moisture in the soil profile reveals wide variation in the near surface organic soils and the annual variation decreases with depth (Figure 4-3). Analysis of summer precipitation events revealed that 1.5 cm of rainfall was necessary before downslope runoff would occur following a period of no precipitation (Kane et al., 1989). This confirms the laboratory measurement of the amount of moisture required to re-wet the organic soils in the spring.

Characteristic curves of the active layer soils were developed in the laboratory using samples collected in the watershed (Figure 4-4 a,b,c, and d). The characteristic curves relate how much moisture a soil will hold against some force such as gravity or extraction by plants. These curves convey much information about a particular soil's physical and hydrologic properties. The amount of moisture which can be held in the soil matrix is primarily a function of soil texture and structure and large differences occur between the organic and mineral soil curves. In the near surface organic soils (0 to 10 cm, Figures 4-4 a and b), the pores are large and once these large pores are empty, a relatively small amount of moisture

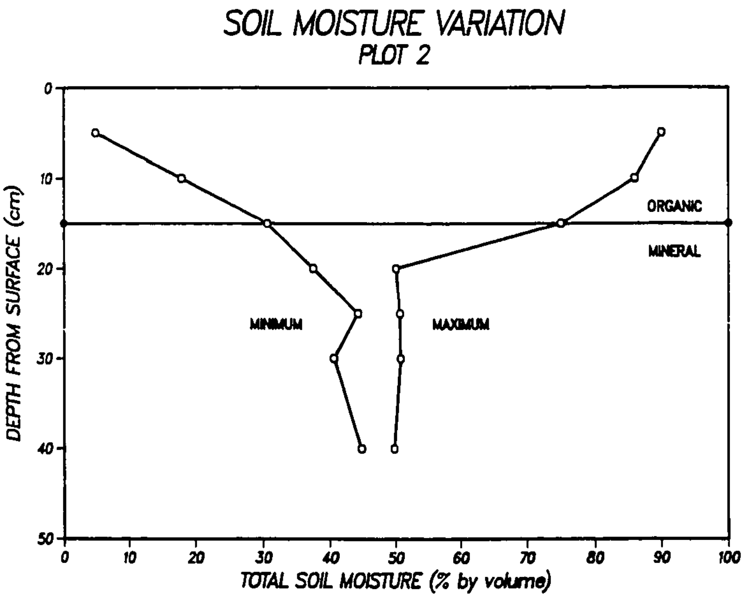

Figure 4-3. Range in fluctuation of total moisture content (ice + water) during the period of study.

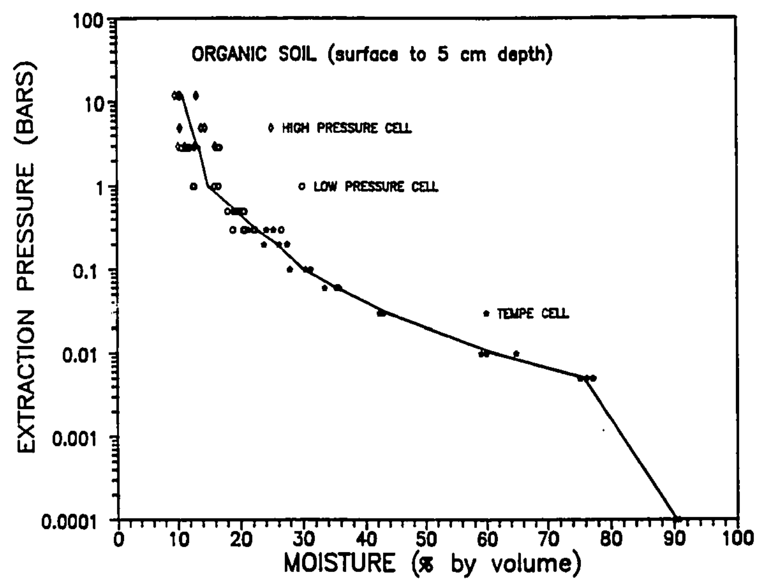

Figure 4-4a. Characteristic curve for the active layer, 0 to 5 cm organic soil.

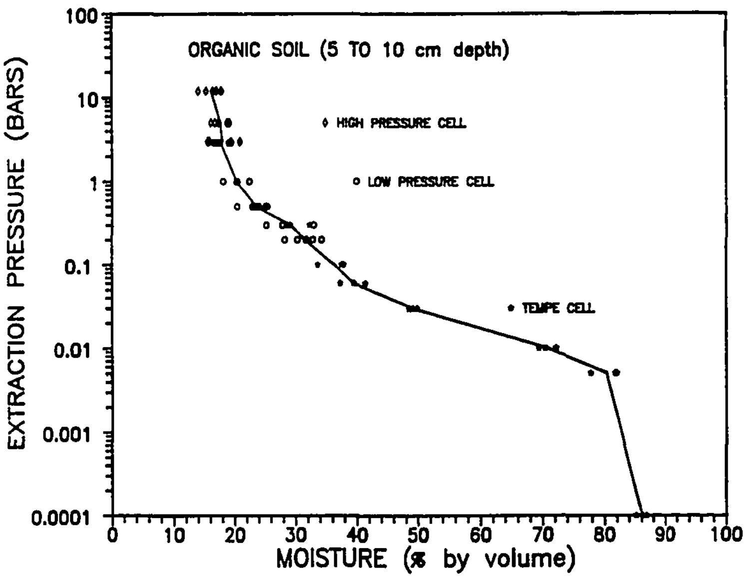

Figure 4-4b. Characteristic curve for the active layer, 5 to 10 cm organic soil.

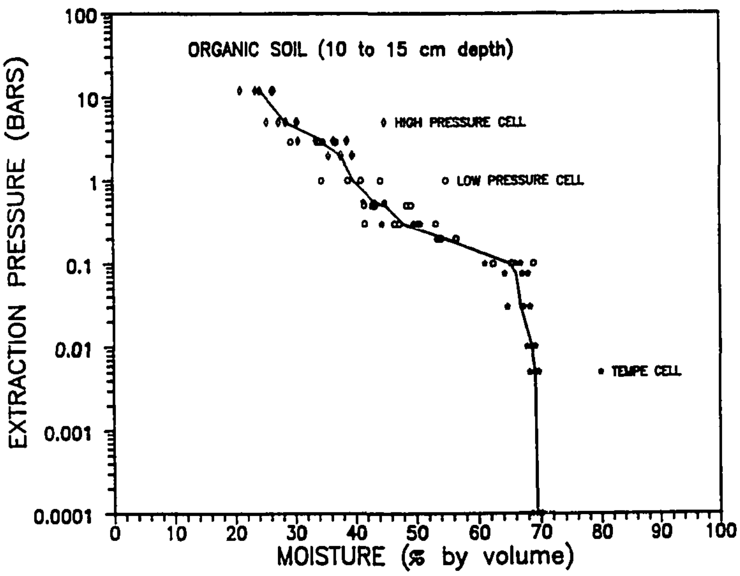

Figure 4-4c. Characteristic curve for the active layer, 10 to 15 cm organic soil.

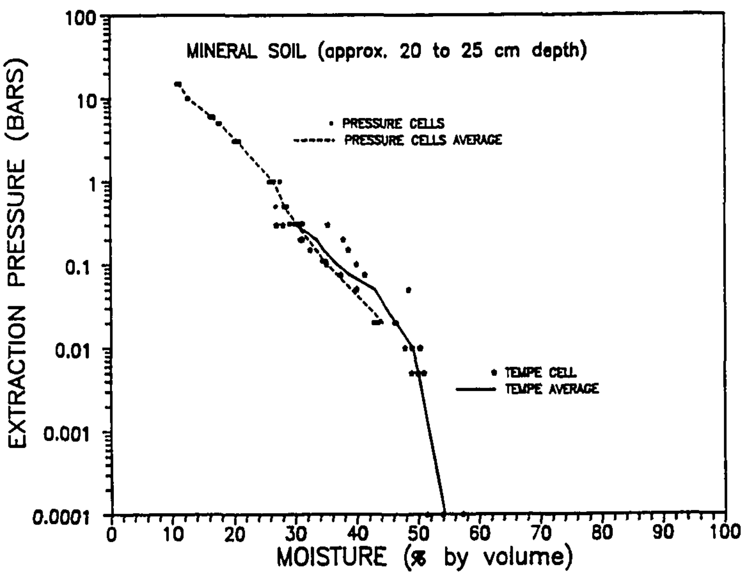

Figure 4-4d. Characteristic curve for the active layer, 20 to 25 cm mineral soil.

remains. In the deeper mineral soils (20 to 25 cm, Figure 4-4 d), there is much more fine grained material and more of the water is adsorbed. Therefore, as the matric suction is increased, the decrease in moisture content is much more gradual. In the natural tundra environment, during the thaw season, the mineral soils remain near saturation and the actual soil water pressure is only slightly negative. The water table is normally below the organic layer, but the surficial organic soils never become greatly dry. Even during periods of low precipitation, these soils do not experience greatly negative soil water tensions.

Due to the high porosity of the organic layer (between 70% and 90%), it was only saturated for brief periods during snowmelt and heavy rain events. However, the mineral soil remained near saturation throughout the study. In calculating downslope (horizontal) movement of water for most of the thaw season, the values of saturated hydraulic conductivity should be utilized (Table 4-4) but the curves of unsaturated hydraulic conductivity should be used for most applications involving vertical moisture movement in the soil profile above the water table (Figure 4-5 a-d).

2) Thermal Properties and Processes

The climate is most simply described as cold. Imnavait

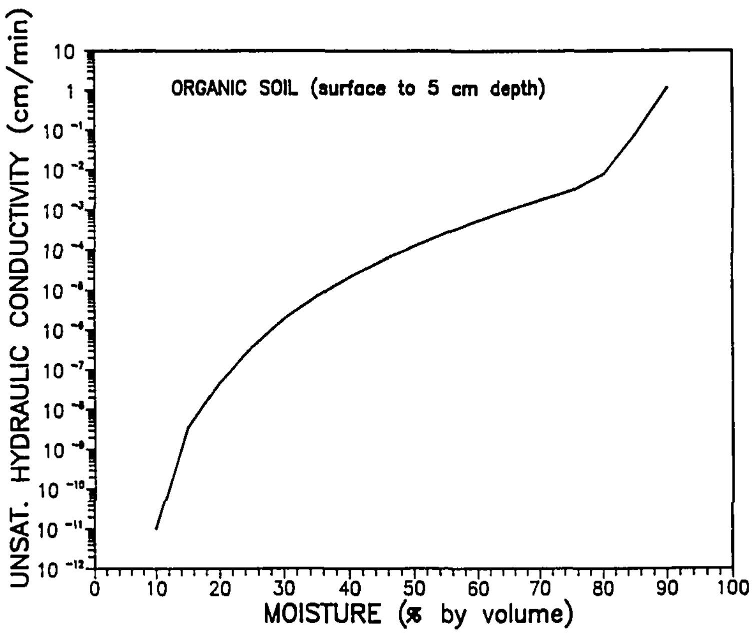

Figure 4-5a. Calculated unsaturated hydraulic conductivity for the active layer, 0 to 5 cm organic soil.

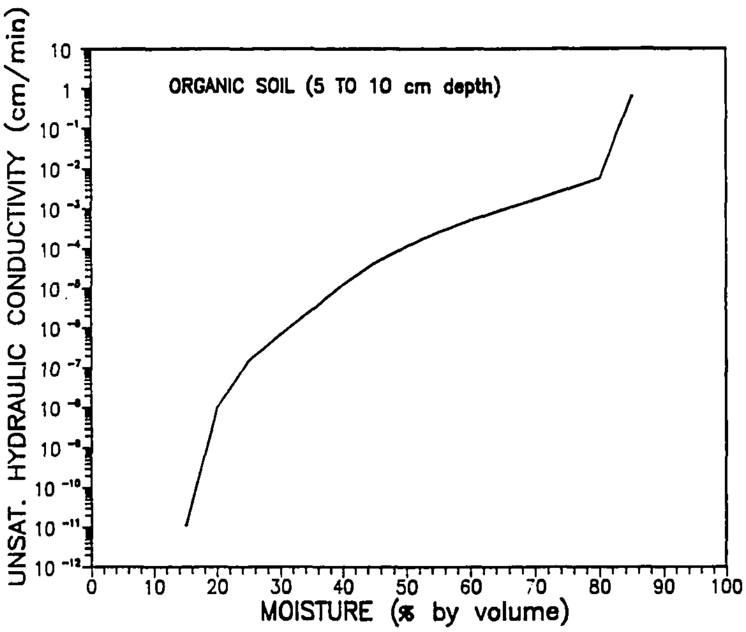

Figure 4-5b. Calculated unsaturated hydraulic conductivity for the active layer, 5 to 10 cm organic soil.

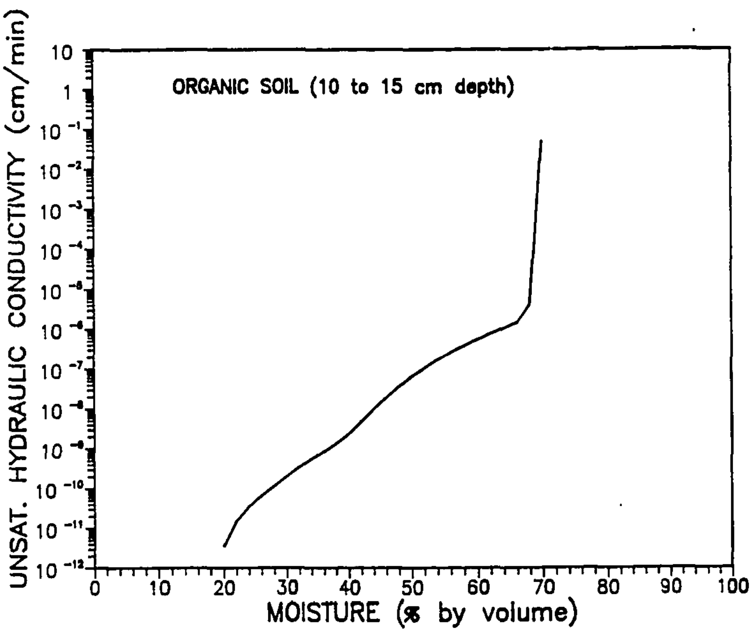

Figure 4-5c. Calculated unsaturated hydraulic conductivity for the active layer, 10 to 15 cm organic soil.

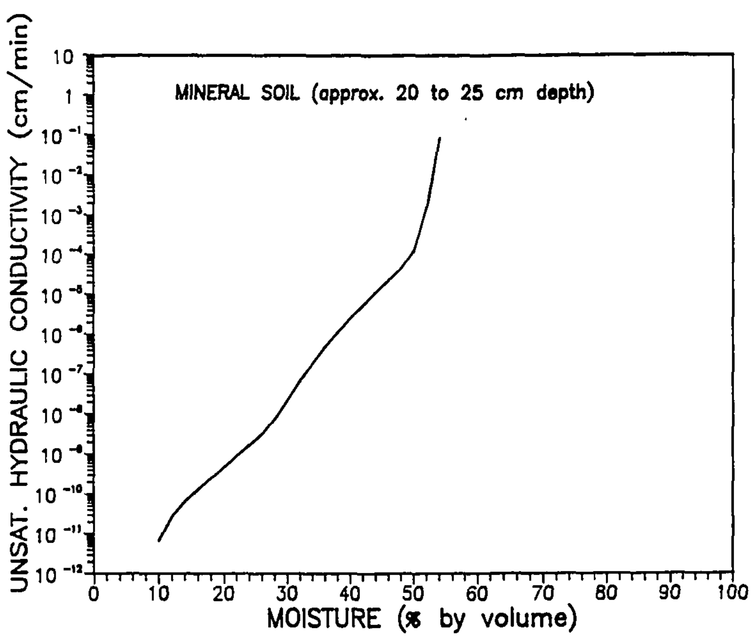

Figure 4-5d. Calculated unsaturated hydraulic conductivity for the active layer, 20 to 25 cm mineral soil.

watershed has a continental climate, characterized by long, cold winters and short, cool summers. During the winter, the climate of the tundra is affected primarily by radiative heat loss and atmospheric circulation (Ohmura, 1981). Net radiation becomes positive near the end of April or May and abruptly increases following snowmelt. Before the onset of spring melt, which normally occurs in May or early June, just prior to the time of maximum incoming solar radiation, the surface is characterized by a uniform high albedo. As the melt progresses, due to the uneven distribution of snow, the albedo of the surface exhibits considerable variability (Liston, 1986). Between the period of snowmelt and snow accumulation, about three months, the surface maintains its lowest annual albedo which results in the maximum annual energy exchange at the surface and greatest diurnal variation. The accumulation of new snow occurs in September and since solar radiation is considerably less at this time of year, the arrival of the new snowcover does not cause the dramatic changes in the surface energy and water balances as during the spring. A layer of snow on the surface of the tundra will serve to insulate the active layer from the cold winter air keeping the subsurface temperatures much warmer than the air, reducing surface heat flux and damping diurnal variation for nearly nine months.

The warmest month is July, averaging 10.8°C, and the coldest month is January, averaging -20.4°C (Table 4-2). During the course of the study, the minimum recorded air temperature was -56°C, the maximum was 34°C. The surface soil temperatures, insulated by the snow in the winter and possessing a relatively high specific heat due to large moisture contents did not experience such extreme high and low temperatures. The warmest surface temperatures coincided with the warmest air temperatures in July, averaging 6.6°C, while the coldest surface temperatures were recorded in March lagging the coldest air temperatures by many weeks and averaging -10.2°C (Table 4-3). The average monthly air temperature and average monthly surface temperature for the period of record are -7.4°C and -3.1°C respectively.

Soil temperatures were measured adjacent to each plot, and soil heat fluxes were measured adjacent to plot 3. In the winter, the air temperatures dropped below -50°C, but the soil, insulated by the snow and warmed by heat transfer from below, remained above -15°C (Figures 4-6 and 4-7). In the summer, the air temperature occasionally rose to 30°C, but the surficial organic soil remained below 15°C, probably due to the large heat sink in the permafrost. Winter and the onset of soil freezing came in September. The active layer cooled to

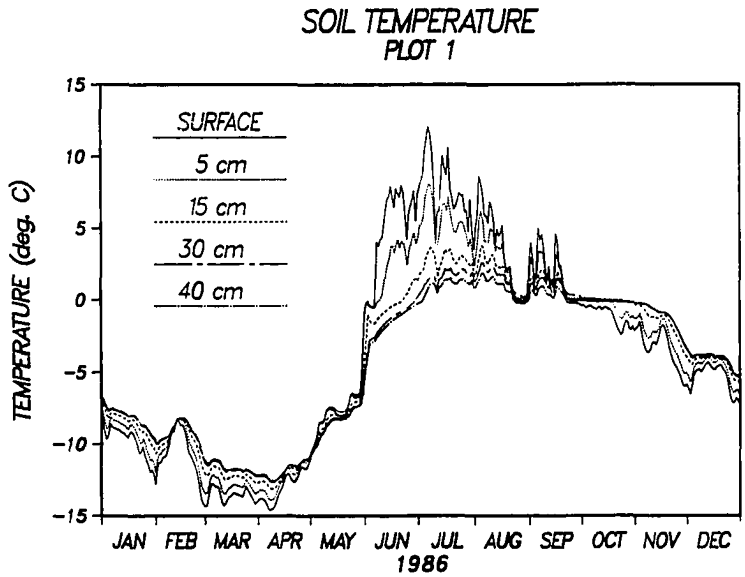

Figure 4-6. Annual variation in soil temperature for several depths in 1986.

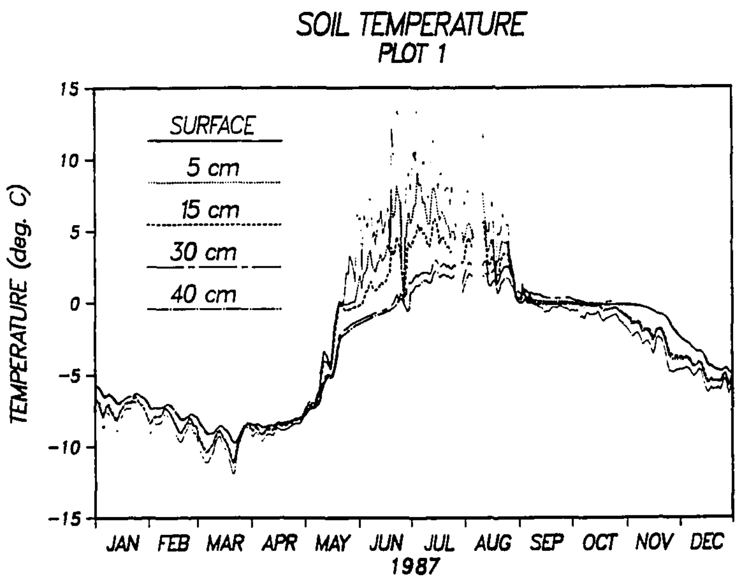

Figure 4-7. Annual variation in soil temperature for several depths in 1987.

the 0°C isotherm within a few days of sustained freezing air temperatures. The mineral soil remained isothermal at 0°C for several weeks, while it progressed through the phase change.

In comparison to the fall freeze-up, as the soil warmed in the spring, the entire active layer did not warm to isothermal conditions at 0°C and remain there as the soil ice melted. Instead, the active layer quickly warmed from the surface downwards. The soil warmed through 0°C and completed the phase change at each depth before the underlying soil warmed to 0°C. This is additional evidence that there was little infiltration of water from the snowmelt. If there had been infiltration, the entire soil profile would have warmed quickly, becoming isothermal at 0°C as latent heat was released when the water froze in the cold soil.

The spring thaw occurs near the summer solstice, incoming solar radiation is quite near its annual maximum, there is no snow on the surface to slow the rate of heat transfer and the surface heat flux is at its annual maximum (Figure 4-8). Autumn snows affect the process of soil freezing. In the Arctic Foothills, precipitation can fall as snow on any day of the year, however it becomes more frequent in late September. In late summer, the snow usually melts soon after it contacts the ground surface drawing heat from the ground and cooling the subsurface. Eventually the surface heat balance becomes negative and snow begins to accumulate. At this point, the snow acts as insulation, reducing the heat flux and the process of soil freezing. One can see from Figure 4-6 that the soil cooled from the top down and from the bottom up during a brief cold period in August. The soil above and below 25 cm cooled to 0°C before the center of the active layer cooled. This demonstrates that the active layer cools from two directions (Mackay, 1983).

Soil heat flux was measured near the soil surface (Figure 4-8), at the organic/mineral interface and in the mineral soil (Figure 4-9). The diurnal variation caused by wide fluctuations of air temperature during the summer was evident at both levels. The instantaneous magnitude and daily amplitude of the heat flux of the organic layer were always greater than in the mineral soil. The mineral soil has less heat flux because the overlying organic soil has a lower thermal conductivity (Figure 4-10), thus the organic layer functions as a layer of insulation for the underlying soil. The organic soil undergoes wider daily fluctuations because it has a lower specific heat than the mineral soil. Even a thin snowpack will greatly dampen the diurnal variation. Snow, falling in late

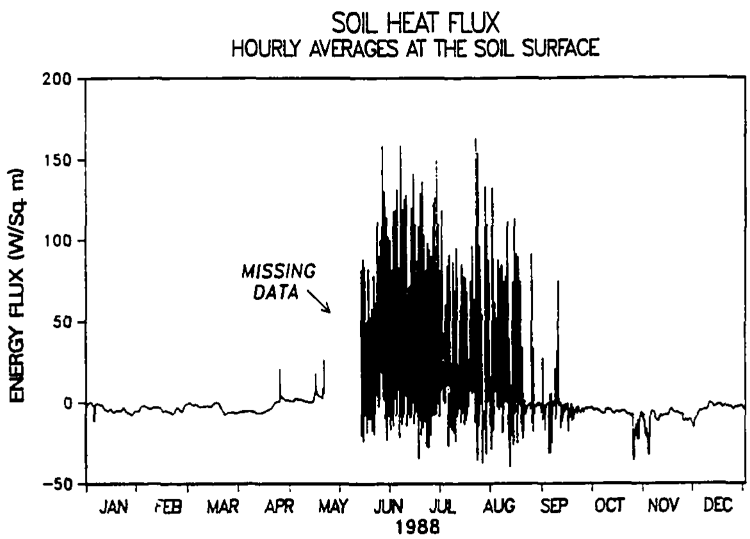

Figure 4-8. Soil heat flux measured hourly at the soil surface in 1988.

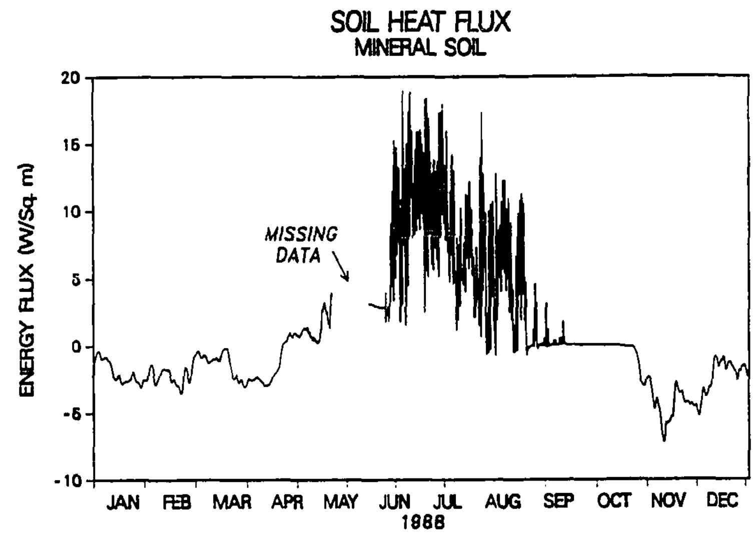

Figure 4-9. Soil heat flux measured hourly at 20 cm depth, in the mineral soil in 1988.

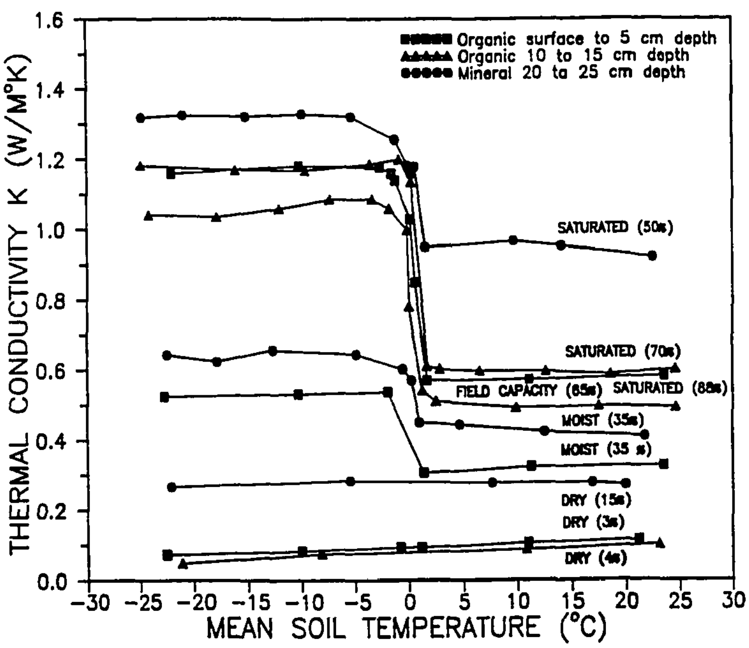

Figure 4-10. Laboratory measurements of effective thermal conductivity of active layer soils as a function of temperature, soil type and moisture content.

August, greatly reduced the soil heat flux and as the snowpack deepened, the diurnal variation was completely suppressed.

The thermal conductivity of the mineral soil was calculated from measurements of soil heat flux and soil temperatures using Fourier's Law of heat flow;

$$Q = -K \frac{dT}{dX} \tag{3}$$

where Q is the heat flux, K is the thermal conductivity, and dT/dX is the temperature gradient. This equation is only valid for heat conduction and consequently only data collected when the soil was frozen and convective heat transfer was insignificant could be included in the analysis.

A plot of soil thermal conductivity, calculated hourly is shown in Figure 4-11. The accuracy of the thermistors was ±0.2°C, so when the temperature difference between the thermistors was less than 0.2°C, the error in measuring the temperature gradient prevented calculation of thermal conductivity. The average value of K is 1.02 W/m°C with a standard deviation of 0.2 W/m°C. These values compare well with curves developed in the laboratory (Figure 4-10) and with values reported in the literature (Sepaskhah and Boersma, 1979; Farouki, 1981).

An error analysis according to the method of Holman (1989) can be performed according to the equation

$$dK = \left[\left[\frac{X \cdot \delta Q}{dT}\right]^2 + \left[\frac{Q \cdot X \cdot \delta dT}{(dT)^2}\right]^2 + \left[\frac{Q \cdot \delta X}{dT}\right]^2 \right]^{\frac{1}{2}} \tag{4}$$

The uncertainty in the thermal conductivity, dK can be determined by selecting typical values of $Q = 6.12$ W/m², $dT = -0.6$°C, and $X = 0.1$ m and using the reported instrument accuracy of $\delta Q = 5\% = 0.31$ W/m², $\delta dT = 0.2$°C, and $\delta X = 0.02$ m. The uncertainty in K at this gradient and heat flux is substantial, $dK = 1.04$ W/m°C. The uncertainty would increase at lower thermal gradients as is evident in Figure 4-11. The effective thermal conductivities of the organic and mineral soils were determined as a function of temperature and moisture content using a guarded hot plate (Figure 4-10). When the organic soil is thawed with moisture content near field capacity, the effective thermal conductivity is about 0.45 W/m°C. The same soil when frozen has an effective thermal conductivity of about 1.0 W/m°C. The mineral soil, when saturated and thawed, has an effective thermal conductivity of about 1.0 W/m°C. The same soil frozen has an effective thermal conductivity of about 1.3 W/m°C. These values compare well with

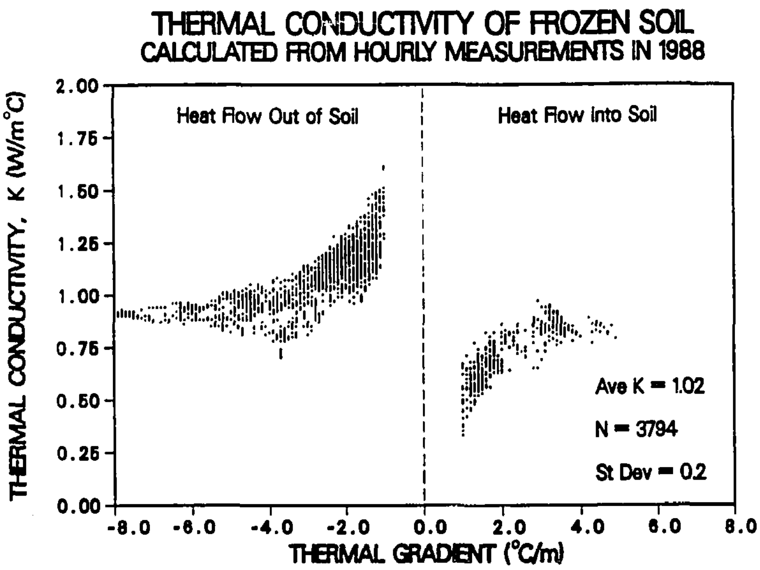

Figure 4-11. Thermal conductivity of mineral soil calculated from soil heat flux and soil temperatures.

published data (Farouki, 1981). A common scenario in early summer is a dry, thawed, organic soil insulating a saturated, frozen, mineral soil. In this situation, the thermal conductivity of the organic soil would be 1/3 to 1/6 the thermal conductivity of the mineral soil. Therefore, the organic soil has more resistance to heat flow in the summer than it does in the winter. It protects the underlying permafrost from melting due to surface heating in the summer, and allows cooling of the subsoil by more effective heat transfer in the winter. However snow will also act as a barrier to heat flow, slowing winter heat loss. The effective thermal conductivity of snow can range between 0.05 to 1.0 W/m°C (Langham, 1981) and with an average depth of 0.5 m, its insulating effect will be more significant than the increased conductivity of the frozen soil.

F) Conclusions

The temporal pattern of soil moisture and temperature in the active layer were consistent year to year, but ranged greatly from season to season. The volumetric moisture content of the organic soil varied by up to 60%, but the mineral soil remained near saturation throughout the year. In spite of the great spatial variability, it is hoped that these values of the material properties are reasonable estimates of the mean. The values of saturated hydraulic conductivity should be used for calculations involving downslope moisture movement. The curves of unsaturated hydraulic conductivity should be used for vertical moisture movement within the soil profile above the water table. During the spring runoff event, subsurface water flows only through the top 10 cm of organic soil. During the major rain events of summer, most downslope flow occurred in the organic layer above the organic/mineral interface. The thermal conductivity of the soils depend upon the phase and amount of the moisture present. Regardless of the temperature or moisture content, the organic soil has a lower thermal conductivity than the mineral soil and will serve as a layer of insulation to the permafrost.

G) Acknowledgments

This research was sponsored by the U.S. Department of Energy in association with the program Response, Resistance, Resilience and Recovery from Disturbance in Arctic Ecosystems. Elizabeth K. Lilly and Catherine G. Egan helped with data collection and reduction and preparation of figures.

V) HYDROLOGY OF IMNAVAIT CREEK, AN ARCTIC WATERSHED

Kane, D.L., L.D. Hinzman, C.S. Benson, and K.R. Everett. 1989. Hydrology of Imnavait Creek, an arctic watershed. Holarctic Ecology, 12:262-269.

A) Abstract

This paper closely examines certain unique features of hydrology in an arctic watershed. Permafrost completely underlies this basin and significantly impacts many hydrologic processes. Percolation to subpermafrost groundwater is restricted by the ice-rich conditions, consequently all water losses occur through evaporation, transpiration or runoff. Soil storage of moisture is very limited, however prior to spring melt or after an extended dry period, approximately 1.5 cm of moisture will be absorbed prior to surficial runoff. Precipitation is a difficult component to measure in a windy environment, especially in the form of snowfall. A Wyoming snow gage located within the watershed appears to give a good estimation of the total snowpack water content at the end of winter.

B) Introduction

An ecological study is presently being carried out at the headwaters of Imnavait Creek to develop a better understanding of the dynamics of an arctic ecosystem.

While all ecosystems are driven by both energy and mass inputs, in an arctic setting this is much more evident. In the United States, no hydrologic studies of an arctic watershed have been made where measurements are continued throughout the winter season. While most soil biologic processes are dormant throughout the winter, this is not true of physical processes such as active layer freezing and thawing and snow accumulation and redistribution by wind.

Several characteristics distinguish arctic watersheds from those in temperate and tropic areas. First, the accumulation and ablation of the snowpack is the dominant hydrologic event of the year (Kane and Hinzman, 1988). While rainfall events can be significant, every spring a major, dramatic ablation runoff event occurs. Secondly, because of the presence of permafrost, the subsurface system that actively plays a role in the hydrologic cycle is very shallow and therefore subsurface storage is severely limited (Dingman, 1975). Thirdly, extreme fluctuations in the various surface energy fluxes occur throughout the year. This is responsible for the freezing and thawing of the active layer and the continued existence of permafrost. While the quantity of annual precipitation may be similar to arid and semi-arid regions and runoff patterns may be likewise similar, the role of the subsurface system is reduced significantly.

These physical processes directly affect the response of the biologic system. This system has a very narrow window of activity and benefits substantially from an immediate start after the ablation period. This immediate response is possible because substantial quantities of snowmelt water are available in the surface organic soils and rapid thawing of the active layer results from the downslope movement of water through the porous organic soils.

C) Setting

Imnavait Creek is a small headwater basin, 2.2+ km^2, located in the northern foothills of the Brooks Range in Alaska (see Figure 3-1). Permafrost completely underlies this region; the active layer that thaws each summer varies in thickness from 25 to 60 cm. The maximum thickness of permafrost is probably between 250 and 300 m near the site (Osterkamp and Payne, 1981). A north draining beaded stream emanates from the watershed. Water flows from this drainage beginning during the snowmelt period in late May until freeze-up in September. A west-facing slope constitutes 78% of the basin area and an east-facing slope accounts for the remaining 22%. Slopes vary from 1 to greater than 13%.

Small, parallel water tracks that flow directly down the slopes channel runoff during snowmelt and rainfall events to the valley bottom.

The soils of the watershed are Histic Pergelic Cryaquepts. The surface organic soils vary from live organic material at the surface to partially decomposed organic matter between the 10 and 20 cm depth. Silt, overlying a glacial till, comprises the mineral soil. Sedge tussocks, mosses and low shrubs are the primary vegetation.

D) Measurement Program

A measurement program was set up to measure both energy and mass fluxes entering and leaving the basin. Very detailed snow surveys were carried out throughout the watershed to ascertain the amount of snow on the ground just prior to snowmelt. During the summer months shielded gages measured rainfall. Due to the ice-rich permafrost, percolation to sub-surface groundwater was extremely limited in this basin. There was no measurable groundwater loss from the basin; water left the basin as evaporation, transpiration, or runoff.

During 1985, continuous measurements were not made of the streamflow leaving the basin. In 1986, a stage-discharge relationship was developed and stage was monitored continuously. The following year, 1987, a flume was also installed in the stream channel.

During 1986 and 1987, a standard evaporation pan was in operation over the summer months. It was installed as soon as thawing conditions allowed in the spring. With low albedos and large quantities of incoming radiation, energy is available for significant amounts of evaporation.

A meteorological station was installed to measure various radiation terms. During the first year, an albedometer, short wave pyranometer and an all-wave net radiometer were installed. The following year, an additional all-wave net radiometer was installed; however, this instrument gives two outputs, one for total incoming radiation and one for total outgoing radiation. This site is located on the west facing slope. On the ridge near the outlet of the basin, another site has been installed where incoming short wave, incoming long wave, reflected short wave and emitted long wave are measured. This site is not manned except for short periods during the winter, snowmelt period and summer. Therefore, these radiometers are only operated from March until September. Rime ice and condensation on the sensors are still major problems. As the daylight hours increase, their effect on measurements is confined mainly to twilight hours.

Air temperature, relative humidity, and wind speed are monitored at two levels continuously throughout the year. Wind direction is also monitored continuously, but at only one elevation. Snowpack temperatures at 10, 20 and 30 cm are measured along with soil temperatures at 5 cm increments from the ground surface down to the bottom of the active layer at four sites located along the hillslope. At various times and depths, heat flux plates have been installed with variable success.

E) Hydrologic System and Processes

Although the environment is harsh and frozen ground prevails both spatially and temporally, watershed hydrologic processes in the Arctic lack some of the complex interactions that exist in other regions. The lack of a deep and extensive groundwater system greatly simplifies that aspect; unfortunately, moisture migration of unfrozen water in the active layer in response to temperature gradients severely complicates the process of groundwater movement. The important concepts of mass transfer can be formulated as a water balance:

$$SM + R - E - R = \frac{dS_s}{dt} \qquad (1)$$

Where,

SM = snowmelt
R = rainfall
E = evapotranspiration
R = runoff
dS_s = soil storage
dt = time increment

Each term in the above water balance equation will be discussed in the following sections.

1) Snow and Rain Inputs

Snowfall input into the basin is difficult to measure. During windy periods, both unshielded and shielded gages under-measure the quantity of snowfall precipitation. Sturges (1986) showed that the Wyoming-shielded gage measured 37 to 53% of actual precipitation between November and March when air temperatures were below freezing and monthly wind speeds were at a maximum. It was concluded that the under-measurement was directly related to the wind as the turbulence caused by the gage deflected the particles around and over the orifice. Rainfall precipitation appeared to be measured adequately. He concluded that undercatch would be more severe in an environment where the wind is more dominant such as the plains of central United States.

Wyoming gages have been maintained for several years by the Soil Conservation Service (USDA) in Imnavait watershed and at many other locations in Alaska. The gage in Imnavait watershed is located on the ridge between Imnavait Creek and the Toolik River. This gage is named after the Toolik River in the S.C.S. publications. Realizing that snowfall precipitation is under-measured, monthly data from this gage and two others on the North Slope are averaged over several years and presented in Figure 5-1. The wettest months are obviously July and August, with May being the driest. This trend appears to be true for the North Slope of Alaska. While precipitation appears to be low during the winter months, the total accumulation of snow over 8 months is significant. The small numbers at each point in Figure 5-1 represent the number of months of data available for calculation in the average.

One process that goes on during the winter is the sublimation of the snowpack; this is a very important process when snow is being redistributed by winds. In this study no measurement of this rate of loss was made. However, by comparing the results of data from the Wyoming gage and measurements of snow on the ground just prior to ablation, it appears that sublimation losses are of the same magnitude as the under-measurement of snow precipitation. This was concluded because the snow on the ground at the end of winter is nearly equal to the results indicated by the Wyoming gage (Table 5-1). Statistical analysis of this data indicates no significant difference at the 90% confidence interval.

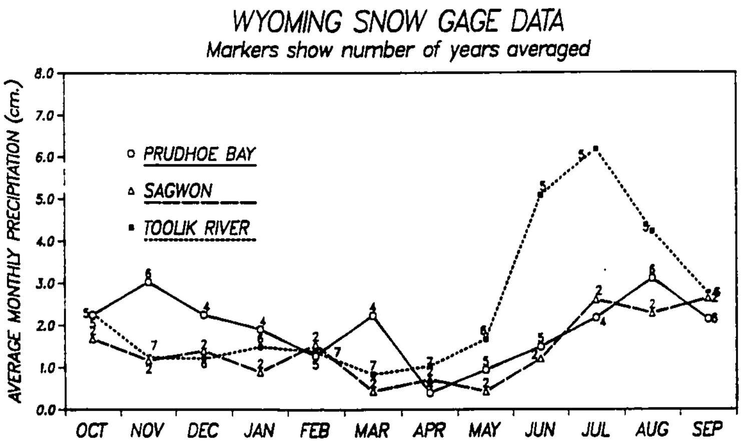

Figure 5-1. Average monthly precipitation at several locations on Alaska's North Slope.

Table 5-1. Comparison of Wyoming gage and snow on ground just prior to ablation.

Year	Wyoming Gage (cm) (water equivalent)	Snow on Ground (cm) (water equivalent)
1985	10.4	10.2
1986	9.9	10.9
1987	10.9	10.8
1988	9.1	7.8
1989	10.9	15.5

This same conclusion could not be reached at the 95% confidence level.

A large percentage of the annual precipitation falls during July and August. During the summer months, snow usually melts within a few days. Winter snow accumulation generally starts around mid-September. Whether runoff is produced from rainfall events during the summer depends upon intensity, duration and antecedent soil moisture conditions. The relationship between summer precipitation and runoff is shown in Table 5-2 for major events. There were several other major runoff events in 1987, but a combination of the facts that the storms overlapped and that snow was mixed with the rain, makes them very difficult to analyze.

2) Soil Storage

The subsurface system, for hydrological purposes, is limited to the shallow active layer. This permits easy access for the purpose of measuring soil temperatures, moisture contents and collecting soil water samples.

Table 5-2. Comparison of rainfall inputs and runoff.

Date	Rain (mm)	Runoff (mm)	5 Day Antecedent (mm)	10 Day Antecedent (mm)
1986				
Jul 27-Aug 6	43.28	23.65	7.93	13.94
Aug 6-Aug 9	19.77	13.39	4.32	39.42
Aug 26-Sept 4	37.9*	23.23	*	*
1987				
Jul 17-Jul 31	57.2	34.6	12.6	12.6

* This was primarily a summer snowfall event.

The permafrost table can be considered as an impermeable surface for water balance computations. Water movement into and storage in the active layer is restricted even more by the seasonal frost. Freezing of the active layer starts in earnest in mid-September and thawing is not initiated until late-May or early June (Figure 5-2). This figure shows the annual variation that can be expected in the temperatures in air at 1.5 m, at the surface of active layer and at a soil depth of 40 cm.

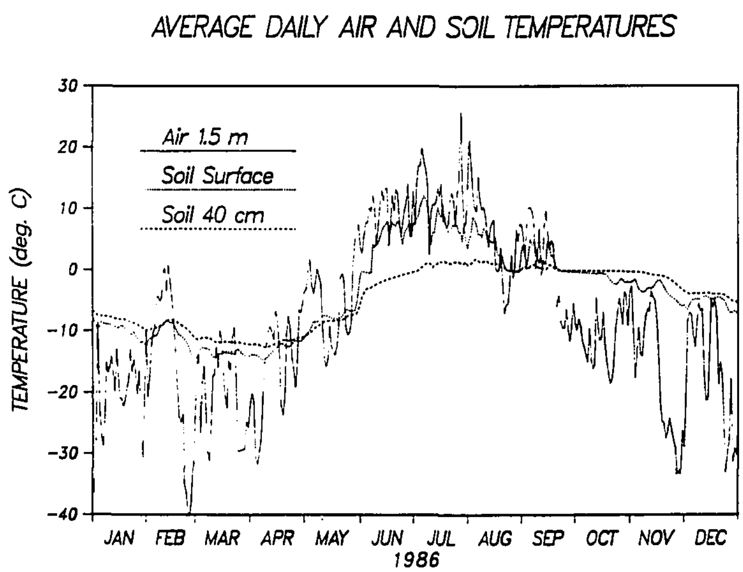

Figure 5-2. Average daily temperatures of air at 1.5 m soil surface and soil at 40 cm depth.

The maximum depth of thaw was only slightly greater than 40 cm in 1986 and in other years did not deviate very much from this value. The maximum summer temperature at the 40 cm depth was about 2°C and the temperature at this depth remained near 0°C for about two months during freeze-up.

The ability of the active layer to store large quantities of groundwater is severely limited because of the shallow depth. And because the mineral soil remains nearly saturated the year around (Hinzman et al., 1990), changes in soil storage take place in the near surface organic soils. The amount of soil storage increase before runoff is produced only depends upon the moisture levels within the active layer. During the winter months some desiccation of the organic soils takes place as soil moisture moves against gravity, in the direction of the colder surface temperatures. Rewetting of these organic soils is necessary before snowmelt runoff is possible. During the summer months evaporation and transpiration are responsible for drying the organic soils.

From laboratory measurements of soil properties and field measurements of soil moisture content, on the average about 15 mm of snowmelt water goes into storage in the active layer. This can be compared with the response of the watershed to rainfall precipitation in the summer (Table 5-3). On three occasions in 1986, between 12.75 and 15.14 mm of rain fell and only insignificant quantities of runoff were produced. Two similar events occurred in 1987.

Table 5-3. Rainfall events that produced minimal runoff.

Date	Precip (mm)	Runoff (mm)	5 Day Antecedent (mm)	10 Day Antecedent (mm)
1986				
Jun 18-Jun 26	12.75	0.8	0.0	0.24
Jul 9-Jul 11	12.75	0.04	0.0	0.24
Jul 13-Jul 21	15.14	0.3	12.75	12.75
1987				
Jun 26-Jun 27	13.7	~0.0	0.0	2.2
Jul 10-Jul 11	12.6	~0.0	0.0	7.5

Quantification of the hydrologic role of the active layer has been minimal. Steer and Woo (1983) examined suprapermafrost groundwater flow and compared it to the volume of surface runoff at a high arctic site. They found that surface runoff was 2.5 times greater than subsurface flow in the active layer. In contrast, Lewkowicz and French (1982) concluded that subsurface flow to the stream was more important than surface. Mackay (1983) studied water movement, particularly snowmelt, into the active layer in response to the temperature gradient and concluded that heave could occur during the summer months. This may have implications in water balance calculations, however at this time the error associated with most water balance measurements in cold regions would preclude this concern.

3) Evaporation and Transpiration

The most difficult and least accurate of all watershed process measurements are those that include water losses back to the atmosphere. This is particularly disturbing because whether one is in the Arctic, temperate or tropic regions of the world, this is the major mechanism whereby water leaves a basin.

With a relatively impervious barrier so close to the ground surface, saturated conditions exist in the active layer near the surface allowing substantial evaporation and transpiration during the summer thawing months. When ablation is occurring, the active layer is completely frozen. However, the near-surface organic soils with high porosity and low moisture contents, readily accept meltwater. Highly decomposed organic soils with high moisture contents may be relatively impermeable. This induces near-surface runoff to shift from surface to subsurface flow. This, coupled with a much lower albedo once the snowcover has ablated, provides both water and energy for evaporation. The porous organic soils absorb moisture and drain quickly in response to spring snowmelt and summer precipitation; however the mineral layer generally is saturated throughout the summer.

Pan evaporation at Imnavait Creek in the summer of 1986 averaged 4.2 mm/day: this compares with 3.4 mm/day in 1987. For this basin, from water balance computations, Kane and Hinzman (1988) estimated the evapotranspiration to average 2.0 mm/day for the months of June, July and August in 1986.

At Prudhoe Bay on the northern coast of Alaska, Kane and Carlson (1973) measured pan evaporation at 2.8 mm/day and lake evaporation at 2.1 mm/day during parts of June and July. Roulet and Woo (1986a) computed average daily losses of 4.3 mm/day for a wetland in continuous permafrost. Woo et al. (1981) using the Priestley and Taylor model calculated the evaporation for July and August near Resolute in the Canadian Arctic for different surface types. They calculated the daily evaporation from a gravel surface to be less than 1 mm/day.

The rate of water loss will depend upon atmospheric conditions suitable for evaporation and transpiration and the availability of water. Under ideal conditions in the Arctic, it appears that daily atmospheric losses could exceed 4 mm/day. The lower limit is near zero, but the actual value will depend upon many surface features such as topography, vegetation, soil type, drainage, etc.

4) Near Surface Drainage and Runoff

Runoff leaving the basin is usually confined to a period of four months. There is no flow in Imnavait Creek until snowmelt and by mid to late September flow has ceased. Because of the very shallow subsurface system, flow can cease in the summer during periods of little or no precipitation. Almost all small to medium sized basins in the Arctic behave in this manner. Significant flows in the stream channel occur both from snowmelt and rainfall events. For this basin and the three years of data collection, snowmelt runoff events have been the most important, both in terms of peak flow and volume of runoff. It is likely that the peak of record for this small basin will be a rainfall generated event. Convective storms, although not as intense as in more southerly regions, can provide rainfall at rates that far exceed the rate of snowmelt. Energy availability for phase change is the limiting factor.

An interesting surface drainage feature in the Arctic is the existence of water tracks. These small drainage channels carry water off of the slopes down to the valley bottom. They generally take the most direct route down the slope but do not connect directly with the stream in the valley bottom. As the slope flattens out in the valley bottom, water moving down the water tracks disperses into numerous, poorly defined channels and slowly makes its way over to the channel of the main stream. Water moves downslope in these water tracks much more rapidly than by subsurface means.

During snowmelt, the role of the snowpack in retarding snowmelt generated runoff is very obvious. Snow, redistributed by the wind, accumulates in both the water tracks and valley bottoms where the melt water collects. At first water seeps through the snow as in any porous media. However it reaches a degree of saturation when both the snow and melt water start to move, cutting a channel through the snowpack. Often the stream channel, which is narrow and shallow as well as filled with ice, is not aligned with the channel cut by the flowing snow and water. Once a channel has been cut through the snowpack and the water in the surrounding snowpack has drained away, there is no mechanism for additional channels to be cut. Therefore, the runoff may flow out of the stream channel for the entire snowmelt period or until the snowpack melts.

Recently, some aspects of the hydrologic role of the snowpack in delaying channel runoff have been examined. Woo (1982a) discussed the impact of snowdrifts across the stream. Kobayashi and Motoyama (1985) examined the time lag in runoff caused by snow depth and stratigraphy for a slope setting. Woo and Heron (1987) developed a generalized sequence of events where snow and ice are important elements of the runoff process. The importance of snow in hindering runoff in this basin can be established by comparing the initiation of runoff to the remaining snow on the ground. Kane et al. (1990a) show that for three consecutive years, the west-facing slope has seen an 80% reduction in the water content of the snowpack before stream runoff begins at the basin outlet. This is significant since this slope represents 78% of the total basin area.

The runoff response of the watershed to rainfall was discussed in an earlier section. Since the subsurface system has a very limited amount of storage, runoff is produced for all storms in excess of 15 mm. However to date, peak values of runoff have always been lower than snowmelt runoff because of the low rainfall intensities.

Many researchers have observed the typical pattern of the snowmelt runoff being the dominant event in the

Arctic. Woo et al. (1981) showed that over three quarters of the recharge of a high arctic lake came from snowmelt. Lewkowicz and French (1982) reported on the lack of rainfall generated runoff over a three year period on Banks Island in the Canadian Arctic. It should be expected that large quantities of snowmelt runoff are generated because of the limited soil storage at that time in the active layer and the extensive period of time that snow accumulates. There is a strong relationship between the water content of the snowpack and the volume of runoff for several small plots constructed on the west-facing hillslope (Figure 5-3). The percentage of runoff increases as the water content of the snowpack increases; this is due to the low upper limit of soil storage and minimal losses due to evaporation. This figure also shows that during periods of light snow accumulation that higher percentages will be lost to storage and evaporation and for water contents less than 35 mm, little or no runoff will be produced.

F) Energy Related Processes

In northern latitudes, one cannot study hydrology without including processes of energy transfer. Incoming solar radiation varies from none during the winter months to a maximum that nearly coincides with

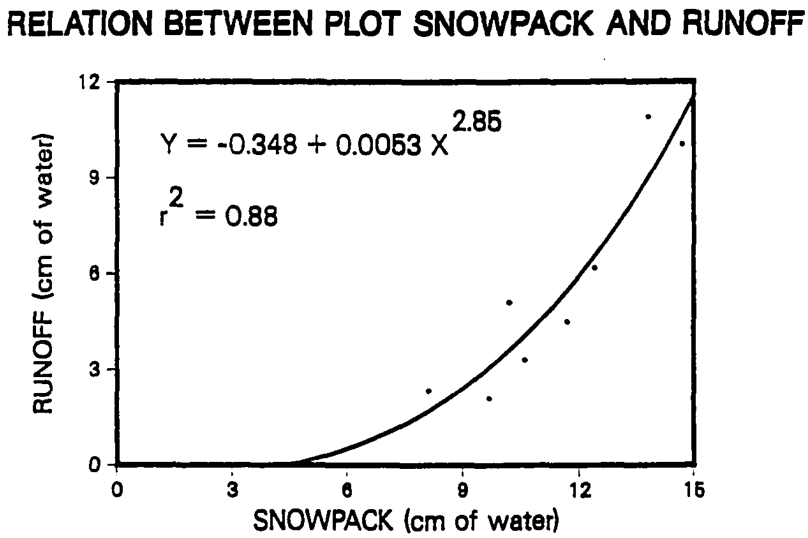

Figure 5-3. The relationship of the initial snowpack water content and subsequent snowmelt runoff from 1985 and 1986 data.

ablation. The rate of snowmelt proceeds according to the net energy at the surface. Also, losses due to evaporation are controlled mainly by the net energy at the surface. The rate of soil thawing depends upon the surface temperature and thermal and moisture properties of the soil.

A simple energy equation can be developed along the lines of the water balance equation:

$$(1 - \alpha)S + L_i - L_e = Q_N = Q_S + Q_L + Q_C \qquad (2)$$

where

α = surface albedo

S = solar radiation

L_i = incoming long wave radiation

L_e = emitted long wave radiation

Q_N = net radiation at surface

Q_S = sensible heat flux

Q_L = latent heat flux

Q_C = ground heat flux

The above equation can be more complicated for cold regions than shown. Lunardini (1981) discusses the increased complexity if heat transfer is coupled with mass transfer.

1) Net Radiation

Radiometers used to measure net radiation from March through August show the expected variation (Figure 5-4) that will lead to ablation and the onset of summer runoff. Because of the large amount of incoming solar radiation in March and April, there are some days with a small net positive energy balance. During the snowmelt period, an obvious increase in the net energy is observed. As soon as the ablation is complete, there is a very significant jump in the net energy balance. Most of this excess energy at the surface is utilized for thawing of the active layer, evaporation and transpiration. During the summer months, a gradual decrease in the excess surface energy is seen as the amount of incoming solar energy diminishes. In early September, snow with its high albedo returns and the energy balance at the surface is again similar to late winter conditions.

2) Surface Albedo

The greatest change in the surface energy balance occurs in response to albedo change during ablation. The change in surface albedo during ablation is shown for three consecutive years (Figure 5-5). It is obvious from this figure, each year after ablation the bare surface values of albedo are consistent. The high peaks

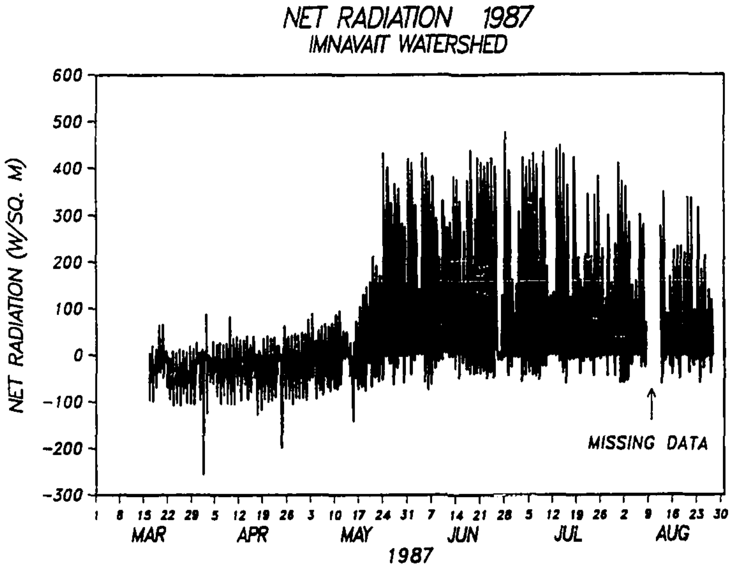

Figure 5-4. Average hourly net radiation from March through August, 1987.

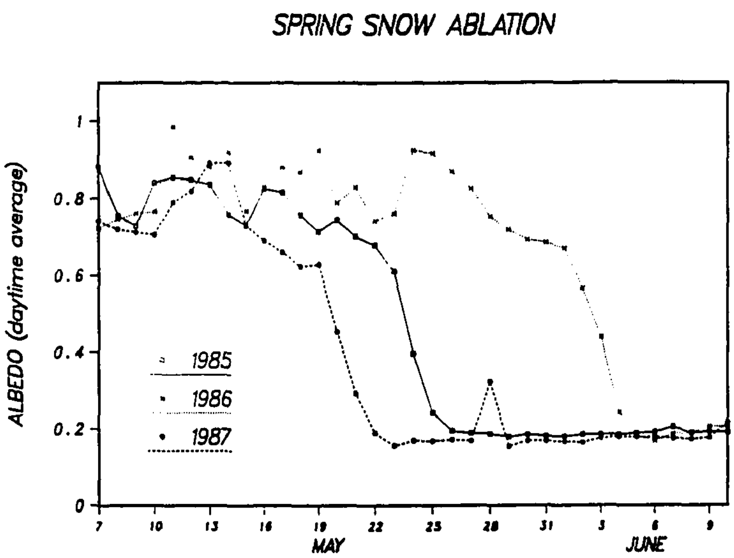

Figure 5-5. Average daytime albedo during the snowmelt period.

observed before the major melt periods are fresh snowfalls that can raise the albedo near a value of 1.0. Figure 5-5 is also very useful for showing the range of dates for the termination of snowmelt, May 22 to June 4. During the summer months, the surface albedo remains fairly constant except for brief periods of snow cover.

After the ablation of snow, the amount of solar radiation reaching the ground increases by a factor of 4. With this surface warming, heat conduction becomes the major heat transfer process in the thawing of active layer.

3) Heat Conduction

Data collected from heat flux plates buried at various depths in the soil profile give a clear picture of the heat transfer processes associated with arctic soils. Seasonal heating at the surface causes temperature gradients resulting in heat transfer by conduction. The thermal properties of the soil are dependent upon the temperature, moisture content, density and soil type. There is an abrupt change in the soil thermal conductivity between frozen and unfrozen soils. For the same moisture content, organic soils at Imnavait Creek have thermal conductivities that increase by a factor of two when frozen. There is also an abrupt change in the thermal conductivity between the organic and mineral soils, primarily because the organic soils retain less soil water and therefore have much lower thermal conductivities. This is very important process because it limits the depth of thaw of the active layer. The most cloud-free period is in June and this coincides with annual maximum incoming solar radiation; however, the ability of the organic soil to conduct heat is significantly reduced because of the low moisture contents that exist in the large pores of the surface organic soils (Hinzman, et al. 1990).

Results from a heat flux plate from early summer until December are shown (Figure 5-6). Prior to ablation and immediately following ablation, the heat flux does not show much daily variation. It is not until the soil is thawed that we see daily variation in the heat flux in response to surface heating. First, the snow cover reflects a large percentage of the incoming radiation and acts as a buffer to heat transfer because of its low thermal conductivity. Then, as the soil is thawing there is phase change of the soil ice. Once the soil thaws, the soil heat flux and the daily variation are potentially at a maximum. During the remainder of the summer there will be a tapering off of the heat flux. For sustained cold spells during the summer, the soil may be isothermal and of course the heat flux is zero.

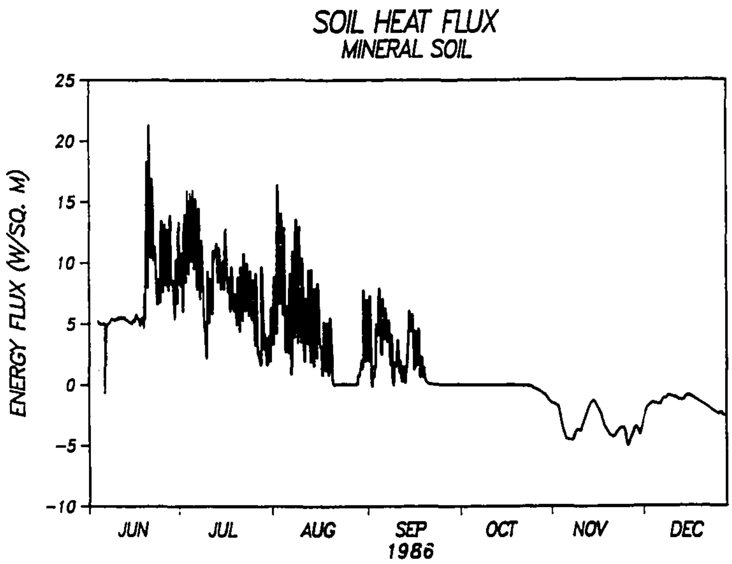

Figure 5-6. Hourly soil heat flux at 20 cm depth from June through December, 1986.

With cooler fall temperatures, the role of radiation in heating the surface soils becomes less dominant and the daily fluctuations disappear.

G) Conclusions

Hydrology in the Arctic differs substantially from that in more temperate regions. The spring snowmelt guarantees that water will be available for the plants each spring at about the same time. This is important because precipitation is then near an annual minimum. Snowmelt is always a dramatic hydrologic event in the Arctic; this is because it coincides with the near maximum incoming solar radiation and the retarding effect the snowpack has on runoff.

The summer hydrology can vary significantly from year to year, depending upon the pattern and magnitude of summer precipitation. The surface organic soils seem to play a much more dominant hydrologic role than the deeper mineral soils. The mineral soils remain near saturation much of the year with phase change being the dominant activity. During dry periods, runoff is minimal or ceases in the stream and water tracks, indicating that subsurface flow contributions from the thawing active layer are minimal, if not insignificant.

H) Acknowledgments

The research reported here was supported by the U.S. Dept. of Energy's Office of Health and Environmental Research, Ecological Research Division as part of the R_4D program in Arctic Tussock Tundra. R.E. Gieck, E.K. Lilly and C.G. Egan assisted in data collection, reduction and preparation of figures.

VI) PERMAFROST HYDROLOGY OF A SMALL ARCTIC WATERSHED

Kane, D.L., and L.D. Hinzman. 1988. Permafrost hydrology of a small arctic watershed. K. Senneset (ed.) Proc. Fifth International Conference on Permafrost, Trondheim, Norway. Tapir, p. 590-595.

A) Abstract

A small 2 km^2 watershed in an area of continuous permafrost has been monitored for three years. Measurement of precipitation, including detailed snow surveys prior to ablation, runoff leaving the basin, and the soil moisture regime in the active layer permit the determination of the basin water balance. Similar measurements were made for four bounded hillslope plots. As expected, spring snowmelt is the major hydrologic event each year, with this water being partitioned into soil storage, evaporation and runoff. Runoff dominates this portion of the annual hydrologic cycle. This is visually evident in the watershed by the large number of water tracks that convey water toward the main stream channel. Moist soil conditions in the fall that produce high ice contents when frozen severely limit infiltration into the mineral soil. Summer precipitation events, although never near the volume of snowmelt, can produce peaks in the streamflow hydrograph if the intensity is high. How fast the basin responds depends upon the antecedent soil moisture conditions. Downslope movement of water is initiated when saturated

conditions develop in the near-surface organic soils since the less permeable underlying mineral soils retard infiltration.

B) Introduction

The hydrology of an arctic watershed exemplifies the accepted principles of watersheds in temperate and tropic regions. However, the predicted response of the hydrologic regime can vary substantially because of environmental and physical differences. The large amount of energy lost during the winter months is responsible for maintaining permafrost quite near the ground surface. This severely limits the role of the subsurface system in the hydrologic cycle.

This paper examines the water balance of an arctic watershed underlain by continuous permafrost. The hydrologic roles of the subsurface active layer, surface or near-surface runoff, and evaporation are discussed in conjunction with snow and rainfall precipitation. Water balances during the spring snowmelt and the period of surface runoff have been calculated based on numerous field measurements.

Snow plays a very important role in arctic watersheds from the viewpoint of hydrology, climatology, and ecology (Woo, 1982a). Snow may accumulate for as long as 9 months in the Arctic and then ablate in a relatively short time, typically 10 days. The ability of the active layer soils to absorb and store snowmelt runoff is limited in poorly drained organic (Roulet and Woo, 1986a) and mineral (Kane and Stein, 1983a) soils by the high ice contents. Evaporation during the snowmelt period and evapotranspiration during the summer months are difficult to quantify. There have been several attempts at measuring evaporation using small lysimeters and evaporation pans (Brown et al., 1968; Dingman, 1971; Addison, 1977; Marsh et al., 1981; Ohmura, 1982a; and Kane and Stein, 1983b). Alternatively, it is possible to determine evaporation from the residual term of an energy balance (Wendler, 1971; Weller and Holmgren, 1974). Finally, there are several quasi-theoretical approaches such as the Thornthwaite method (Brown et al, 1968; Dingman, 1971; and Kane and Carlson, 1973) and Priestley and Taylor (Rouse and Stewart, 1972; Rouse et al., 1977; Wright, 1981; and Woo et al., 1983). Ohmura (1982a) also used both an aerodynamic profiling method and the Bowen ratio method.

The most important factor controlling summer evapotranspiration in the Arctic for tussock tundra is the moisture level in the active layer. In very few studies has this parameter been measured continuously throughout the year; yet this measurement is critical to understanding the response of the watershed to hydrologic mass and energy inputs.

Precipitation measurements in the Arctic can be characterized as sparse, of short duration, and of very poor quality. This is particularly true for snowfall precipitation, for which the water content is typically underestimated by a factor of 2 to 3 by standard gages when windy conditions prevail (Benson, 1982).

Probably the most reliable hydrologic measurements made in the Arctic are those of runoff. Again, there is the complaint that there are too few streamflow gaging stations and the quality of the data is questionable during periods of ice and high flows.

C) Site Description

Imnavait Creek is a small, 2.2+ km^2 watershed located in the northern foothills of the Brooks Range of Alaska, U.S.A. (Figure 3-1). Permafrost underlies the entire watershed and the active layer thickness ranges from 25 to 60 cm. Soils with organic surface horizons of varying thickness and bulk density mantle the mineral sub-soil and are classified as Histic Pergelic Cryaquepts. The primary vegetation consists of sedge tussocks, mosses and low shrubs. The stream drains northward between east and west-facing slopes that dominate the basin. The stream is a typical beaded stream and no significant bodies of free-standing water exist. Numerous water tracks, flowing directly down the slopes, carry water to the base of the slope. From the base of the slope, no well-defined channel exists from the water tracks to the stream, and water takes a dispersed path across the valley bottom. Tussock tundra and mosses dominate the slopes and the valley bottom, so water moving downslope alternates from surface to subsurface flow. Near-surface flow is prevalent during snowmelt, and may exist during prolonged periods of rainfall.

D) Water Balance

Numerous techniques can be used to determine the water balance of a watershed, hillslope or plot (Van der Beken and Herrmann, 1985). Woo (1986) presents a water balance equation to define the various hydrologic relationships for a basin underlain by permafrost:

$$P + (Q_m + Q_i)/\rho_w\gamma + Q_{in} = E + Q + dS_D + dS_T + dS_A \quad (1)$$

where
P = rainfall
$Q_m/\rho_w\gamma$ = quantity of snowmelt
$Q_i/\rho_w\gamma$ = quantity of ice melt in active layer
ρ_w = density of water
γ = latent heat of fusion
Q_{in} = inflow
E = evaporation or evapotranspiration
Q = outflow
dS_D = change in depression storage

dS_T = change in detention storage
dS_A = change in active layer storage

Each term in Equation 1 represents a flux or a storage change with units of depth over time for a unit area. This equation can be simplified and it can take many forms depending on the physical system and the assumptions made. This paper looks at the water balance during snowmelt, over the summer period, and during individual rainfall events, and in addition will look at the water balance at the scales of both the plot and watershed.

For the snowmelt period, Equation 1 can be simplified to the following form:

$$Q_m/\rho_w \gamma = E + Q + dS_A \qquad (2)$$

Rainfall was insignificant in this study during ablation. This is assuming there was no change in storage terms, except active layer storage. From laboratory measurements of soil properties and field measurements of moisture levels, one could determine the volumetric changes in soil moisture, i.e. the addition of water to the soil from snowmelt. Daily snowmelt and runoff were measured in the field directly. Therefore, estimates of evaporation during the snowmelt period can be made.

For the summer water balance the equation is quite similar to Equation 2:

$$P = E + Q + dS_A \qquad (3)$$

From measurements of field soil moisture levels, it appears that prior to freeze-up each fall the amount of water in storage within the active layer is virtually the same each year. This means, on an annual basis, that both changes in soil storage and melting of ice in the active layer can be ignored. Simply, inflow is equated with outflow.

Very few hydrologic studies in areas of continuous permafrost have measured enough elements of the hydrologic cycle to calculate a water balance. Brown et al. (1968) studied the summer hydrology of a small drained lake basin near Barrow, Alaska. They reported the response of the watershed to individual storms and concluded that summer rainfall was offset by evapotranspiration during a normal year. Marsh and Woo (1979) reported on data from four basins near Resolute, N.W.T., Canada. They showed that 67-81% of the annual precipitation left the basin as runoff and 13-23% as evaporation. Lewkowicz and French (1982) measured the runoff from four plots on Banks Island in the Canadian Arctic Archipelago during snow ablation. Three of the plots were located in areas of drifting snow. Woo

(1983) completed an annual water balance for McMaster Basin near Resolute, N.W.T., Canada for a number of years. Only small changes in storage occurred; evaporation was minimal, and runoff dominated. Roulet and Woo (1986a and b) determined the annual water balance for a peat wetland near Baker Lake, N.W.T. Water moving downslope from a lake was the main hydrologic activity in this basin.

For all of the above studies, it is difficult to make comparisons among the data. The soil types range from coarse grained soils to peat, and some have no surface vegetation. During snowmelt some soils in the active layer were capable of storing meltwater, while in other settings soil storage was nil. In almost all cases, the snowmelt event during break-up was the major hydrologic event of the year in terms of runoff peaks and volumes. The compelling reason is that permafrost close to the ground surface severely limits storage regardless of the soil type.

E) Discussion

Measurements of snowfall and rainfall were made continuously within the Imnavait Creek watershed. A Wyoming shielded precipitation gage operates throughout the year and two standard 8 inch (20 cm) shielded gages complement the Wyoming gage during the summer (Figures

6-1a and b). In addition, at winter's end, detailed surveys of snow on the ground are made using an Adirondack snow corer. Liston (1986) has shown that the Wyoming gage gives a good indication of the snow on the ground at the end of the winter. It is known that moisture is lost from the snowpack throughout the winter due to sublimation. Sturges (1986) has shown that Wyoming gages, although shielded, under-measure the total snowfall. Therefore, it appears that the error of the Wyoming snow gage is comparable to the cumulative sublimation during the winter. For water balance computations, the water content of the snow on the ground just prior to ablation was used.

The pattern of rainfall for 1986 and 1987 (Figures 6-1a and b) illustrates both low intensities and low quantities. There is a general trend of increasing rainfall during the summer months. Four arrays of time domain reflectometry (TDR) probes were installed along the slope to measure the unfrozen water content of the active layer. TDR is an indirect method of measurement whereby the dielectric properties of the soil are measured (Stein and Kane, 1983). The annual variation in the unfrozen water content at selected depths for one of the profiles (Figure 6-2) indicates that the unfrozen water content increases as the active layer thaws and as

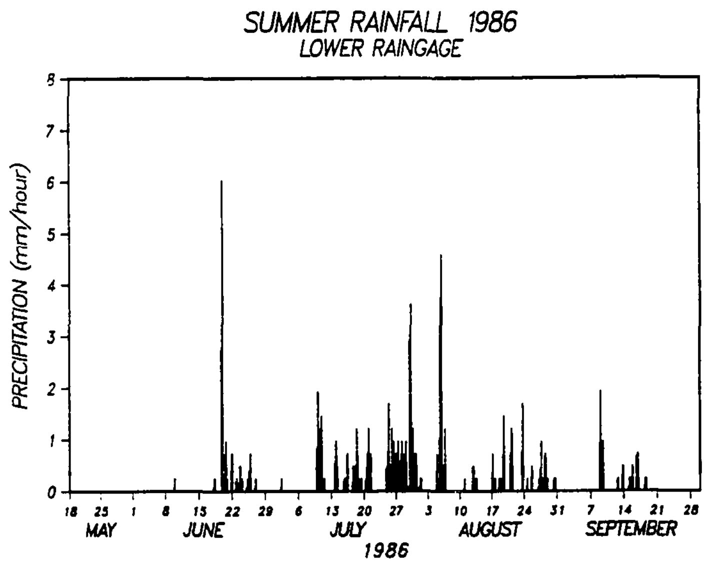

Figure 6-1a. Pattern of summer rainfall at Imnavait Creek Watershed in 1986.

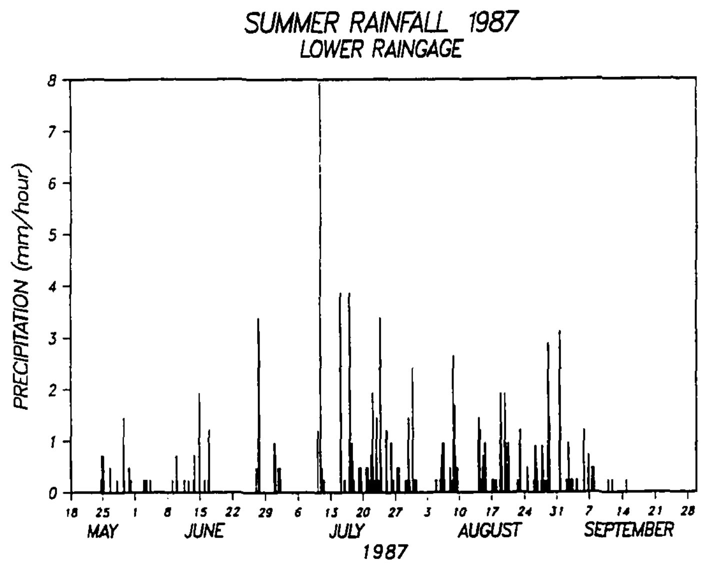

Figure 6-1b. Pattern of summer rainfall at Imnavait Creek Watershed in 1987.

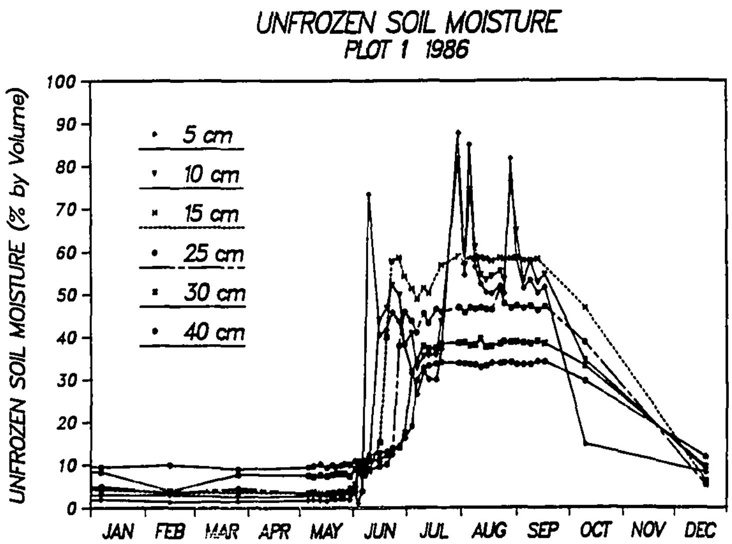

Figure 6-2. Unfrozen soil moisture variation at selected depths, 1986.

the rainfall shown in Figure 6-1 a and b occurs.

The soils above 15 cm depth in Figure 6-2 are highly organic, with mineral soils at greater depths. Once the mineral soil thaws at a given depth during the summer, the moisture content remains practically constant. This is also true when data from a number of years are compared. A plot of the unfrozen water content profile in late summer (Figure 6-3) shows that there is very little variation from year to year for the mineral soil, but there is some variation at the two upper depths in the organic soil. The actual values depend upon the length of time since rainfall or late summer snowmelt. By spring this layer of organic soil is quite dry, except where it is highly decomposed and has a density approaching that of the mineral soil. So, there is virtually the same quantity of water in active layer storage each spring prior to ablation. A major assumption in the annual water balance calculation is that there is no change in storage from year to year.

Runoff was measured both in four plots constructed on the hillslope and in the Imnavait Creek basin. The plots were 89 m^2 in size and bounded with heavy ($\approx$1 mm) plastic to prevent overland and subsurface flow into the plots. Runoff above the mineral soil was collected in a collection gutter and was routed to a holding tank

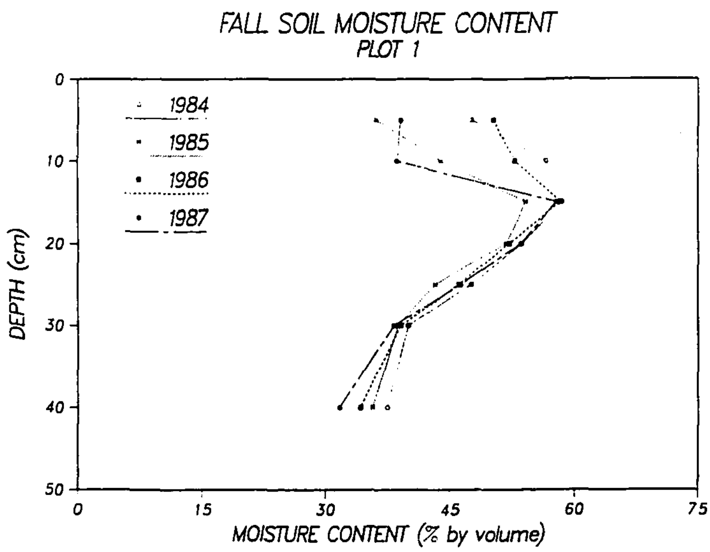

Figure 6-3. **Typical soil moisture profiles just prior to seasonal freezing.**

equipped with a water level recorder. The plots were uniformly spaced from near the base to near the top of the west-facing slope and were continuously monitored during snowmelt. For the summer months, they have been monitored for some individual storms.

The watershed outflow was measured continuously in 1986 and 1987 (Figures 6-4a and b), allowing annual water balances to be determined. A stage-discharge curve was constructed for the stream in 1986, and a flume was installed in the channel in 1987. Spring snowmelt is usually the peak annual runoff event because of both the limited storage in the active layer and the significant quantities of snow that accumulate during the winter. Runoff during the summer months is minimal, except during substantial rainfall events that produce near-surface runoff.

A comparison of pan evaporation with summer precipitation (Figures 6-5 a and b) shows that potential evaporative losses exceed precipitation. While there is a gradual decline in pan evaporation over the summer, precipitation increases in late July and August.

Water balances were computed for two periods, spring snowmelt and annual. For the snowmelt period, computations were made for both the plots and the basin using three years of data (Figures 6-6 a, b and c). The

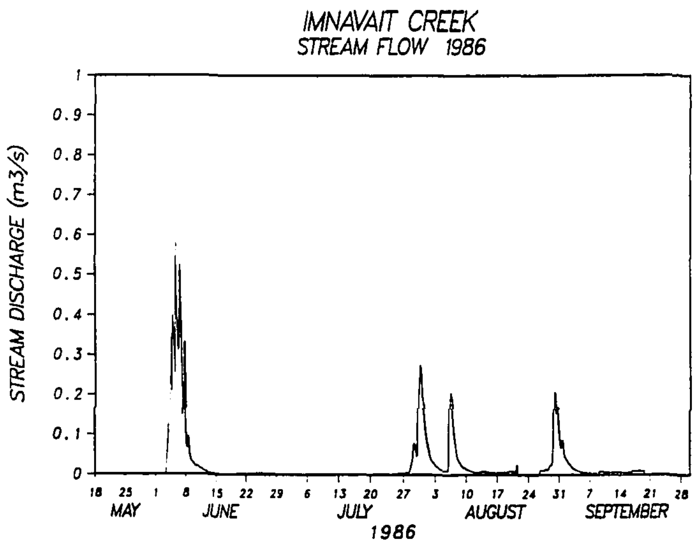

Figure 6-4a. Snowmelt and rainfall generated runoff from Imnavait Creek, 1986.

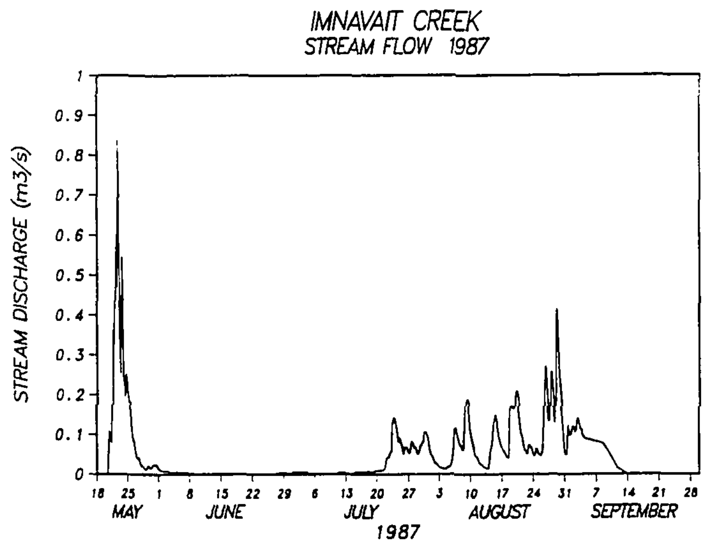

Figure 6-4b. Snowmelt and rainfall generated runoff from Imnavait Creek, 1987.

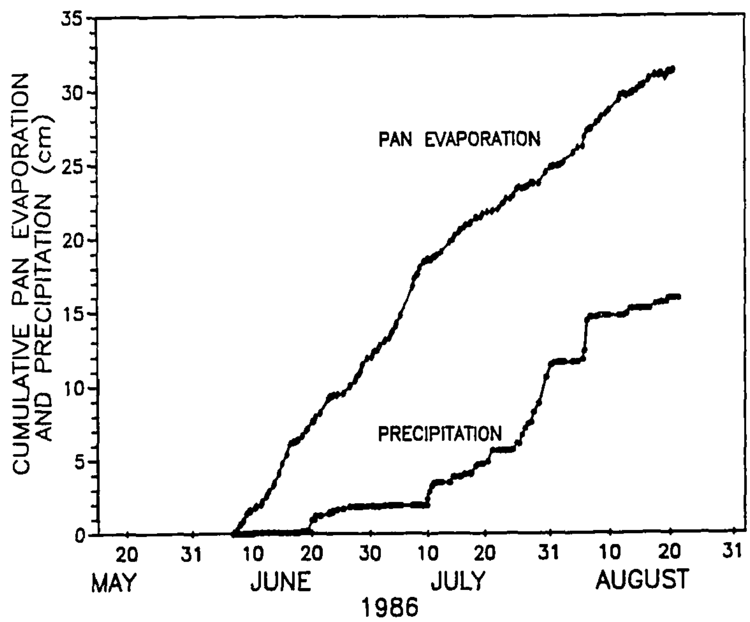

Figure 6-5a. Cumulative pan evaporation and summer rainfall precipitation in 1986.

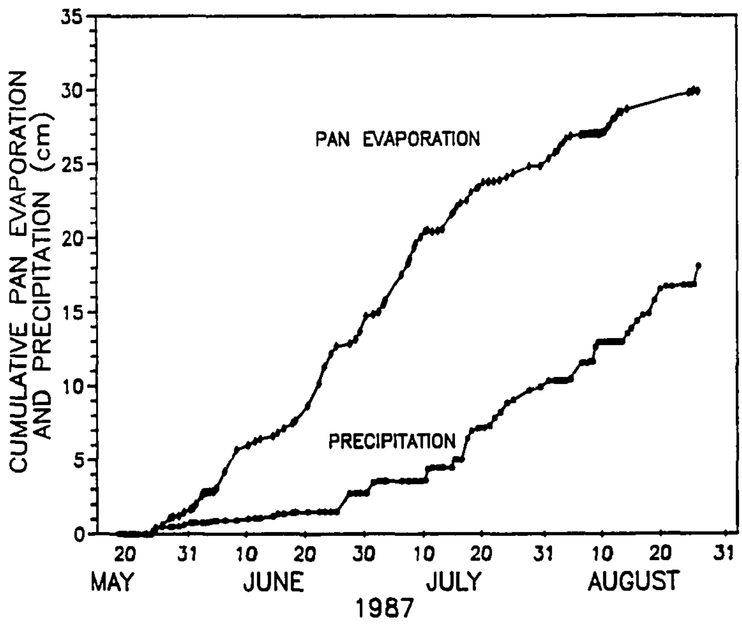

Figure 6-5b. Cumulative pan evaporation and summer rainfall precipitation in 1987.

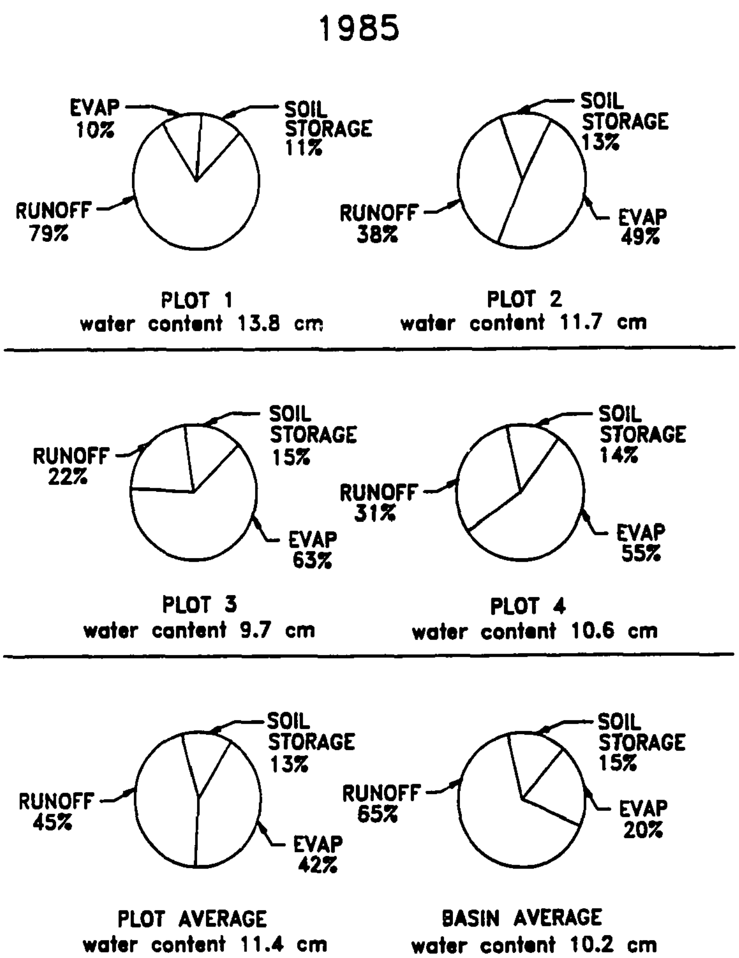

Figure 6-6a. Charts illustrating the partitioning of snowmelt into runoff, storage, and evaporation in 1985.

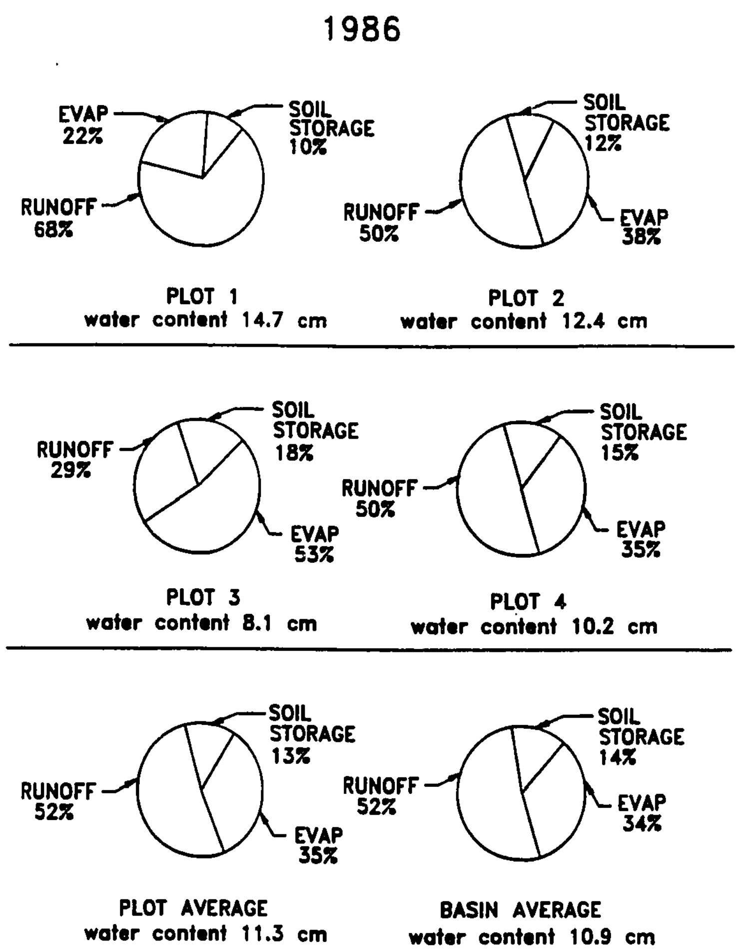

Figure 6-6b. Charts illustrating the partitioning of snowmelt into runoff, storage, and evaporation in 1986.

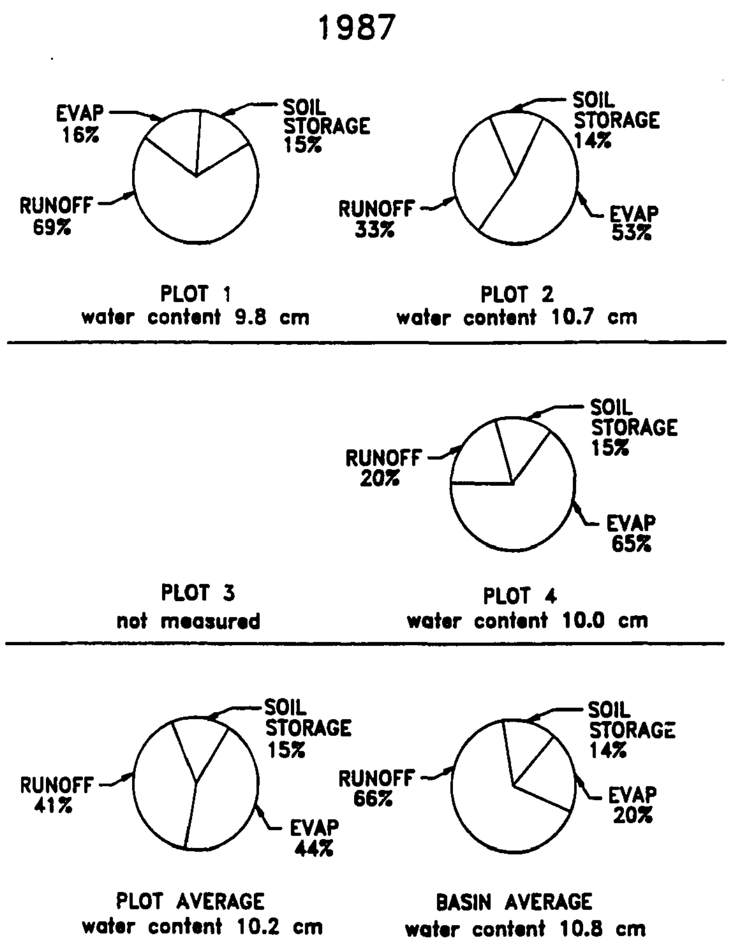

Figure 6-6c. Charts illustrating the partitioning of snowmelt into runoff, storage, and evaporation in 1987.

annual water balance was calculated for the basin only in 1986 and 1987 (Table 6-1). The runoff measurements for the plots are not continuous over the summer, so annual balances cannot be performed for them. During 1985, a breakdown of the water level recorder during a major rainfall event prevented similar calculations.

Table 6-1. Annual water balance of Imnavait watershed.

Year	Snowpack Water Content (cm)	Summer Precip (cm)	Total Precip (cm)	Snowmelt Runoff (cm)	Summer Runoff (cm)	Total Runoff (cm)	Evapo-transpiration (cm)	Pan Evap (cm)
1986	10.9	16.3	27.2	5.7	6.2	11.9	15.3	31.0
1987	10.8	27.2	38.0	7.1	17.9	25.0	13.0	32.0

F) <u>Conclusions</u>

Considerable variation exists among the four runoff plots in regards to percentages of runoff and evaporation losses during snowmelt. Although there are some differences between the plots in terms of slope and active layer soil structure, the most important factor is the distribution of snow on the ground. The plots with the highest snowpack water equivalent also have the highest percentage of runoff. Likewise, those plots with the least snow have the highest percentage of evaporation. Where snow is thin, vegetation extending through the snow and bare ground enhances radiation absorption and surface heating. This provides considerable energy to drive evaporation, whereas the surface albedo remains quite high for the thicker snowpacks. Comparison of the same plots over the three years of study shows that there is much less variation temporally than spatially.

In 1986, snowmelt runoff represented 48% of the annual runoff, while the snowpack water content represented 40% of the annual effective precipitation (snow on ground just prior to ablation plus summer precipitation). Although the water content of the 1987 spring snowpack was nearly equivalent to the 1986 spring snowpack, the snowpack water content was only 28% of the annual effective precipitation and snowmelt runoff was also only 28% of the annual runoff. Runoff is the dominant process during the snowmelt period, and evapotranspiration dominates during the dryer summers.

Summer runoff events are controlled by the antecedent moisture conditions of the organic soils in the active layer because once thawed, the mineral soils remain near saturation throughout the summer months. Runoff from the plots ceases shortly after snowmelt or rainfall events. During summer periods of little or no precipitation, runoff is virtually zero in the stream. Water from the thawing active layer which could provide runoff during the summer does not appear to be an important process in Imnavait Creek watershed. The water that is made available by the thawing active layer is probably used by plants or lost through evaporation. A comparison of the pan evaporation data and the computed evapotranspiration (Table 6-1) shows that insufficient water is available for evapotranspiration to proceed at its potential rate.

G) Acknowledgments

The research reported here was supported by the U.S. Department of Energy's Office of Health and Environmental Research Ecological Research Division as part of the R4D program in Arctic Tussock Tundra. This research project was quite labor intensive and several individuals contributed to experimental design, data collection and analysis. Thanks are due to Dr. Kaye R. Everett, Dr. Carl S. Benson, Catherine G. Egan, Robert E. Gieck, Elizabeth S. Lilly, Glen E. Liston, Bertram Ostendorf, Michael R. Lilly, Steven J. Hastings, Knut Sand, and Matthew D. Zukowski.

VII) SNOW HYDROLOGY OF A HEADWATER ARCTIC BASIN

Kane, D.L., L.D. Hinzman, C.S. Benson and G.E. Liston. 1990. Snow hydrology of a headwater arctic basin. Paper submitted to Water Resources Research.

A) Abstract

The hydrologic cycle of an arctic watershed is dominated by such physical elements as snow, ice, permafrost, seasonally frozen soils, wide fluctuations in surface energy balance, and phase change of snow and ice to water. Hydrologic data collected over a five year period (1985 - 1989) from an arctic watershed in northern Alaska are presented. Processes related to snow accumulation, redistribution, ablation, evaporation and subsequent snowmelt generated runoff are discussed. The total water content of the snowpack at the end of winter has comprised between 28 and 40% of the annual precipitation. Redistribution of snow on the ground by the wind is a major factor in increasing the snowmelt runoff as compared to runoff from an area with uniformly distributed snow. Much of the redistributed snow accumulates in the valley bottom along the stream and also along depressions on the hillslopes. These depressions are small surface drainage channels that are referred to as water tracks. Partitioning of the snowmelt into runoff, evaporation and increased soil-water storage in the active layer was carried out on the

plot and watershed scale.

B) Introduction

The physical presence of the snowpack as a buffer to heat transfer, the rewetting of the soil from snowmelt and the change in surface albedo following snowmelt are three processes that strongly influence summer biologic response in the Arctic. The accumulation of snow from late September to May and subsequent melt guarantees that there will be no moisture deficit for plants in spring. The snowmelt hydrology has been closely studied the last five years (1985-1989) in a small 2.2+ km^2 watershed located near Toolik Lake, Alaska. The objective of this study was two-fold: first, to quantify the processes in the Arctic associated with snow accumulation and ablation, and secondly, to integrate these results with the products of other studies to understand more fully the ecology of this cold region system.

While studies of the nature described here have been performed for many years in more temperate climates, in-depth hydrologic studies in the Arctic are limited. There are several reasons for this: lack of resource development, harsh climate, lack of suitable equipment for working in such environments, and possibly ignorance of the role that hydrology plays in arctic ecology.

True hydrologic research in the Alaskan and Canadian Arctic began in the early 1970's in response to oil and gas development. However, most of these early studies concentrated on only one or two aspects of the hydrologic cycle, and in most cases this was surface runoff. Many of the early studies were not even initiated until after snowmelt had been completed. Data collection in this environment was nearly impossible at remote sites in the winter; recent advances with solid state electronics have significantly reduced this problem.

The need for arctic hydrologic data is typical of other areas: water supply, flood forecasting, instream flow requirements, snow loads and the design of various hydraulic structures. More important however, may be the need to understand the interaction of the hydrology and ecology. It is the vegetation that sustains animal life and also protects the permafrost and thawed surface soils from excessive erosion. The growing season is very short in the Arctic and therefore it is very important that these plants get an early start in the spring. Soil water from snowmelt each spring provides sufficient water to initiate this growth.

While hydrologic measurements are made throughout the year, only results from the snowmelt period will be presented in this paper. This is because the major hydrologic event each year is snowmelt and this is an area of limited study.

C) Study Area

Imnavait Creek (Figure 3-1) is a small headwater basin located between the Toolik and Kuparuk Rivers in the northern foothills of the Brooks Range (Lat 68° 30', Long 149° 15'). The north draining basin trends from a non-incised valley at the headwaters to an incised valley at the measured outlet. Headwater hillslopes are around 10% on the west facing slope and slightly greater than 1% on the east facing slope. This is in contrast with >13% west facing slope and >7% east facing slope at the outlet. Most of the field measurements were made in the center of the basin on the west facing slope where the slope averaged 10%. The west facing slope constitutes 78% of the basin area.

A small beaded stream (Figure 7-1) drains the basin. The beads are small ponds of about 1 to 3 m in diameter caused when ice-rich permafrost soils melt. These beads are connected by a small channel typically 30 cm wide and 30 cm deep. During snowmelt the flow completely inundates the stream and in some cases because of high density snow, the flow is not even in the channel. No

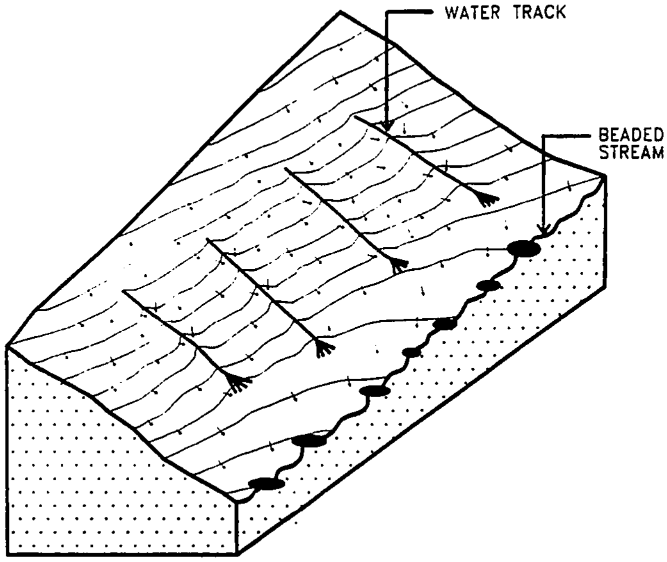

Figure 7-1. **Cross-section of hillside displaying water tracks and beaded stream.**

flow exists in the channel during the winter months; however, the beads/ponds along the stream are filled with water that can exceed 2 m in depth, with the deeper ponds remaining partially unfrozen at depth throughout the winter.

Water moving off the slopes collects in small water tracks that carry the surface flow toward the stream. These water tracks (Figure 7-1) are highly visible on the slopes and convey flow downslope perpendicular to the elevation contours. At the base of the slope and before entering the stream, they become quite diffused. Over 130 such water tracks on the hillslopes have been identified in this watershed from aerial photography. Only during snowmelt and major rainfall events does flow exist in these water tracks. These water tracks are very efficient at moving water downslope, the travel time being much less than the alternative of flowing over and through the surface organic soils.

This area is completely underlain by permafrost to a depth of 250 to 300 m (Osterkamp and Payne, 1981). During the summer, the active layer thaws to a depth of about 50 to 60 cm along the water tracks, 40 to 50 cm near the ridgetops and 25 to 35 cm in the valley bottoms. The soils tend to be better drained near the tops of the slopes and less well-drained near the valley bottom. Organic soils overlay mineral soils. This is particularly important because large changes in thermal and hydraulic properties exist over extremely short vertical distances (Hinzman et al., 1990). The thickness of the organic soils increases from 10 to 15 cm on the ridges to 25 cm or more in the valleys.

D) Field Instrumentation

Numerous meteorologic and hydrologic measurements have been made in Imnavait Creek. To accurately partition the components of the water balance into runoff, evaporation and active layer soil-water storage, a detailed snow survey was performed to determine the water content just prior to ablation. This included data collected at several hundred points on snow depth and equivalent water content. During the ablation period 40 measurements were made daily on the west facing slope to quantify the daily snowmelt. A Wyoming snow gage, operated by the USDA, Soil Conservation Service, is located within the watershed. In addition, two standard 8 inch (20 cm) orifice rain gages with wind shields are operated during the summer months.

Four 90 m^2 runoff plots were constructed on the west facing slope. These plots were built such that water moving downslope, both on the surface and through the organic soil, could be collected. The rationale for these plots was to determine both the quantity and the quality of runoff generated from different positions on the slope. The water flowed into a trough at the base of the plots and then into a tank instrumented with a water level recorder. All of the plots were bounded with ≈1 mm thick plastic down to the permafrost table to prevent water from moving into the plots.

Alongside each runoff plot, thermistor strings were installed at 5 cm increments from the surface down to the permafrost table. Three thermistors were also installed at 10, 20 and 30 cm above the ground to measure snowpack temperatures. To complement the temperature readings, soil moisture measurements were made at essentially the same soil depths and locations. The method of time domain reflectometry (TDR) was used to measure the unfrozen moisture content. The TDR method is an indirect measurement of soil moisture; dielectric properties of the soil are measured and correlated with the moisture content (Stein and Kane, 1983). Two heat flux plates were installed adjacent to plot 3, one at the organic/mineral soil interface and one at 20 cm depth in the mineral soil.

Numerous meteorological measurements were made near the runoff plots. These including: air temperature, relative humidity, precipitation, wind velocity, wind run, wind direction and various radiation components. Radiation measurements of incoming shortwave, reflected shortwave, total net radiation, total incoming radiation, total emitted radiation and albedo were measured from late March or early April until September. During the summers of 1986, 1987, 1988 and 1989, evaporation pan measurements were recorded.

During the first year (1985), runoff leaving the watershed was estimated from discharge measurements and water level measurements. Both cup type and electromagnetic current meters were used to measure discharge; the latter being preferred when the flow exceeds the channel capacity and ultimately flows through the vegetation. In 1986, an H flume was installed at the outlet of the basin to improve the estimate of discharge.

E) Physical System

During the ablation period, Imnavait Creek watershed is a rather simple hydrologic system to study. Both surface and subsurface storage components play a limited role in this hydrologic system. While the upper layer of organic soils is somewhat desiccated during the early winter and infiltration into this layer can occur, the mineral soils at depth are generally found to have ice contents near saturation, and infiltration is essentially zero. The mass balance of the watershed during snowmelt can be expressed as:

$$R = P - E - dG - dS \qquad (1)$$

where
R = quantity of runoff
P = precipitation, both rainfall and snowmelt
E = evaporation or evapotranspiration
dG = change in volume of storage in the active layer
dS = change in surface storage

For this arctic watershed, it is physically possible to measure the change in the water stored in the subsurface system. The changes in the active layer occur near the surface where they are easily measured. The presence of permafrost, which is ice-rich near the permafrost table, is responsible for this condition. Since permafrost extends to great depths (>250 m) and isolates the surface hydrologic system from any deep subsurface system, no deep instrumentation needed to be installed. Woo (1986), in a review paper on permafrost hydrology, discusses the applicability of the above equation in arctic watersheds.

Surface storage in large ponds or lakes does not occur in this basin. However, storage in numerous shallow, naturally occurring depressions does take place. Also, temporary storage or damming does occur during ablation by the existing snowpack in the water tracks and along the stream channel in the valley bottom.

If the snow cover was equally deep and had a constant water content, then quantifying the melt rate would be easier. Unfortunately, there was considerable spatial variability in the snowpack distribution which was directly due to wind events that may or may not accompany snowfall. The greatest snow depths were found in and along the stream channel and water tracks, as these represented surface depressions, and on the lee slopes (Liston, 1986). Both the ridge tops and the windward slopes were sources for most of the redistributed snow.

At the time of snowmelt, evaporation is a very viable process. Since ablation occurs in late May and early June, it corresponds to a period of near-maximum incoming solar radiation. Transpiration at this time is quite minimal and is assumed to be absent. The final form of the water balance equation is:

$$R = P_m - E - dS \qquad (2)$$

where
R = total runoff from snowmelt
P_m = total melt from the snowpack
E = total evaporation from ground surface
dS = increase in soil moisture storage in the active layer from the onset of melt until snowmelt runoff is complete.

The units used are depth of water over the watershed area.

F) Results and Discussion

1) Snow Accumulation

In Imnavait watershed, snow started to accumulate in late September or early October and the maximum snowpack water content was reached just prior to ablation in mid May. Benson (1982) has shown that attempts to accurately measure snowfall precipitation on Alaska's Arctic Slope with standard National Weather Service precipitation gages has been futile. The Wyoming gage has been shown to be much better at measuring precipitation in a windy environment than the traditional gages (Clagett, 1988).

No significant mid winter melt occurs in this basin. The distribution of the water content in the snowpack on an east-west transect across the valley for four years is shown in Figure 7-2. Also shown in this figure is the topography and position of the four runoff plots. During the winter, the snow cover of this area experiences significant redistribution by the wind, resulting in water equivalent depths ranging from 0 to 150 cm. The snow water equivalent over Imnavait Creek watershed is shown as a function of area in Figure 7-3. Higher than average water contents are consistently found in catchments such as the valley bottoms (Figure

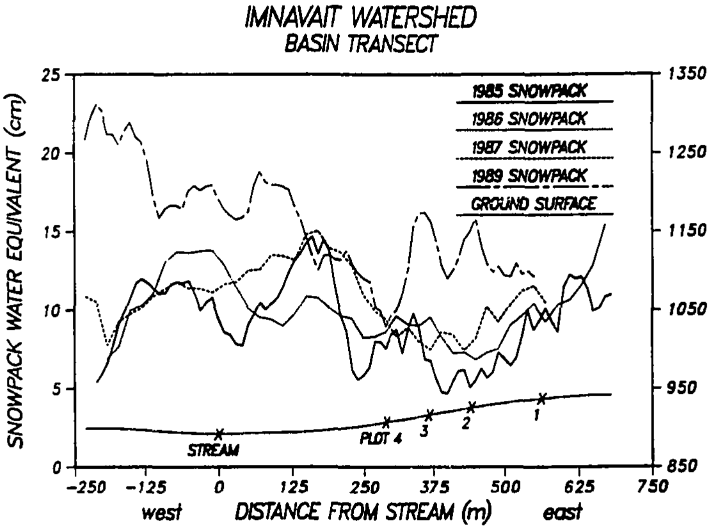

Figure 7-2. 5 sample running average of snow water equivalent and ground surface elevation along a transect across watershed, distance is measured from stream.

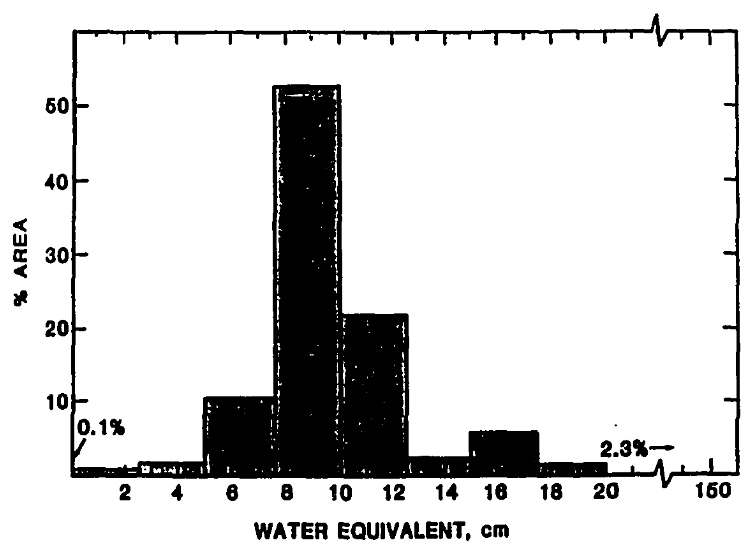

Figure 7-3. Snowpack water equivalent as a function of aerial coverage of watershed in 1985.

7-2) and water tracks. Hydrologically, this is quite significant because it places the snow very close to the drainage channels. This not only results in higher soil moisture levels in these high deposition areas, but produces a higher percentage of runoff during snowmelt than would have resulted if a uniform snowpack had existed.

Snow surveys show that the rolling hills in the area also develop snow accumulations as much as 65% greater on lee slopes of only 2 to 3 degrees. Woo (1982a) has shown that similar patterns of snow distribution are seen in the Canadian High Arctic near Resolute. The depth and water content has been found to be least on the hilltops, followed by the flat areas and slopes, with gullies and valleys holding the most snow.

Typical properties of the snowpack are illustrated in Figure 7-4 at various points within the study site for two different time periods. The effect of the wind in producing high density slabs and the development of depth hoar at the soil-snow interface because of moisture and heat transfer from the active layer into and through the snowpack is evident. Two of the profiles displayed were collected only one day apart but are from different locations with greatly different properties giving an indication of the variability of

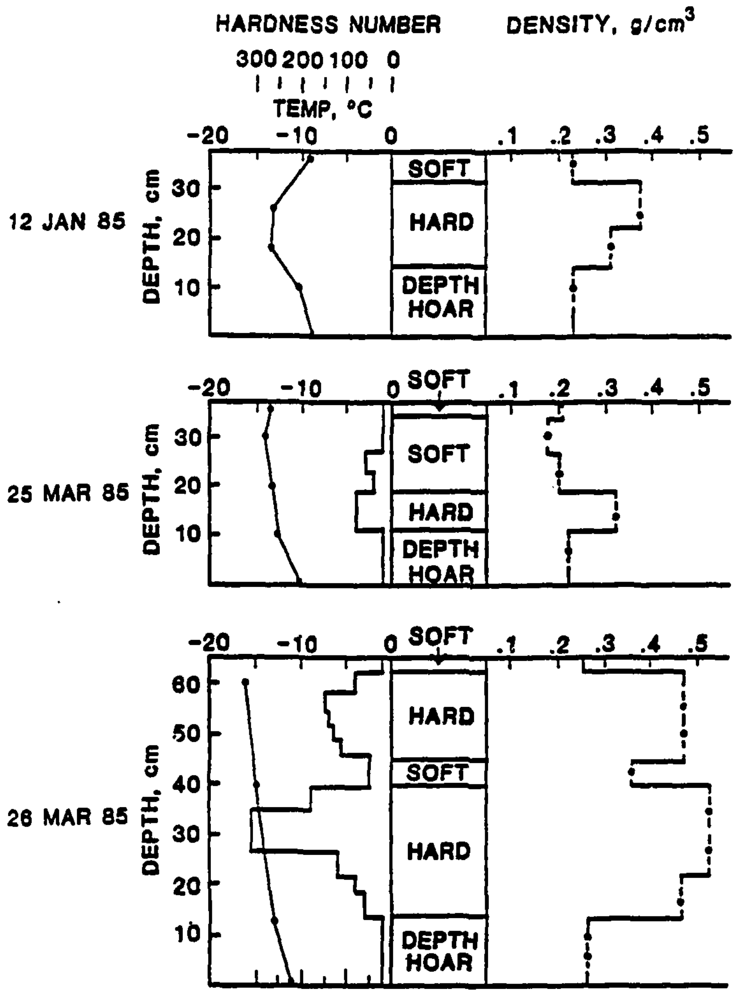

Figure 7-4. Snowpack properties on three dates during the winter of 1985.

the snowpack. Santeford (1978) and Woo (1982b) have both looked at the magnitude of upward vapor transport during the winter months from the active layer soils and related this flux to the development of depth hoar.

While this method of quantifying snow water equivalent is appropriate to get a measure of the maximum water content on the ground, it is not very useful as an indicator of the total water equivalent of the snowfall throughout the winter. A Wyoming snow gage is located within the watershed. Although it is an improvement over standard gages used, it has been shown that these gages do not perform well in windswept environments (Sturges, 1986). Sublimation from the snowpack and blowing snow occurs, and the data from the Wyoming shielded gage (Kane et al., 1989) for three years was essentially equal to the basin average as measured on the ground just prior to ablation. This implies that not all of the snowfall was collected and this corroborates the conclusions of Sturges (1986). The field data indicate that the amount of undercatch by the Wyoming gage is about equal to the amount of sublimation (Kane et al. 1989).

Each year, estimates from field measurements were made of the average water content of the snowpack in the basin: 10.2 cm for 1985, 10.9 for 1986, 10.6 for 1987, 7.8 for 1988 and 15.5 for 1989. Although the average water content was quite similar for the first three years in the basin, the distribution varied significantly because of the wind conditions each year. This is supported by the data in Figure 7-2.

2) Snow Ablation

The true onset of ablation each year for the five years of this study varied from early May until the end of May. Both in 1985 and 1987, some snow ablated in early May; however, these events produced no measurable runoff. In each case, cold air masses moving down from the North arrested this process. An examination of the albedo each year clearly shows the demise of the snowpack (Figure 7-5). The pattern of ablation was strongly influenced by the direction of the predominant winds. If the winds came from the North, air temperatures were much lower and incoming radiation was utilized for sensible heating of the atmosphere. When winds prevailed from the South, coming over the Brooks Range, air temperatures were much warmer and incoming radiation was responsible for ablation. Rouse (1984c) has shown that a similar process was observed for offshore and onshore winds at Churchill on Hudson Bay.

There was a general trend in the albedo each year from daily averages around 0.8 for snow covered ground to 0.2

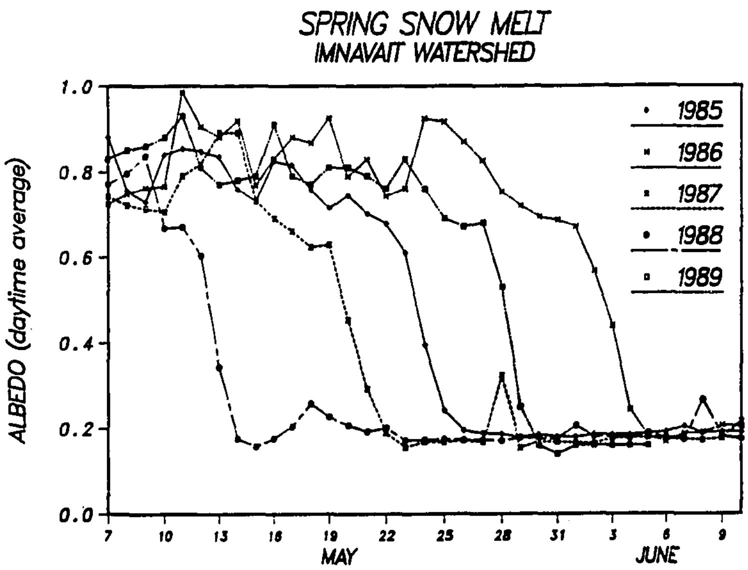

Figure 7-5. Average daytime albedo during the springmelt period of 1985, 1986, 1987, 1988 and 1989.

as the ground becomes snowfree (Figure 7-5). This rapid decline of the albedo occurred in a relatively short span of 3 to 4 days, closely followed by very significant surface heating: surface temperatures as high as 30° to 40°C were measured during mid day.

At the onset of ablation, the snowpack is generally several degrees below freezing at the base. This results in some refreezing of meltwater as it migrates down through the snowpack. The hydrologic significance of this has not been evaluated, but one study (Woo et al., 1982) found extensive basal ice in high arctic snowpacks.

The rate of snowmelt for the west facing slope is depicted in Figure 7-6 a - e for each year. Forty measurements of the water content of the snowpack were taken each day and daily averages were obtained. By taking the differences of the water content of the snowpack for successive days, the daily ablation was derived.

3) Soil Water Storage in the Active Layer

The first meltwater moving down through the snowpack encounters temperatures below freezing and refreezes. The released latent heat is very effective at warming the snowpack. This same process is repeated in the

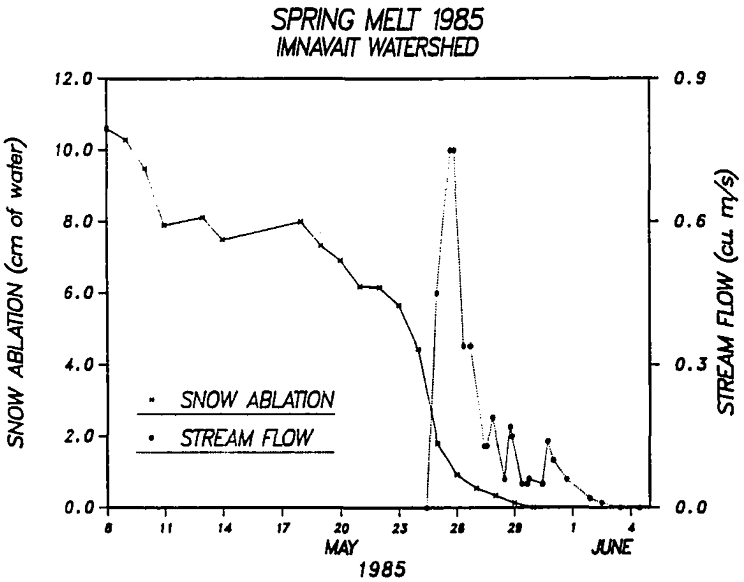

Figure 7-6a. Spring snowpack ablation and consequent stream flow in 1985.

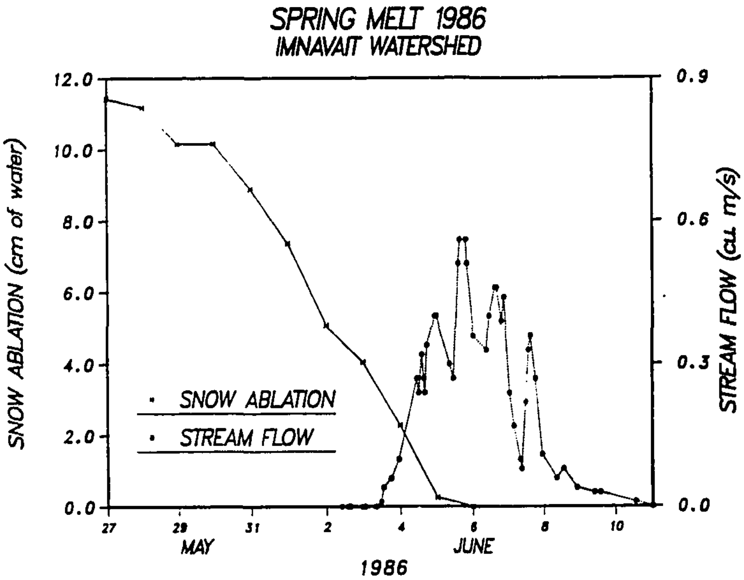

Figure 7-6b. Spring snowpack ablation and consequent stream flow in 1986.

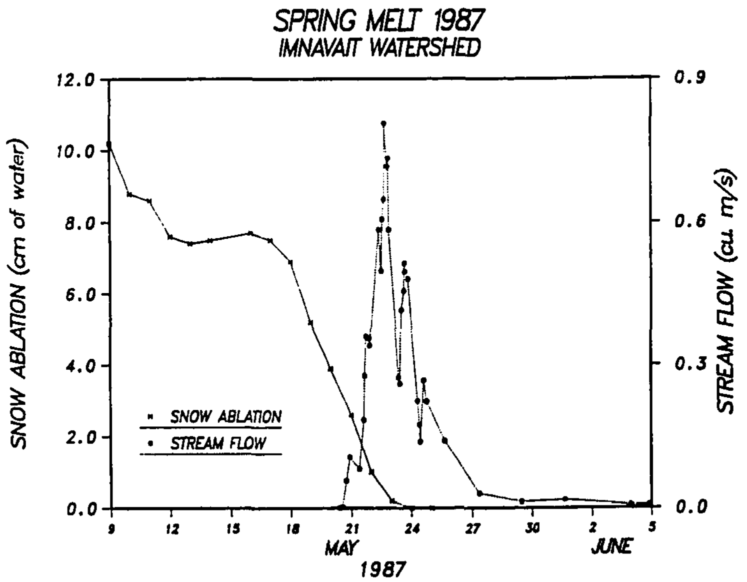

Figure 7-6c. Spring snowpack ablation and consequent stream flow in 1987.

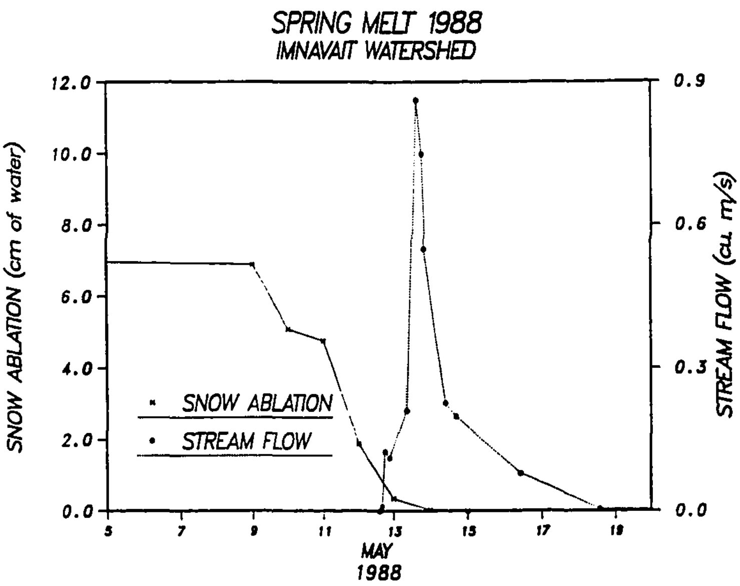

Figure 7-6d. Spring snowpack ablation and consequent stream flow in 1988.

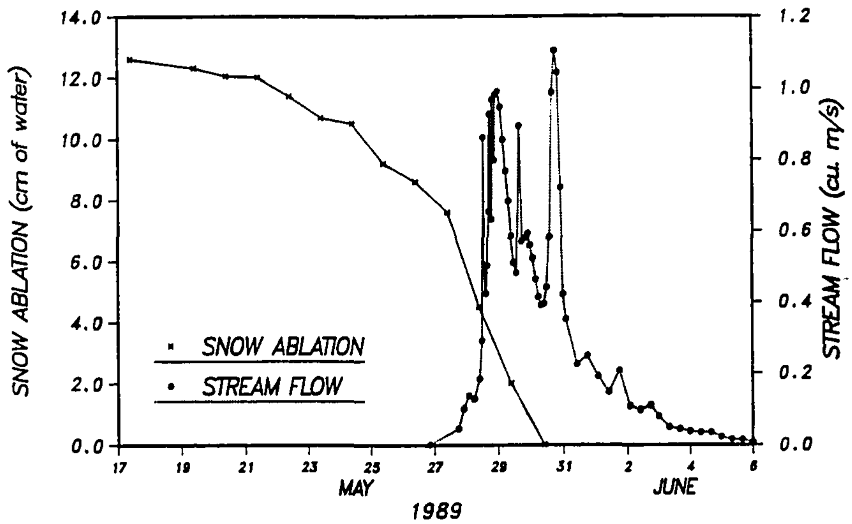

Figure 7-6e. Spring snowpack ablation and consequent stream flow in 1989.

upper organic soils in the active layer. These organic soils, with low bulk densities and large pores, are prone to desiccation over the winter months. However, the mineral soils directly beneath the organic soils are near saturation prior to freeze-up and develop numerous ice lenses during freezing. Ice-rich soils have relatively low infiltration rates (Kane and Stein, 1983a); consequently, there is very little infiltration into the mineral soils during ablation. This was confirmed with the time domain reflectometry measurements.

Both laboratory and field measurements have been made to ascertain the water holding characteristics of these soils. Characteristic curves have been developed for various depths and time domain reflectometry has been used to measure the unfrozen water content in the field (Hinzman et al., 1990). From these measurements, it has been concluded that about 1.5 cm of water goes into storage before runoff is produced, representing an average over the watershed. This has been supported by data collected during the summer. It has been observed during relatively dry periods following ablation that rainfall events of around 15 mm do not produce significant runoff (Kane et al., 1989).

4) Evaporation

Evaporation from the melting snow is a very difficult component of the hydrologic cycle to estimate for arctic watersheds. This paper does not consider this subject in detail; in the water balance equation this was the unknown term that was determined.

Theoretical and experimental work in Finland and Sweden (Lemmelä and Kuusisto, 1974; Bengtsson, 1980; and Kuusisto, 1984a) on evaporation from snow covers has shown that evaporation (sublimation) is very small during the winter and increases during snowmelt. They primarily conclude that during the winter there is not energy available for evaporation/sublimation and that during snowmelt, high relative humidity limits evaporation. In this case, the relative humidity is much lower during ablation. Warm southerly air that has flowed over the Brooks Range to reach this area is relatively dry. Also, water moving downslope over snow free areas encounters surfaces with rather low albedos and high surface temperatures that provide available energy for evaporation.

5) Runoff

Measurements of runoff were made both at the plot and watershed scales. The initiation of snowmelt runoff from the four plots each year is shown in Table 7-1 and for the watershed in Figure 7-6.

Table 7-1. Date of initial flow from runoff plots.

	1985	1986	1987	1988	1989
Plot 1	May 21	June 2	May 20		May 25
Plot 2	May 21	June 1	May 20	May 11	May 25
Plot 3	May 21	May 31	May 19	May 11	May 25
Plot 4	May 22	May 31	May 19	May 11	May 25

Runoff in the stream always commenced 2 to 3 days after the initial runoff from the west facing slope. In fact, there was no flow in any portion of the stream channel and then, in the matter of a few hours, the entire stream channel was flowing, with the initial flow originating near the headwater of the basin. Woo and Sauriol (1981) and Woo and Heron (1987) discuss the hydrologic role of snow in the drainage channel. These results agree that runoff, retarded by snow in the valley, can be stored as unfrozen water in the snowpack for several days. Once a channel has been eroded through the snowpack, all of the stored water is rapidly released. This is one reason why there is often a continuous rise to the peak outflow. Another reason is if the runoff is retarded for enough time, when stream runoff is initiated, it may correspond with the peak daily snowmelt. The runoff data from 1985 shows this quite clearly.

6) Water Balance

Partitioning the various components of the hydrologic cycle for the purpose of performing a water balance is not new, but the technique continues to be a very useful tool in hydrologic analysis. New approaches in water balance computations are covered in depth in a recent compilation of papers (Van der Beken and Herrmann, 1985).

For both the 90 m^2 study plots and Imnavait watershed as a whole, water balances (Table 7-2) were carried out during the snowmelt period each year. The water balance equation presented earlier was used to estimate the evaporation from the watershed. The only storage term included in the water balance was the amount of water going into storage in the upper 10 to 15 cm of the active layer. There was some surface storage, but this was minimal in this basin and was neglected. The role of storage in the snowpack in the valley bottom has been discussed, but since this balance is carried out from the initial snowmelt through the end of snowmelt runoff in the stream, it does not influence the results of this calculation. It does, however, influence the shape of the resulting runoff hydrograph.

The quality of the data in Table 7-2 is variable. There

Table 7-2. Partitioning of the melting snowpack.

	Initial Water Content, cm	Runoff %	Evaporation %	Soil Storage %
		1985		
Plot 1	13.8	79	10	11
Plot 2	11.7	38	49	13
Plot 3	9.7	22	63	15
Plot 4	10.6	31	55	14
Plot Average	11.4	45	42	13
Basin Average	10.2	65	20	15
		1986		
Plot 1	14.7	68	22	10
Plot 2	12.4	50	38	12
Plot 3	8.1	29	53	18
Plot 4	10.2	50	35	15
Plot Average	11.3	52	35	13
Basin Average	10.9	52	34	14
		1987		
Plot 1	9.8	69	16	15
Plot 2	10.7	33	53	14
Plot 3	10.2	*	*	*
Plot 4	10.0	20	65	15
Plot Average	10.2	41	44	15
Basin Average	10.8	66	20	14
		1988		
Plot 1	7.2	+	+	+
Plot 2	7.3	16	63	21
Plot 3	8.4	43	39	18
Plot 4	6.9	33	45	22
Plot Average	7.4	31	49	20
Basin Average	7.8	50	31	19
		1989		
Plot 1	13.5	*	*	*
Plot 2	11.6	34	53	13
Plot 3	11.7	34	53	13
Plot 4	13.7	*	*	*
Plot Average	12.6	34	53	13
Basin Average	15.5	61	30	10

* No measurements made for this plot because of leakage.

\+ No measurements collected

is considerable variation in the snowcover, both because of the wind and the surface relief of the tussock tundra. This makes it difficult to make good estimates of the snowpack on the plots. Also, the measurements are made adjacent to the plots, not directly on them. In 1987, some early melt had taken place prior to the first snowpack measurements on May 9, therefore, the estimates of the snowpack water contents may be a little low.

Runoff data were measured very accurately by the collection system used. The only problem was when water flowed over or under the barrier of the plot. This occurred in 1987 for plot 3 and 1989 for plots 1 and 4. The spring melt of 1988 was so intense and fast, time did not allow all four plots to be monitored so plot 1 was neglected. The estimates of soil storage are based on laboratory measurements of the amount of water that can be retained by the upper 10 cm of organic soil and measurements of unfrozen soil moisture levels with the time domain reflectometer.

Evaporation on the plot scale can be seen to be very important. Where the snowpack is light such as windswept areas, more vegetation protrudes through the snow and a larger percentage of the ablated snow is lost to evaporation. On the watershed scale, evaporation is still important but not as dominant. In 1985, 1987 and 1988, 2.0, 2.2 and 2.4 cm of water respectively were lost due to evaporation. In 1986, the evaporation was 3.7 cm; during this year the melt period was quite late. In 1989, 4.6 cm were lost as evaporation. The snowmelt was quite late this year and also had the highest snowpack water content and required the longest time period of active melt.

Numerous plot and watershed studies have been carried out in arctic regions to develop an understanding of various hydrologic processes. Lewkowicz and French (1982) examined snowmelt runoff from four plots of variable size in an area of continuous permafrost on Banks Island in northern Canada. They found considerable variation in surface and subsurface flow from these plots that they attributed to snow distribution and meteorological conditions at the time of melt. They observed that those plots with the greatest snow accumulations also had the highest percentage of runoff. This conclusion can be reached from the data in Table 7-2. The main physical reason for this was evaporation can only proceed at a certain rate that is controlled by meteorological parameters and there is a limited amount of subsurface storage in the active layer.

Woo (1982a) points out that the energy balance dictates the availability of energy for the ripening and melting of the snowpack, but that the coldness of the snowpack and the soil active layer causes considerable delay in the hydrologic response of runoff. Calculating the water balance over the time period did not cause any problems in this regard; however, in a model where you are interested in predicting the temporal distribution, it is very important (Hinzman and Kane, 1990).

There have been similar water balance studies carried out in the Canadian High Arctic (Woo, 1983, Marsh and Woo, 1979), but the physical systems were substantially different. Their studies were done in an area where snowfall constitutes 80 to 85% of the total precipitation, snowmelt starts later and lasts much longer, active layer storage is more severely restricted than at our site, soils are coarse sands and gravel, and vegetation is very sparse.

G) <u>Conclusions</u>

The three major factors in determining the quantity of snowmelt runoff are the quantity and distribution of the snowpack, the soil condition, and the meteorological conditions at the time of ablation. Redistribution by the wind of the shallow arctic snowpack results in above average water contents in the valley bottoms, water tracks, and the lee side of ridges. This guarantees a high runoff volume and uneven soil moisture levels throughout the watershed. Snowmelt and subsequent runoff in the Arctic is the most dynamic hydrologic event every year and it ensures that soil moisture will be available for immediate use by vegetation.

The onset of ablation is controlled by direction of air flow. The southerly flow of warm air complements the incoming solar radiation causing a rapid snowmelt event. A shift in the wind direction so that cold air flows off the Arctic Ocean will stop the ablation process, particularly early in the melt when the albedo is still high.

For the watershed snowpack, runoff varied from 50 to 66%, evaporation varied from 20 to 34%, and storage increases constituted 10 to 19%. Considerable variation existed spatially as evidenced by the results of the runoff plots.

H) <u>Acknowledgments</u>

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VIII) MODELING THE SNOWMELT HYDROLOGY OF A HEADWATER ARCTIC BASIN

Hinzman, L.D. and D.L. Kane. 1990. Modeling the snowmelt hydrology of a headwater Arctic basin. Submitted to Water Resources Research.

A) Abstract

Lack of hydrologic data in the Arctic, particularly during snowmelt, severely limits modeling strategy. The snowmelt hydrology of Imnavait watershed, a small headwater basin on the North Slope of Alaska, was studied with particular reference to improving modeling capability. Spring snowmelt in Imnavait watershed is a very brief event, usually lasting about 10 days. Peak flow normally occurs within 36 hours of the beginning of stream flow. All downslope movement of water occurs within the top 10 cm of the highly organic soil mat. Snow damming of snowmelt runoff and non-uniform snow distribution are important processes which must be considered in the modeling process of small watersheds. These unique characteristics of arctic hydrology affect the methodology and performance of the model. HBV, a hydrologic model, was used in an investigation of the hydrologic regime during the spring snowmelt period. The analysis of five spring melt events has shown that HBV can adequately predict soil moisture, evaporation, snow ablation and accumulation, and runoff. It models the volumes of snowmelt runoff well, but more data are

needed to improve the prediction of snowmelt initiation. Use of HBV as a predictive tool is dependent upon the quality of the meteorologic forecast data.

B) Introduction

The arctic ecosystem is a delicate and fragile environment where even slight disturbances can upset the thermal regime, resulting in thermal degradation to the extent it will remain visible for years. As part of an ongoing research project to determine the "Response, Resistance, Resilience and Recovery from Disturbance in Arctic Ecosystems" this project was initiated to quantitatively describe the hydrology of an arctic watershed. One objective of this research was to determine if the hydrologic processes which occur during the spring snowmelt could be adequately simulated with a computer model in spite of the extreme variability that exists in the Arctic in terms of soil properties, snow distribution and snowmelt processes. To aid in predicting the hydrologic response of the system, the Swedish HBV model was selected to simulate the hydrology of the Imnavait watershed. This model was originally developed by Bergström (1976) of the Swedish Meteorological and Hydrological Institute.

HBV has been used effectively to model streamflow in Sweden, Norway (Bergström, 1976), Finland (Vehviläinen and Motovilov, 1988), Columbia (Häggström et al., 1988) and in Alaska (Sand and Kane, 1986), but has never been used to model the hydrology of a basin completely underlain by permafrost. The ultimate goal of this application is hydrologic forecasting for basins with no or very limited data and for basins where surface disturbances have been imposed, however that has not yet been attempted. Such predictions are quite difficult because they not only require accurate models, but also models whose components can be related to the characteristics of the catchment. Additionally, the effect of the disturbance upon the hydrologic processes must be understood and correlated with confidence to the parameters of the model. The purpose of this research is to identify and properly incorporate the distinctive characteristics of arctic hydrology into the modeling process.

An arctic watershed at 68°30′ north latitude and 149°15′ west longitude was selected for this study (Figure 3-1). It is a small basin at the headwaters of Imnavait Creek and is approximately 2.2 km^2. The watershed is completely underlain by permafrost, the thickness is believed to be between 250 and 300 m (Osterkamp and Payne, 1981). The active layer, the depth of soil which thaws each summer, is between 25 and 60 cm. The active layer depth depends upon surface conditions and drainage characteristics. The soils are a complex of Pergelic and Histic Pergelic Cryaquepts consisting of 10 to greater than 20 cm of organic material overlying glacial drift. The vegetation is primarily sedge tussocks, mosses and low shrubs.

C) Data Quality

The quality and quantity of hydrologic data available for analysis varies greatly across Alaska. In northern Alaska, there is a sparsity of short-term data, and essentially no long-term data. Measurements of air temperature and precipitation tend to be collected near the only road bisecting the North Slope of Alaska or near the coast. These few stations have relatively short periods of record and the quality of the precipitation data is usually poor. Although the North Slope occupies an area of 210,000 km^2 (14% of Alaska) there are only 8 U.S. Weather Service stations, 7 of which are located along the coast, and 5 permanent U.S. Geologic Survey stream gaging stations. In comparison to most stations in the U.S., these sites have only a very brief record. Recent oil development related activities in the North has led to increased interest in understanding arctic climatology and permafrost hydrology and several research projects have improved upon our knowledge of these processes and made valuable contributions to the existing data base.

The amount of precipitation recorded in the Arctic is usually underestimated due to problems with the instrumentation. Standard gages operated by the U.S. Weather Service are not shielded from the wind and therefore undercatch the total precipitation by varying amounts depending upon windspeed. This undercatch is particularly high for snowfall which amounts to 1/3 to 1/2 of the total yearly precipitation. Precipitation can occur as snowfall in any month. During windy conditions, the water content of snowfall is typically underestimated by a factor of 2 to 3 (Benson, 1982). The U.S. Soil Conservation Service maintains a series of Wyoming snow gages across the North Slope. These gages are a substantial improvement over the standard unshielded gage, but still suffer from some undercatch. A related study (Kane et al., 1989) for three years has shown that the total amount of undercatch of winter snowfall was about equal to the total amount of moisture lost as sublimation from the snowpack. Consequently, the total snowfall measured by the gage compared quite well with the water content in the spring snowpack on the ground. Undercatch of the gage and sublimation are both directly related to the windspeed, however undercatch occurs during precipitation events;

sublimation occurs primarily between these precipitation events and throughout the entire winter. It is probably a coincidence that these independent processes approach the same value after 8 or 9 months of accumulation.

Although continuous recording stations are sparse, measurements of runoff are probably quite reliable during the summer months. Accuracy falls in early spring when high flows are accompanied by snow and ice along the channel sides and bottom and ice jams with subsequent backwater effect. Most streams and rivers in the Arctic are reported to have zero flow in late winter. However some rivers have continuous base flow from groundwater sources throughout the year (Sloan et al., 1975). Huge deposits of aufeis can develop at these sites. Such flows are very difficult to measure and consequently are not recorded as winter flow, but are included as melt the following summer. No aufeis developed in Imnavait watershed. The amount of runoff that will be generated from snowmelt is dependent not only upon the snowpack water content but also upon the distribution of snow in relation to the stream (Kane and Hinzman, 1988).

Evaporation is a difficult variable to measure in any climate. A related study (Kane et al., 1990a) has shown that the evaporation during the spring ablation can vary from 20% to 34% of the total snowpack water content for this watershed. Benson (1969) reported that evaporative loss in mid-winter is minimal, however loss to sublimation by strong winds can be significant. He also stated under some conditions during snowmelt, when evaporation is enhanced by 24 hour daylight and vegetation protruding through the surface, evaporation rates may approach that of a free water surface. Weller and Holmgren (1974) calculated evaporation through an analysis of the energy balance. They estimated during the snowmelt period mean rates of evaporation are 4.5 mm/day. The processes associated with evaporation appear to be highly variable with season and location (Lemmelä and Kuusisto, 1974; Bengtsson, 1980). It is generally agreed that relatively little evaporation occurs during mid-winter but can be quite high during spring melt.

In this study, the accumulation of the snowpack was monitored by periodic snow surveys throughout the winter and continuous measurements from the Wyoming snow gage. The water content in the snowpack prior to spring melt was determined through intense snow surveys across the entire watershed. Snowpack ablation was observed by daily snow surveys. Precipitation was measured by a Wyoming snow gage and during the summer by two shielded tipping bucket rain gages, however no significant amount of rain fell during any of the spring melt events. Two complete meteorologic stations measured the important parameters of air, snow and soil temperatures, relative humidity, wind speed and direction, and the components of longwave, shortwave and net radiation. All instrumentation was connected to Campbell Scientific 21X data loggers.

Streamflow was measured with a Montedoro Whitney electromagnetic current meter. The stream was gaged frequently in the spring when snow and ice upset the stage-discharge relationship and occasionally throughout the summer. A flume was installed to monitor the flow in the stream channel and a Stevens F type water level recorder measured the stage from which the discharge was computed.

Important physical properties of the soil profile such as hydraulic conductivity, bulk density, porosity, and thermal conductivity were measured in the laboratory (Hinzman et al., 1990). The highly variable nature of the active layer, both spatially and temporally, significantly affects the surficial hydrology (Kane and Hinzman, 1988; Hinzman et al., 1990). A complete analysis of the thermal regime of Imnavait watershed has also been completed (Kane et al., 1990b). Analysis of the spring melt hydrologic regime is presented in Chapter VII (Kane et al., 1990a) and a general analysis of the annual hydrologic regime is presented in Chapter V (Kane et al., 1989).

D) Model Structure

HBV is a conceptual runoff model designed for continuous calculation of runoff using a lumped parameter representation (Bergström, 1976). After calibration the model is designed to require little meteorologic input. The model has 41 parameters of which 26 are used to describe the basin or input data while the rest collectively describe snow accumulation and ablation, soil moisture accounting, and generation or transformation of the hydrograph. Several of these parameters are empirical values which are determined in the calibration procedure and can not be determined through analysis of the physical system. Input data are observations of air temperature, precipitation, and monthly estimates of evapotranspiration. The time steps are usually one day, however 5 spring melt events have been modeled utilizing two hour increments with good results. Air temperature is used to calculate snow ablation or accumulation and is necessary for this application.

The model will calculate the runoff component for several basins and add the contribution of each utilizing a damping parameter or the Muskingum routing method. The basins can also be subdivided into elevation zones and each elevation zone can be further sub-divided into vegetation zones. Each subbasin may have unique values of many parameters such as percolation, soil moisture limits, and recession coefficients. This watershed is treated as a single basin with only one elevation zone since there is only a 90 m range in elevation.

The model can be best understood by examining each of its important components individually (Figure 8-1). The separate sections include routines for snowmelt, soil moisture accounting, control of runoff response, and the transformation function.

The routine which calculates snowmelt and refreezing is based upon a simple degree-day method. A threshold temperature which is usually close to 0°C defines the temperature at which snow will melt. This threshold temperature also determines if the precipitation will be treated as rain or snow and thus snow accumulation can be monitored. During spring melt, the threshold temperature functioned to control the initiation of simulated snowmelt runoff. The degree day procedure is a temperature index method which does not consider the

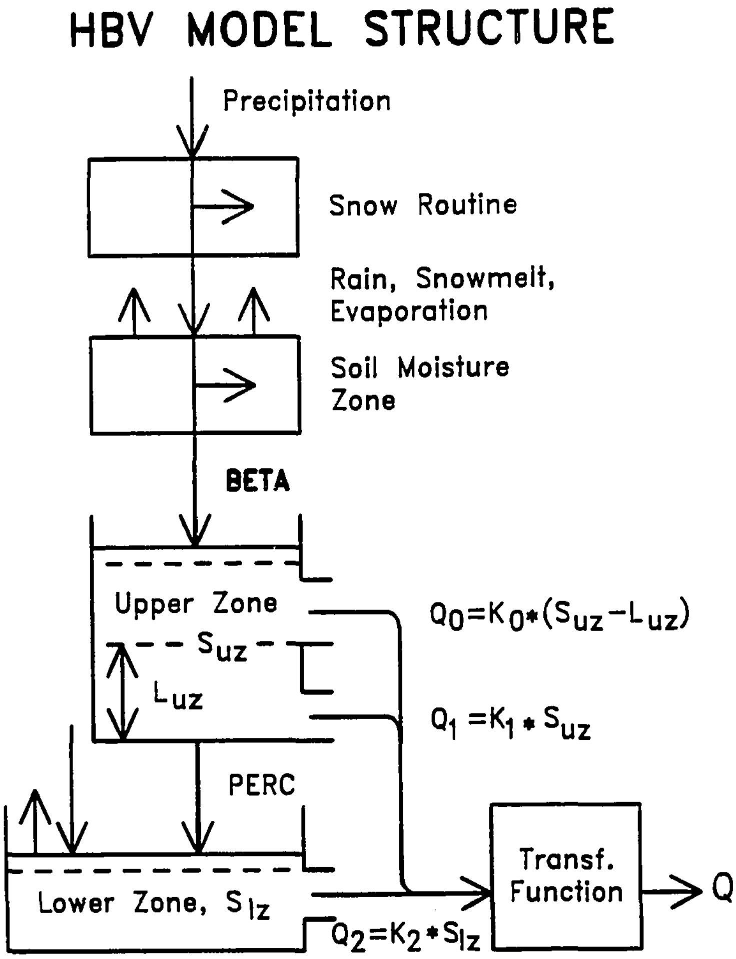

Figure 8-1. Conceptual flowchart of HBV runoff model.

effects of radiation or convective heat transfer directly. Snowmelt is calculated through the formulation:

$$M = CFMAX * (T - TT) \qquad (1)$$

where
M = snowmelt, mm
CFMAX = degree day factor, mm/°C
T = mean daily temperature, °C
TT = Threshold Temperature, °C

This equation is functional over the particular time step. CFMAX has been called the degree day factor and is simply the amount of melting which occurs per degree of positive air temperature within one time step, i.e. one degree day or one degree hour. CFMAX was determined through parameter optimization. TT is the factor in the equation which specifies at what temperature snow will begin to melt. TT is also called the critical temperature or equilibrium temperature. When the air temperature is equal to TT, the net heat exchange between the air and the snow surface is assumed to be zero (Kuusisto, 1984b).

Water from precipitation or snowmelt is partitioned into evaporation and infiltration in the soil moisture routine (Figure 8-1). The soil moisture routine is the main component controlling the generation of runoff. This routine is based upon the storage capacity of water in the soil, evapotranspiration from the soil, and a response function describing the amount of runoff generated from precipitation. The routine will have the effect that the contribution to runoff from rain or snowmelt is small when the soil is dry and greater for wet conditions. The predicted evapotranspiration also decreases as the soil dries.

The rapid and slow characteristics of near surface runoff are created through an upper and a lower reservoir. Yield of soil water from the soil moisture routine will be added to storage in the upper zone. The upper zone in this application is meant to correspond to the highly porous organic soil near the surface (0-10 cm). The parameter L_{uz} (see Figure 8-1) corresponds to the less porous, but still highly organic soil below the surface (10-20 cm). The lower zone reservoir represents the subsurface mineral soils and also can include the effects of rain and evaporation from lakes. Conceptually, some water from the upper zone will percolate to the lower zone. The response function is a runoff generation routine which transforms excess water from the soil moisture zone to runoff. When yield from the soil routine exceeds percolation, the soil moisture level will rise and discharge will be generated from the upper zone. This represents drainage through more superficial channels, giving rise to a slower process of runoff. At a storage level in the upper zone exceeding the upper zone limit, more rapid drainage will occur, thus the rapid component. The lower zone represents the groundwater storage of the catchment contributing to base flow. As mentioned previously, in this watershed groundwater base flow was not significant and did not enter into the modeling process during snowmelt.

HBV has the capability for hydrologic forecasting. After calibration, the model must be run up to the first day of the forecast period. The model can make short-range predictions based upon meteorologic forecasts or long-range predictions utilizing meteorologic data for corresponding dates from the previous years available (usually 10 to 20 years). Long-range predictions are not possible for this application as the long-term data does not exist. Hydrologic forecasting of the basin over the short-term is quite easy, but is only as good as the meteorologic predictions will allow. On the North Slope of Alaska, the U.S. Weather Service has less than one meteorologic observation station per 20,000 km^2, therefore such predictions are usually of dubious value.

The HBV model was compared to similar hydrologic models by the World Meteorological Organization (1986) and performed quite well in spite of its simple structure and limited data requirements. After a comparison of other models designed for continuous streamflow simulation (Singh, 1988b; Renard et al., 1982), it is apparent that HBV is particularly well suited for analysis of watersheds in the High Arctic. Unlike many models, HBV can consider snow accumulation and ablation on any day depending upon current weather conditions. It does not consider interception or reservoir routing, which are usually not important in arctic hydrology. The components of arctic hydrology which are most important, i.e. snowmelt, soil moisture accounting, surficial runoff and channel routing, receive paramount consideration in HBV. Many of the parameters describing the watershed can be determined through parameter optimization and thus the model will be usable in arctic watersheds which normally lack any data regarding hydrologic characteristics. The reservoir type structure of the soil zones is very well suited to the strongly stratified, shallow soil layers commonly found in the Arctic. The most significant advantage of HBV is due to the method of actuating parameters. Each parameter is input with an effective date, thus those parameters which change throughout the year can be logically varied without stopping the simulation. Parameters which vary with time include those affecting soil moisture storage which are dependent upon the depth of thaw. The values used for each parameter are included in Table 8-1.

Table 8-1. Parameters used in HBV model for springmelt.

			BASIN PARAMETERS
AREA	May 1	2.20	Area of subbasin (km^2)
ELEV	"	910.00	Elevation of each subbasin (m)
VEG	"	1.00	Proportion of subbasin of vegetation type
LAKE	"	0.00	Proportion of subbasin covered by lake
PELEV	"	930.00	Elevation of precipitation station
TELEV	"	930.00	Elevation of temperature station
QFACT	"	1.00	Conversion factor of input data to m^3/s
CP	"	1.00	Station weight for precipitation (areal)
CT	"	1.00	Station weight for temperature (areal)
CQ	"	1.00	Station weight for discharge (temporal)
EVAP	"	4.50	Potential evapotranspiration (mm/day)
PCORR	"	1.00	Correction factor applied to precipitation
PCALT	"	10.00	Increase in precipitation per 100 m (%)
TCALT	"	10.00	Temperature Lapse rate per 100 m (°C)
CEVP	"	1.00	Correction factor applied to EVAP values
LAREA	"	0.00	Area of lake at outlet of subbasin (km^2)
ICEF	"	135.00	Last day with ice on lakes
EVDAY	Not Used		Days for which EVAP data are valid
REPL	"		Station to replace missing data
ELAG	"		Evaporation lag time (days)
QADD	"		Additional flow from another subbasin
QRED	"		Reduction of computed discharge
PATH	"		Routing from subbasin to subbasin
WQREL	"		Stage/discharge relation number
FELEV	"		Elevation of forecasted data
SUBQR	"		Indicates which discharge station/subbasin
			SNOW ROUTINE PARAMETERS
SFCF	May 1	1.00	Correction factor applied to snowfall
CFMAX	"	3.50	Melting factor (mm/°C)
CFR	"	0.08	Refreezing factor
TT	"	-1.9 to 0.10	Threshold temperature, °C (see text)
			SOIL ROUTINE PARAMETERS
FC	May 1	40.00	Field Capacity (mm)
LP	"	7.00	Limit for potential evapotranspiration (mm)
BETA	"	2.00	Parameter in soil routine (see Figure 8-1)
			RESPONSE FUNCTION PARAMETERS
PERC	May 1	0.00	Percolation from upper to lower zone.
UZL	"	15.00	Limit for high recession, K0, (mm)
K0	"	0.90	High recession coefficient in upper zone
K1	"	0.30	Lower recession coefficient in upper zone
K2	"	0.00	Recession in lower zone
			TRANSFORMATION FUNCTION PARAMETERS
MAXBAS	May 1	1.00	Base in transformation function (days)
BLAG	Not Used		Translation of computed hydrograph
DAMP	"		Damping parameter in routing routine

The model is calibrated by trial and error. Five to ten years of observed daily discharge are preferred for calibration, but this much data are quite rare for arctic watersheds. The calibration technique generally relies upon three criteria; visual inspection of simulated and observed hydrographs, a continuous plot of the accumulated difference between simulated and observed hydrographs, and the explained variance, r^2. The validity of using the explained variance, or Nash-Sutcliffe coefficient, as a comparative index of performance of the model was discussed by Martinec and Rango (1989). The explained variance is calculated as:

$$r^2 = 1 - \frac{\sum_{t=1}^{n} [Q_{sim}(t) - Q_{obs}(t)]^2}{\sum_{t=1}^{n} [Q_{obs}(t) - \overline{Q_{obs}}]^2} \qquad (2)$$

Q_{sim} = simulated discharge (m^3/s)
Q_{obs} = observed discharge (m^3/s)
t = time variable (days or hours)
n = number of time steps
$\overline{Q_{obs}} = \frac{1}{n} \sum_{t=1}^{n} Q_{obs}(t)$ (m^3/s)

In addition to the above criteria, plots of computed evapotranspiration, snowpack accumulation and ablation and soil moisture aid in calibration.

E) Results

Data collected over a 5 year period were used in the analysis. The spring snowmelt events were analyzed using 2 hour and 1 day time steps for 1985 to 1989.

From the hydrologic viewpoint, each year was quite different presenting a wide range in conditions and processes, thus offering a good test of the flexibility of the model. The character of the snowmelt events is displayed in a plot of snowpack ablation (Figure 8-2). There is considerable range in initial snowpack water contents, timing of snowmelt initiation, and snow distribution throughout the watershed. The total daily melt is determined from daily snow surveys (Figure 8-3). Snow is redistributed by strong winds throughout the winter. It is normally deposited in the valley bottom, in small water tracks, and on the lee side of slopes. The amount of redistribution varies from year to year depending upon the number and magnitude of the wind events. Snow deposited in the valley bottom retards streamflow for several days after the onset of spring melt depending upon snowpack depth and density (Kane et al., 1989). This snow damming is an important process which must be considered in stream flow modeling in the Arctic.

When the daily measured snowmelt was compared with the average daily air temperature, it was apparent that much of melt was due to other factors such as radiation and convection, and could not be explained by a simple degree day formulation (Figure 8-3). These measurements

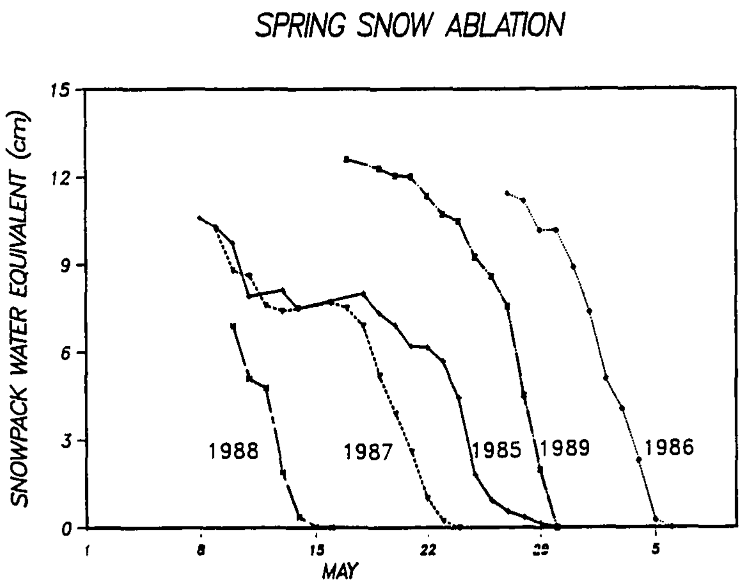

Figure 8-2. Process of snowpack ablation for five spring melt events.

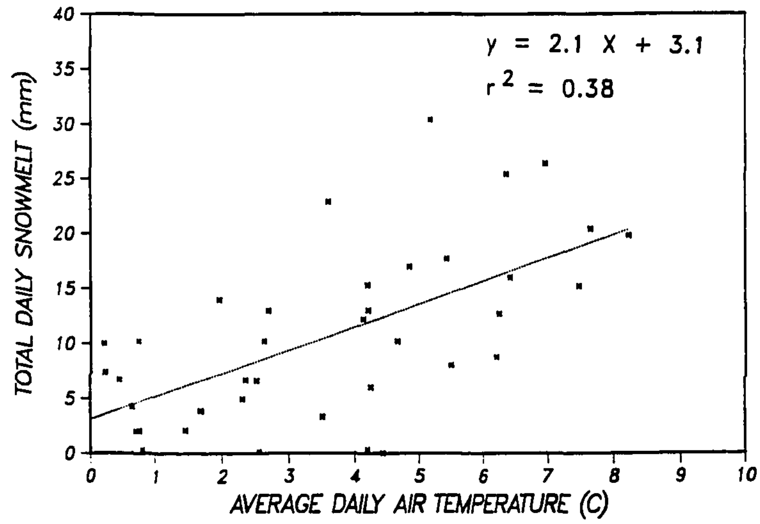

Figure 8-3. Relation of daily snowmelt and average daily air temperature.

of snowmelt reveal that CFMAX was 2.1 mm/°C and TT was 3.1°C (CFMAX was the slope and TT was the Y intercept of the line of best fit). A sensitivity analysis determined that 3.5 mm/°C was consistently the best value given to CFMAX in all simulations in which all parameters were optimized to give the best prediction of discharge. The value of 2.1 was determined through the classic degree day definition (Figure 8-3), that is the amount of melt which occurs per one degree of positive air temperature, while the value of 3.5 was actually determined from the amount of water yield per degree day.

TT was different for each spring melt event, and was different in simulations where the parameters were optimized to fit the observed snowmelt to computed snowmelt as compared to optimizations of discharge and was also different for different time steps, i.e. 1 day or 2 hour (Table 8-2). The obvious importance of TT in the model performance and uncertainty in its independent determination remain the biggest problem in using HBV for snowmelt runoff studies. A sensitivity analysis has shown that TT is one of the most important parameters affecting snowmelt simulation. The importance of TT on the performance of the model is particularly obvious in a plot of the r^2 as a function of TT (Figure 8-4). Values for the degree day factor and the threshold temperature can be estimated for a given region, but must be checked for reliability.

Table 8-2. Performance of model and selected threshold temperature.

Year	Time Step	Threshold Temperature (°C)	r^2	Accumulated Difference (mm)
1985	2 hours	0.30	0.86	-7.1
1986	"	-0.70	0.81	12.7
1987	"	-0.40	0.89	-2.6
1988	"	-1.90	0.82	-0.4
1989	"	-0.30	0.80	13.2
1985	1 day	0.25	0.81	-7.2
1986	"	3.30	0.77	11.7
1987	"	0.20	0.93	7.7
1988	"	-1.80	0.91	-0.2
1989	"	-0.35	0.77	13.6

TT performs two primary functions which are physically related, but still independent. TT determines the minimum air temperature at which snow will begin to melt. It also controls the timing at which snowmelt runoff will begin to flow. As discussed in Kane et al. (1990a), approximately 80% of the snowpack will ablate before runoff begins due to snow damming in the valley bottom. Consequently, TT is adjusted to improve one mechanism at the expense of another so the optimum TT for discharge calculations will not be the same as the optimum TT for snowmelt calculations. The physics of snowmelt and snowmelt runoff have been studied intensely at Imnavait watershed (Liston, 1986; Kane and Hinzman,

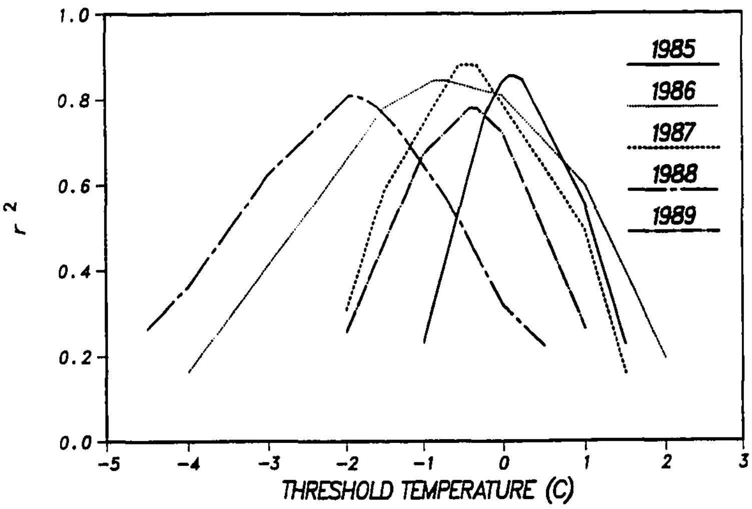

Figure 8-4. Performance of model as a function of threshold temperature for five simulations using two hour time steps.

1988; Everett and Ostendorf, 1988; Kane et al., 1989) and have been shown to be a complicated interaction of many mechanisms. The degree day representation is not an entirely satisfactory emulation of reality, however the simplicity of the degree day method is preferred in arctic watersheds where more complete data sets are not available.

The initiation of snowmelt on the North Slope is strongly dependent upon the presence of convective air masses transported to the north over the Brooks Range (Kane et al., 1990a). Longwave emittance from low clouds and fog accelerates melt. Shortwave radiation is very near the seasonal maximum during spring melt, and although important does not vary greatly year to year. Also important is the energy required to warm the snowpack and surface organic layer to isothermal conditions prior to melt. This process could be included in the model by incorporating an accounting of the surface energy balance (Motoyama, 1986). An improvement to the model would be to change the snowmelt routine to include a choice of the degree day method or a more physically based energy balance method such as that presented by Vehviläinen (1989). The effect of model complexity on snowmelt calculations is discussed by Bergström (1986).

The results of modeling each snowmelt event were included because close examination of each individual event reveals many of the strengths and limitations of the HBV model. In 1985, snowmelt began early but low temperatures stopped the melt after approximately ¼ of the snowpack had melted (Figures 8-2 and 8-5). After 3 days of cold weather, the snowmelt event resumed again. Although significant wind events had produced very substantial wind slabs, the snow was distributed more evenly across the watershed when compared to later years. The year 1986 was marked by a late spring melt; temperatures during snowmelt were quite high and incoming solar radiation was very near the seasonal maxima (Figure 8-6). Extensive snow redistribution left large quantities of high density snow in the valley which retained much snowmelt water before runoff began. In 1987, spring snowmelt was much earlier, spring temperatures were lower (Figure 8-7). The spring snow distribution of 1987 was quite similar to that of 1985 yielding confirmation to idea that the density of snow in the valley bottom controls stream damming. The 1988 snowmelt event was indeed unusual, truly testing the capability of the model (Figure 8-8). The snowpack was very thin throughout the watershed. The air temperature became very warm in early May and the thin snowpack ripened quickly. The entire snowmelt event lasted only

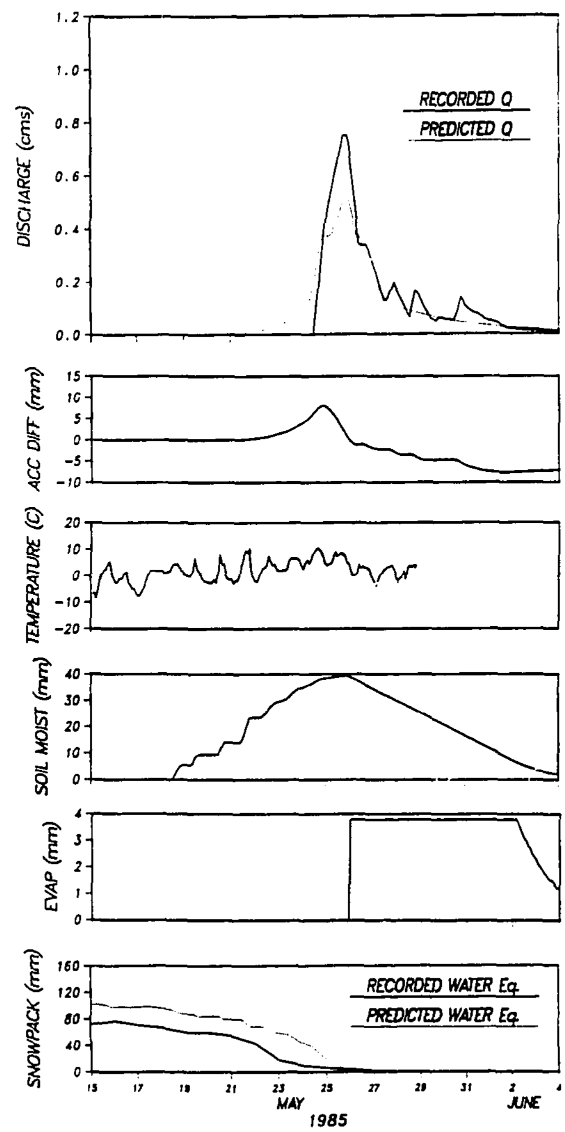

Figure 8-5. Observed and simulated hydrologic processes at Imnavait watershed during the 1985 spring melt event, accumulated difference between observed and simulated flow, measured air temperature, predicted soil moisture, evapotranspiration, and measured and predicted snowpack ablation.

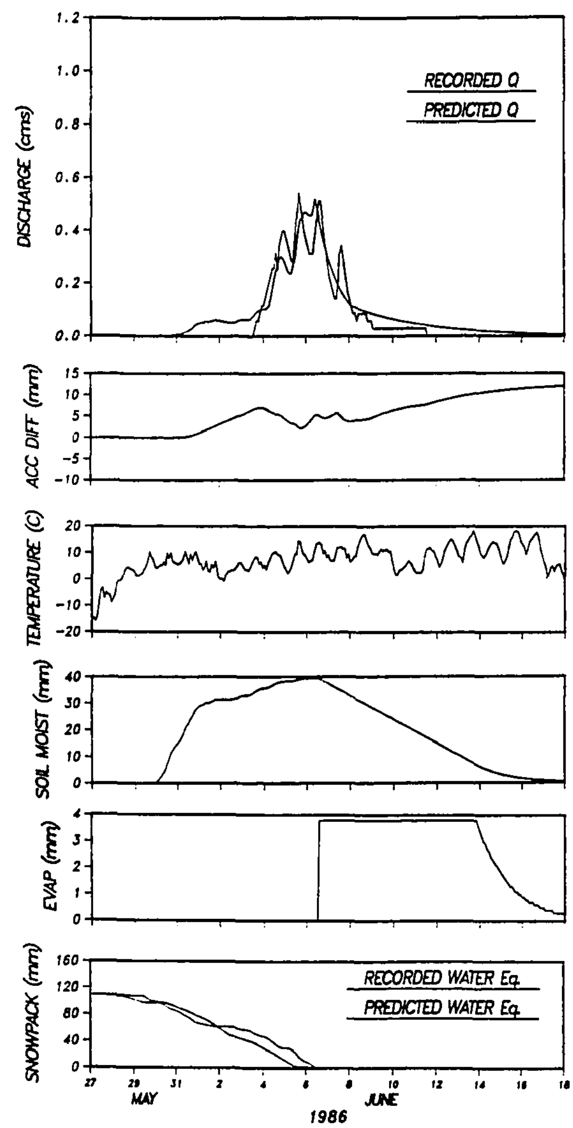

Figure 8-6. Observed and simulated hydrologic processes at Imnavait watershed during the 1986 spring melt event, accumulated difference between observed and simulated flow, measured air temperature, predicted soil moisture, evapotranspiration, and measured and predicted snowpack ablation.

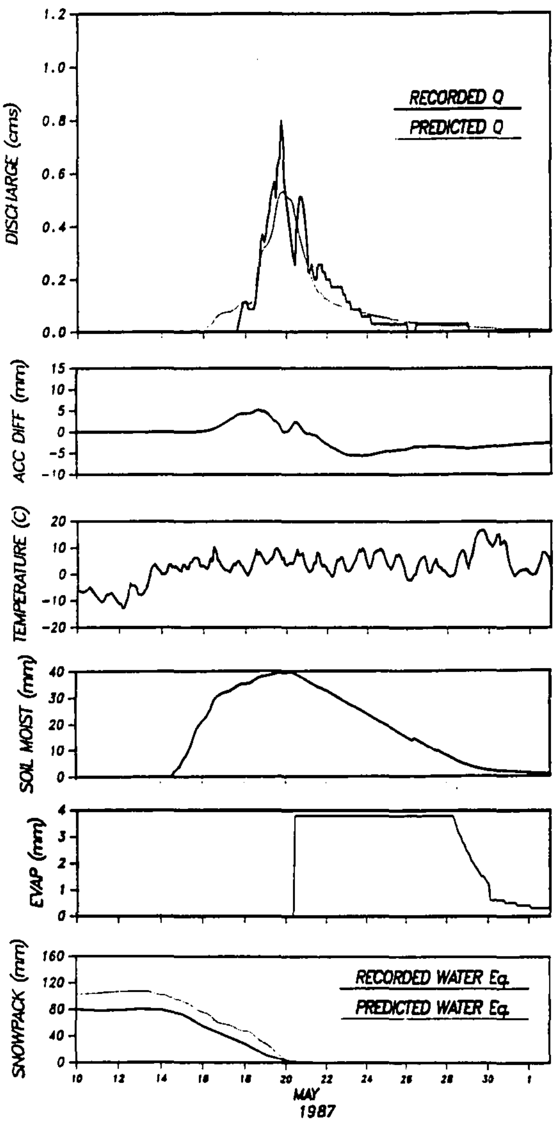

Figure 8-7. Observed and simulated hydrologic processes at Imnavait watershed during the 1987 spring melt event, accumulated difference between observed and simulated flow, measured air temperature, predicted soil moisture, evapotranspiration, and measured and predicted snowpack ablation.

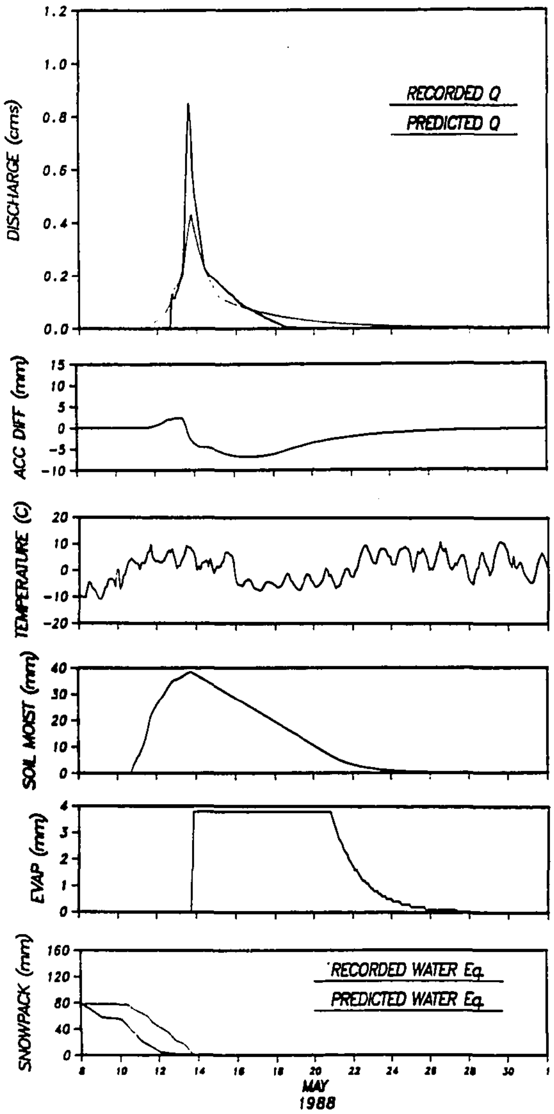

Figure 8-8. Observed and simulated hydrologic processes at Imnavait watershed during the 1988 spring melt event, accumulated difference between observed and simulated flow, measured air temperature, predicted soil moisture, evapotranspiration, and measured and predicted snowpack ablation.

1 week. The meager snowpack in the valley bottom provided little resistance to stream flow and thus snow damming was not a significant factor in the runoff process.

The snowpack water content of 1989 was the greatest during the period of the study generating the greatest total volume of runoff and the largest peak flow (Figure 8-9). Extremely high density snow in the valley bottom caused substantial damming and unusually large slush flows caused many fluctuations in discharge. In 1987, there was significant diurnal temperature variation (Figure 8-5), the consequences of which will be higher accuracy in snowmelt calculations based upon smaller time increments. The amount of melt on a day with a wide temperature variation is linear with respect to the mean, but the refreezing depth is not (Bengtsson, 1982). Days with large temperature variations will have more melt than days with small temperature variations if they have the same mean. The air temperature dropped below freezing many times during the snowmelt event so refreezing of the snowpack was an important factor to consider. The effect of diurnal temperature variation and the importance of refreezing of the snowpack is discussed by Bengtsson (1982). HBV can take refreezing into account through a refreezing factor, CFR. On

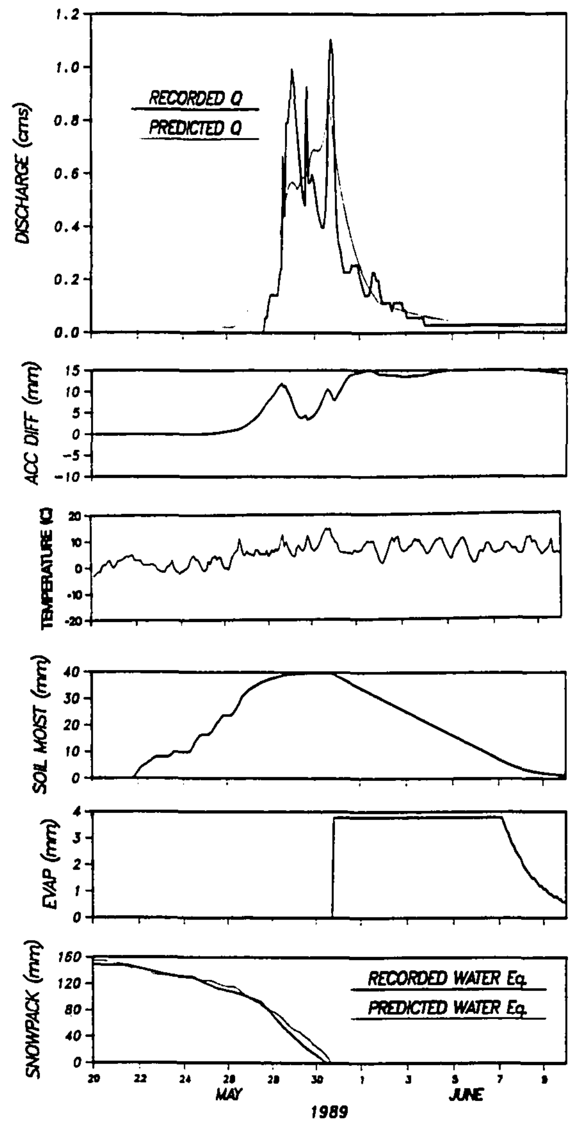

Figure 8-9. Observed and simulated hydrologic processes at Imnavait watershed during the 1989 spring melt event, accumulated difference between observed and simulated flow, measured air temperature, predicted soil moisture, evapotranspiration, and measured and predicted snowpack ablation.

several nights, significant amounts of melt occurred due to a low, thick overcast which caused substantial melt via longwave radiation, in spite of cool air temperatures. This radiative melt could not be accounted for in this model in its present form.

The curves of snowmelt calculated by HBV did not always correspond well with measurements (Figures 8-5 to 8-9), the initiation of simulated streamflow is primarily dependent upon TT. This is attributed to an important mechanism of cold regions hydrology which HBV does not consider. Snow damming will result in a delay of the initial streamflow until the force of the ponded meltwater in the valley bottom can overcome the material strength of the snow. To achieve the best fit of modeled discharge to observed discharge, it was necessary to select a different TT than the optimum TT for calculation of snowmelt. In this application, TT actually served two related but different functions. A better approach would be to include another routine which could impact the timing of initiation of runoff. It does seem possible to simulate snow damming, as it is primarily a property of snow depth, snow strength and melt rate.

The performance of the model decreased as the fluctuations in the hydrograph increased. This was due to rapidly changing meteorologic conditions such as wide diurnal temperature variation. Using smaller time increments improved the performance of the model especially at those times when when conditions changed rapidly. An exception to this trend occurred in cases where larger time steps tended to smooth the observed hydrograph data. The plots of the accumulated difference (Figure 8-5 to 8-9) aid in determining where the model fits well and when adjustments need to be incorporated. In every simulation, the modeled discharge occurred before the observed discharge began. Normally the simulated peak discharge was less than the observed discharge, the only exception occurring in 1986.

The performance of the model during each modeling scenario is summarized in Table 8-2. The accuracy of the model at estimating the volumes of water in the spring melt water balance is compared to measured values in Table 8-3.

Table 8-3. Observed and simulated water balances.

	Measured				Modeled Hourly			Modeled Daily		
Year	Snowpack Water (cm)	Snowmelt Runoff (cm)	Soil Storage (cm)	Evapo-trans (cm)	Snowmelt Runoff (cm)	Soil Storage (cm)	Evapo-trans (cm)	Snowmelt Runoff (cm)	Soil Storage (cm)	Evapo-trans (cm)
1985	10.2	6.6	1.5	2.1	6.1	0.9	3.1	6.0	2.2	1.8
1986	10.9	5.7	1.5	3.7	6.5	1.5	2.5	6.7	1.7	2.2
1987	10.8	7.1	1.5	2.2	6.4	1.5	2.6	7.4	3.1	0.9
1988	7.8	3.9	1.5	2.4	3.3	1.5	2.4	3.8	1.6	2.2
1989	15.5	9.4	1.5	4.6	10.5	2.5	1.5	11.2	2.2	1.8

F) Conclusions

Accurate modeling of the hydrology of a catchment in any environment is a difficult task. Describing the hydrology associated with snowmelt in an arctic watershed is equally complicated. The physical differences of arctic ecosystems cause markedly different responses to snowmelt events. Given measurements of initial snowpack water content and air temperature, the model functioned well in predicting streamflow timing and volumes. The model was calibrated well and functioned well in all scenarios with all the parameters the same between simulations except the threshold temperature, TT. TT was very important in controlling both snowpack ablation and snow damming. Snow damming was an important consideration in the timing of initiation of snowmelt runoff and much more information is required to quantify its effect. The importance of snow damming may decrease in large watersheds.

G) Acknowledgments

This research was sponsored by the U.S. Department of Energy's Office of Health and Environmental Research, Ecological Research Division as part of the R_4D program in Arctic Tussock Tundra. Elizabeth K. Lilly, Michael R. Lilly, and Robert E. Gieck contributed much to this project through their assistance in data collection, reduction and in the modeling effort.

IX) THERMAL RESPONSE OF THE ACTIVE LAYER IN A PERMAFROST ENVIRONMENT TO CLIMATIC WARMING

Kane, D.L., L.D. Hinzman, and J.P. Zarling. 1990. Thermal response of the active layer in a permafrost environment to climatic warming. Accepted for Publication in Cold Regions Science and Technology.

A) Abstract

Global warming is occurring; the only question is what will be the magnitude of the temperature change and the temporal and spatial distribution? Existing models predict that the greatest change from present climatic conditions will happen in the polar regions. In the Arctic, continuous permafrost exists and climatic warming could have severe consequences. In this paper the consequences of global warming on the active layer are examined. Soil temperature data were collected over a four year period at a field site near Toolik Lake, Alaska. A finite element, two dimensional, heat conduction model with phase change was used to predict soil temperatures at the site. After verification that the model could be used with confidence to predict the thermal regime, various climatic warming scenarios were used as input into the model over a fifty year simulation. The impact of climatic warming on the thickness of the active layer was estimated.

B) Introduction

While global warming will result generally in an increase in near surface soil temperatures, in areas of permafrost this slight rise in temperature will result in phase change of soil ice to water. This means that the active layer, that layer of soil at the surface that thaws seasonally, will increase in thickness. The amount of increase will be directly related to the amount of warming, the length of time since warming began and the complicating effect of vegetation. In dry areas, which become drier, vegetation may disappear and thus heighten annual soil temperature fluctuations. In some areas, vegetation may increase and eventually provide more insulation, decreasing annual soil temperature fluctuations. With sufficient warming, eventually an unfrozen layer called a talik will form between the seasonally frozen layer and the permafrost. In the extreme case where the permafrost is relatively warm, it will completely disappear relatively soon after warming begins. Even in the High Arctic, where permafrost temperatures are quite low, climatic warming will have a noticeable effect. The rate and magnitude of the response is time and temperature dependent and can not be determined analytically due to changing boundary conditions. Computer models are the only technique to accurately examine the problem and forecast the environment which we may face.

For the engineer who must design transportation facilities or site specific structures in areas of permafrost, the uncertainty as to the ground thermal regime introduces considerable risk into the design process. Most structures built in areas of continuous permafrost are constructed upon pilings which allow airflow under the building. Even in frozen soil, the support strength of the piling is dependent upon temperature of the soil around the piling. Above freezing, the piling becomes less effective as a support member. Thawing of near surface ice-rich permafrost in coastal areas with subsequent subsidence, coupled with possible sea level rise, will produce further difficulties. Alteration of hydrologic processes, particularly in areas of continuous permafrost, may pose additional problems. There are numerous other possible consequences of global warming in cold regions that are not elaborated on here.

The purpose of this research was an attempt to ascertain what effect climatic warming may have on the near surface thermal regime. This research was not an attempt to predict what the magnitude of global warming will be. Projections of global warming from a number of sources were examined and various scenarios were selected to use in a numerical model to provide an estimate of the impact on the active layer thickness.

C) Global Warming

Confirmation of global warming from meteorologic data is clouded by the fact that a weak signal emanates from a noisy record (Wigley and Jones, 1981). However, Hansen and Lebedeff (1987) report that global warming between 0.5 to 0.7°C has occurred in the past century. The main evidence for global warming is that concentrations of certain gases in the atmosphere are increasing and these gases are transparent to shortwave radiation, but not transparent for the re-emitted longwave radiation. Gases such as carbon dioxide, methane, nitrous oxide, chloro-fluorocarbons, ozone and water vapor exhibit these properties. Changes in the concentrations of these gases since 1850 are summarized by Ramanathan (1988). The concentrations of most of these gases are increasing, but in terms of absolute increases, carbon dioxide is the highest.

Numerous approaches to evaluate the impact of man on the climate of this planet have been tried. Concern for this problem dates back to 1916 (Ellsaesser et al, 1986). There are also concerns that secondary effects will cause further warming. Warming will result in less of the ocean covered with ice and a reduction in the spatial and temporal distribution of snowcover. This will result in a lower value of the surface albedo and additional surface warming. Because of the large heat capacity of the water in the oceans and the efficient transfer of heat within the water body through circulation, warming of the atmosphere will be delayed. Many other feedback mechanisms exist (Ramanathan, 1988) that need to be studied.

In an attempt to come to grips with the problem of global warming, large general circulation models (GCMs) have been developed and simulations made (Schlesinger, 1984; Schlesinger and Mitchell, 1985). These models are very complex, require considerable computer capability, produce results in response to changes in the composition of atmospheric gases and generally use a very coarse 3-dimensional network of nodes. Although these models are quite complex, numerous limitations still prevail: some processes are not well understood, some processes are modeled with simplifying assumptions, input data for some parts of the world are lacking and uncertainties are generated in the results because of poor parameterization.

Presently, most of the models show that in the Arctic, climatic warming will be 2 to 3 times greater than in more temperate regions (Schlesinger and Mitchell, 1985). The magnitude of this change varies from model to model. Input into these models for areas above 60° latitude is very limited. These models also predict that warming will be the greatest during the winter. These models predict that the temperature in Northern Alaska should increase between 2 and 4°C in July and 8 and 14°C in January for a doubling of CO_2 (Schlesinger and Mitchell, 1985).

Lachenbruch and Marshall (1986) reported that analysis of field data in northernmost Alaska shows a warming trend over a period from the last few decades up to a century ago in the near surface soil temperatures on the order of 2°C. Osterkamp et al. (1987) reached a similar conclusion for arctic Alaska. These conclusions were primarily based on analysis of deep soil temperature profiles using heat conduction theory.

D) Physical Setting

In arctic regions, the active layer plays an important role in almost all physical, biological and chemical processes. The active layer starts out completely frozen prior to snow ablation in late May and reaches a maximum thaw depth of around 40 to 50 cm in mid-August (Kane et al, 1989). Ice-rich soils near the permafrost table are responsible for producing saturated conditions in the active layer. In the area around Toolik Lake (Figure 3-1), Alaska, the active layer consists of 10 to 15 cm of organic soils over mineral soils (Hinzman et al., 1990). The organic soils may be partially saturated during snowmelt and during periods of heavy rain, but for most of the summer the water table resides in the underlying mineral soil. Water to sustain plant life is provided from this zone.

Permafrost thickness in this area is on the order of 250 to 300 m (Osterkamp and Payne, 1981). Thermal conductivity of near surface soils as a function of depth and volumetric soil moisture content have been determined (Hinzman et al., 1990). Generally, the mineral soils have the highest thermal conductivities, with the organic soils being lower. For the range of temperatures measured in the field, the thermal conductivity is essentially constant with a step increase in this property as the soil freezes. The magnitude of this step decreases as the moisture content decreases; at low moisture contents it is nearly zero. Soil thermal conductivities at depth were estimated because samples below the active layer were unavailable. The geothermal heat flux and geothermal gradient were determined by comparison with data collected by

Osterkamp (personal comm.) in a deep well 15 km from Imnavait watershed.

E) Model Formulation

Numerous models have been developed for the purpose of predicting the soil thermal regime. Early physically based models, for the sake of simplicity, concentrated on predicting temperature response due to the process of conduction. As computational methods improved, later models were improved by adding in heat transfer due to convection and phase change. It has been concluded in other studies that heat transfer in frozen soils is mainly by conduction (Lachenbruch and Marshall, 1986; Nixon, 1975).

The strategy used to predict the response of the active layer to climatic warming was to select a model that would adequately handle heat conduction and phase change. This model would need to be fairly efficient, such that long simulations of fifty years or more could be completed in a reasonable amount of time. A two dimensional heat diffusion model with phase change developed by Goering and Zarling (1985) was used. This model is based on the solution of the following equation:

$$\delta/\delta x\ [K\ (\delta\theta/\delta x)] + \delta/\delta y\ [K\ (\delta\theta/\delta y)] = C\ (\delta\theta/\delta T)$$

Where;

x = coordinate in horizontal direction, [m]
y = coordinate in the vertical direction, [m]
θ = temperature, [°C]
C = heat capacity, [J/(m^3°C]
K = thermal conductivity, [W/(m°C)]
T = time, [s]

This equation of transient heat flow is solved using a finite element technique known as the Galerkin weighted residual process.

Four years of soil temperature data of the active layer were collected at a field site near Toolik Lake, Alaska (Figure 3-1). The test of this model was to see how well, over this four year period, the model above could simulate the temperatures of the active layer. The upper boundary in the model was the ground surface and the lower boundary was at 30.5 m depth. The measured surface temperatures were used to drive the model at the surface and at the lower boundary a varying geothermal heat flux was specified. The quantity of geothermal heat entering the lower boundary was equated directly to the temperature gradient at the lower boundary. As the permafrost warmed, the geothermal gradient decreased from a maximum value of 0.0505 W/m^2. Snowcover is an important factor in the thermal regime of the permafrost and the active layer. Heat transfer through the snowpack is not a consideration in this analysis because the surface temperature was measured at the ground surface. Goodrich (1982), using a finite element model, but with a different solution technique, examined the impact of the snowcover on the ground thermal regime.

Linear triangular elements of varying size were used in the computations (Figure 9-1), with the elements smaller near the surface and increasing in size at depth. There are two reasons for using smaller size elements near the surface; the greatest temperature changes occur there and the soils and their properties are more variable there.

F) <u>Soil Thermal Regime</u>

Three factors play a very dominant role in the near surface soil temperatures in arctic Alaska. First, the snowcover that exists for more than eight months insulates the ground and therefore minimizes heat loss to the atmosphere. This ensures that minimum soil temperatures will not fall below about -15°C. During the summer months, the insulative properties of the organic soils act as a buffering layer to heat gain. Whereas the soil surface may reach fairly high temperatures, at the 10 cm depth, they rarely get above 5°C. Finally, because the water table may be quite close to the surface at times of snowmelt and intense rain, evaporative cooling reduces soil warming at a time when the organic soils have a relatively high thermal

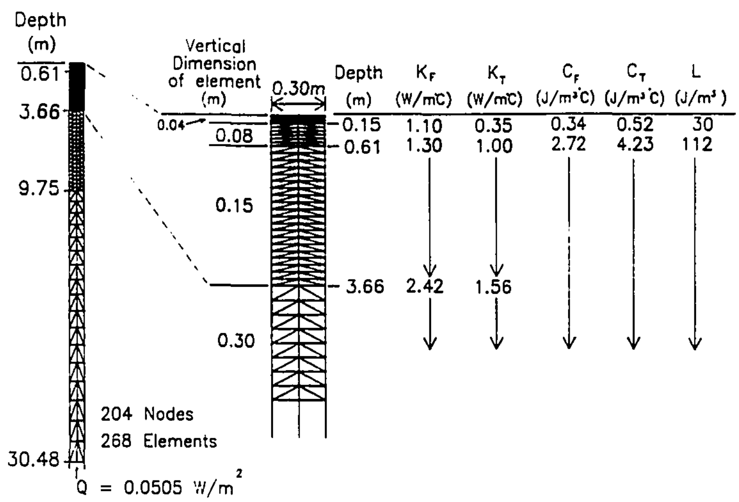

Figure 9-1. Construction of finite element grid and distribution of thermal properties.

conductivity. The temperature regime of the active layer throughout the year is illustrated by Hinzman et al., 1990. Active layer soils reach minimum temperatures in late March or early April. Gradual warming follows this period with accelerated warming during snowmelt. By the time that ablation is complete, about 10 cm of the organic soils are thawed, but the mineral soils are still frozen. Gradual warming and thawing of the active layer occurs throughout the summer with the maximum depth of thaw occurring between early and mid August. Gradual sensible cooling of the active layer follows until the soil approaches isothermal conditions near 0°C. The soil remains isothermal through phase change and then gradual cooling below the freezing point proceeds throughout the winter. Measured average daily surface temperatures monitored over a four year period show the pattern of annual soil temperature variation (Figure 9-2).

The characteristics of the annual surface temperature cycle are very important to this modeling process because the surface temperature represents the primary driving mechanism of the model. A curve fitting the surface temperature could be thought of as sinusoidal, however there are several distinguishing features. The snow during the winter months acts as a layer of insulation to the underlying permafrost shielding the

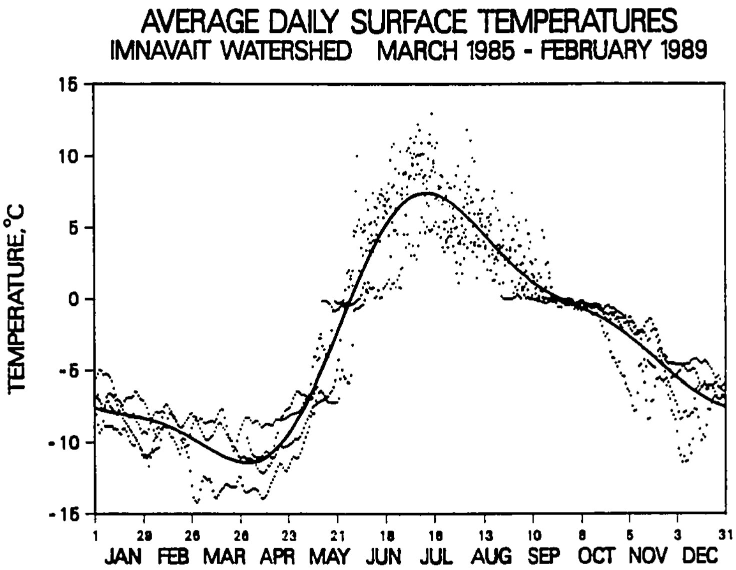

Figure 9-2. Plot of average daily surface temperatures for four years with curve of best fit.

surface from the extreme cold. Therefore the amplitude of the curve in the winter is much less than the summer amplitude. The snow also delays the timing of the minimum soil temperature in the winter, so the phase lag in the winter is later than a summer phase lag, i.e. the time difference between the minimum or maximum temperature and solstice. Considerable soil warming occurs near the time of summer solstice, when incoming radiation is near its maximum, thus the surface experiences rapid warming within a few weeks. In the fall, during the soil freeze, the radiation is still significant, but much less. Early winter snowfall plays a part in the active layer cooling by melting as it initially falls, drawing energy from the surface and gradually cooling the active layer. The net result is a lower heat flux out of the surface yielding a slower, more gradual cooling. It was assumed that these same characteristics would be maintained with the advent of climatic warming.

G) Modeling Strategy

The approach used to develop estimates of the increase in active layer thickness due to global warming was to first apply the model to the upper 30.5 m and run it for the four years of field data, 1985 through 1988. This would verify the applicability of this model and at the same time, give an indication of the validity of the assumption that conduction was the main mechanism for the transfer of heat. Thermal conductivities of the soils were determined in the laboratory using a guarded hot plate for varying soil temperatures and moisture contents (Hinzman et al., 1990). Heat capacities and latent heats of the soils were estimated from empirical techniques based upon the soil properties and moisture content. The average daily surface temperatures shown in Figure 9-2 were used to drive the model for the four year period.

The results of this four year simulation are shown in Figure 9-3. The simulated soil temperatures track the measured soil temperatures quite well for the entire test period. However the simulations do lag behind the measured values by at most 1°C each spring at the onset of soil thawing. This difference can be attributed to the additional heat transfer mechanism of convection that is not incorporated into this model. Throughout the winter moisture migrates to the surface in response to a strong thermal gradient leaving open pores in the organic soils which will allow limited infiltration during the spring snowmelt. The infiltrating meltwater increases the soil temperature sooner than the model would predict.

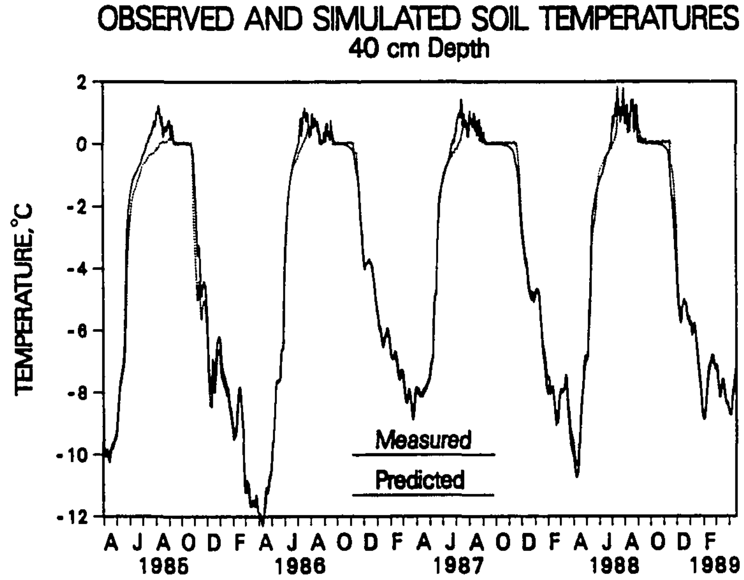

Figure 9-3. Measured and simulated soil temperatures for four years at 40 cm depth.

The ramifications of neglecting convective heat transfer will be negligible when the soil is frozen. When the soil is thawed, the model will slightly under-predict the soil temperature and consequently under-predict soil thaw depths. The length of time and amount of under-prediction of temperature are both relatively small, so the estimate of thaw depths will not be grossly in error. Convective heat transfer will become progressively more important as the soils warm and remain unfrozen longer.

After the simulations were complete with measured surface temperatures and the model was providing satisfactory results, the model was used to simulate the response of the active layer for periods up to 50 years. Since no surface temperature data reflecting global warming was available for this simulation, it was generated. This was not a simple task because, as described earlier, the annual cycle of surface temperature does not follow a simple cosine curve, as is often assumed. To maintain a semblance of reality, it was necessary to create a data set which emulated the present surface temperature and then superimpose a gradual warming trend to that initial temperature curve. A curve approximating the initial surface temperatures was determined by fitting a truncated Fourier series to the four years of surface data (Figure 9-2).

The equation is:

$$\begin{aligned}\theta(T) = -3.05 &- 7.83 * \cos(2\pi ((T - \phi) / 365)) \\ &+ 0.38 * \sin(2\pi ((T - \phi) / 365)) \\ &+ 0.76 * \cos(4\pi ((T - \phi) / 365)) \\ &- 2.73 * \sin(4\pi ((T - \phi) / 365)) \\ &+ 1.21 * \cos(6\pi ((T - \phi) / 365)) \\ &+ 0.70 * \sin(6\pi ((T - \phi) / 365))\end{aligned}$$

Where
θ = temperature, [°C]
T = day of the year, 1 to 365
ϕ = phase lag, 43.9 days

Four different scenarios of global warming were examined: 2, 4, 6, and 8°C of gradual warming for up to a 50 year period. The curve fitting the surface temperature data in Figure 9-3 was adjusted for each climatic warming scenario. This was done by adding a linear function of the particular scenario to the above equation. For example, for a 4°C warming trend with 1 day time steps accumulated over 50 years, the surface temperature would be calculated thus:

$$F(T) = \frac{4 * T}{365 * 50}$$

where T = day number from 1 to 18250 (365 days * 50 years)

In this form, the predicted surface temperatures used to drive the model would still not emulate realistic surface temperatures in autumn when the soil should refreeze. The superposition of the linear relation will at this time level off at a point F(t) above the phase change temperature of 0°C, which is of course not possible. A third routine was introduced to the climate generator which became effective only during freeze-up. This equation took effect at the point of inflection of the soil cooling curve, maintaining the current slope until reaching the phase change temperature. The surface would remain at that temperature until the temperature curve $\theta(T) + F(T)$ dropped below 0°C and again became the controlling equation. The initial surface temperature curve and the final curve after 50 years of 4°C gradual warming is presented in Figure 9-4. Note each of the important characteristics of the surface temperature are maintained and the projected scenario does realistically emulate a constant temperature increase.

This approach is not consistent with some projections of climatic warming (Schlesinger and Mitchell, 1987). These models predict that the greatest temperature change will occur in the winter and the difference will be less in the summer. This analysis assumes that the change will occur uniformly throughout the year and steadily over the 50 year period.

The next step was to run the model for the simulations, up to 50 years, with the generated surface temperatures for the 4 different climatic warming scenarios. The soil profile temperatures were calculated and the

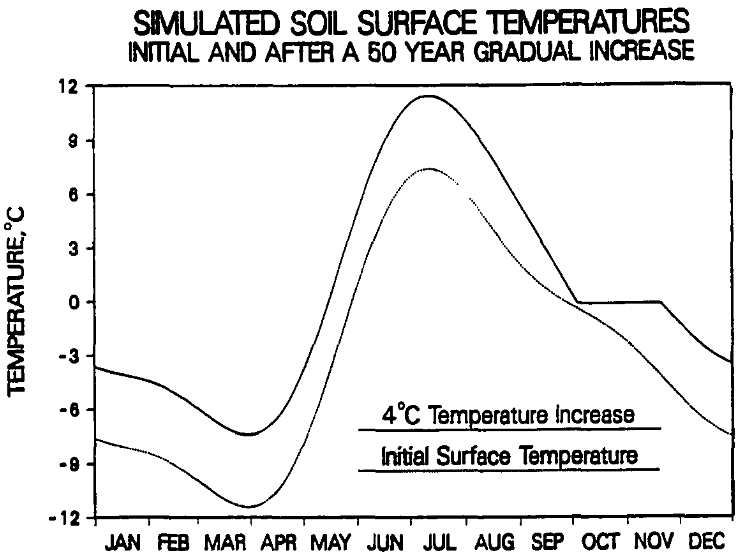

Figure 9-4. Difference between current ground surface temperatures and after year 50 warming 4°C.

maximum depth of thaw for each year was recorded.

H) Results

The change in the surface temperature over the 50 year simulation is shown Figure 9-4 for a 4°C increase in the mean annual surface temperature. Not only is there a change in the daily mean temperature, but the length of the thaw season increases and that of the freezing season decreases. Similar changes are observed for the 2, 6 and 8°C scenarios in direct proportion to the magnitude of the temperature change.

In Figures 9-5 a and b, the soil temperature conditions prior to climatic warming are shown in a whiplash curve and in curves that show the temperature at selected depths throughout the year. The whiplash curves in Figures 9-6 a, b, c, and d show the degree of warming that will occur from the global warming scenarios. It is obvious that the first two cases result in gradual warming of the permafrost and an increase from 50 cm to 70 and 90 cm for the active layer thickness in 50 years. For the last two cases, the active layer increases to values greater than 120 cm, at which time a talik begins to form. This means that there is a zone that does not refreeze the following winter. When this talik starts to form, numerical oscillations occur in the solution and the model no longer predicts the correct rate of

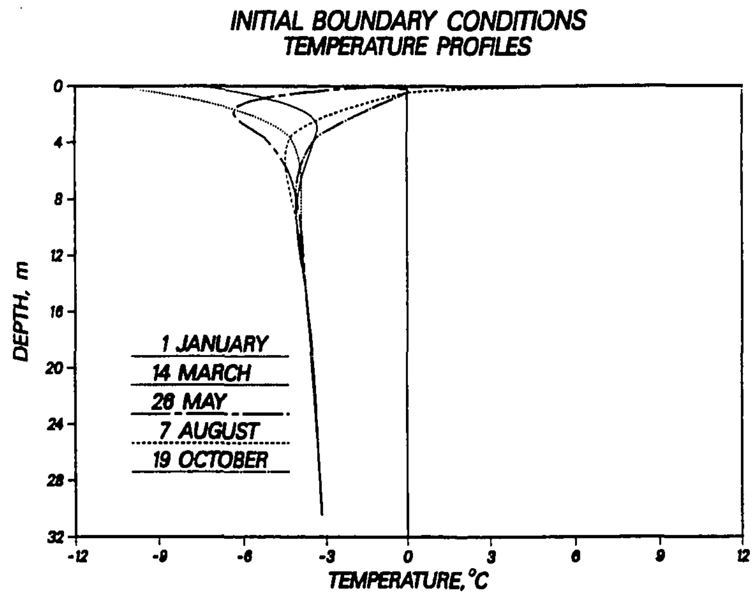

Figure 9-5a. Initial soil temperature profiles illustrated by whiplash curves on selected dates before the onset of climatic warming.

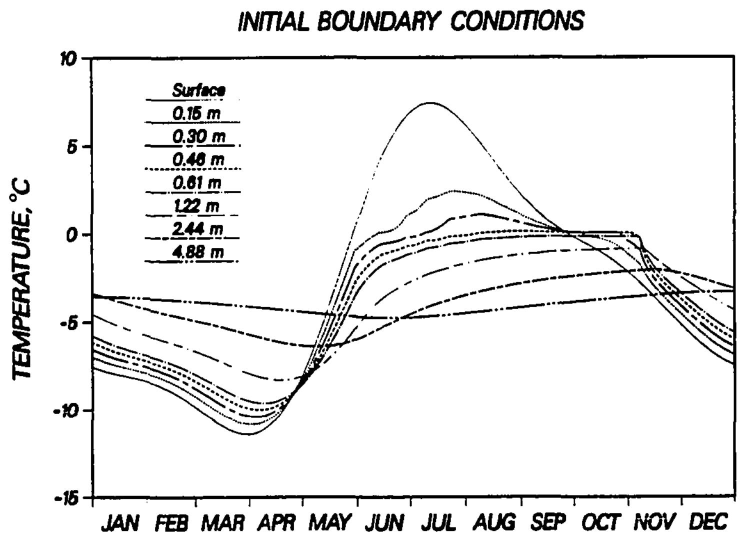

Figure 9-5b. Initial soil temperatures illustrated by curves at selected depths near the ground surface before the onset of climatic warming.

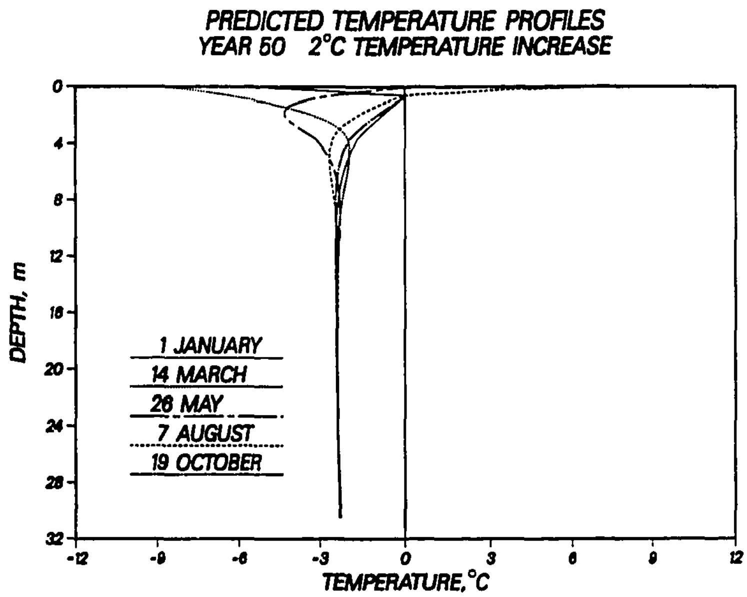

Figure 9-6a. Whiplash curves illustrating soil temperatures at the end of the 2°C warming scenario.

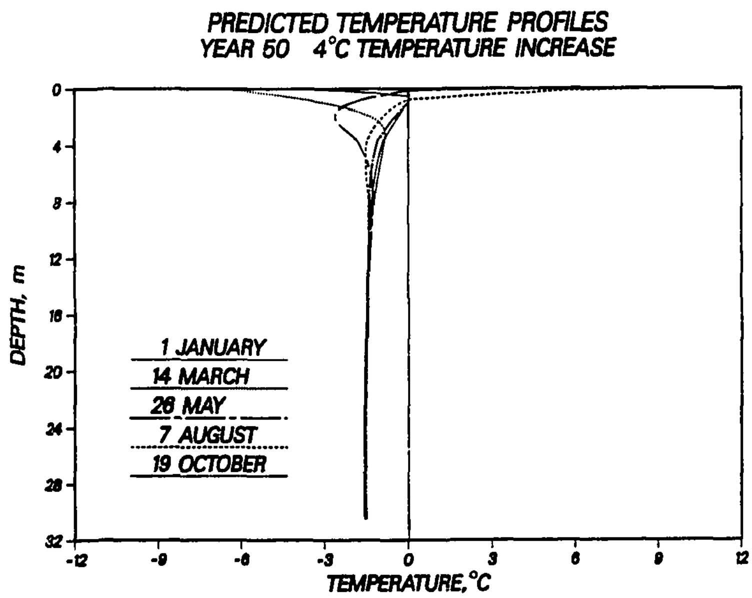

Figure 9-6b. Whiplash curves illustrating soil temperatures at the end of the 4°C warming scenario.

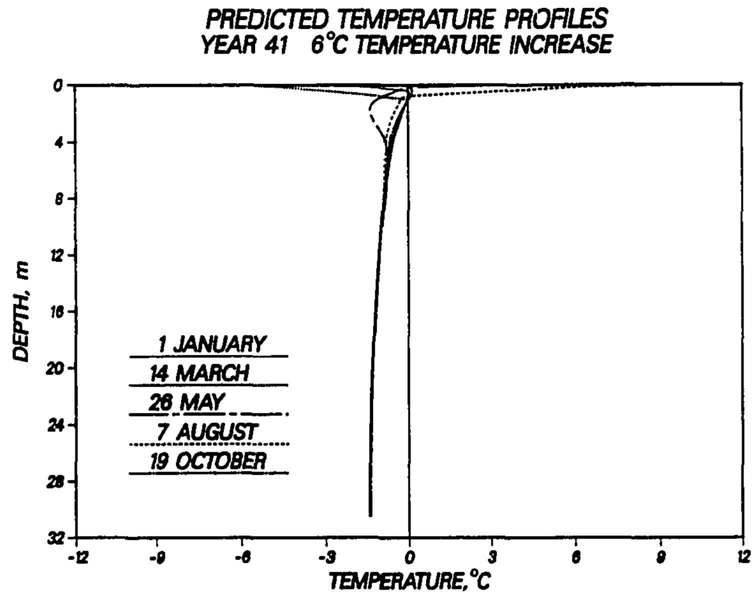

Figure 9-6c. Whiplash curves illustrating soil temperatures at the end of the 6°C warming scenario.

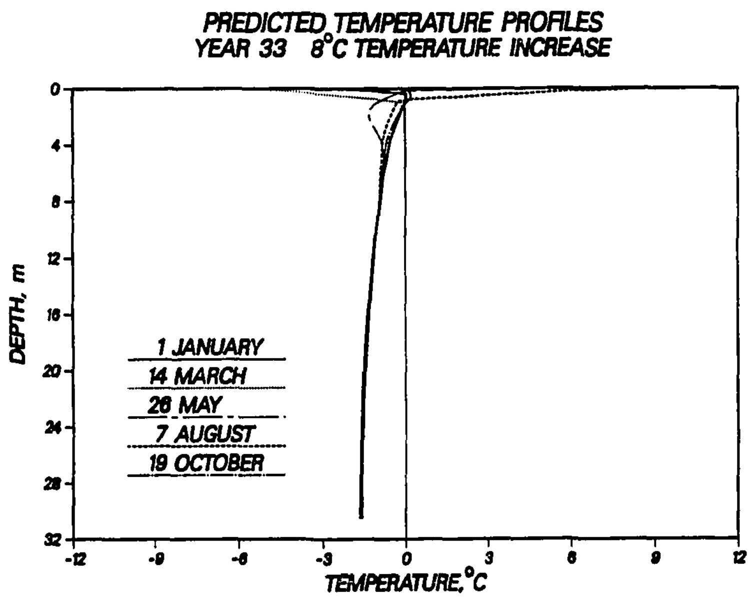

Figure 9-6d. Whiplash curves illustrating soil temperatures at the end of the 8°C warming scenario.

thaw and therefore the simulations were stopped after 41 and 33 years for the 6 and 8°C scenarios respectively. If the amount of global warming raises the surface temperature by more than the difference between the minimum permafrost temperature and the freezing point of water, the permafrost will eventually melt. Because permafrost exists from 0°C in areas of discontinuous permafrost to temperatures as low as -10 to -12°C in areas of continuous permafrost, any amount of global warming will also result in an areal reduction in the amount of permafrost in Alaska.

The predicted increase of the active layer thickness in response to the global warming scenarios is shown (Figure 9-7). Over the fifty year period, a gradual increase in the active layer thickness occurs for the 2 and 4°C warming cases. At the same time that the active layer is increasing in thickness, thawing at the base of the permafrost will occur. A limitation of the model becomes evident as portions of the soil profile began thawing and approach isothermal conditions at 0°C. There is no temperature gradient and in fact at 0°C the soil can be either frozen or unfrozen. The level of the thaw depth is determined as the depth at which the temperature is above 0°C. The amount of phase change is determined through apparent heat capacity calculations rather than an accumulation of the enthalpy.

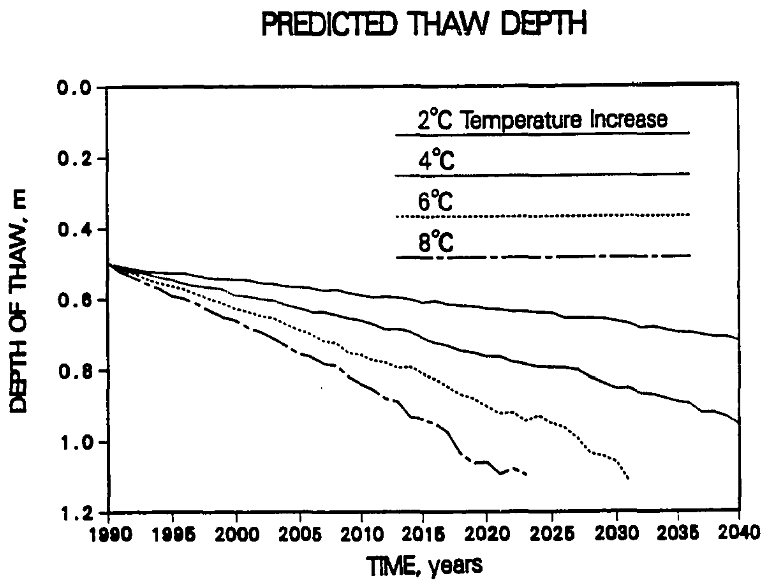

Figure 9-7. Increased depth of thaw of the active layer in response to each climatic warming scenario over 50 years.

The method used is the Dirac delta function (O'Neill, 1983) which was used to account for the latent heat in the global heat capacitance matrix. This finite element formulation of latent heat allocates the energy to the element nodes based on their distance from the freeze/thaw interface. The temperature gradient across the element is required for this formulation and caused numerical difficulties when part of the soil profile goes isothermal after prolonged warming. For this reason, the runs were terminated when isothermal conditions were reached. It is perceived that most standard finite element programs would experience similar numerical difficulties when extremely small thermal gradients exist.

I) Discussion

In this paper, the increase in depth of the active layer in response to various scenarios of climatic warming have been determined. It was concluded that a model of heat conduction with phase change would be suitable for predicting the thermal response of the active layer to warming at the upper boundary layer. The next question then is this critical or important? One major concern is how arctic hydrology may be modified by global change.

The present hydrologic system in arctic Alaska has the following major components: precipitation (snowmelt and rainfall), active layer storage, evaporation, transpiration, and runoff. In arctic watersheds, the permafrost at the surface has a high ice content and is relatively impermeable. In the hydrologic cycle, subsurface processes are confined to the active layer.

With climatic warming, not only will the various components of the hydrologic cycle change, but melting of the permafrost from the top down will physically alter the subsurface system. As the active layer thickness increases in response to global warming, available storage in the active layer will increase. During snowmelt and major summer rainfall events, the active layer is saturated. Also, during most of the summer, the mineral soil beneath the surface organic soil is very close to saturation. In terms of the plants, this is important because a substantial amount of available soil water is within 10 to 15 cm of the ground surface.

Assuming that the pattern and amounts of precipitation remain the same during the next 50 years, then it is the distribution of this water into runoff, evapotranspiration, and active layer storage that is important. Most GCM models predict that there will be changes in both the pattern and quantity of precipitation. It is an accepted fact that the present global models need considerable improvement in the area of hydrology. Presently we have a good understanding of the hydrology including the ratios of precipitation, runoff, and evapotranspiration for some areas in the Arctic. If there is an increase in runoff, then in-place hydraulic structures may not be designed with sufficient capacity to handle the flows.

Thermal erosion which will accompany substantial increases in active layer depths will yield significant amounts of sediment into surface runoff. Increased sediment loads will probably affect channel morphology and possibly the lifespan and effectiveness of hydraulic structures.

What happens in terms of arctic hydrology in response to global warming depends upon the vegetation. If the active layer increases in thickness and the annual precipitation remains fairly constant, then the water table in the active layer will fall. Properties of these soils (Hinzman et al., 1990) show that the organic soils readily give up water to tensions as low as 10 cm of water. If the plants cannot adapt and draw water up from greater depths, then they will become stressed and disappear. They may eventually be replaced by other plants. Presently, the plants are responsible for holding in place the organic soils and also transpiring large quantities of water back to the atmosphere. Kane et al. (1989) have shown that between 34 and 56% of the annual precipitation falling on Imnavait Creek returns to the atmosphere and that the remainder leaves the basin as runoff.

Manabe and Wetherald (1986, 1987) have looked at the reduction in summer soil wetness in response to doubling CO_2. They have concluded from a global climate model that most of the arctic regions of the world will undergo a reduction in soil wetness. For northern Alaska, a decrease in soil wetness during June, July and August is predicted to be about 10%.

If there is a decrease in the vegetation cover, then the percentage of runoff leaving the basin should increase at the same time the potential for erosion and sediment transport will substantially increase.

In the 6 and 8°C scenarios, there was a point where the active layer did not completely refreeze. This occurred when the maximum active layer thickness reached slightly more than 120 cm. The soil moisture conditions were not varied during the simulations. If the amount of soil water decreases during the simulation, then the thickness of the active layer could be greater than 120 cm and still freeze back the next winter. How much the active layer could increase would be directly related to the reduction in the moisture content and the amount of water going through phase change.

J) Conclusions

Climatic warming on a global scale is taking place; the question is what will be the magnitude of the change. The oceans, as large heat sinks, can attenuate the thermal effect of large increases in greenhouse gases. It is expected that polar regions will be most sensitive to climatic changes, with the total warming being 2 to 3 times greater than in tropical regions. While most regions of the world will undergo significant changes in their hydrology as we know them today, areas with permafrost will undergo a physical transformation. This transformation will be a decrease in permafrost thickness or in the case of relatively warm permafrost, complete degradation. At the same time, active layer thicknesses will increase. Both of these events are significant because in most cases, high ice content permafrost which is relatively impermeable is being replaced through phase change with a more permeable medium where the water is not tied up as ice.

The use of the heat conduction model in this paper to predict active layer thicknesses is the first step in ascertaining the possible ramifications of climatic warming in a tundra region. Results show that an increase in the average annual soil surface temperature of 2°C increases the active layer thickness at Imnavait Creek by about 20 cm after 50 years. A 4°C change increases the active layer thickness by 40 cm and the remaining two scenarios of 6 and 8°C greatly increase the active layer thickness and eventually produce a talik.

It is obvious from the model results that even a slight climatic change will produce significant changes in the hydrology of the arctic areas of Alaska. Large changes in average surface temperature will cause substantial changes in the geomorphic, hydrologic, and biologic regimes. These changes may be imperceptible from year to year, and the lack of long term hydrologic and meteorologic data in the Arctic will prevent us from seeing deviations from the past hydrologic time series. Further consideration should be given to the effect of climatic warming prior to extensive development in areas underlain by permafrost.

K) Acknowledgments

This research was supported by the U. S. Department of Energy's Office of Health and Environmental Research, Ecological Research Division. The authors would like to thank the following individuals for their assistance; Robert E. Gieck, Elizabeth K. Lilly, and Clarke A. Pelz.

X) THERMAL AND HYDROLOGIC PROCESSES IN AN ARCTIC WATERSHED AND THEIR RESPONSE TO CLIMATIC WARMING

Hinzman, L.D. 1990. Thermal and hydrologic processes in an arctic watershed and their response to climatic warming.

A) Abstract

The heat and mass transfer processes which comprise the thermal and hydrologic regime were monitored continuously from March 1985 until September 1989 in Imnavait watershed. These continuous measurements have given us a better understanding of the physical processes which determine the character of an arctic watershed. The state of the hydrologic regime is dependent upon the thermal regime. The amount of snow accumulation, ablation, snowmelt or precipitation runoff, evaporation and rainfall are all controlled by the atmospheric processes and energy input. The energy exchange in these hydrologic processes is a significant component of the total energy balance. The hydrologic and thermal regimes interact to such an extent that neither can be fully understood without considering the other. Soil moisture content and phase strongly affect the soil thermal properties, such as the thermal conductivity, volumetric latent heat and specific heat. The soil thermal regime strongly affects the soil hydrologic properties. The hydraulic conductivity and

field capacity of a soil profile vary drastically with depth. The depth of the active layer, a result of the thermal regime, affects the amount of soil moisture storage, the magnitude of base flow, even the shape of the recession curve. Strong temperature gradients in the soil profile cause a desiccation of the surface organic soils over the winter. The consequences of a manmade or environmentally induced alteration in the thermal regime can have dramatic and perhaps dire effects on the hydrologic regime and vice versa.

The implications of global warming reach beyond warmer air temperatures, milder winters and longer summers. The potential effects of climatic warming on the hydrologic regime of an arctic watershed were explored with respect to physical changes in the active layer and the resultant changes in the components of the annual water balance and the nature of the hydrologic cycle. With the advent of climatic warming, the annual depth of thaw in the permafrost will increase, affecting the amount of soil moisture storage, the depth to the water table in the active layer, even the shape of the runoff hydrograph. The gradual thawing of the active layer was simulated using TDHC, a heat conduction model which incorporated phase change. The results of four possible scenarios of climatic warming were input into HBV, a hydrologic model to elucidate the effects on the hydrologic regime. The results indicate an earlier spring melt event, greater evaporation, greater soil moisture storage, and a potential for severe moisture stress on current vegetation types in early summer unless the precipitation pattern changes. The amount of free water in the soil will largely depend upon precipitation patterns and amount.

B) Introduction

Most research on the Arctic Slope of Alaska has been completed along the Arctic Coast or in a survey along the only road which bisects the Slope. Many of these studies were ecological and few of them extended beyond the short summer. No previous information exists for the annual hydrologic and thermal processes occurring in the interior foothills. The focus of this project has been to understand the annual hydrologic sequence of events and the energy gradients which drive them and to determine their response to disturbance.

Data were collected over a five year period at Imnavait watershed located near Toolik Lake in the foothills on the north side of the Brooks Range at 68°30′ north latitude and 149°15′ west longitude (Figure 3-1). The physiography is typical of the foothills with low rolling piedmont hills and tussock tundra. The vegetation is described in Haugen (1980). The watershed is completely underlain by permafrost to a depth of about 250 to 300 m (Osterkamp and Payne, 1981). The active layer depth ranges from 25 to 60 cm. The highly stratified soils have been classified as Histic Pergelic Cryaquepts and strongly influence both the hydrologic and thermal regimes (Hinzman et al., 1990). An organic mat 10 to 20 cm thick overlies a fine grained till. The configuration of the soil impacts the thermal regime through the insulative properties of the organic mat. The thermal conductivity of this layer is 1/2 to 1/6 that of the underlying mineral soil (Hinzman et al., 1990). The albedo of this surficial layer is also higher than the mineral soil, consequently if this layer is removed heat flux at the air soil interface will increase. The configuration of the soil also impacts the hydrologic regime through distinct differences in hydrologic properties. The hydraulic conductivity of the organic layer is 10 to 1000 times greater than that of the mineral soil (Hinzman et al., 1990). Consequently water from snowmelt or summer rainfall will move downslope above the organic mineral interface. This is important in considering source of transported nutrients or in calculating travel times.

An examination of the potential impact of climatic warming must be based on a firm understanding of the processes and properties which currently exist. One must be able to accurately simulate the present thermal and hydrologic processes before one can extrapolate a change in the system. The model selected to simulate a changing condition must be firmly based in physics and not merely a statistical model based upon empiricisms which cannot be used outside the range of calibration data. It is also important to correctly relate the parameters of the model to the natural factors of the environment so an alteration to the system may be properly incorporated into the modeling process. The process of determining the effect of climatic warming on the hydrologic regime necessarily begins with a discussion of the present environment and the process of modeling the thermal and hydrologic regimes.

C) Data Collection

Field work was initiated in the autumn of 1984 at the time when the depth of thaw of the active layer was near the maximum. Four 90 m^2 plots were installed along a traverse of the west facing slope. The purpose of these plots was to measure the amount of snowmelt or precipitation runoff at various locations along the slope (Figure 3-2). They were bounded with heavy ($\approx$1 mm) plastic down to permafrost to prevent water from upslope from entering the plot. A gutter system at the bottom of the plot collected the runoff and channeled the water to a holding tank mounted with a water level recorder, thus the volumes and rates of runoff could be monitored. The snowpack adjacent to each plot was measured periodically throughout the winter and daily during the snowmelt event. Two meteorological stations were maintained in the watershed, one on the ridgeline near the outlet of the watershed and one between plots 1 and 2 (Figure 3-2). The radiation components of incoming and outgoing shortwave and longwave and net radiation were measured along with air temperature, relative humidity, barometric pressure, wind speed and direction. Radiation was only monitored from about vernal equinox through autumnal equinox. Precipitation was measured using two shielded tipping bucket raingages and a Wyoming snow gage. The snowpack was measured periodically throughout the winter and daily during snowmelt with and Adirondack snow sampler. Potential evaporation was measured in the summer using a standard 4 foot (1.3 m) evaporation pan. Thermistors were installed adjacent to each plot in 5 cm increments above permafrost in the active layer and at 0, 10, 20 and 30 cm above the soil in the snow. Soil heat flux plates were installed at the soil surface, in the soil profile at the organic/mineral interface and at 20 cm in the mineral soil near plot 3. Time domain reflectometry

(TDR) probes were inserted next to the soil thermistors at 5, 10, 15, 20, 25, 30, and 40 cm to measure the unfrozen soil moisture content in the active layer. The meteorological sensors and thermistors were sampled and recorded on Campbell Scientific 21X dataloggers. Stream flow was frequently measured using an electromagnetic current meter and stage was continuously recorded using a water level recorder mounted near a 1 meter H type fiber glass flume. The rating curve provided with the flume was corroborated using the measurements of discharge and stage. Soil samples were collected at the site and returned to the laboratory for determination of thermal and hydrologic material properties.

D) <u>Thermal Regime of Imnavait Watershed</u>

The energy input into a system drives all processes within that system. Although it is usually possible and reasonable to study a particular mechanism for its own merits, a better understanding of the system can be developed if we understand the comprehensive pattern of energy exchanges. In the broadest sense, the consequence of position in the Arctic will mean receiving much less solar radiation annually than the lower latitudes. Due to conservation of energy, the surface energy gains and losses must equal zero. The negative net radiation balance in the Arctic impacts the wind circulation patterns of the world as energy is convected northward from lower latitudes to balance the energy equation. On an annual or longer time scale, a dynamic equilibrium of the surface energy balance is maintained by the aggradation or degradation of permafrost.

The distinct variation in the annual radiation budget affects every aspect of the hydrologic and thermal regime. The arctic winter is characterized by low incoming solar radiation and a uniform high albedo yielding little energy to the active layer. The harsh winter climate hampers year round measurement of radiation, but simple calculations reveal that Imnavait watershed will receive no direct solar radiation between December 5 and January 8. There will still be several hours of diffuse radiation throughout the winter, but the energy yield from the low sun angle throughout the winter is minimal. In Imnavait watershed about 77% of the annual sunlit hours occur between March 21 and September 21, but if one considers the low intensity of winter radiation, the proportion is even more skewed toward summer dominance. With reduced incoming shortwave radiation, emitted longwave radiation is the dominant term of the winter radiation balance. Even in the summer, the incoming solar radiation is somewhat attenuated by the atmosphere due to the low solar angle.

The annual maximum solar angle, which occurs on the summer solstice, is still less than 45° above the horizon. The average annual air temperature in the watershed (from May 1985 to April 1989) was -7.4°C (Figure 10-1) and the since the radiation balance totaled to a net loss, the ground remains in a perpetually frozen state.

The shortwave radiation is strongly affected by the atmospheric conditions, being greatly reduced on cloudy days, and by sun angle, being greatest near solstice and solar noon (Figure 10-2). Reflected shortwave is naturally most dependent upon incoming shortwave, and secondly upon the albedo (Figure 10-3). The albedo varies as a function of surface condition and sun angle, but responds primarily to presence of snowcover (Figure 10-4). The snow does not melt in the winter and therefore the albedo remains near 0.8 from October to May dropping to 0.2 after snowmelt. The snow ablation is also evident in a plot of emitted longwave radiation (Figure 10-5). Longwave radiation is emitted from the snow or soil surface as a function of temperature and from the atmosphere as a function of cloud cover and temperature. It is difficult to separate reflected and emitted longwave radiation, so these are commonly considered collectively. The arithmetic sum of incoming

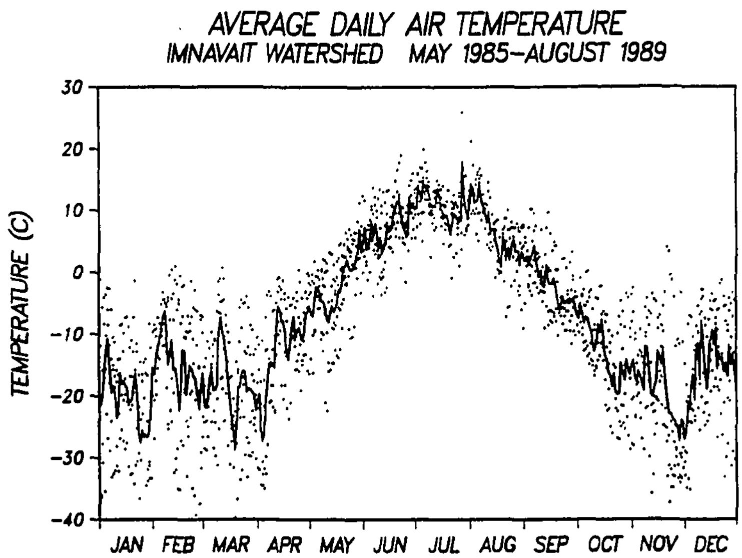

Figure 10-1. Average daily air temperature 1985 - 1989 with line connecting daily averages.

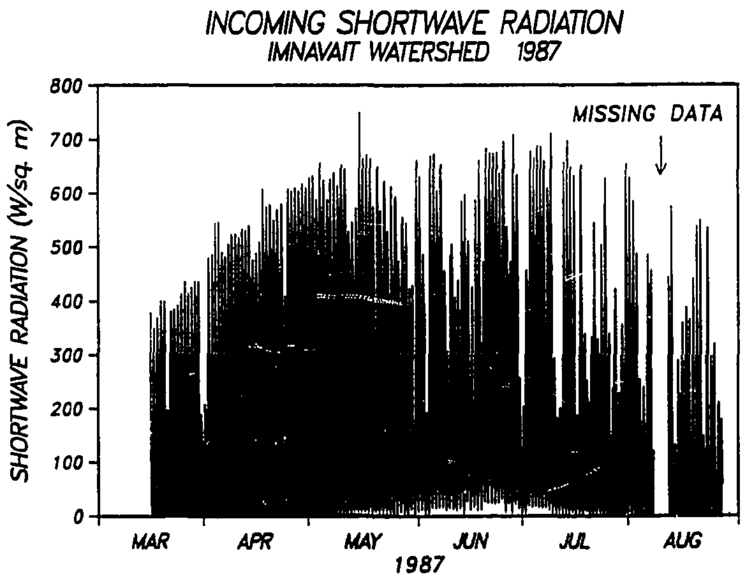

Figure 10-2. Incoming shortwave radiation measured hourly in 1987.

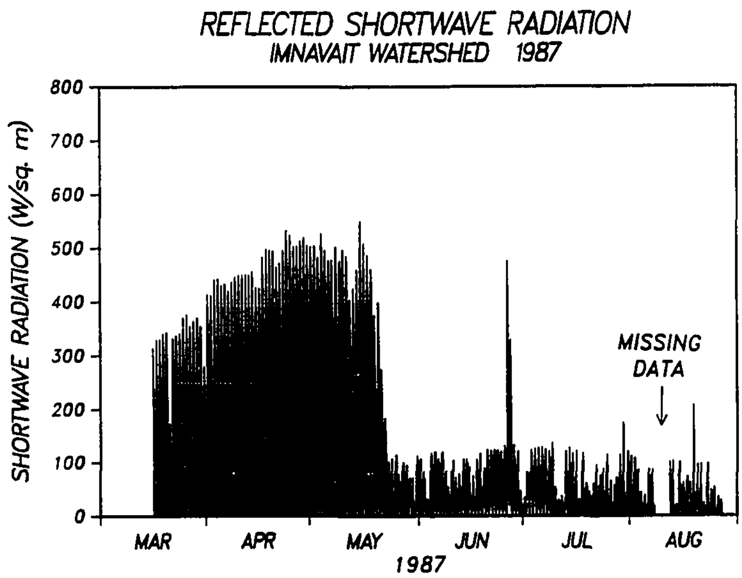

Figure 10-3. Reflected shortwave radiation measured hourly in 1987.

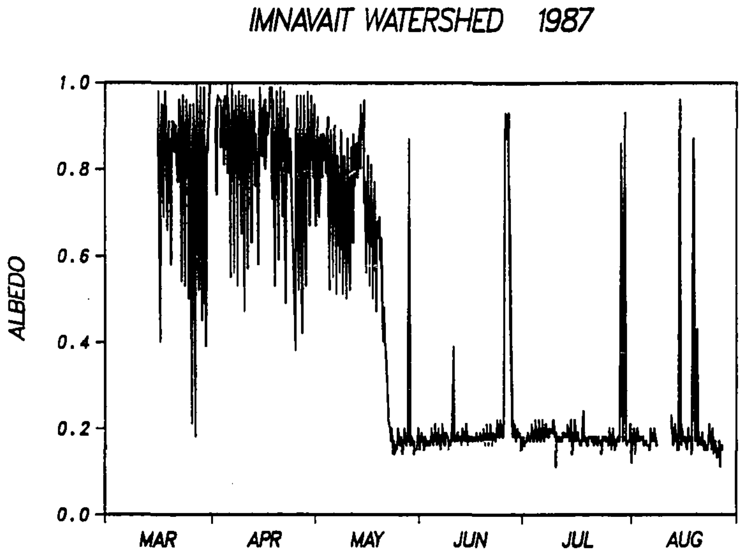

Figure 10-4. Albedo measured hourly in 1987.

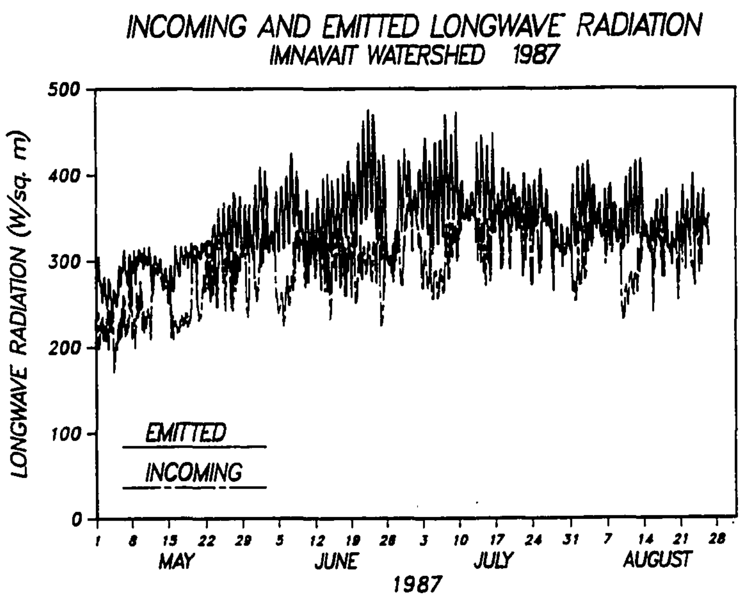

Figure 10-5. Incoming and emitted longwave radiation measured hourly in 1987.

shortwave and incoming longwave minus reflected shortwave and emitted longwave is the net radiation and is a dominant component of the summer energy balance. The net radiation becomes positive during the day in March and increases substantially after snowmelt (Figure 10-6) as the albedo sharply decreases and the amount of reflected radiation decreases.

The net radiation is just one component, albeit the dominant constituent, of the summer energy balance. An energy balance calculated at the surface also consists of the energy transferred due to conductive, convective and evaporative heat loss, plus any energy transferred in melting or freezing of snow. The surface energy balance can be described as:

$$Q_N + Q_H + Q_C + Q_E + Q_M = 0 \qquad (1)$$

where Q_N = Net Radiation
Q_H = Convective Heat Transfer
Q_C = Heat Conduction into Soil
Q_E = Energy utilized for Evaporation, Transpiration, Condensation and Sublimation
Q_M = Energy utilized for Snowmelt

Energy transferred by rain or running water and that used for biologic activity is not included in the analysis. Convection of sensible heat is the energy transferred between the air and the soil or snow surface. Unlike radiation, this energy flux is not normally measured directly.

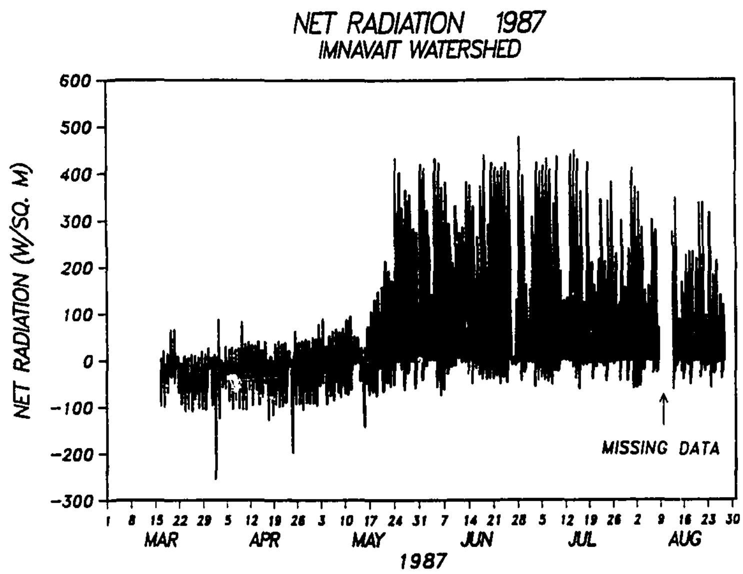

Figure 10-6. Net radiation measured hourly in 1987.

It is usually calculated as a function of the temperature gradient, wind speed and surface conditions:

$$Q_H = C_p * \rho_a * D_{h(n,s,u)} * (T_a - T_{sur}) \qquad (2)$$

where C_p = specific heat of air
ρ_a = density of air
T_a = temperature of air
T_{sur} = temperature of the surface
D_{hn} = neutral convective transfer coefficient
= $k^2 u_z / (\ln (z / z_o))$
D_{hs} = stable convective transfer coefficient
= $D_{hn} / (1 + (\sigma * Ri))$
D_{hu} = unstable convective transfer coefficient
= $D_{hn} * (1 - (\sigma * Ri))$
Ri = Richardson number
= $(g * z * T_{del})/(u_z^2 * (T_a + 273.15))$
u_z = wind speed at height z, m/s
z = height of the wind speed measurement, m
z_o = roughness length, m
= $\exp[((u_{z2} * \ln z_1) - (u_{z1} * \ln z_2))/(u_{z2} - u_{z1})]$
σ = empirical constant = 10
g = gravitational constant
k = von Karman's constant = 0.41
T_{del} = temperature difference between evaporative surface and height z = $T_a - T_{sur}$

The terms of Q_n, Q_s and Q_c can be calculated directly from measurements taken in the field. However, the convective heat transfer term (Q_h) was not calculated directly from field measurements because the heat exchange coefficient ($D_{hn,s,u}$) was derived by assuming the momentum exchange coefficient (D_m) between the atmosphere and ground to be equal to the heat exchange coefficient (Szeicz et al., 1969) or:

$$D_m = D_h \qquad (3)$$

This relationship is only valid for neutral (isothermal) conditions. A correction must be applied to account for the stability of the air just above the ground surface. This is possible using the either the Monin-Obuchov factor or, as in this paper, the Richardson number (Price and Dunne, 1976).

Effective surface temperatures for the Q_h calculation (Equation 2) were obtained by back calculation using the emitted terrestrial longwave radiation. Daily heat exchange coefficients were adjusted to compensate for air stability based on the air temperature profile between the surface and 10 meters using D_{hn} for neutral, D_{hs} for stable and D_{hu} for unstable conditions. The average surface roughness length, z_o, was determined from wind speed profiles between 1.5 and 10 meters. The roughness length was estimated at 0.02 m by averaging calculations made from several hundred wind profile measurements during near neutral conditions. This value falls well within the reported values for short grasses (Szeicz et al., 1969) and yields reasonable results. Energy advected with precipitation, Q_a was assumed to be negligible.

Conductive heat transfer is the energy passed particle to particle within the soil matrix. Since the energy budget is calculated in a plane at the surface, it is not necessary to consider the latent heat effects of freezing or thawing the active layer. Conduction (Q_c)

can be calculated from Fourier's Law:

$$Q_C = - K \frac{dT}{dX} \qquad (4)$$

where K = thermal conductivity
$\frac{dT}{dX}$ = thermal gradient

The true evapotranspiration from a system is difficult to measure so a variety of techniques have been developed to calculate reasonable estimates. Evapotranspiration can be calculated from the remainder term in a water balance, which is discussed later, or from the remainder term in an energy balance:

$$Q_E = Q_N + Q_H + Q_C + Q_M \qquad (5)$$

Several methods of calculating evapotranspiration from meteorologic variables and certain empirical surface condition parameters abound in the literature. The technique selected is the Priestley Taylor equation (Priestley and Taylor, 1972). This is a simplified approach which estimates evapotranspiration based upon the net radiation and the air temperature (Rouse and Stewart, 1972):

$$E = \alpha * (0.406 + 0.011 * T_a) * Q_N \qquad (6)$$

where E = evapotranspiration
α = surface control factor
T_a = air temperature
Q_N = net radiation

The importance of each of the above terms varies considerably throughout the year. The rates of all energy fluxes are reduced during the winter for a variety of reasons. Net radiation is reduced due to lower incoming solar radiation. Conductive heat transfer is reduced due to the insulating effect of the snow layer. Evapotranspiration is normally considered to be quite low during the winter due to energy constraints (Ohmura, 1982a), however the loss of snow to sublimation can be high during large wind events. Convective heat transfer is affected the least, but can be lower due to the smoother surface of snow and smaller temperature gradients between the snow and the air.

Heat conduction into the soil can be calculated using Fourier's Law or it can be measured by a heat flux plate. The heat flux at the soil surface (Figure 4-8) as measured by a heat flux plate, largely mirrors the net radiation (Figure 10-6). The daily variation of the surface heat flux is completely damped out in winter due to the insulative properties of the snowpack and low variation in energy input from radiation. The daily amplitude and instantaneous magnitude of heat flux decreases with depth in the soil (Figure 4-9). Each spring and autumn, the measured heat flux approaches zero as the soil progresses through phase change. The amount of heat transfer at the surface is undermeasured by a heat flux plate during phase change due to its construction. A heat flux plate requires a thermal gradient to register an energy flux. As the soil water freezes in September, energy is drawn from the warm soil to the colder air and the soil temperature remains at 0°C as latent heat is released nearly balancing the heat lost at the surface. Even though energy is transferred across the boundary, the small thermal gradient prevents its measurement by a heat flux plate.

Collectively, each of the above energy transfer mechanisms determine the character of the surface energy balance. The composition and magnitude of the summer surface energy balances calculated in weekly increments for 1987, 1988 and 1989 are presented in Figures 10-7, 10-8 and 10-9. Solar radiation is the primary source of energy during the summer months, especially after the spring snowmelt period. During spring melt, the effect of convective heat transfer is especially important to complement the radiative heat transfer. The contribution of energy advected over the Brooks Range and transferred via longwave radiation or convective heat transfer is of primary importance in determining the initiation and rate of snowmelt. The fundamental sinks of energy are warming of the snow and soil, snowmelt, evaporation, and thawing of the active layer. In the summers of 1987, 1988, and 1989, ET consumed the

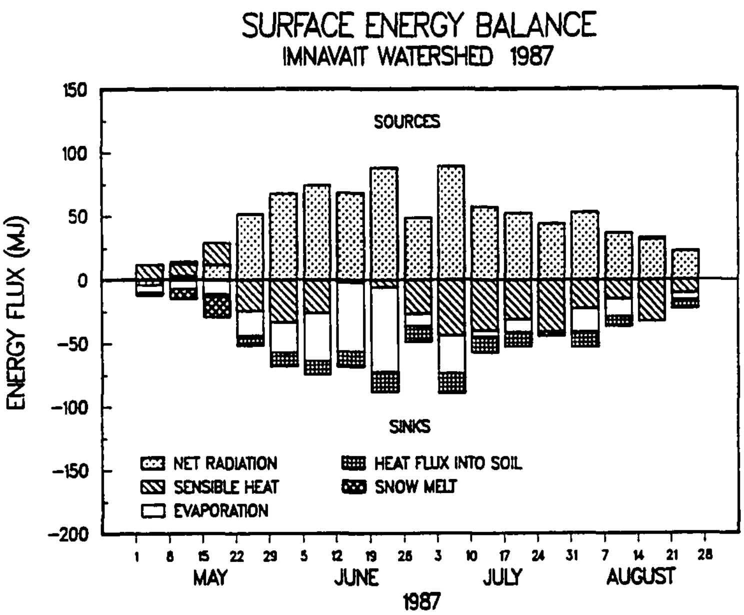

Figure 10-7. Energy balance calculated weekly at the surface for the summer of 1987.

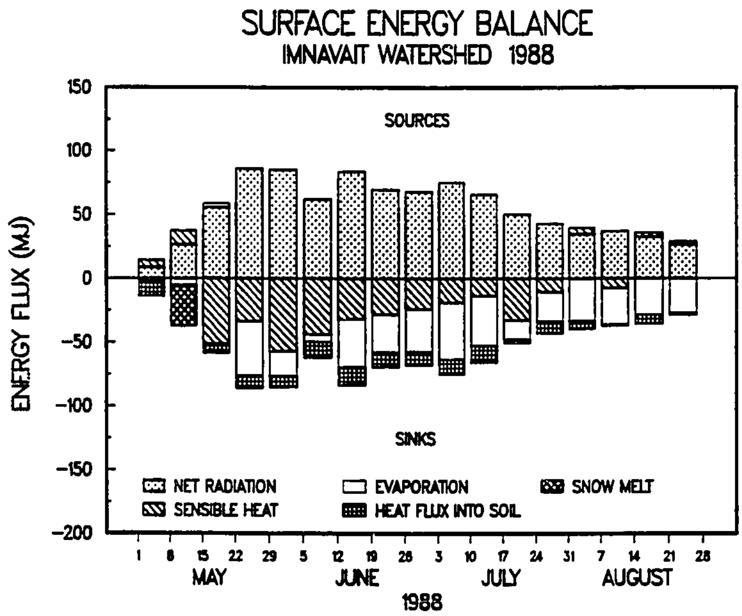

Figure 10-8. Energy balance calculated weekly at the surface for the summer of 1988.

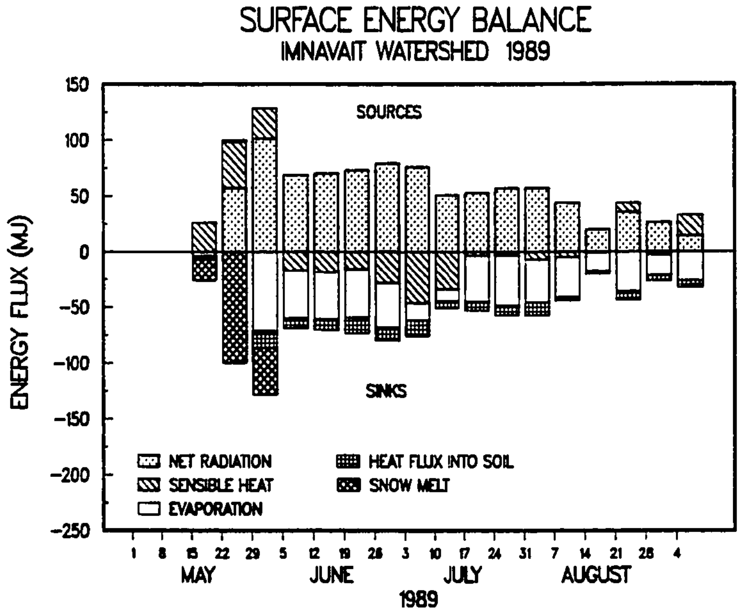

Figure 10-9. Energy balance calculated weekly at the surface for the summer of 1989.

equivalent of about 39%, 46% and 65% of the net radiation respectively. As mentioned previously, net radiation is also a loss term in the winter energy balance. During mid summer, the surface is frequently warmer than the air due to radiative heating and sensible heat is lost by convection from the warmer surface to the cooler air.

The thermal regime of the arctic biome is largely manifested in the soil temperatures, though extremes and rates of heat transfer are not obvious. A plot of soil temperatures from 1985 through 1988 is presented in Figure 10-10. There are several distinguishing features which are present in each year of data. Although air temperatures normally sustain their annual minimum in January or February, the annual minimum soil temperature occurs in late March or April. The soil experiences rapid warming of 6 or 7°C within a few days in late May or June when solar radiation and soil heat fluxes are near their annual maximum. The daily and hourly variability of each layer is greatest in the summer and also decreases with depth. In September and October, the soil begins to freeze, passing through a period of isothermal conditions. The amount of time required to freeze the active layer is greater than the time required to thaw that depth in the spring due to lower rates of heat transfer at the surface. In October, the

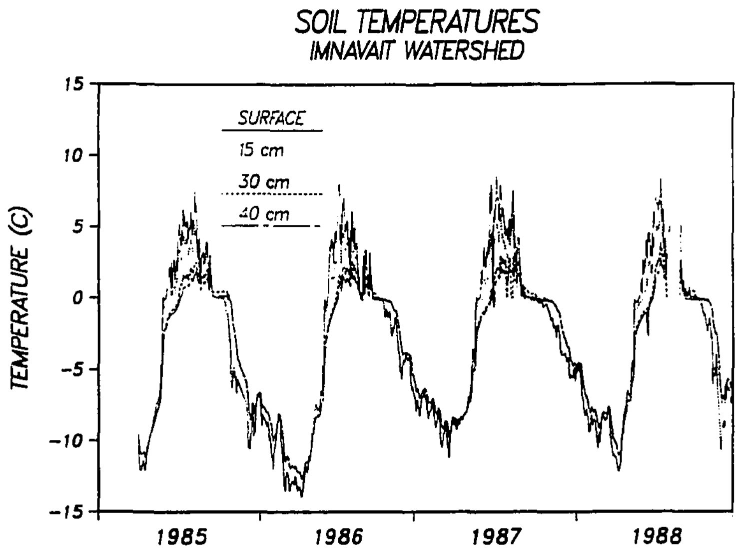

Figure 10-10. Soil temperatures measured at several depths from 1985-1988.

amount of incoming radiation is much lower than the amount incoming during spring thaw but the net radiation balance is still positive. The primary reason the rates of heat transfer are lower is due to the presence and function of autumn snow fall. Early season snow will usually melt soon after alighting upon the surface drawing energy from the warmer soil surface to melt the snow. As the surface quickly cools to 0°C, and the snowpack begins to accumulate, the rate of heat loss slows as the snow again acts as insulation.

The snow temperatures (Figure 10-11) graphically display the length of the accumulation season and large thermal gradients which exist in the snowpack due to its low thermal conductivity. At the end of the accumulation season, snow depths can range from a few centimeters on wind swept ridgetops to over a meter on ponds in the valley bottom; an average depth over the watershed is about 50 cm. The depth, density and distribution of the snow is largely a function of the wind and topography. In this region the winds are primarily katabatic which are the result of downslope drainage of more dense air from the Brooks Range to the south (Figures 10-12 a-l). However large wind events can originate from any direction causing extensive drifts and hard wind slabs throughout the watershed. The snow tends to accumulate in the valley bottom, in water tracks, and on the lee

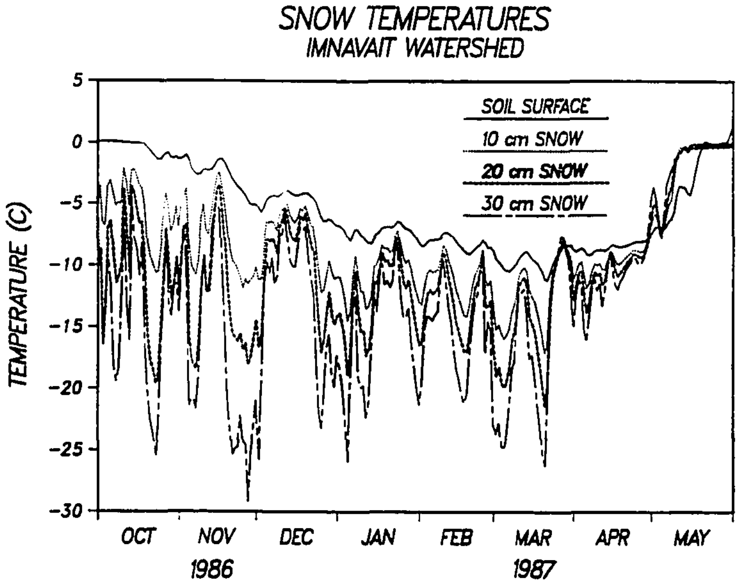

Figure 10-11. Temperatures measured in snow and at the soil surface during the 1986/1987 winter.

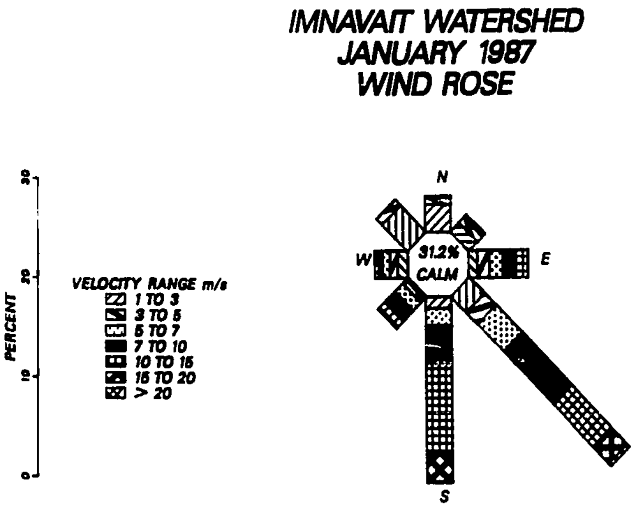

Figure 10-12a. Wind rose showing January distribution of wind direction and magnitude.

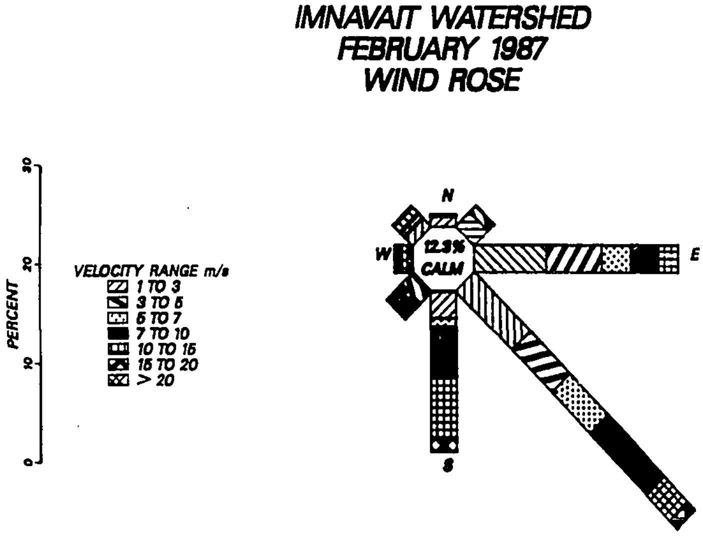

Figure 10-12b. Wind rose showing February distribution of wind direction and magnitude.

IMNAVAIT WATERSHED
MARCH 1987
WIND ROSE

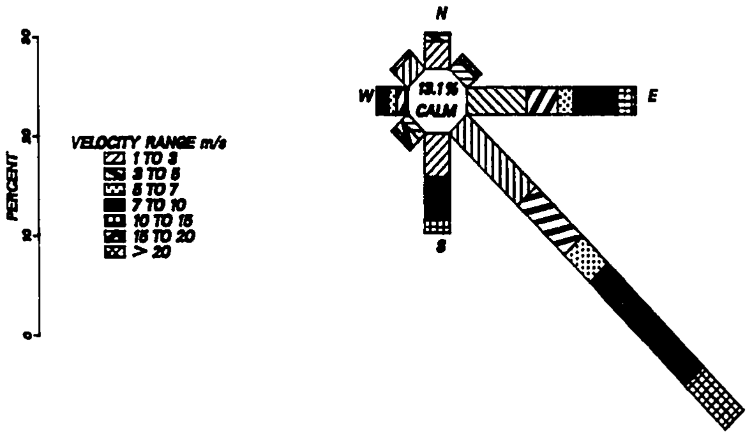

Figure 10-12c. Wind rose showing March distribution of wind direction and magnitude.

IMNAVAIT WATERSHED
APRIL 1987
WIND ROSE

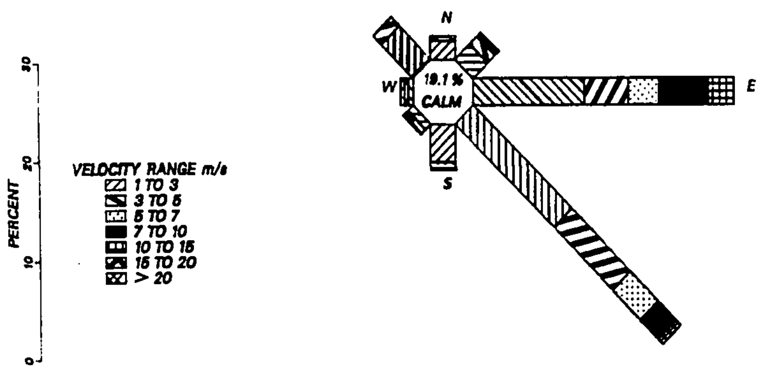

Figure 10-12d. Wind rose showing April distribution of wind direction and magnitude.

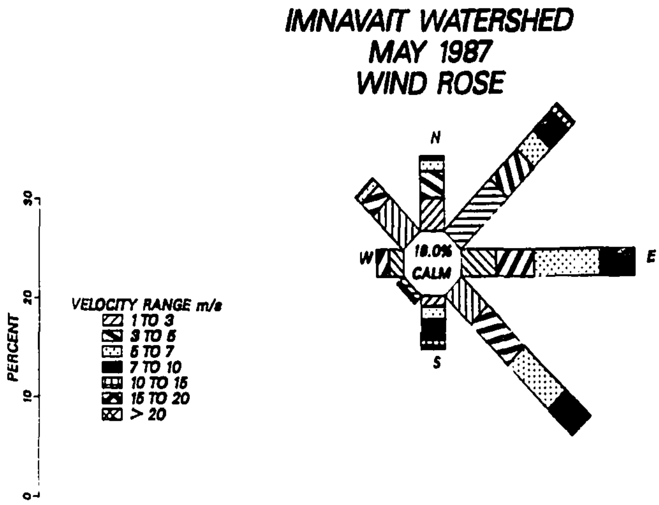

Figure 10-12e. Wind rose showing May distribution of wind direction and magnitude.

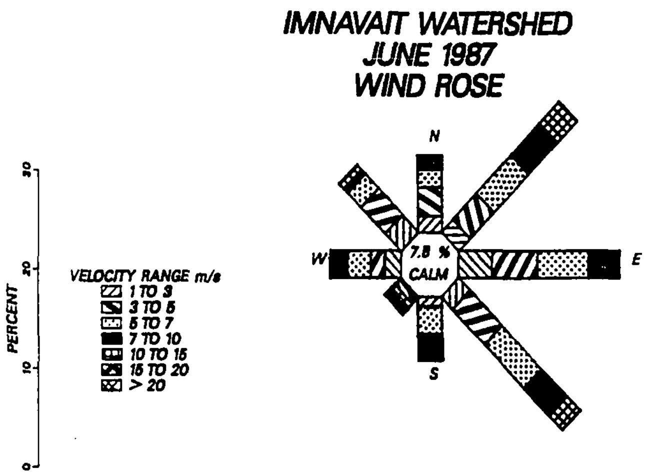

Figure 10-12f. Wind rose showing June distribution of wind direction and magnitude.

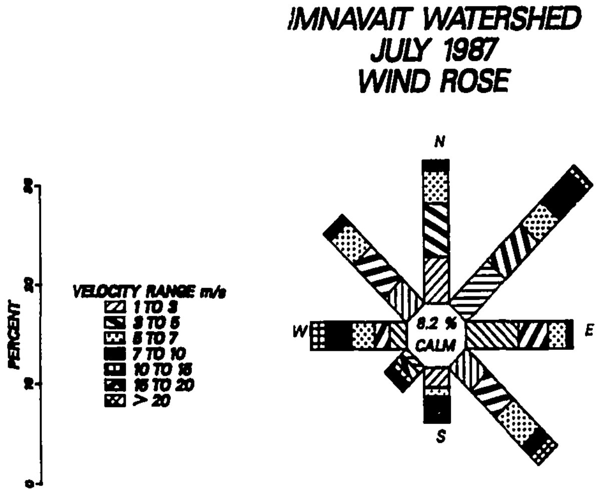

Figure 10-12g. Wind rose showing July distribution of wind direction and magnitude.

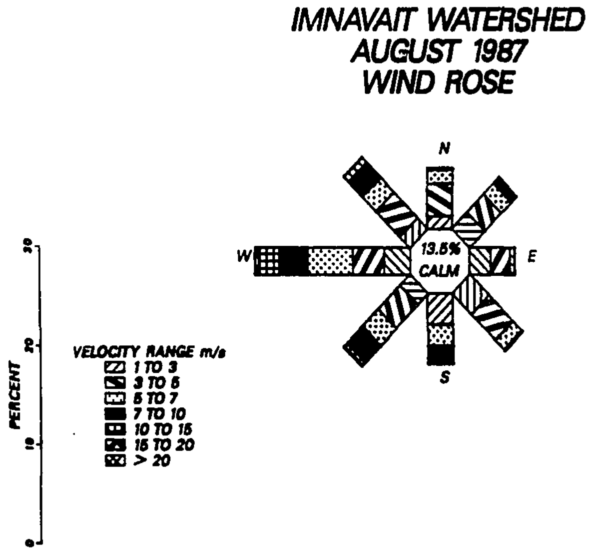

Figure 10-12h. Wind rose showing August distribution of wind direction and magnitude.

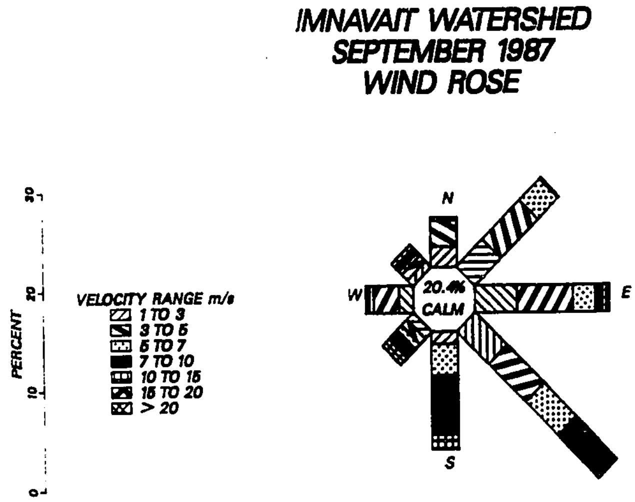

Figure 10-12i. Wind rose showing September distribution of wind direction and magnitude.

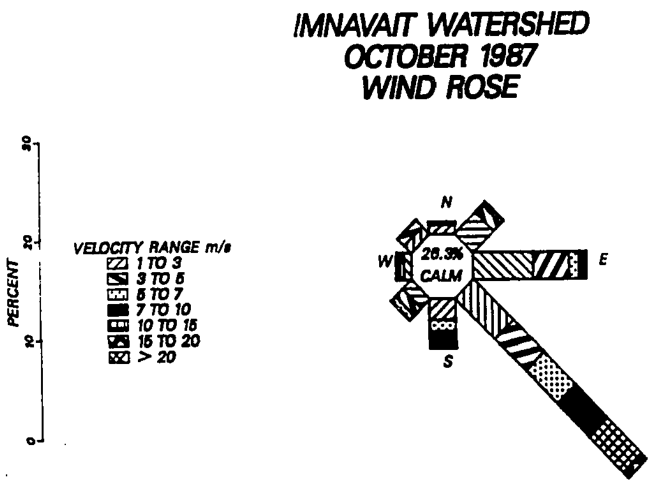

Figure 10-12j. Wind rose showing October distribution of wind direction and magnitude.

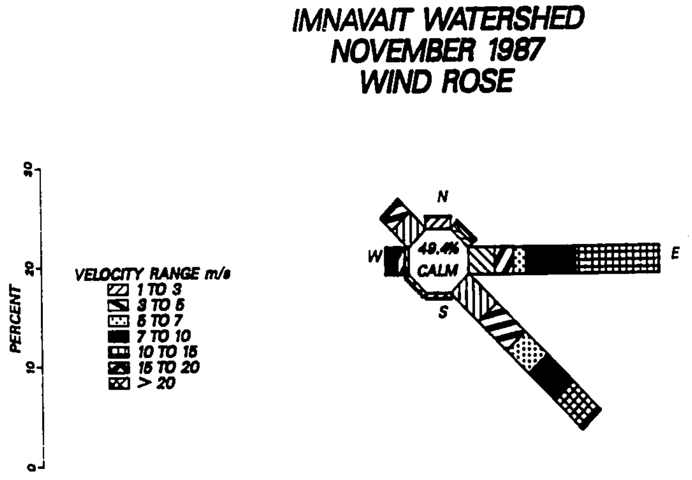

Figure 10-12k. Wind rose showing November distribution of wind direction and magnitude.

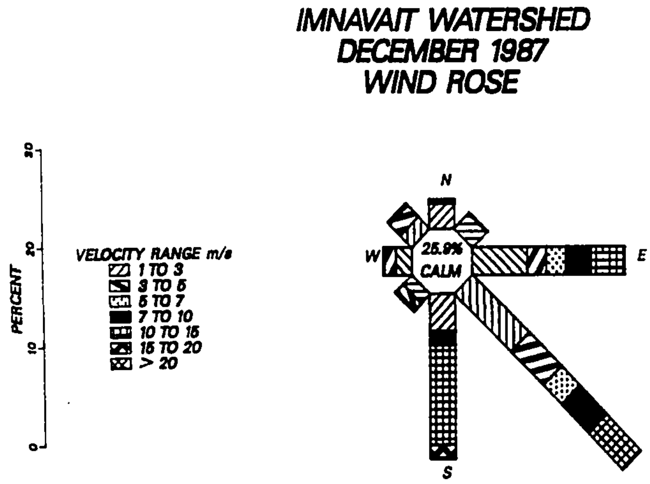

Figure 10-12l. Wind rose showing December distribution of wind direction and magnitude.

slopes of ridges as dictated by wind patterns (Figure 7-2). The direction and force of the larger wind events varies from year to year, but consistency of the predominant wind yields similar snow distribution each year. The variable density of the snow slabs will influence the spring runoff in a manner which will be discussed later. Liston (1986) studied the characteristics of the snowpack and the distribution of snow across Imnavait watershed and related them to the wind patterns and morphology of the watershed.

Greater depths of the snowpack strongly influences the thermal regime of the underlying soil by increasing the thermal resistance to heat flow. This has significance in moderating the extremely cold temperatures and establishing certain locations which characteristically develop deeper snowpacks and usually do not experience the severe cold. This will impact the biota which may survive the arctic winter under a deep snowpack and consequently may influence where certain plants grow or where small rodents choose to hibernate. In turn, taller and denser plants which grow in these protected areas may trap more snow.

Figure 7-2 relates much information about the snow distribution and variability, but the data can yield more information when it is displayed in tabular form

(Table 10-1). The data in Table 10-1 were collected with an Adirondack snow sampling tube. In 1985 and 1989, the samples were collected in 10 meter intervals and in 1986 and 1987, the data were collected in 20 meter intervals. The snow sample transect was not completed in 1988 due to the very early and rapid melt. N indicates the total number of samples.

Table 10-1. Statistical description of the snowpack water equivalent in 1985, 1986, 1987, and 1989.

Year	Mean	St Dev	Median	Min	Max	N
	(cm)	(cm)	(cm)	(cm)	(cm)	#
1985	9.3	3.7	9.4	0.0	17.3	94
1986	9.9	3.3	9.8	3.0	16.0	47
1987	8.8	3.3	8.1	3.6	21.6	47
1989	15.7	4.6	15.5	7.1	30.5	92

The onset of spring snow melt varies greatly depending upon the relative depth of the snowpack and current meteorologic conditions. The date of initial melt has ranged from May 8 through June 1, with the average length of time required for complete ablation being only 10 days. The daily surface energy balance during the snowmelt periods are shown in Figures 10-13 to 10-15.

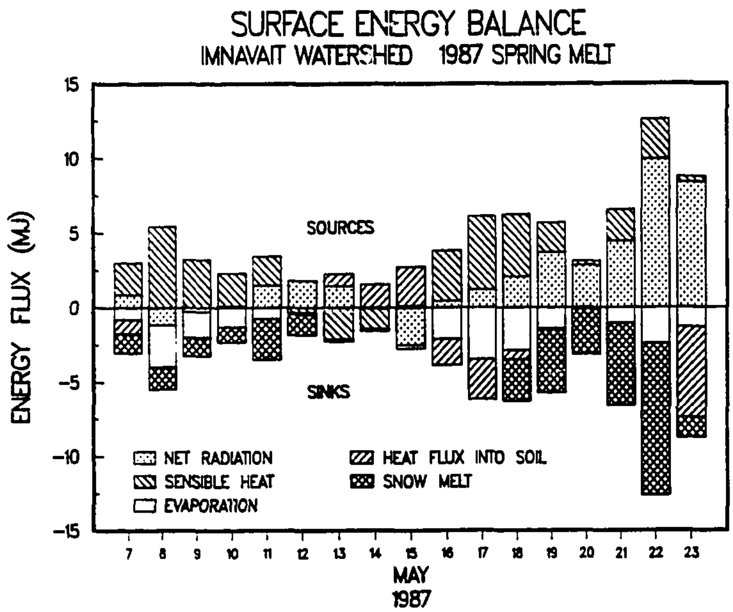

Figure 10-13. Energy balance calculated daily at the surface during the spring snowmelt in 1987.

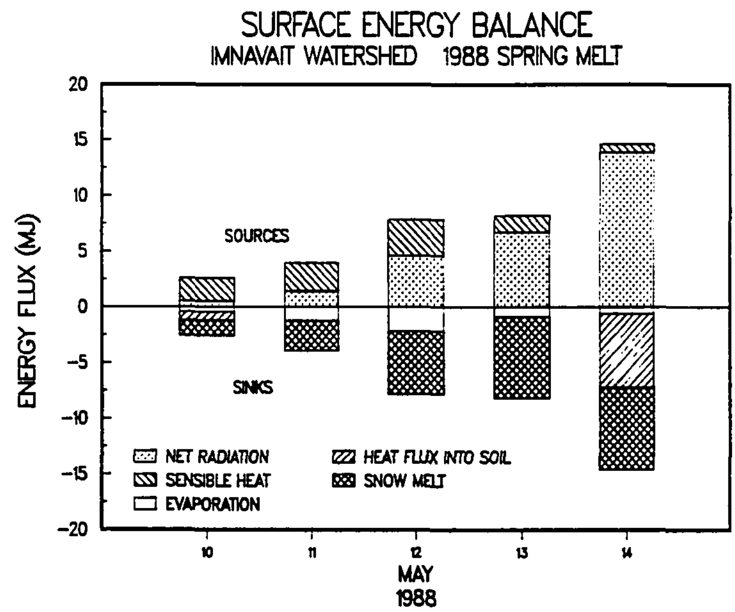

Figure 10-14. Energy balance calculated daily at the surface during the spring snowmelt in 1988.

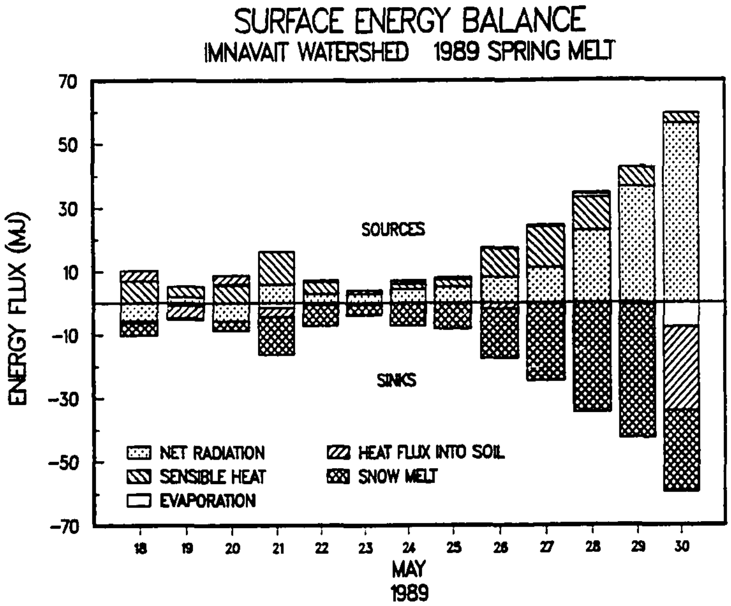

Figure 10-15. Energy balance calculated daily at the surface during the spring snowmelt in 1989.

Net radiation is again the dominant component, but convective heat transfer can be substantial and often triggers the melt initiation. Even though incoming shortwave radiation is very high, near the annual maximum, no melt will occur until the convective heat transfer becomes positive, complementing the radiative flux. The mechanisms of snowmelt are very complicated in Imnavait watershed due to the extreme variability in snowpack thickness and uneven surface topography. Within a few days of sustained melt, the entire watershed becomes a patchwork of snowcovered and bare tundra. Liston (1986) measured the radiative temperatures of a bare tundra surfaces surrounded by snow and found the typical temperatures of the tundra ranging from 15 to 42°C, while the bordering snow was isothermal at 0°C. Complex patterns of stable and unstable air will arise due to wildly varying thermal gradients. Longwave radiation emitted from snowfree ground will warm the overlying air and turbulent transfer will act to accelerate the melt of the surrounding snow. The west facing slope melts sooner than the rest of the watershed because it retains less snow than the east facing slope and valley bottom (Figure 7-2) and also receives more direct solar radiation in the afternoon when melting is at its peak.

The surface energy balance is the only physically correct method of calculating snowmelt at a point (Figures 10-13 to 10-15), and it works well with some restrictions, usually data limitations. Many other methods have been developed due to the intricacies of measuring and calculating all of the interactive mechanisms of heat transfer and the intense data requirements. These will be discussed in more detail later. The ultimate problem in modeling snowmelt from a surface energy balance for Imnavait watershed is the great spatial variability in snow depth, which affects albedo, longwave emittance and sensible heat. This in turn affects the energy balance and consequently affects the melt rate. The energy balance as presented in Figures 10-7 to 10-9 and 10-13 to 10-15 were calculated from instrumentation on the west facing slope. The measured and calculated snowpack ablation (Figure 10-16) also represent processes occurring only on the west facing slope. A truly comprehensive formulation of watershed snowmelt runoff would entail measuring the surface energy balance at many points throughout the watershed and compiling the results, but this is not possible with available resources.

E) Hydrologic Regime of Imnavait Watershed

Hydrology in the Arctic exemplifies the accepted hydrologic principles developed in more temperate

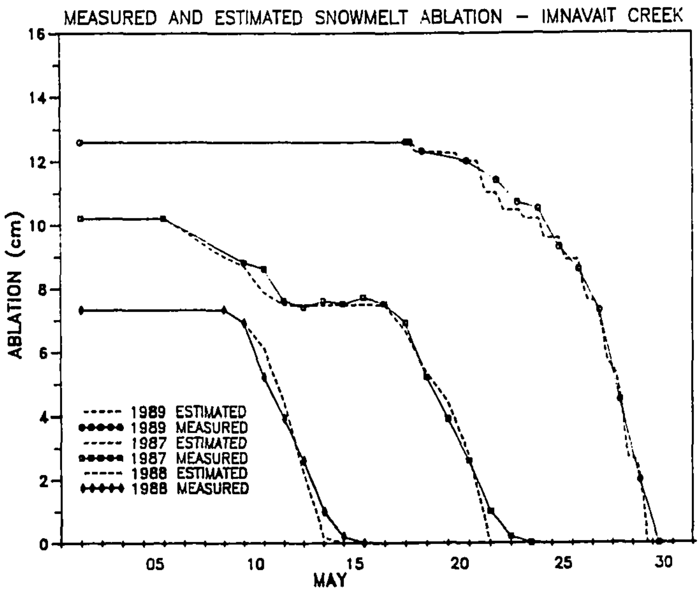

Figure 10-16. Snowmelt, observed and calculated from an energy balance.

regions, however the extreme environmental differences cause marked differences in response to various rainfall or snowmelt events (Woo, 1986). The permafrost effectively eliminates the role of the deep subsurface system in the hydrologic cycle. The ice-rich mineral soils at the permafrost table act as a barrier to prevent infiltration from snowmelt or summer rains and thus the contribution to base flow from below the permafrost table at Imnavait Creek is zero. Therefore, the presence of permafrost actually simplifies the hydrologic system. Although deep springs do provide water for base flow throughout the year in some places on the North Slope of Alaska, Imnavait Creek is isolated from a subsurface water source. Since none of the annual water balance is lost as groundwater recharge, all water leaving the basin is either through near-surface runoff or evapotranspiration.

The highly variable nature of many of the features of the arctic biome complicates quantitative analysis of its hydrology. Snow distribution is extremely variable both across the watershed and between different years (Liston, 1986). The snow tends to be deposited in valley bottoms and on the lee side of slopes (Figure 7-2), however the orientation of the wind slabs is dependent upon the direction of the strong winds and the density of the drift is dependent upon the magnitude of wind event. The distribution and density of the snowpack will strongly affect the runoff processes in several ways. First, due to the formation of snowdrifts, the snowpack water content can vary by a factor of 2 to 3 within a spatial distance of a few meters. The result will be a fast melt in the areas of thin snowpack, and the development of bare patches with considerable edge effect around the drifts during melting. The second consequence of snow redistribution in the valley bottom is the mechanism of snow damming. A thick snowpack of high density will function as a dam, holding back the initiation of stream flow until the force of the water in the snow can overcome the bonding strength of the snow. Another consequence of snow redistribution is due to the accumulation near stream channels and water tracks yielding a higher proportion of runoff and less evaporation than would occur if the snow were uniformly distributed.

Runoff plots were installed at four places along a transect of the slope to determine the effects of position (Figure 3-2). The west facing slope frequently had large differences in the snowpack as one proceeded down the slope (Figure 7-2). The variable snowpack strongly affected the runoff processes as shown in Figures 10-17 to 10-21. Plots with a thin snowpack

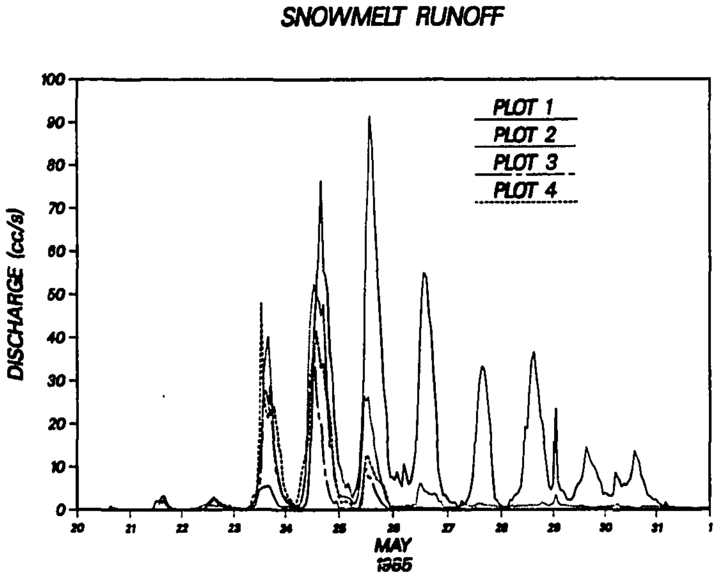

Figure 10-17. Spring snowmelt hydrographs from runoff plots in 1985.

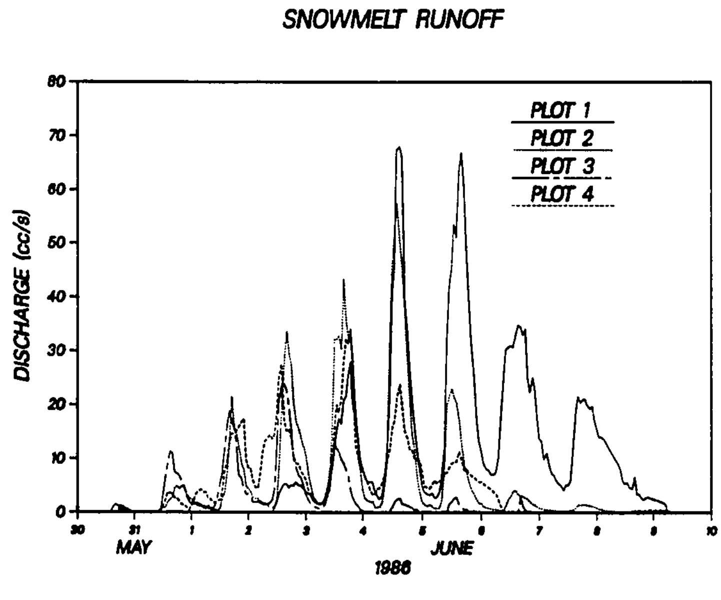

Figure 10-18. Spring snowmelt hydrographs from runoff plots in 1986.

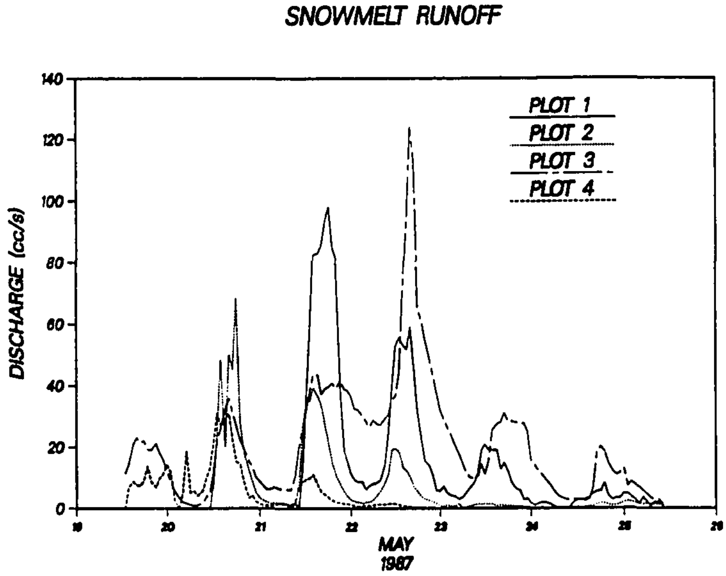

Figure 10-19. Spring snowmelt hydrographs from runoff plots in 1987.

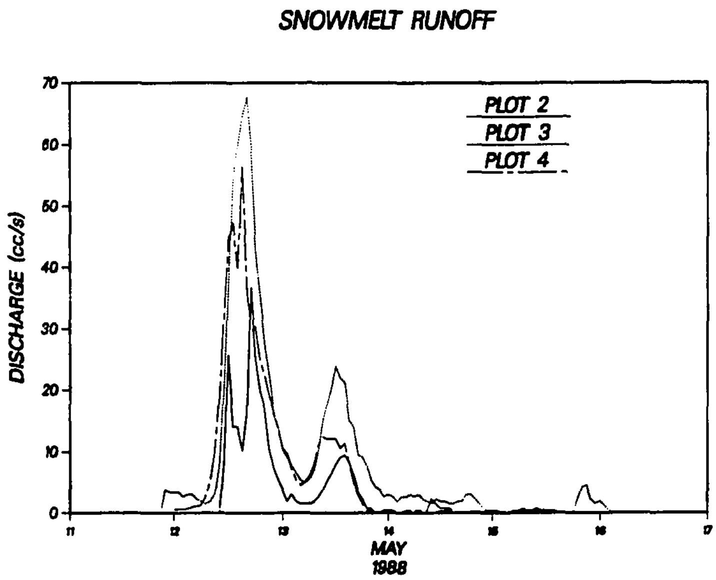

Figure 10-20. Spring snowmelt hydrographs from runoff plots in 1988.

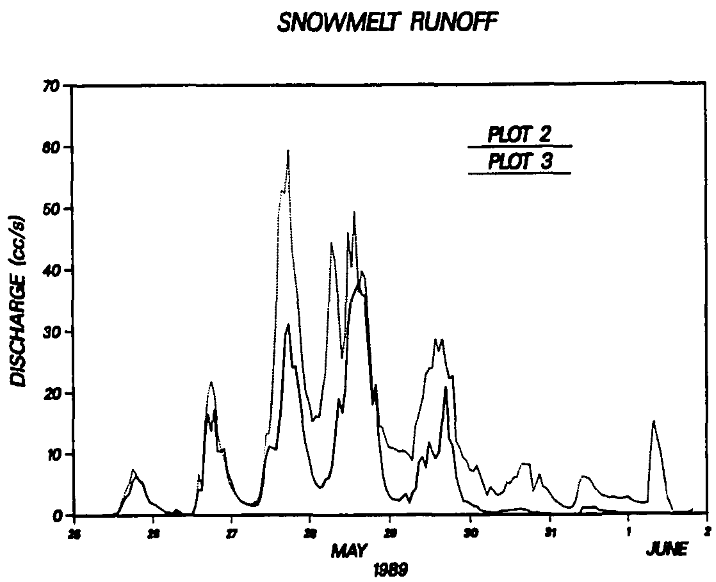

Figure 10-21. Spring snowmelt hydrographs from runoff plots in 1989.

started sooner, probably due to a lower albedo, lower cold content in snowpacks of lesser depth, greater shortwave absorbance by underlying soil, and longwave emittance from shrubs protruding through the snow surface. Plots with the highest snowpacks had the greatest runoff volumes and the greatest peak flows.

The runoff plots provided valuable information on the apportionment of the spring water balances (Figures 6-6a - c, 10-22 and 10-23). The snowpack of each plot and the basin average was determined prior to snowmelt initiation. Runoff was measured directly. The proportion of moisture going into storage in the desiccated organic layer was determined in the laboratory to be 1.5 cm. This estimate has been confirmed in the field (Kane and Hinzman, 1988). Evaporation was calculated as the remainder term in the water balance equation:

$$E = P - R - dS \tag{7}$$

where E = evaporation
P = precipitation or snowmelt
R = runoff
dS = change in soil moisture storage

The initial snowpack on each plot ranged between 7.2 and 14.7 cm and the basin average ranged between 7.8 and 15.5 cm. The broad range was very helpful in

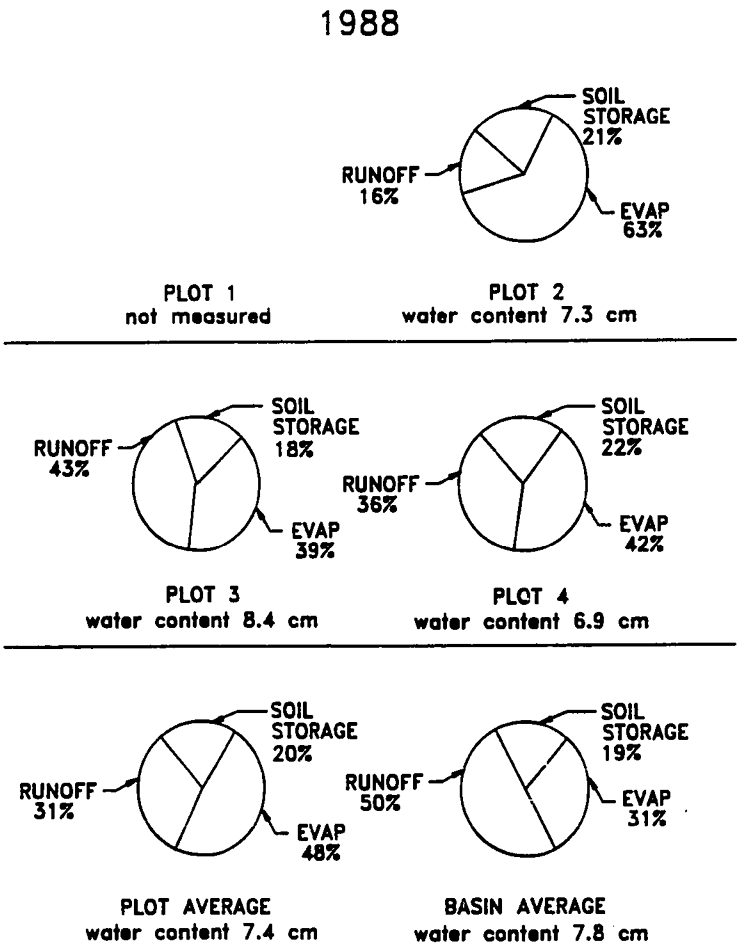

Figure 10-22. Charts illustrating the partitioning of snowmelt into runoff, storage, and evaporation in 1988.

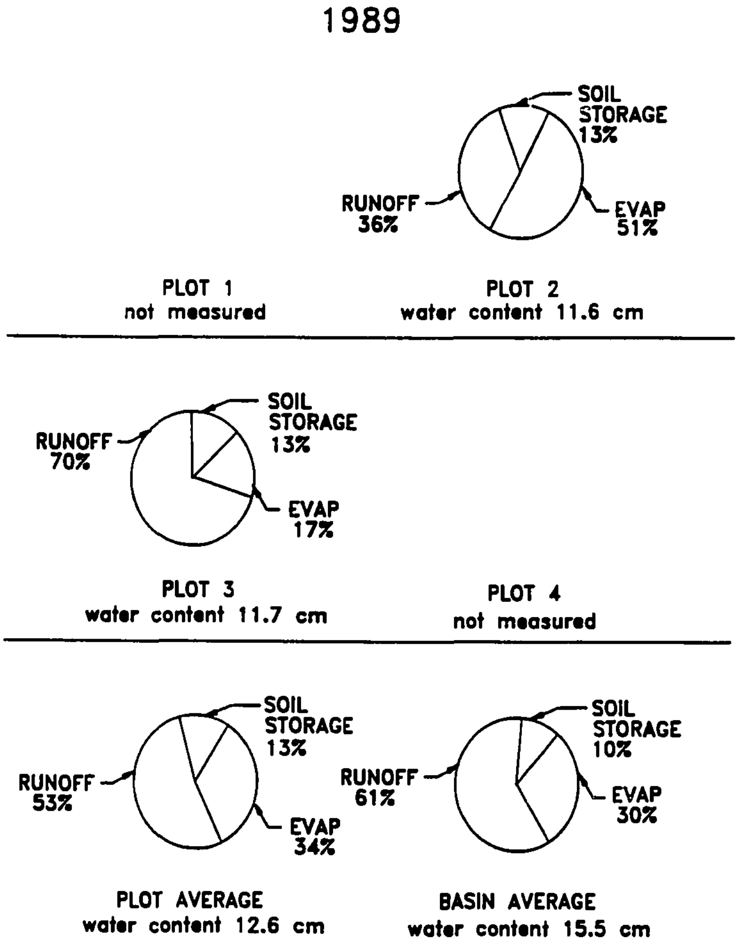

Figure 10-23. Charts illustrating the partitioning of snowmelt into runoff, storage, and evaporation in 1989.

delineating important processes. Evaporation was highly variable depending upon the initial snowpack water content, proportionally increasing as the snowpack decreased. On the basin scale, the amount of runoff was dependent primarily upon the amount of snow, but the distribution of the snow was important in the determining the fraction going into runoff or evaporation. Examination of Figure 7-2 and Figures 6-6 a - c, 10-22 and 10-23 reveal when more snow was deposited in the valley bottom, near the stream, runoff was greater and evaporation was lower. The processes associated with spring melt and runoff are discussed in more detail in Kane et al., 1989 and Kane et al., 1990a.

The intensity of the arctic snowmelt is obvious from the rates of runoff recorded (Figures 10-17 to 10-21). The lag time required in downslope migration of meltwater is quite small in these soils. The process of downslope movement of water is as variable as the soil surface. During the spring melt event, most downslope movement of water occurs in the top 10 cm of the organic layer (Hinzman et al., 1990). A plot of unfrozen moisture content (Figure 4-1) clearly shows that the moisture content at 5 cm increases abruptly as melt water begins to infiltrate, but the moisture content below this layer did not substantially increase until the snowmelt event was complete and the depth of thaw proceeded below 10 cm. As the surficial organic layer becomes saturated, an intricate pattern of inter-tussock flow begins. Flow will revert from surface to subsurface depending upon the soil conditions, topography and depth of thaw. The highly porous moss layer and well developed inter-tussock channels allow rapid downslope movement. The hillside is drained by small water tracks which are nearly parallel to each other approximately 50 m apart. The channels are small and somewhat nebulous during the dry season, but their effect is to drain the excess water from the hillslope quickly, resulting in a rapid runoff response to precipitation events (Kane et al., 1990a).

As the melt season progresses and the depth of thaw increases, the amount of available moisture and moisture storage increases. In effect, the parameters describing the runoff processes and soil moisture properties are time dependent and must be adjusted accordingly. This has significant implications in modeling the hydrologic regime, especially if one considers the effects of disturbance. As the summer continues, the ice-rich soil melts releasing excess water and keeping the mineral soil near saturation throughout the year (Figure 4-1). The moisture content of the organic soil varies greatly throughout the year. Prior to spring melt, the organic soil is quite desiccated and will absorb up to 1.5 cm of meltwater before downslope runoff can occur (Kane and Hinzman, 1988). The moisture content of the surface organic soils vary by up to 60% by volume throughout the summer (Figure 4-3). The organic soils saturate and dry quickly in response to summer rains or snowmelt events, however the subsurface soils remain saturated and the moisture content is unaffected by precipitation events on the surface (Hinzman et al., 1990).

The drainage is a beaded ephemeral stream which flows from spring melt until freeze-up in September. The spring runoff event is usually the dominant hydrologic event of the year (Kane and Hinzman, 1988); an event that produces the annual peak flow and about 40% of the total annual runoff volume. Streamflow essentially ceases after extended periods of low precipitation which may occur in late May or June (Figures 6-1a and 6-1b). Precipitation normally occurs as a slow drizzle although very localized convective storms can yield high rainfall intensities for short time periods. July and August are normally the months of greatest precipitation. The average annual precipitation is only about 32 cm, two thirds of which falls in late June, July and August. The cumulative daily precipitation and the average monthly total are shown in Figure 10-24; this plot represents the average of seven years data. Although

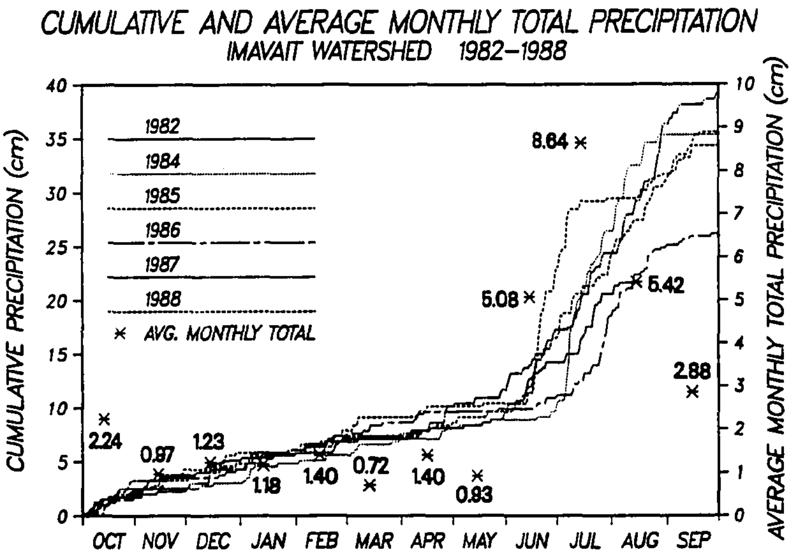

Figure 10-24. Cumulative daily precipitation measured by a Wyoming snow gage from 1981 - 1989, with average monthly total precipitation shown.

there is an expected variability, the characteristic distribution of precipitation centers around very wet autumns and very dry conditions during the rest of the year. This can probably be attributed to the presence of the polar icepack which covers the primary source of moisture. The period of the most open water occurs in late July and August (Labelle et al., 1983) and this corresponds nearly with the time period of greatest precipitation. Moisture coming from the open Pacific Ocean to the south must cross two mountain ranges and about 800 miles before arriving on the North Slope. Although this does occur, it is usually not a major path of precipitation. Alternatively during the ice-free months, moisture may flow east from the Bering Sea and only cross the Brooks Range.

As previously discussed, evaporation is quite low during the winter months increasing abruptly after snowmelt (Figures 10-25 and 10-26). Actual evapotranspiration is usually between the amounts of pan evaporation and precipitation. Actual evapotranspiration is probably close to pan evaporation immediately following snowmelt due to the high soil moisture contents, high amounts of incoming solar radiation and the low atmospheric relative humidity; evaporation is probably occurring at a rate near its potential. As the summer continues, and precipitation increases, actual evaporation will proceed

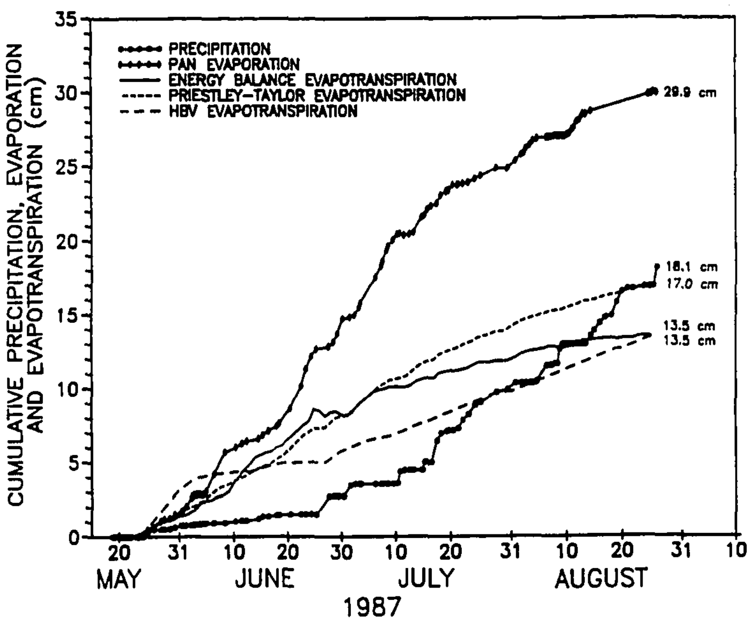

Figure 10-25. Cumulative pan evaporation, precipitation and calculated evapotranspiration during the summer of 1987.

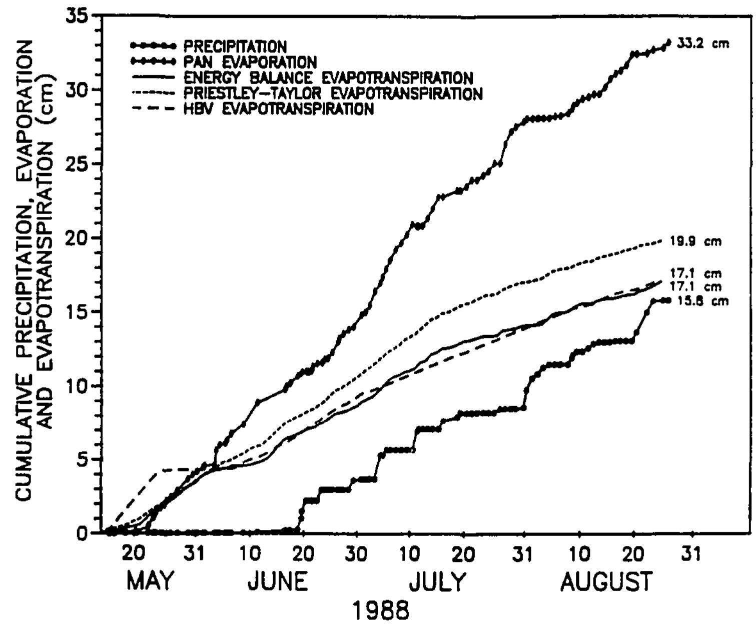

Figure 10-26. Cumulative pan evaporation, precipitation and calculated evapotranspiration during the summer of 1988.

at a rate less than its potential. Evapotranspiration estimated by the energy balance method and the Priestley Taylor equation compare quite well with estimates determined from the water balance method (Table 10-2).

Table 10-2. Spring and annual water balances.

Year	Snowpack Water (cm)	Summer Precip (cm)	Total Precip (cm)	Snowmelt Runoff (cm)	Summer Runoff (cm)	Total Runoff (cm)	Evapo-Trans (cm)	Pan Evap (cm)
1985	10.2	25.1	35.3	6.6	*	*	*	*
1986	10.9	16.3	27.2	5.7	6.2	11.9	15.3	31.0
1987	10.8	27.2	38.0	7.1	17.9	25.0	13.0	32.0
1988	7.8	25.2	33.0	3.9	7.2	11.1	21.9	33.0
1989	15.5	25.7	41.2	9.4	7.8	17.2	24.0	42.0

* - data unavailable

F) Simulation of the Thermal Regime

To fully understand the current thermal regime, it was necessary to develop a procedure to accurately model the present thermal conditions utilizing data collected over a four year period in Imnavait watershed. Soil samples were collected at the site and their material properties were measured in the laboratory (Hinzman et al., 1990). Some thermal properties were calculated from the soils' material properties according to the method described in Zarling et al. (1989). Thermal conductivity was measured in the laboratory for different soil types and at different moisture contents (Hinzman et al., 1990). Using data collected in a deep well 15 km southwest of

Imnavait watershed, it was possible to determine the depth of the permafrost and the geothermal gradient (Osterkamp, personal comm.).

The TDHC model is a two dimensional, heat conduction model which incorporates the effects of phase change through a Dirac delta function (Goering and Zarling, 1985). This model is based on the solution of the following equation:

$$\delta/\delta x\ [K\ (\delta\theta/\delta x)] + \delta/\delta y\ [K\ (\delta\theta/\delta y)] = C\ (\delta\theta/\delta T)$$

Where;
- x = coordinate in horizontal direction, [m]
- y = coordinate in the vertical direction, [m]
- θ = temperature, [°C]
- C = heat capacity, [$J/(m^3 °C$]
- K = thermal conductivity, [W/(m°C)]
- T = time, [s]

This equation of transient heat flow is solved using a finite element technique known as the Galerkin weighted residual process. The top 30.5 m of soil depth was modeled using a grid with a fixed mesh of 268 linear triangular elements formed from 204 nodes (Figure 9-1). The model was driven by soil surface temperatures and the lower boundary was specified as the geothermal heat flux. This value would gradually decrease as the lower boundary approached phase change and would equal zero when the gradient equaled zero, i.e. when the soil on both sides of the boundary was isothermal. The model has the capability to use the appropriate frozen or thawed thermal property on a portion of an element depending upon the amount of the advance of the freeze front through an element. Use of TDHC for determining active layer thickness as influenced by climatic warming was described in Kane et al. (1990b).

The consequence of neglecting convective heat transfer, i.e. the heat transferred through moving water, will cause the model to underpredict the true thaw depth. In frozen and/or fine grained soils, heat transfer from particle to particle, conduction, is the dominant heat transfer mechanism, however in the surface organic mat, convective heat transfer due to running water can be the primary heat transfer mechanism. This error may increase as the thaw depth increases.

The ability of TDHC to simulate the present soil thermal regime was checked by entering actual soil surface temperatures and comparing simulated subsurface temperatures with measured subsurface temperatures (Figure 9-3). The model performed extremely well during the freezing period. Accuracy always lapsed during the spring snowmelt event, but was still within 1°C of the actual temperature. This is probably due to infiltration of melt water into the soil pores which had been desiccated over the winter.

G) Simulation of the Hydrologic Regime

Before one can simulate a future response to climatic change, one must accurately simulate the present hydrologic system. Data collected over a four year period were used in the analysis. During the winter the main hydrologic activities include snowpack accumulation and redistribution, sublimation and desiccation of the surface soils. In the Arctic, there are no significant midwinter thaw events. Data from the summer season, including spring melt, were analyzed using 1 day time steps from 1986 to 1989. The winter processes were accounted for as initial conditions.

The HBV model (Bergström, 1976) was used to simulate all hydrologic processes. This model was selected because it was a conceptual model based primarily upon hydrologic processes not upon empirical relationships. The model has 41 parameters, most of which were physically based, however some were determined during the calibration procedure and could not be calculated through an analysis of the physical system. The model used 26 of these parameters to describe the basin or the input data, while the rest collectively described snow accumulation and ablation, soil moisture accounting, and generation or transformation of the hydrograph. It calculated runoff continuously using a lumped parameter approach. It also had the capability to forecast runoff given estimates of temperature, precipitation and monthly estimates of ET. The time steps were usually one day, however two hour increments were used with good results (Hinzman and Kane, 1990).

This model is especially well adapted to use in a stratified system such as tundra soils because of the model structure (Figure 8-1). The separate snow routine calculates whether precipitation is in the form of rain or snow and from this determines accumulation or infiltration appropriately. This particular routine is especially important for several reasons; in the Arctic, snowfall can occur on any day of the year, and with the onset of climatic warming, precipitation events which occur as snowfall now may occur as rainfall in the future. Snowmelt is calculated through a degree day formulation. The soil moisture routine considers how much moisture can be held in the soil matrix. As the thickness of the active layer increases, the field capacity of the soil will also increase. The amount of evapotranspiration is also calculated in this routine. As the active layer increases, the most dynamic portion of the model will be the upper and lower zone. Conceptually, the top portion of the upper zone relates to the very porous organic mat. The region below this,

L_{uz} is related to the less porous but still highly organic soils below the organic mat. The lower zone is conceptually related to the frozen soils below the permafrost table. As the active layer thickens, L_{uz} will thicken and become more important in the formation of runoff and in storing soil moisture. Eventually the lower zone may influence the runoff hydrograph.

The summer of 1986 was quite dry and spring snowmelt accounted for about half the total volume of runoff. The three summer events were quite distinct and provided good opportunity to analyze events not complicated by overlapping recession curves. The first two events after the spring melt were caused by rainfall and the third event was a summer snowmelt event. Again, HBV does have an advantage over other similar models in that it does consider the state of summer precipitation, i.e. snow or rain, and determines runoff appropriately.

It proved somewhat difficult to adequately model the 1987 summer events as some precipitation fell as snow or snow combined with rain. Also, some of the storms were immediately followed by other storms and the recession curves for several events overlapped. The summer of 1988 was very wet and was again difficult to analyze due to extensive overlapping of rainfall and snowmelt events. The 1988 snowmelt event was intense and brief because the snowpack was very thin and the temperatures were quite high. Stream runoff rose to a single peak and receded rapidly. Almost in contrast, the following spring snowmelt event in 1989 was completely different. The snowpack was substantially greater than the previous 4 years (Table 10-1). The density in the valley bottom was quite high due to significant wind events and the resultant effect was substantial stream damming. For the first time, the stream dams broke in sections yielding 3 days with peak flows each higher than the previous.

The values of each model parameter and the associated effective dates are listed in Table 10-3. Data collected during the summer of 1986 were the primary calibration data set. Some parameters were changed after later data sets were analyzed. An important objective during analysis was to determine how the model parameters relate to the basin characteristics. In northern basins, where soil frost is an annual occurrence, the hydrologic properties of the active layer vary logically and predictably throughout the year. In a non-permafrost basin, a model must account for changing rates of infiltration as the soil goes from frozen to unfrozen (Sand and Kane, 1986). In a permafrost watershed, the hydrologic properties of the

Table 10-3. Parameters used in HBV model for snowmelt and summer simulations.

			BASIN PARAMETERS
AREA	May 1	2.20	Area of subbasin (km^2)
ELEV	"	910.00	Elevation of each subbasin (m)
VEG	"	1.00	Proportion of subbasin of vegetation type
LAKE	"	0.00	Proportion of subbasin covered by lake
PELEV	"	930.00	Elevation of precipitation station
TELEV	"	930.00	Elevation of temperature station
QFACT	"	1.00	Conversion factor of input data to m^3/s
CP	"	1.00	Station weight for precipitation (areal)
CT	"	1.00	Station weight for temperature (areal)
CQ	"	1.00	Station weight for discharge (temporal)
EVAP	Jan	0.00	Potential evapotranspiration (mm/day)
	Feb	0.00	"
	Mar	0.00	"
	Apr	0.00	"
	May	4.50	"
	Jun	2.50	"
	Jul	1.50	"
	Aug	1.50	"
	Sep	1.50	"
	Oct	0.00	"
	Nov	0.00	"
	Dec	0.00	"
PCORR	May 1	1.00	Correction factor applied to precipitation
PCALT	"	10.00	Increase in precipitation per 100 m (%)
TCALT	"	0.60	Temperature Lapse rate per 100 m (°C)
CEVP	"	1.00	Correction factor applied to EVAP values
LAREA	"	0.00	Area of lake at outlet of subbasin (km^2)
ICEP	"	135.00	Last day with ice on lakes
EVDAY	Not Used		Days for which EVAP data are valid
REPL	"		Station to replace missing data
ELAG	"		Evaporation lag time (days)
QADD	"		Additional flow from another subbasin
QRED	"		Reduction of computed discharge
PATH	"		Routing from subbasin to subbasin
WQREL	"		Stage/discharge relation number
FELEV	"		Elevation of forecasted data
SUBQR	"		Indicates which discharge station/subbasin
			SNOW ROUTINE PARAMETERS
SFCF	"	1.00	Correction factor applied to snowfall
CFMAX	"	3.50	Melting factor (mm/°C)
CFR	"	0.08	Refreezing factor
TT	"	-1.80 to 2.95	Threshold temperature, °C (see text)
TT	After spring melt	1.00	"
			SOIL ROUTINE PARAMETERS
FC	"	40.00	Field Capacity (mm)
FC	June 15	30.00	"
LP	May 1	7.00	Limit for potential evapotranspiration (mm)
BETA	"	2.00	Parameter in soil routine (see Figure 8-1)
BETA	July 30	0.50	"
			RESPONSE FUNCTION PARAMETERS
PERC	May 1	0.00	Percolation from upper to lower zone.
UZL	"	15.00	Limit for high recession, K0, (mm)
UZL	June 15	30.00	"
K0	May 1	0.90	High recession coefficient in upper zone
K1	"	0.30	Lower recession coefficient in upper zone
K2	"	0.00	Recession in lower zone
			TRANSFORMATION FUNCTION PARAMETERS
MAXBAS	"	1.00	Base in transformation function (days)
BLAG	Not Used		Translation of computed hydrograph
DAMP	Not Used		Damping parameter in routing routine.

active layer change between spring melt and summer precipitation events. Four parameters were changed during the summer, the threshold temperature, TT, the field capacity, FC, the upper zone limit, UZL, and BETA, a parameter in the soil moisture routine. If the physical structure of the active layer changes due to climatic warming, these properties to may change proportionally.

The timing of change in hydrologic properties of the active layer is due to the rate of thaw which is a function of meteorologic conditions and can vary from year to year. To make the model easier to apply to future data sets and other basins, parameters were not changed as a function of meteorologic conditions because these data are normally not available. The parameters instead are changed on a specific date, which is the same every year. By varying the effective date of the parameter change from year to year it is possible to improve the model performance, but that would make the model less useful in predicting flows where limited data are available. All the parameters were the same for each data set analyzed except for the threshold temperature, TT. TT is a factor in the snow routine which specifies at what temperature snow will begin to melt. It also functions to control when streamflow will begin and can thus act as some measure of snow damming.

The temperature at which snow will begin to melt, according to a degree day formulation, should be about the same from year to year. However, the effect of snow damming is broadly variable depending upon snow distribution and density in the valley bottom. The significance of TT and effect it has on the modeling process is discussed in great detail in Hinzman and Kane (1990).

Plots of the simulated and actual hydrologic processes are presented in Figures 10-27 to 10-30. In all cases, the model performed well (Table 10-4). For any hydrologic model, accuracy within 25% is considered quite good. HBV not only modeled streamflow well, but also gave good estimates of daily evapotranspiration and soil moisture levels. Not surprisingly, HBV performed best at modeling simple discharge events such as the snowmelt or rainfall events of 1986 (Figure 10-27) or the snowmelt event in 1988 (Figure 10-29). The degree day formulation of snowmelt is not as accurate as the energy balance method presented in Figure 10-16, but the lack of accuracy in calculating snowpack ablation is also due to the fact that the parameters of the model were optimized to best simulate discharge and not ablation (Hinzman and Kane, 1990). Perhaps the model could be improved by developing a routine which would allow

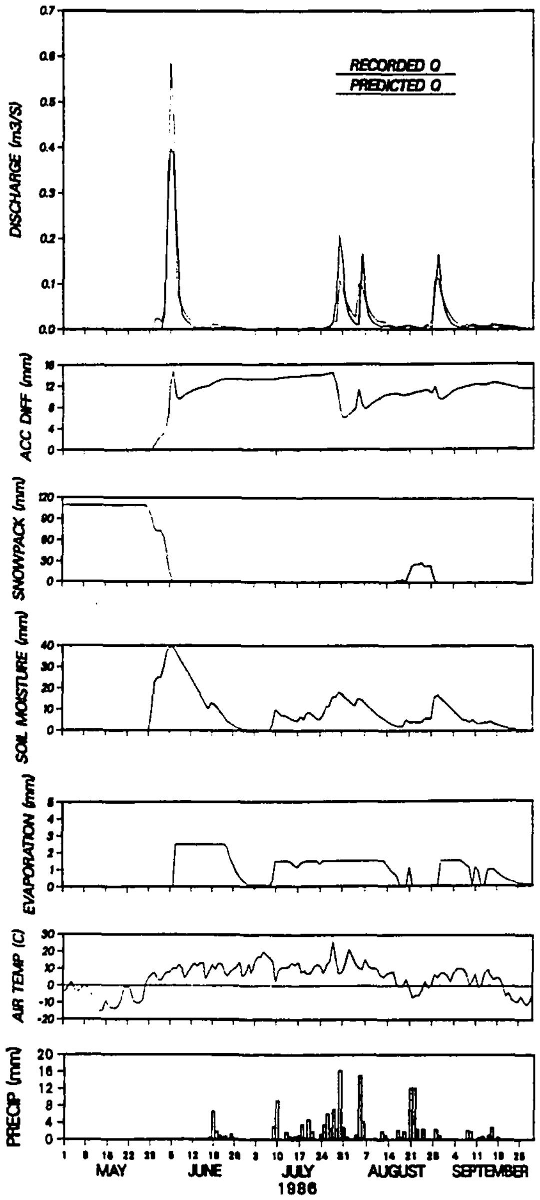

Figure 10-27. Hydrologic processes in 1986, stream flow observed and simulated by HBV model, accumulated difference between observed and simulated flow, predicted snowpack ablation, soil moisture, evapotranspiration, and measured air temperature and precipitation.

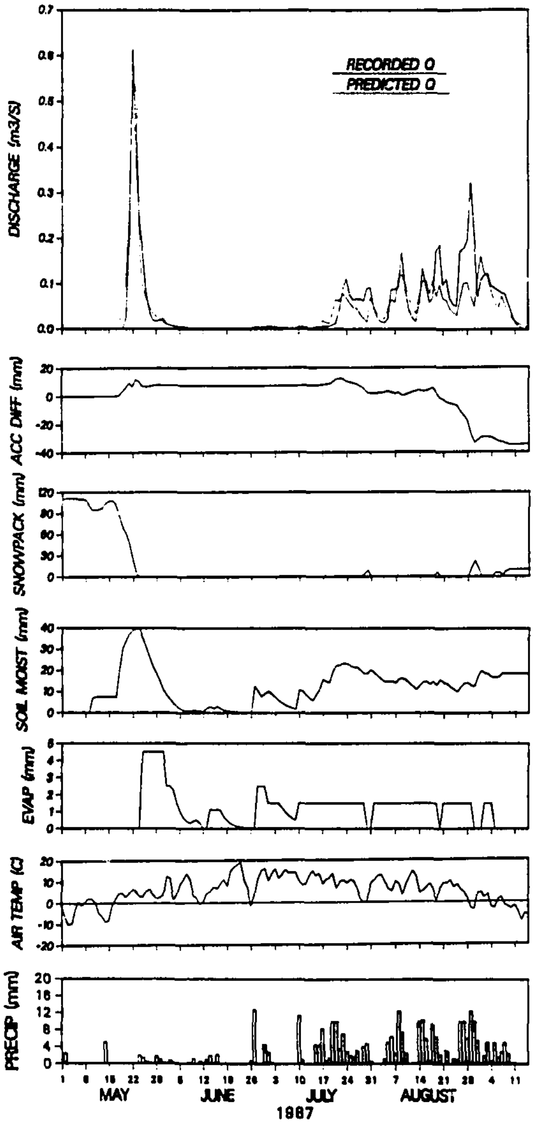

Figure 10-28. Hydrologic processes in 1987, stream flow observed and simulated by HBV model, accumulated difference between observed and simulated flow, predicted snowpack ablation, soil moisture, evapotranspiration, and measured air temperature and precipitation.

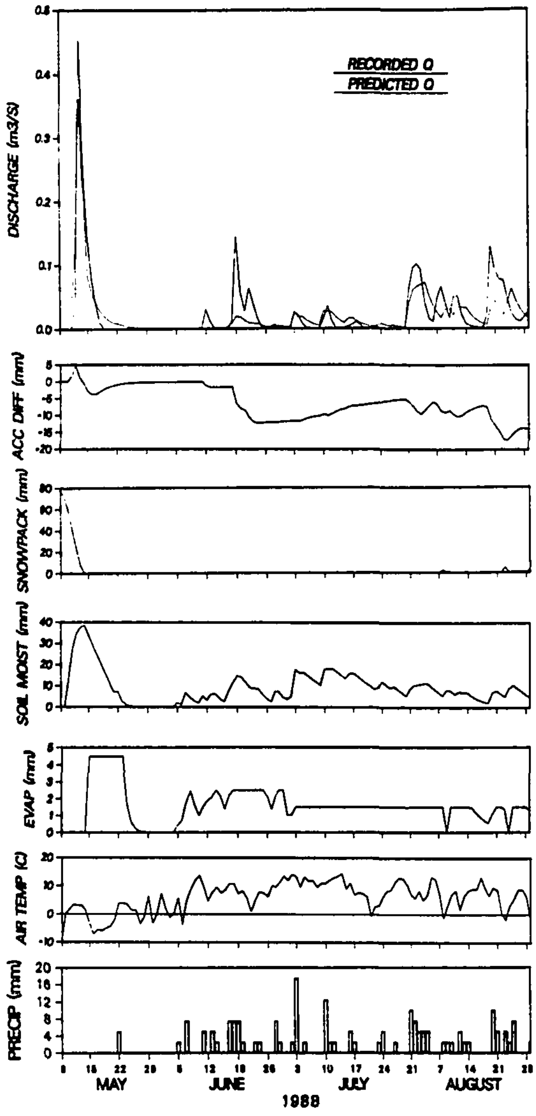

Figure 10-29. Hydrologic processes in 1988, stream flow observed and simulated by HBV model, accumulated difference between observed and simulated flow, predicted snowpack ablation, soil moisture, evapotranspiration, and measured air temperature and precipitation.

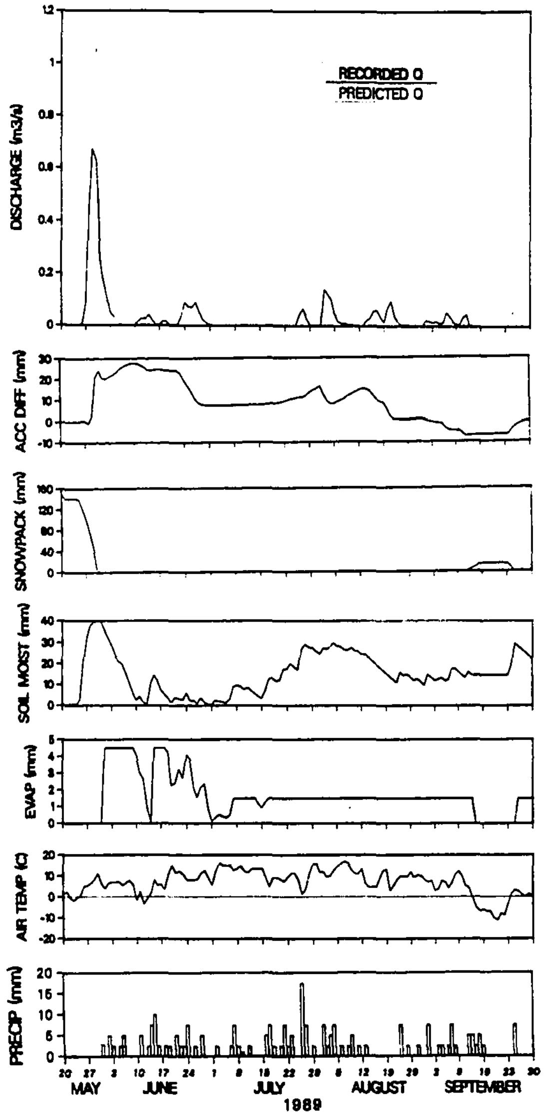

Figure 10-30. Hydrologic processes in 1989, stream flow observed and simulated by HBV model, accumulated difference between observed and simulated flow, predicted snowpack ablation, soil moisture and evapotranspiration, and measured air temperature and precipitation.

snowmelt calculations based upon the energy balance method.

Table 10-4. Performance of model and selected threshold temperature.

Year	r^2	Time Step	Threshold Temperature (°C)
1986	0.82	1 day	2.95
1987	0.77	"	0.15
1988	0.78	"	-1.80
1989	0.70	"	-0.30

H) The Potential Effect of Climatic Warming

International concern about the possibility of global warming has prompted extensive research on the potential response of the environment to the steady increase in atmospheric carbon dioxide (CO_2) and other radiatively active gasses, such as methane (CH_4), ozone (O_3), nitrous oxide (N_2O) and several other synthetic chemicals such as chlorofluorocarbons ($CFCl_3$ or CF_2Cl_2). Most research reports begin with the analogy of these gasses acting as a greenhouse, allowing solar shortwave radiation to pass through, but absorbing infrared radiation emitted from the surface of the earth, trapping some of this energy within the troposphere (Ramanathan, 1988). The net effect is hypothesized to be a warmer, wetter troposphere and a cooler stratosphere. The full consequences of such changes to our environment are still not known. The research completed up to now has been primarily of three types: 1) modeling studies to determine the atmospheric response to increased concentrations of radiatively active gases, 2) field studies to discover evidence that climatic warming is occurring, or 3) cerebral exercises to hypothesize the ramifications of climatic warming.

Global circulation models indicate that with a doubling of CO_2, the entire world will warm, with the greatest warming occurring in the high latitudes (Ramanathan, 1988). Some models indicate that the air temperatures in the Arctic could increase by 3 to 6°C (Schlesinger and Mitchell, 1987). The atmosphere above polar latitudes contains relatively little water vapor, consequently CO_2 is more important in longwave absorption (Etkin, 1989). The existence and potential degradation of the polar icecaps also reinforces a positive feedback mechanism by reducing the surface albedo and absorbing more shortwave radiation. The same mechanism will occur with decreasing snowcover due to shorter winters.

Researchers have known for many years that the concentration of atmospheric CO_2 was increasing but lacked conclusive proof that the consequence of CO_2

build-up would be climatic change (Weller, 1984). Recent evidence now indicates that climatic warming has already begun. Direct evidence comes from analysis of temperature records from around the world. Analysis of data since 1860 indicates a global mean temperature increase of 0.5 to 0.7°C (Hansen and Lebedeff, 1987). Analysis of meteorologic data has indicated that the climate of Alaska has warmed significantly over the past century (Osterkamp and Lachenbruch, 1989). Landscape features which form only in association with permafrost and then according to an upper temperature limit are indicative of climatic change (Washburn, 1979/1980). Probably the most conclusive evidence of climatic warming comes from analysis of deep well temperatures in permafrost. Lachenbruch and Marshall (1986) found a widespread warming of 2 to 4°C which occurred over the last few decades to a century on Alaska's North Slope.

There is good cause for great concern about climatic warming in the Arctic. In northern regions, climatic warming will manifest itself in ablation of permafrost. In areas of continuous permafrost, this will occur as a deepening of the active layer, i.e. the layer near the surface which experiences annual freeze/thaw. In areas of discontinuous permafrost, where the ground temperature is within a few degrees of freezing, one could expect eventual and complete degradation of the frozen ground. Complete thaw can cause massive surface deformation as ice-rich soil melts and extensive thermal erosion may occur. Even without complete thaw, warming of frozen ground will substantially reduce bearing capacity of frozen ground (Smith, 1988). Many structures constructed on permafrost are built on pilings. A piling designed for a soil temperature of -4°C will experience a 70% loss in load capacity at -1°C (Esch and Osterkamp, 1989). Mass wasting or slope failure of ice-rich soils will lead to a plethora of problems. Buildings, roads and airports which were constructed on permafrost and designed utilizing temperature data collected over the previous 30 years will require substantial maintenance or perhaps may have to be abandoned.

The hydrologic response to climatic warming in the Arctic is of critical importance and may have global implications. The contribution of CO_2 from the terrestrial biomass may be greater than that from fossil fuel sources (Woodwell et al., 1978). Currently the tundra acts as a major reservoir of carbon in peat (Schell, 1988). It is possible that climatic warming may stimulate growth of tundra plants increasing the amount of carbon in storage (Prudhomme et al., 1984). It is also possible that warmer soil temperatures will increase the rate of oxidation of organic compounds in the soil resulting in a decrease in the amount of stored carbon (Billings et al., 1982; Billings et al., 1983). The response of the tundra biome will largely depend upon the amount of soil moisture. A wet tundra will continue to store carbon, a dry tundra will release substantial amounts of CO_2 and CH_4 to the atmosphere (Houghton and Woodwell, 1989).

A large uncertainty in quantifying the hydrologic response lies in the amount and temporal distribution of the predicted increase in precipitation (Schlesinger and Mitchell, 1987). Woo (1989) speculated that the arctic nival regime (hydrology dominated by snowmelt) will give way to a pluvial regime (hydrology dominated by rainfall) with the advent of a warmer climate. Evaporation is likely to increase, the magnitude of which will determine the volume of surface runoff, again depending on precipitation. The spring snowmelt event will began earlier in the summer instead of near solstice when radiation is near its annual maxima probably resulting in a less intense melt. Surficial runoff will remain important, but degradation of permafrost will increase the soil water storage capacity.

Shafer and Skaggs (1983) believe that CO_2 induced climatic change will significantly impact the occurrence and accumulation of snowfall but will not impact overland flow or channel flow. They also state that climate change would alter evapotranspiration, however the magnitude (or even detectability) of the effect is uncertain and would be highly variable and location specific.

Prediction of the eventual character of hydrology in a changed world is extremely difficult. The hydrologic regime is defined by the interaction of many processes including precipitation, evapotranspiration, infiltration, soil storage and runoff. The thermal regime affects each of these and as the amount of energy input into the system changes, so each of these processes will change. Therefore, determination of the hydrologic response to climatic warming must begin with a realistic simulation of the thermal regime.

I) Simulation of the Effects of Climatic Warming

The exact timing and magnitude of climatic warming is still unknown, so four reasonable scenarios were selected. Analysis of these scenarios indicated the possible response of the hydrologic regime. The occurrence of an abrupt temperature increase was possible and has been discussed in the literature (Ramanathan, 1988) but appears less likely than a steady increase. Gradual warming trends of 2, 4, 6, and 8°C (cumulative) at the soil surface over the next fifty years were selected for the simulations. Most global circulation models indicate the temperature increase will not be equal throughout the year, but rather will be greater in the fall and winter. To avoid attempting to simulate a response which can not be predicted with certainty, a steady increase throughout the year was emulated, however the actual effect may be greater warming in the winter than in the summer (Figure 9-4). The procedure used to generate the surface temperature data is described in Chapter IX.

One primary influence of climatic warming will be an increase in the annual depth of thaw. The effect of a gradual warming trend will steadily increase the active layer thickness and the associated soil moisture storage capacity. The predicted response of the active layer to the selected warming scenarios is displayed in Figure 9-7. In the 6° and 8°C scenarios, when the average annual temperature had increased by 5°C, the annual thaw depth reached 1.12 m depth. At this point a talik, or a layer of soil which did not refreeze each winter, began to form. Then, there were two phase change fronts and isothermal soil between them. These extremely low thermal gradients were difficult to mathematically constrain and numerical instabilities developed. For this reason, the 6 and 8°C scenarios were stopped after 41 and 33 years respectively. This implies that when the average surface temperature increases by approximately 5°C in this area, the permafrost will began to completely degrade. These predictions were used to determine the effects of climatic warming on the hydrologic regime. The most substantial effect of increasing active layer thickness will be increased soil moisture storage. The parameter in HBV which controls the amount of precipitation which can be held in the soil matrix is the field capacity (FC). Assuming the ratio of FC to active layer depth is a constant, the field capacity was increased in proportion to the increase in thaw depth. In addition to increasing the field capacity, the average daily air temperature was increased by 2, 4, 6, or 8°C. Most global circulation models also predict an increase in precipitation in arctic regions in response to climatic warming. As with so many features associated with climatic warming, the amount and distribution of precipitation changes are largely unknown. From the literature (Etkin, 1989), it appears that a 20% increase is a reasonable estimate. Unless the polar ice pack melts, the current precipitation distribution with greatest amounts occurring in the autumn will probably be close to the projected scenario. For the purpose of comparison, simulations were executed with 0, 15, 20 and 30% increase in daily precipitation and initial snowpack water content. The 1986 data were used with the appropriate changes to the air temperature, precipitation and field capacity (FC). The resulting response of the hydrologic system was compared to the current conditions.

The consequences of 2 and 4°C increases in temperature were compared with current conditions in Figures 10-31 a to e. Each of the simulations in the analysis in Figures 10-31 a to e included a 20% increase in precipitation. The difference in the response was considerable. As the temperature increased, snowmelt occurred sooner and peak flows were less, as Woo (1989) had predicted. The final summer event, which before the increase in temperature had been a snowmelt event, became a rain and snow event with lower peaks and a broader runoff distribution. A very important observation was the increase in evaporation with increasing temperature. This impacted the total amount of runoff and the amount of soil moisture. As temperature increased, soil moisture decreased.

To determine the effect of increasing precipitation, the 4°C temperature increase scenario was selected and the

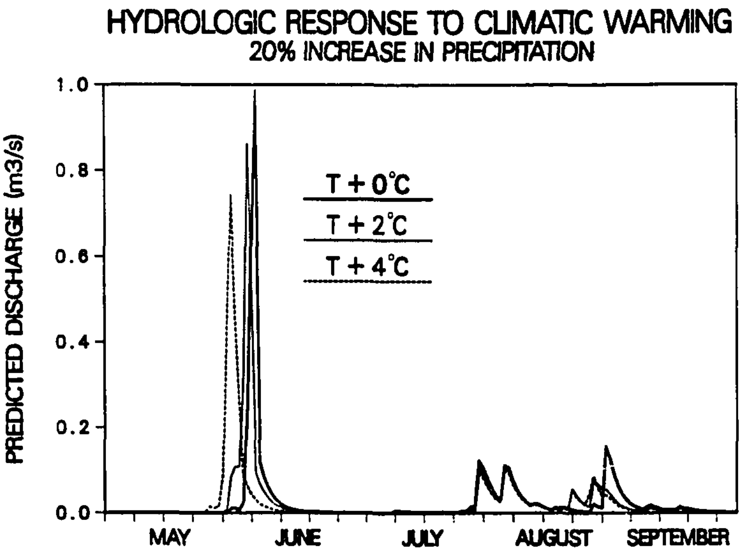

Figure 10-31a. Predicted discharge simulated by HBV with 0, 2, and 4°C temperature increase.

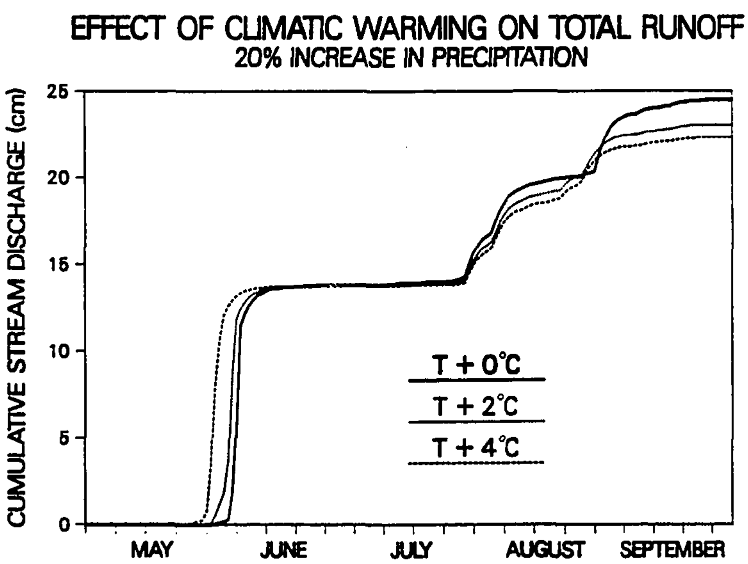

Figure 10-31b. Cumulative discharge simulated by HBV with 0, 2, and 4°C temperature increase.

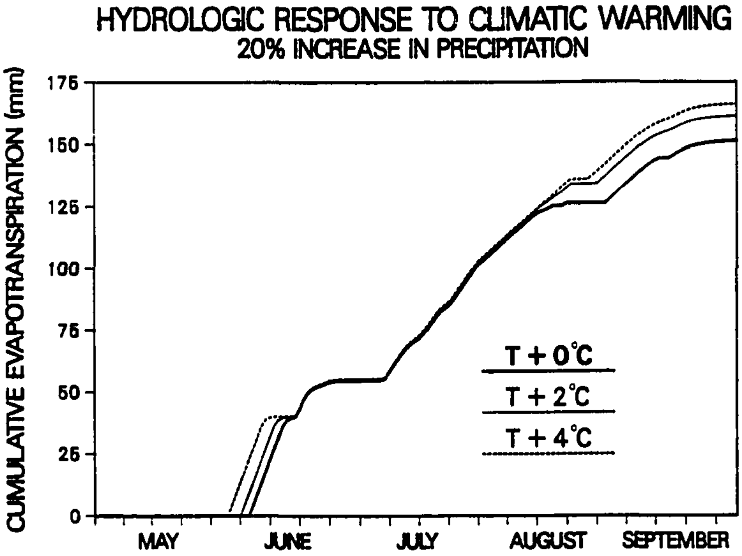

Figure 10-31c. Cumulative evapotranspiration simulated by HBV with 0, 2, and 4°C temperature increase.

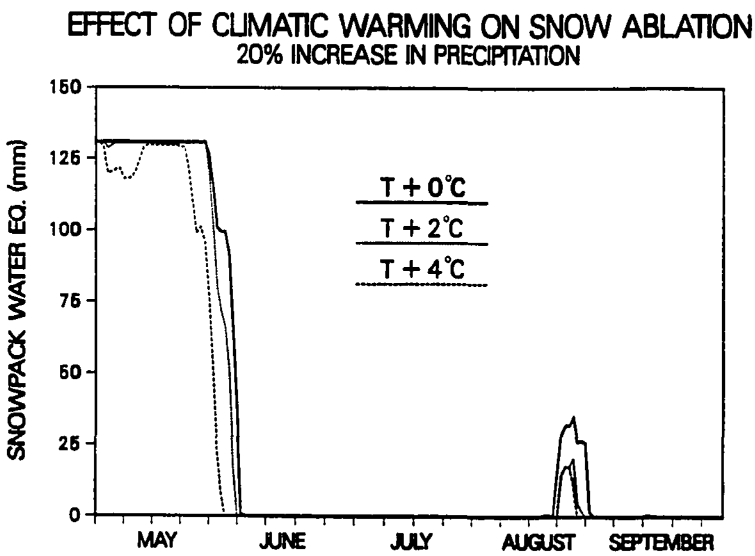

Figure 10-31d. Snow ablation simulated by HBV with 0, 2, and 4°C temperature increase.

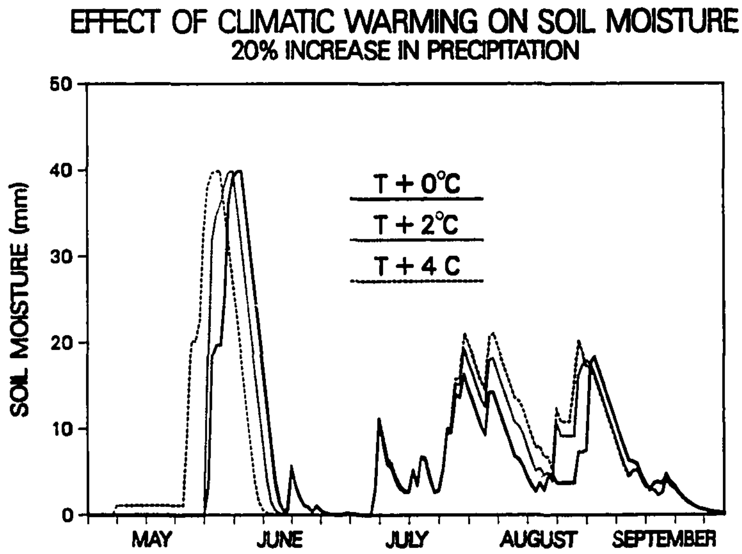

Figure 10-31e. Soil Moisture simulated by HBV with 0, 2, and 4°C temperature increase.

response to 0, 15, 20 and 30% precipitation increase was determined (Figures 10-32 a to e). As expected, greater precipitation yields greater discharge. The peak flows associated with snowmelt were not significantly impacted, however the time required for snowmelt initiation, ablation and runoff recession increased. Peaks associated with rainfall events were greater with greater precipitation. Soil moisture was significantly impacted yielding wetter soils for longer periods of time under conditions of higher precipitation. Evaporation would also increase due to greater availability of moisture. The increases in evaporation and soil moisture were not enough to offset an increase in total volume of runoff.

J) Conclusions

Many important processes associated with arctic hydrology are still not understood, especially in regard to their response to disturbance. More than any other place, the thermal and hydrologic regimes define the permafrost environment. The response of the environment cannot be learned by studying a particular aspect of the arctic biome, but must encompass the entire physical system.

This investigation began with intensive measurements of the thermal and hydrologic processes in Imnavait

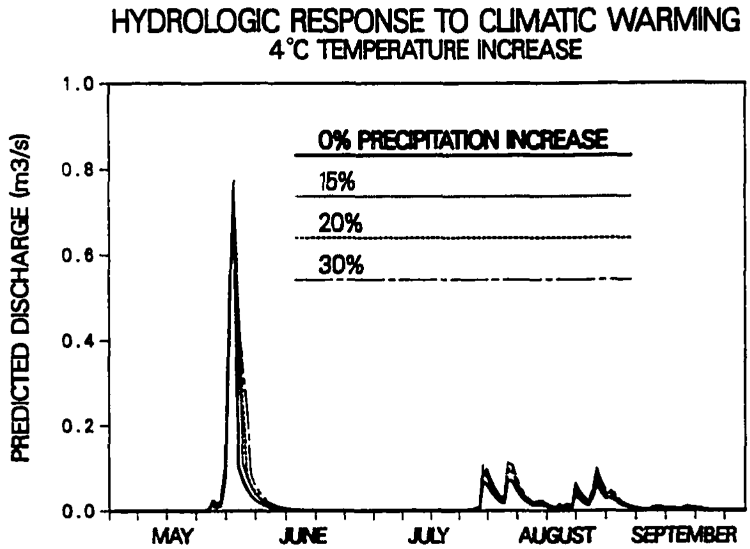

Figure 10-32a. Predicted discharge simulated by HBV with 0, 15, 20, and 30% increase in precipitation.

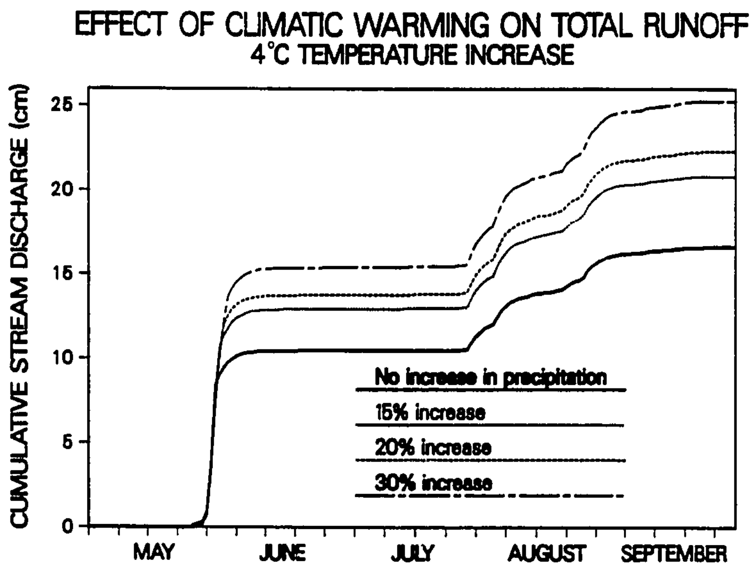

Figure 10-32b. Cumulative discharge simulated by HBV with 0, 15, 20, and 30% increase in precipitation.

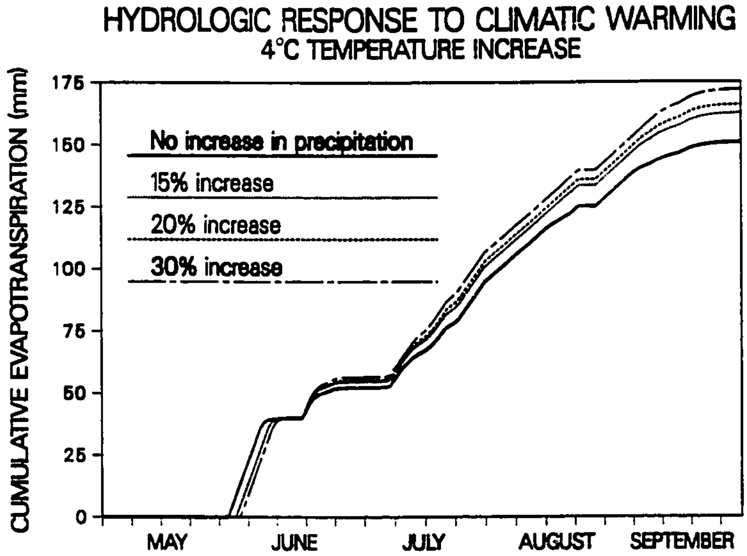

Figure 10-32c. Cumulative evapotranspiration simulated by HBV with 0, 15, 20, and 30% increase in precipitation.

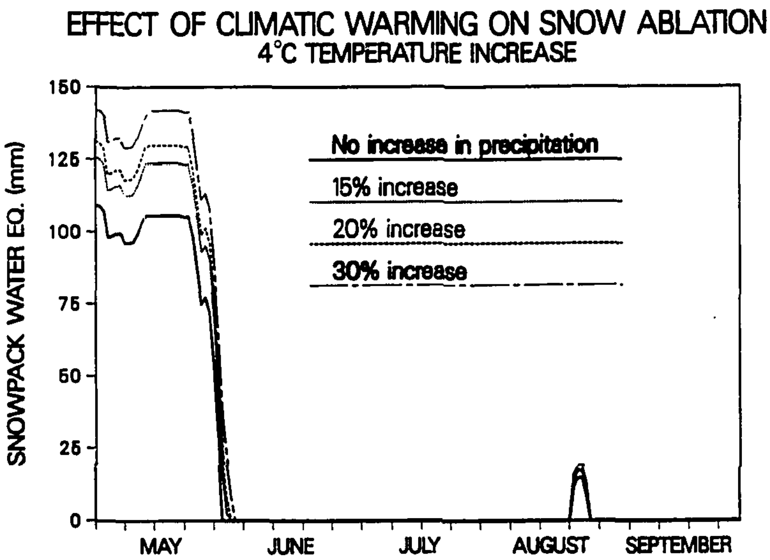

Figure 10-32d. Snowpack ablation simulated by HBV with 0, 15, 20, and 30% increase in precipitation.

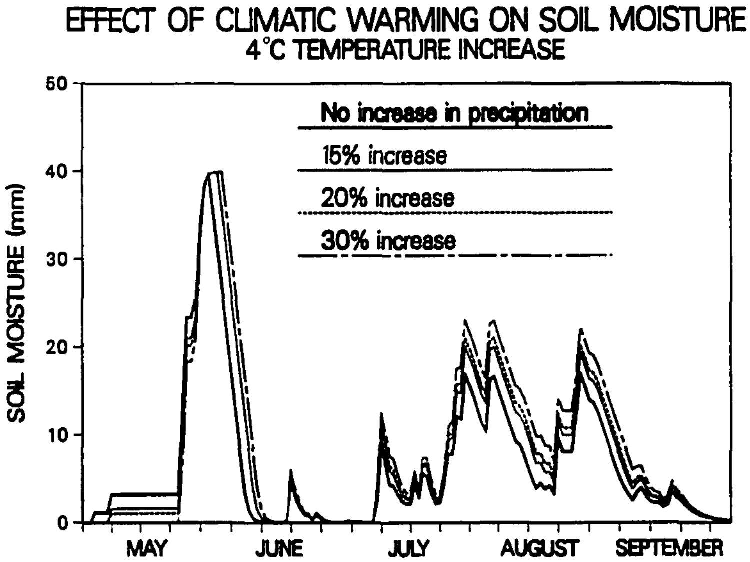

Figure 10-32e. Soil Moisture simulated by HBV with 0, 15, 20, and 30% increase in precipitation.

watershed. The springmelt is the dominant hydrologic event of the year, usually lasting only about 10 days, but normally releasing about half of the annual surface runoff and registering the annual peak flows. Total yearly precipitation averages about 32 cm, with the highest rates occurring in July; about 1/3 to 1/2 occurs as snowfall. Since no surficial water may percolate through the ice-rich permafrost into subpermafrost groundwater, evaporation vies with runoff as the major mechanism of water loss from the watershed. Net radiation is the most important energy source for snowmelt, evaporation and melting of the active layer. Net radiation is low or negative throughout the winter, increasing abruptly after snowpack ablation. Convective heat transfer is especially important during snowmelt; when it complements radiation, substantial melt may occur.

An analysis of the effect of climatic warming on the soil thermal regime in Imnavait watershed reveals that even a slight increase in average annual surface temperature will impact the depth of thaw and temperature of the soil profile to considerable depth. The thermal modeling results indicate that an increase of only 5°C in the average annual surface temperature would be sufficient to eventually completely degrade the permafrost in this area. The increase in active layer thickness would impact many of the hydrologic processes including rates of runoff and soil moisture storage. Hydrologic modeling indicates that increasing the temperature in the watershed will induce earlier spring melt events with lower peaks and broader recessions. Summer events which today would occur as snow events would become rain or rain mixed with snow. As in most hydrologic processes, the amount of soil moisture is largely dependent upon the response of precipitation to climatic warming. Increased temperature with no increase in precipitation will yield lower soil moisture levels.

There are still many unknowns involved in predicting the future climate and its effects, consequently this approach still has many shortcomings. As more research is completed and we learn more about the timing and magnitude of temperature and precipitation changes, estimates of thaw depths, soil moisture contents, evapotranspiration, and streamflows will improve. The estimates of these processes are probably conservative and demonstrate that the effect of climatic change on the hydrologic regime will be significant and must be considered.

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XI) SUMMARY

This thesis presented the results of an extensive examination of the thermal and hydrologic processes in an arctic watershed over a five year period. The central hypothesis of this research is that one may deduce the response of the hydrologic and thermal processes to perturbations in a watershed system given information on the other relevant processes. An understanding of these interactions would enable predictions of certain processes based upon the knowledge of the other important processes. The objective then was to identify the important processes, determine what physical parameters control them and then determine how and why they vary and to what magnitudes.

The approach used to satisfy these objectives was initially one of physical measurements. There have been relatively few studies in the Alaskan Arctic where the hydrologic cycle was studied throughout the year; most studies only address summer processes and usually only a few components of the hydrologic cycle. It was essential to determine the important pathways of water movement and quantify the volumes involved. Since snowmelt represented the greatest single hydrologic event of the year, much of my effort was concentrated during this period. Evaporation vied with runoff as the primary mechanism of water loss from the watershed, so extensive effort was dedicated to determining these amounts. Most of the physical changes occurred within the narrow confines of the active layer, consequently physical processes occurring there were monitored continuously. The most dynamic portion of the active layer was the near surface organic layer which varied from 10 to 15 cm in thickness. Most downslope water movement occurred within this layer and the rates of heat transfer to the underlying permafrost were strongly affected by the insulative properties of this organic soil.

The mechanisms of energy transfer, their annual variation and their magnitudes relative to each other are perhaps the most important parameters influencing this arctic biome. Although the winter period is dominated by a net radiation loss, the summer represents a period of high incoming shortwave radiation and 24 hours of daylight. Energy advected from the South over the Brooks Range is a primary process instigating the initiation of spring snowmelt. Conduction is the dominant mechanism of heat transfer within the soil matrix except during the infiltration of meltwater into the organic layer at the initiation of snowmelt.

Measurement of the material properties of the active layer was a mandatory prerequisite to quantifying the hydrologic or thermal processes. These include determining the thermal and hydraulic conductivities, porosity and bulk density as a function of depth in the soil profile. It was also important to identify the temperature dependence of these parameters for both frozen and unfrozen cases.

Utilizing the extensive data set of measurements, I was able to characterize this watershed in terms of the watershed energy and water balances. Accurate measurements of processes such as snow accumulation, snowmelt, rainfall, and runoff allowed postulation of their relationships, but a clearer perception of these processes and their interactions was possible through application of computer models. By recreating the runoff hydrograph using measurements of basin parameters, snowmelt or rainfall and air temperature, one may more fully understood many of the unique features of arctic hydrology such as the rapid response time of watershed runoff to rainfall events, the very limited soil moisture storage, the impact of summer evaporative processes and the mechanisms of snow damming. Modeling the heat transfer in the soil demonstrated the dominance of conductive heat transfer and the effect of infiltration of spring meltwater. The models thus served as research tools to delineate current conditions and the importance and magnitudes of involved processes.

The thermal regime was simulated using TDHC, a two dimensional heat conduction program. This model also has a routine for determining the heat transfer in phase change. Utilizing measurements of the material properties and active layer temperatures permitted fine calibration of the model. The hydrologic regime was simulated using HBV, a continuous runoff model which considers snowmelt, rainfall, soil storage, infiltration, evaporation and runoff. The model was calibrated with measurements of watershed runoff.

Successful application of a hydrologic and a thermal model allowed further extrapolation of knowledge gained from measurements of the thermal and hydrologic processes. Confident that the models accurately represented the hydrologic and thermal regimes, I was able to artificially impose a warming trend and a physical change in the active layer properties and then determine the response of the thermal and hydrologic regime.

Climatic warming presents an imposing problem to scientists everywhere. The threat it poses in the Arctic is accentuated for several reasons. The temperature increase is expected to be greatest in the higher latitudes and the ramifications of this warming may be more devastating there due to the melting of permafrost. Determination of the magnitude of climatic warming in the Arctic was not within the realm of this study, however examination of several reasonable scenarios does permit exploration of the consequences. Increased precipitation will probably accompany climatic warming in the Arctic confounding the effect of a temperature increase. The interaction of changing hydrologic and thermal processes presents a complex problem which would be difficult if not impossible to analyze without the use of detailed computer programs.

Several scenarios of climatic warming were examined, including a 2°, 4°, 6° and 8°C increase in temperature with 0, 15, 20 or 30% increase in precipitation. The most obvious and perhaps significant response to climatic warming was an increase in active layer depth. Other changes worth noting were warming of the entire soil profile, increased soil moisture storage, increased evaporation and variable response in runoff depending upon the scenario.

One limitation of this approach was that the models were not dynamically coupled throughout the warming scenarios. The changes in the active layer as determined through the thermal analysis were input into the hydrologic simulations. Another difficulty which was not completely overcome is the effect of coupled heat/moisture transfer. Moisture migration in response to a thermal gradient will affect the moisture holding capacity of the soils and also determine the soils' thermal properties. Convective heat transfer was not considered in the thermal modeling, resulting in an under-estimate of thaw depth. A better method would be to develop one model which could calculate conductive and convective heat transfer in the soil and moisture movement through and over the soil. The best possible model would be a combined thermal and hydrologic model, basing heat transfer in the soil, snow melt calculations and evapotranspiration on a complete surface energy balance. The hydrologic processes would be modeled using a physically based formulation which considers changes in hydrologic properties in response to external stimuli. Such a model would be extremely complex and may not be realistically possible, but would be certainly valuable and well worth the effort of development.

Further limitations of this approach lie in the uncertainty of the actual response of the climate to the increase in "greenhouse gases," i.e. the temporal distribution of increased temperature and the exact temporal and spatial change in precipitation distribution. All research into the potential response of the Arctic to climatic warming is rooted in the knowledge or ignorance of the degree of warming and the amount of precipitation. Our scientific community must strive towards determining what the effect of increasing greenhouse gases will be.

This study has made substantial contributions to the current understanding of arctic hydrology. The predictions of the response of the hydrologic and thermal regimes to climatic warming are my best estimates based on currently available knowledge. As estimates of the magnitudes of temperature and precipitation changes improve, predictions of the response of the arctic biome can also be improved.

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