NASA

Technical Memorandum TM 79720

Geomorphic Processes on the North Slope of Alaska

(NASA-TM-79720) GEOMORPHIC PROCESSES ON THE NORTH SLOPE OF ALASKA (NASA) 30 p HC A03/MF A01 CSCL 08G

N79-23561

Unclas

G3/46 25921

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MARCH 1979

National Aeronautics and
Space Administration

Goddard Space Flight Center
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ABSTRACT

Three physiographic provinces comprise the North Slope of Alaska: the Arctic Mountains, the Arctic Foothills and the Arctic Coastal Plain Provinces. The features and processes in the Arctic Coastal Plain, a zone of continuous permafrost, are stressed in this paper. The evidence for and mechanisms of the geomorphic cycle are discussed starting with frost cracks. Frost cracks may form polygonal ground which leads to low-centered ice wedge polygons in areas having ice-rich permafrost. As the low-centered ice wedge polygons enlarge due to thermal erosion they may evolve into thaw lakes which are largely oriented in a northwest-southeast direction on the Arctic Coastal Plain. Eventual drainage of a deep lake may result in a closed-system pingo. Evidence of the various stages of the geomorphic cycle is ubiquitous on the Alaskan Arctic Coastal Plain and indicates the ice content of the permafrost in some areas.

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GEOMORPHIC PROCESSES ON THE
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I. INTRODUCTION

The North Slope of Alaska is an excellent region in which to analyze the processes and forms characteristic of areas underlain by permafrost. The North Slope is composed of three major physiographic provinces as expounded by Wahrhaftig (1965): the Arctic Coastal Plain, the Arctic Foothills and the Arctic Mountains. More specifically, the North Slope extends from the crest of the peaks in the Brooks Range northward to the Arctic Ocean coast (Figure 1) and is within the region of continuous permafrost (Figure 2). Many of the features that will be discussed also form in areas which have seasonally frozen ground, however only selected prominent features of the continuous permafrost zone of the North Slope of Alaska are discussed in this paper. The Arctic Coastal Plain will be stressed because it is the province which best exemplifies the geomorphic features indicative of permafrost. It will be shown that the evolution of suprapermafrost features on the coastal plain is brought about by the interaction between surface and subsurface processes unique to areas underlain by permafrost, and further, that many of the features, given time, evolve into one another in the sense that the destruction of one form means the formation of another. This evolutionary process is dependent upon the interactions among soil characteristics, topography, elevation, climate, snowcover, slope and orientation.

Thaw lakes and former lake basins are ubiquitous on the coastal plain and assume great geomorphological significance. Aufeis or overflow river ice is also significant because its formation and development is a manifestation of local hydrologic, geologic and geomorphic conditions as well as regional climate. Thaw lakes and aufeis will be stressed.

II. PHYSICAL GEOGRAPHY OF THE NORTH SLOPE

Permafrost has been defined as any surface or ground material that has had a temperature of 0°C or less for two or more consecutive years. Three zones of permafrost have been recognized:

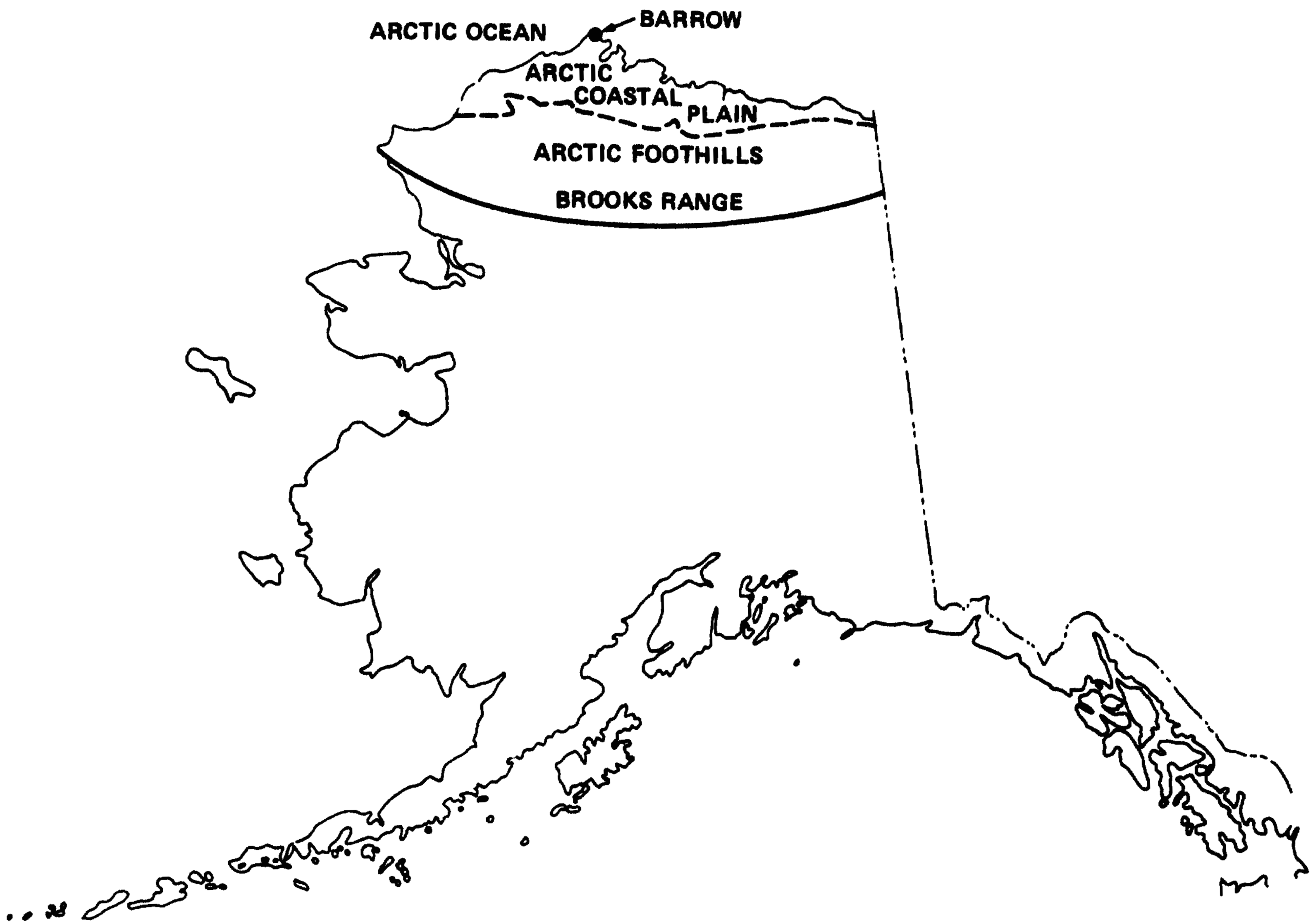

Figure 1. Arctic Coastal Plain

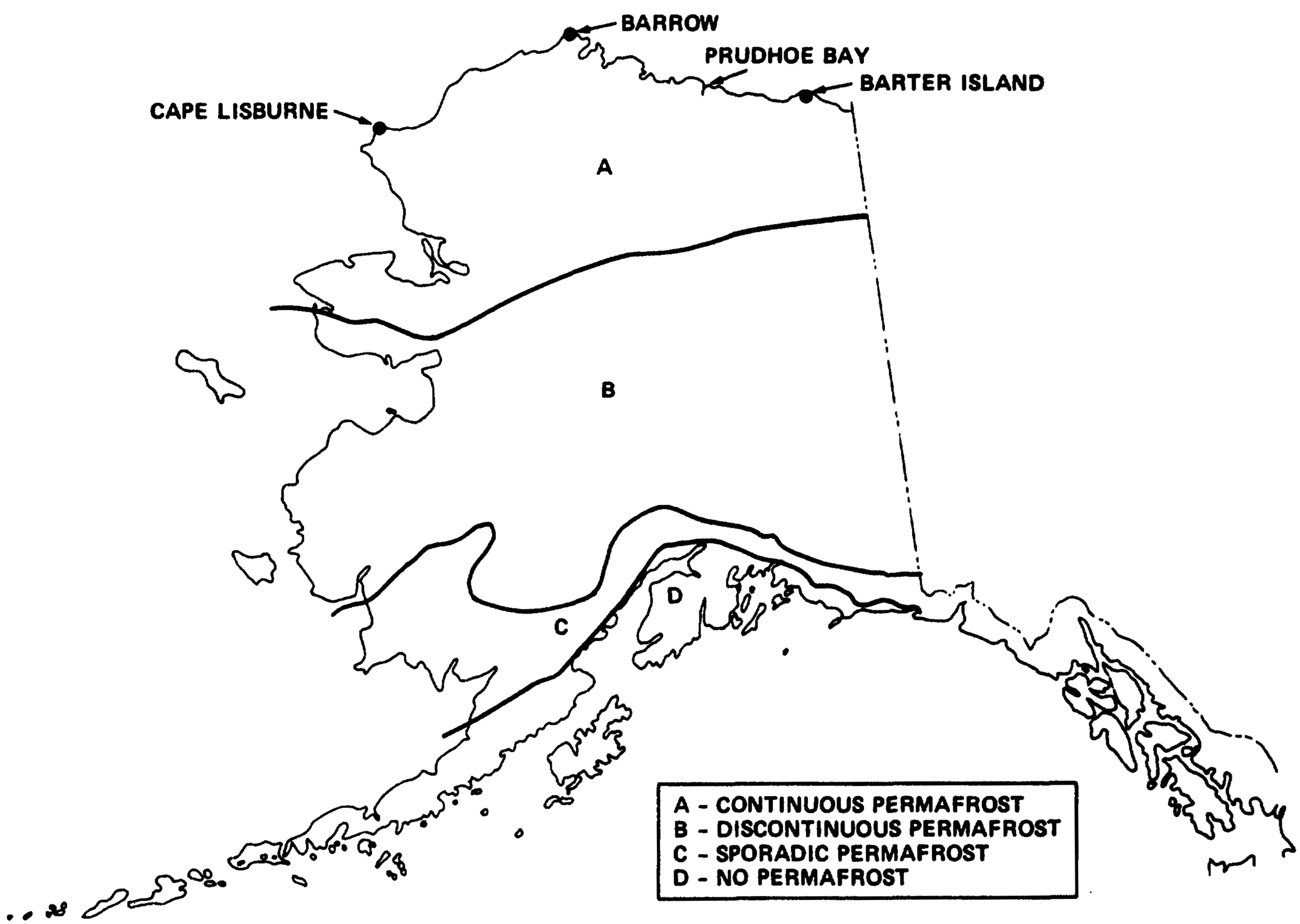

Figure 2. Pleistocene Glaciers and Permafrost in Alaska (after Wahrhaftig, 1965)

continuous, discontinuous and sporadic (Figure 2). In the continuous zone the entire area is underlain by permafrost; an average annual temperature of -5.0°C or less is required to maintain continuous permafrost (Black, 1976). In the discontinuous zone areas free of permafrost are found. In the zone of sporadic permafrost, permafrost is often only found where local factors such as exposure, snowcover and vegetation allow a thermal regime conducive to the perpetuation of permafrost. For example, in the sporadic zone north facing slopes may be underlain with permafrost and south facing slopes which receive more insolation may be permafrost-free.

Even in areas of continuous permafrost, zones of unfrozen ground may exist. These are known as taliks and often exist beneath deep lakes and rivers (Price, 1972). Lakes deeper than ~2 m in Arctic Alaska do not freeze to their beds thus leaving water in the lake beds throughout the winter; this water melts the permafrost below the lake by thermal erosion. Even shallow lakes, <2 m deep, defined as those which freeze to their beds, influence the thermal regime of the ground below because of the presence of water during the summer. Ground temperatures beneath shallow lakes may be up to 3°C greater than surrounding ground temperatures (Holmquist, 1975).

Continuous permafrost underlies the Arctic Coastal Plain to depths of up to 600 m (Holmquist, 1975), however considerable variations in the thickness of permafrost exist. Along the Arctic Ocean coast, thicknesses of $\leq$250 m are common because the influence of the ocean moderates the climate and inhibits permafrost development. It is interesting to note that permafrost exists offshore on the Arctic Ocean floor despite the warming influence of the ocean water (Lewellen, 1970; Osterkamp and Harrison, 1977). The mean annual temperature for the bottom of the Beaufort Sea is ~-1.3°C as opposed to surface mean annual temperatures of ~-9.0°C for adjacent land areas (Lewellen, 1974).

Permafrost is overlain by a zone of seasonally thawed ground, the active layer, which varies in thickness depending upon local conditions but is generally <0.5 m in thickness (Lewellen, 1970) and attains a maximum thickness of ~1.2 m (Kane and Carlson, 1973) on the coastal plain of Alaska. Soils composed of peat and organic matter have shallower active layers than those composed of sands and gravels because infiltration of water and transfer of heat is facilitated through pore spaces in sand and gravel sediments allowing greater depths of thaw. Fine, sandy permafrost soils are the most deeply thawed (Everett and Parkinson, 1977).

The soils of the North Slope are mineral with some poorly-drained organic soils (Kane and Carlson, 1973). The efficacy of the soil forming process in Arctic Alaska decreases with increasing elevation (Brown, 1968) as the production and decomposition of organic material is hampered by lower temperatures.

The Arctic Coastal Plain has very little relief. The surface is characterized by polygonal features, low shrubby vegetation and an abundance of standing water including thousands of lakes formed from the thawing of ground ice. These lakes are oriented in a direction perpendicular to the direction of the prevailing winds. The 'oriented lakes' and their remnent lake basins dominate the tundra surface, covering up to 90% of the land surface in some locations (Holmquist, 1975). The sediments of the coastal plain are unconsolidated clay, silt, sand, gravel and peat which range from a few meters to 60 m in thickness (Holmquist, 1975). The Arctic Coastal Plain was never glaciated because of the sparsity of precipitation.

The Arctic Foothills, just south of the coastal plain, rise to an elevation of 900–1000 m and contain evidence of former glaciation such as glacial drift. The Brooks Range is an extension of the Rocky Mountain system and has peaks up to 3000 m in elevation (Holmgren, 1975). A few small valley glaciers exist in the eastern part of the mountain range, however evidence of former, more extensive glaciation is present (Figure 2).

Climatologically the North Slope may be considered a desert because precipitation is so low, 13–25 cm per year (Kane and Carlson, 1973). Most of the precipitation occurs between June and October and most is in the form of snow. Surface storage of freshwater is abundant and exists in the form of snow, lakes, and river storage in the form of aufeis deposits.

The precipitation can be in the form of snow in any season but is generally in the form of rain during the short summer. A snowpack with a maximum thickness of 30–40 cm (in April) developes each year (Holmgren et al, 1975) and is constantly redistributed and hardpacked by the wind. (Prevailing winds are bimodal, ENE and WSW and blow quite constantly.) In the Brooks Range a thicker snowpack of ~85 cm developes each year (Brown, 1966) due to greater precipitation in the mountains.

Variations in mean annual temperature and precipitation on the North Slope are dependent upon latitude, elevation and continentality. The temperatures of the coastal stations are generally cooler in the summer than farther inland because of the influence of the open ocean. But during the winter the sea ice extends to the coast and the climate of the coast can approximate that of locations farther inland (Brown, Haugen and Parrish, 1975) because the presence of sea ice inhibits exchange of heat between the ocean and the air which would otherwise modify the winter temperatures near the coast.

Continuously operating meteorological stations are few on the North Slope and most of those that do exist are located on the coast: Barrow, Barter Island and Cape Lisburne. Along the coast, the incidence of cloudiness and fog is greater than farther inland although much of the North Slope is cloudy due to winds blowing inland from the Arctic Ocean (Maykut and Church, 1973). Cloudcover on the coastal plain is at a maximum (8.1 tenths) between May and November when leads in the near shore sea ice are common, and at a minimum (4.9 tenths) between the months of December and April when the sea ice is against the coast and few leads are present. There exists a uniformity of temperature and cloudcover conditions along coastal portions of the coastal plain.

The effect of continentality is measureable only a few kilometers inland from the Arctic Ocean. The coastal areas maintain a snowcover several weeks longer in the spring than sites 50–70 km inland (Benson et al., 1975). All months at Barrow have an average temperature of <4°C with February being the coldest and July being the warmest month (Maykut and Church, 1973).

The mean annual temperature and precipitation for Barrow is -12.2°C and 11.5 cm respectively, and for Anaktuvuk Pass in the Brooks Range -10.6°C and 28.4 cm (Holmquist, 1975). The climate of the North Slope is severe enough to allow the perpetuation of continuous permafrost, i.e. $\leq$-5.0°C mean annual temperature.

The melt period on the North Slope begins from mid-May to early June in the mountains and moves rapidly northward onto the coastal plain (Holmgren et al., 1975) with snow and ice breakup occuring latest on the coast of the Arctic Ocean. Shortly after melt the air temperature becomes positive due largely to the dramatic rise in albedo which accompanies melting.

Albedo changes are important to an understanding of the climate because the albedo and thus the receipt of insolation on the North Slope can change very rapidly with the onset of melt. Albedo may decrease from 75 to 10% in 3 to 5 days with the melting of snow (Maykut and Church, 1973) and therefore the date of snowmelt is very important to the total annual receipt of solar radiation. As the albedo decreases more net radiation, Rn, is received. An early snowmelt such as occurred in 1964 at Barrow will result in greater annual totals of Rn as shown in Figure 3.

III. A GEOMORPHIC CYCLE ON THE ARCTIC COASTAL PLAIN OF ALASKA

Ground ice can exist as various features representing various modes of formation. Pore ice exists in coarse sand and gravel sediments whereas segregated ice is typically found in fine-grained organic soils. Foliated or vein ice develops in thermal contraction cracks over thousands of years. Pingo ice can form either by upwelling of ground in a shallow lake basin or by updoming of the surface as groundwater under hydrostatic pressure is forced to the surface. All of these features develop in areas of ice rich permafrost.

Several authors have written about the thaw lake cycle on the Arctic Coastal Plain (Sellman et al., 1975; Hussey and Michaelson, 1966; Carson and Hussey, 1962) but have not given much

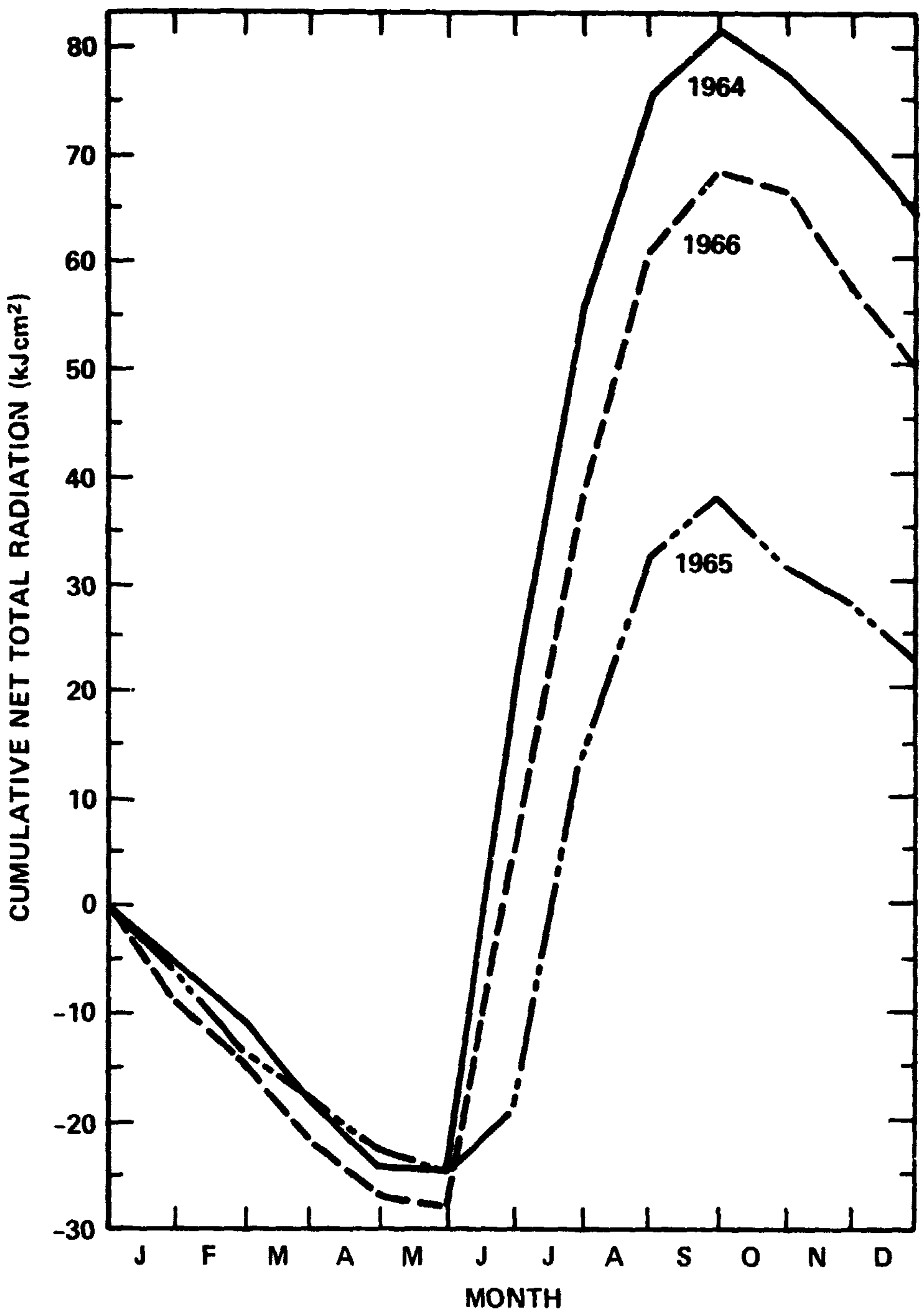

Figure 3. Cumulative Net Total Radiation at the Surface at Barrow, 1964–1966 (from Maykut and Church, 1973)

recognition to the cyclicity inherent in many of the permafrost related processes and features including frost cracks, low- and high-centered ice wedge polygons and pingos. Interactions among these phenomena will be discussed in this section because of their prevalence and dynamic behavior on the North Slope.

Extreme surface cooling causes thermal contraction of the tundra leading to tension cracks or frost cracks which form in a polygonal pattern (Figure 4). Water which is common on the tundra during the melt season seeps into frost cracks and subsequently freezes. Expansion of up to 9% may occur with the freezing of water, increasing the size of the crack. More water seepage occurs during the following summer and further expansion of the crack occurs upon freezing. This process may repeat itself each year in the same crack causing annual enlargement (Hussey

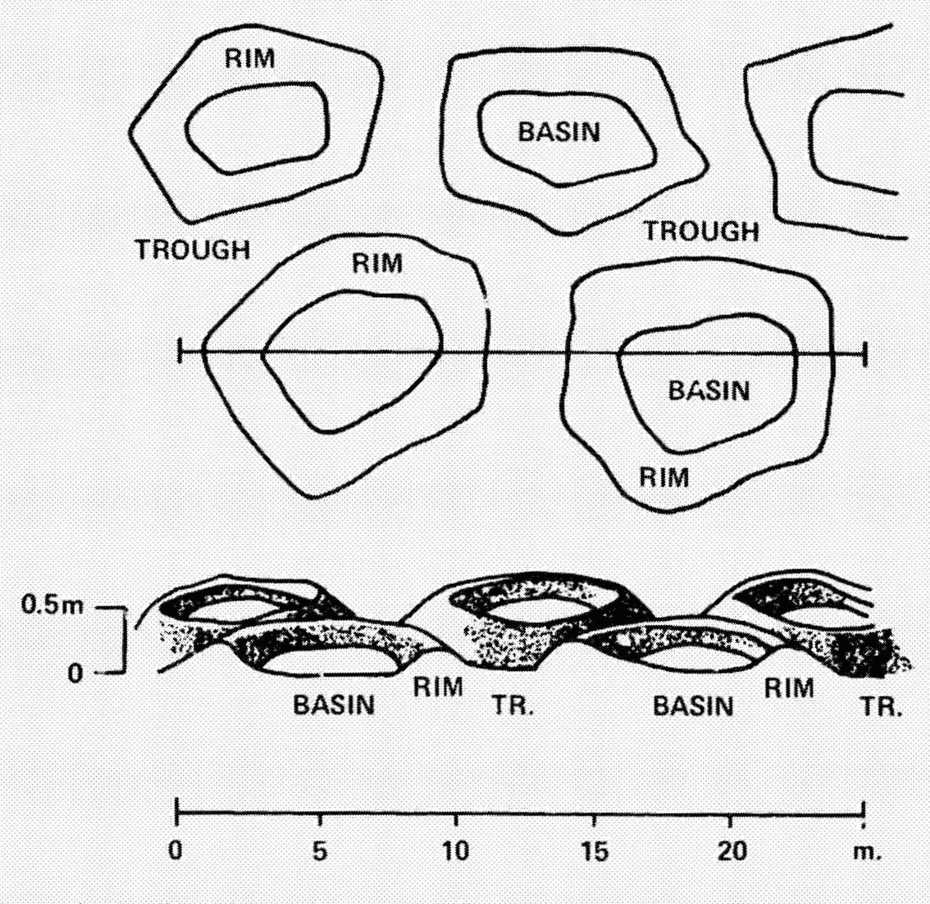

Figure 4. Idealized Cross-Section and Plan View of Polygon Microrelief Near Barrow, Alaska (from Brown et al., 1975).

and Michaelson, 1966). Many of these cracks are surface features and do not have much subsurface expression. When water seepage occurs in shallow cracks the water will not remain frozen throughout the year, however if the crack extends beyond the active layer an ice wedge polygon may form (Shumskii, 1964). Diameters of ice wedge polygons are 2 or more times greater than the depth of the cracks (Embleton and King, 1975). Ice wedges can be 1 cm to 3m wide and 1-10m deep. The diameter of the entire polygon may be up to 100m (Price, 1972).

The permafrost table, the upper limit of permafrost, impedes subsurface drainage under the ice wedge polygon because of its impermeability. Thus water can remain in the crack in the active layer and freeze (Brown, 1966). Ice wedge polygons develop gradually over a period of years as increments of water which later turn to ice are added into the crack each year. The cracks may later be filled with sediments (Price, 1972) forming fossil ice wedges which indicate the existence of former permafrost.

Sod or rock ridges may surround polygonal forms and are either sorted or nonsorted (Kachurin, 1964). Sorted polygons do not require permafrost for their development and are generally smaller in size than ice wedge polygons, which are usually not sorted or are poorly sorted and only develop in permafrost areas. The formation of sorted polygons is dependent upon the annual freeze/thaw cycle in the Arctic and the diurnal freeze/thaw cycle in mid-latitude and -elevation regions. The degree of sorting is dependent upon the amount of soil moisture, the composition of material and the frequency of freeze/thaw cycles. Degree of sorting will increase with an increase in soil moisture. Soil with a high moisture content is most conducive to frost heaving from expansion of water during freezing. If a soil consisting of mixed particle sizes is saturated and the rate of freezing is slow, a large number of sizes can be sorted out, conversely if soil freezes rapidly no sorting will take place (Embleton and King, 1975).

Freeze/thaw cycles which enhance sorting are far more frequent in areas such as the Colorado Rocky Mountains where the temperature may fluctuate around 0°C diurnally (in some seasons), than on the North Slope where the air temperature does not fluctuate around 0°C frequently. The sorting process, driven by frost heaving and thrusting, is not caused by the 9% expansion of water upon freezing alone. Frost heave is also enhanced by ice crystal or needle ice growth which forms perpendicular to the cooling surface. Needle ice exerts pressure and can move objects in the direction of crystal growth. This enhances heaving as water is drawn into the freezing plane by molecular cohesion (Price, 1972). Needle ice is not of great importance in the high Arctic because its development is dependent upon frequent diurnal freeze/thaw cycles which are not characteristic of the North Slope. Nocturnal freezing and daily melting are responsible for needle ice development.

Frost thrusting exerts pressure mainly in the horizontal direction where heaving exerts pressure in the vertical direction. As a result of both of these mechanisms upward displacement and sorting of subsurface objects occurs. Upward movement of coarse stones may occur when freezing and thawing occurs from the surface; the fine particles move downward (Embleton and King, 1975). During upward expansion of moist ground upon freezing, enclosed stones are carried with the soil. Prior to collapse upon thawing, fine silts, sands and clays infill voids created at the base of the stones and as a result the stones do not return to their original positions but are slightly elevated (Price, 1972).

Polygonal features are not restricted to cold regions. Many form in deserts as a result of dessication cracking. However the largest polygonal forms are associated with ice wedges (Price, 1972). Polygons that are associated with ice wedges are usually nonsorted or poorly sorted. Sorted polygons, smaller than the nonsorted variety, have a border of coarse material with a center of fine material. Both sorted and nonsorted polygons occur on relatively flat terrain. Features found on slopes are usually sorted or nonsorted stripes which are elongated as a result of downslope movement.

Not all of the polygonal features discussed have raised centers or ridges; some have little or no relief. Peripheral ridges form as a result of ice wedge growth causing thrusting and upturning of strata (Hussey and Michaelson, 1966). Four or five sided actively growing ice wedge polygons have depressed centers and are termed low-centered ice wedge polygons.

Low-centered ice wedge polygons may evolve into high-centered ice wedge polygons in the following way. Thawing and erosion of the ice in the peripheral ridges may lead to depressed troughs in place of the raised ridges thus the centers become higher than the borders (Price, 1972). These high-centered polygons are associated with ice wedges that are no longer growing and are often associated with the zone of discontinuous permafrost (Figure 5). Accumulation of ice and organic matter in the polygon centers leads to the development of peaty soils (Brown, 1966). Peat causes a decrease in the thickness of the underlying active layer by providing insulation against the penetration of summer warmth.

As shown, the low-centered ice wedge polygons may develop into high-centered ice wedge polygons, on the other hand they may form ponds in their depressed centers from standing water. The ponded water is warm relative to the frozen ground causing melting of the ice in the ground during the summer; this erosion of the ground by melting continues and ultimately forms a thaw lake. On the coastal plain of Alaska thaw lakes form in unconsolidated sediments; preferential erosion induces a preferred NW-SE orientation. This orientation is approximately perpendicular to the direction of the bimodal (NE-SW) prevailing winds (Livingstone, 1954). Erosion of these lakes can occur very rapidly, up to 12 m per year (Lewellen, 1965). The lakes 'migrate' across the tundra but their depth becomes relatively stable after a certain depth (~3 m in the Barrow area) is attained (Hussey and Michaelson, 1966). They may coalesce and eventually drain either partially, leaving a marsh, or fully, leaving a remnent lake basin. Lakes which freeze to the bottom each year often have polygonal cracks in their beds due to thermal contraction (Hussey and Michaelson, 1966). After draining, polygonal ground may thus be present (Brown, 1968) thus completing the cyclic process from ice wedge polygons to thaw lakes, as seen in Figure 6.

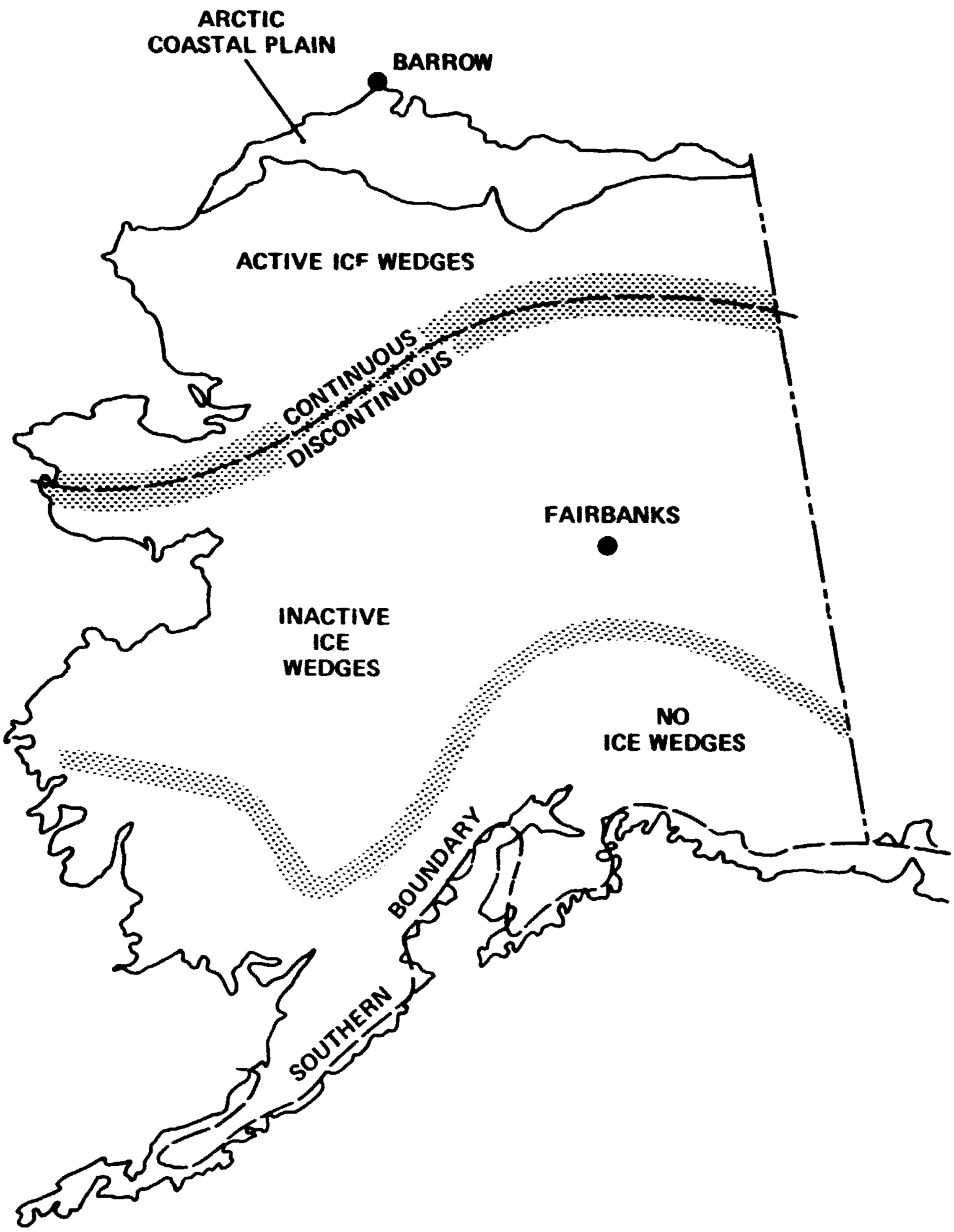

Figure 5. Map of Alaska Showing Occurrence of Permafrost and Activity of Ice Wedges (after Brown, 1968)

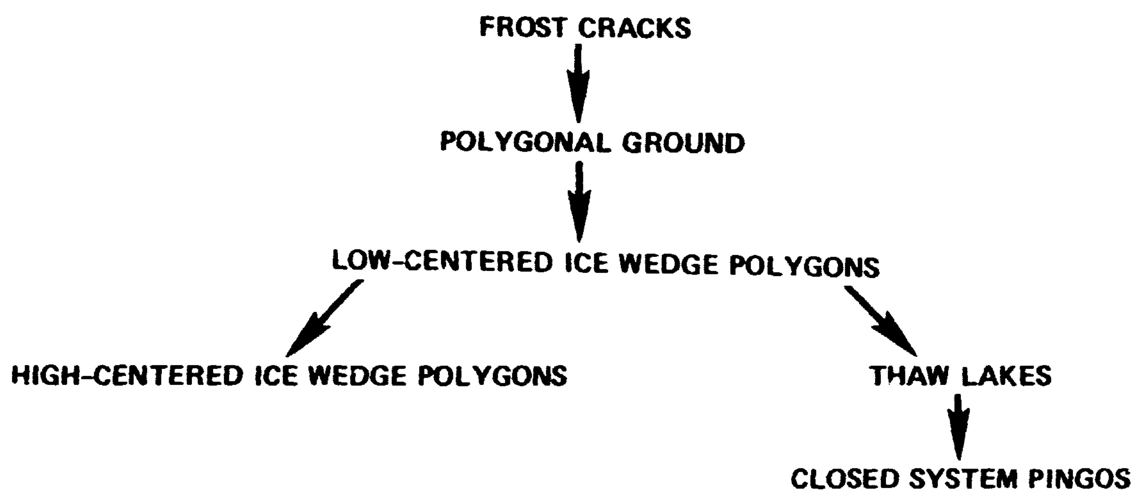

Figure 6. Scenerio for Evolution of Forms in Permafrost Terrain

Thousands of thaw lakes are present on the Arctic Coastal Plain. These lakes form in unconsolidated sediments composed of fine-grained silts and sands which are easily eroded by wind-induced wave action. As previously shown these lakes go through a cyclic pattern of formation, migration and drainage. Once thaw ponds have developed, prevailing winds produce waves that induce shore erosion on the banks perpendicular to the wind direction (Kaczorowski, 1977). Livingstone (1954) has mathematically postulated the cause of orientation of the lakes in the direction normal to the prevailing winds (Figure 7). His theory states that a thaw depression begins as a relatively circular pond; the crosswind shores are scoured out by longshore currents more rapidly than the downwind shores causing the long axis to be at right angles to the prevailing wind direction. As the lake basin continues to develop, more ground ice melts due to the warm lake water in contact with frozen basin sediments. Water temperatures of up to 21°C (70°F) have been measured in these lakes (Carson and Hussey, 1962). As a lake becomes deeper the rate of thermal erosion of the lake bed decreases because the water is cooler owing to the greater quantity of water in the lake.

Density and degree of orientation of thaw lakes increases to the north on the coastal plain, in an area of fine grained sediments and low relief. In this area lake migration can be very rapid,

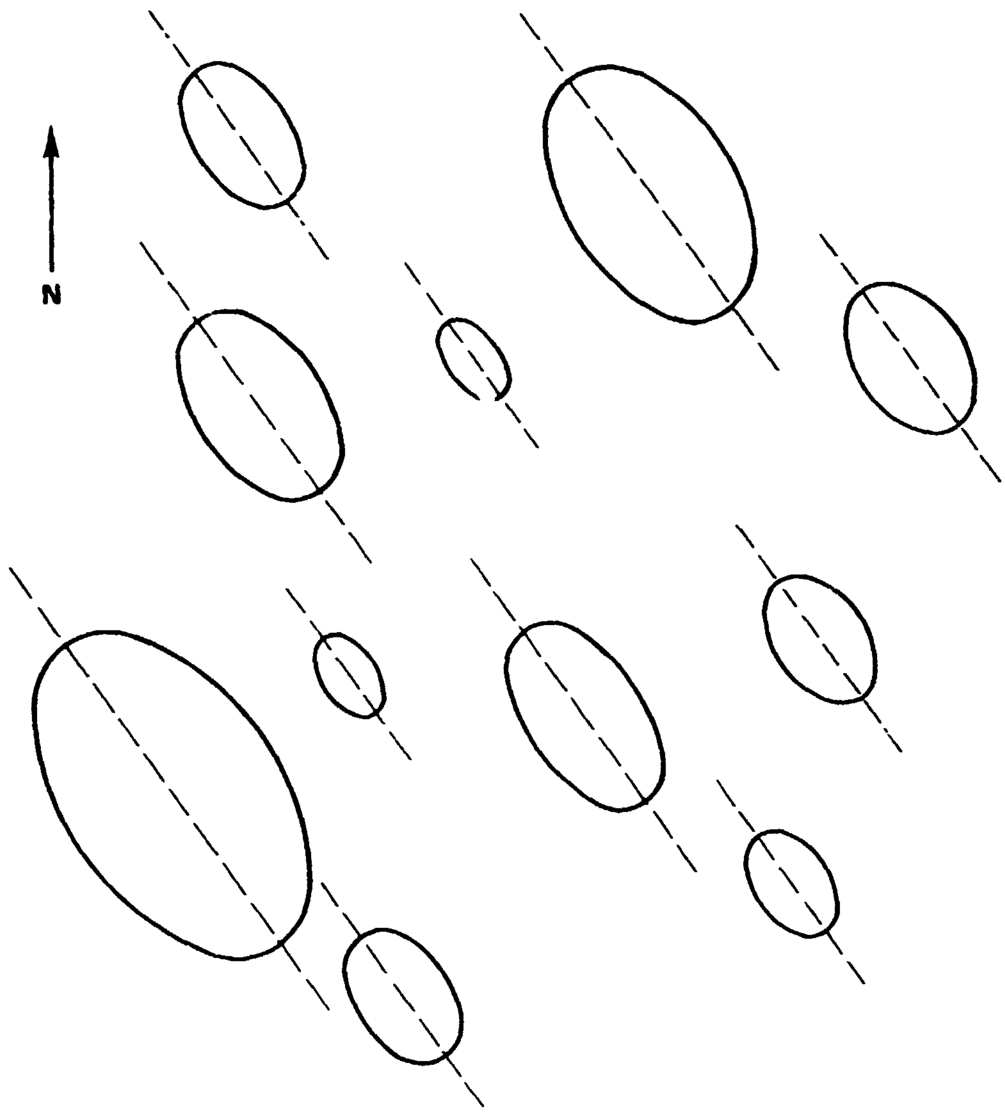

Figure 7. Oriented Lakes (from Kaczorowski, 1977)

~9 m during summer storm conditions (Black and Barksdale, 1957) and an average of 1 m per year (Livingstone, 1954). Livingstone (1954) and other investigators have proven that the processes that oriented the lakes (wind-induced thermal erosion of unconsolidated frozen sediments) continue to operate. The lakes are not features formed by processes no longer in operation as has previously been postulated (Black and Barksdale, 1957); present-day erosion and migration have been measured and quantified for many of the lakes.

A drained lake is a good place for a closed system pingo to develop (Black, 1976). Water is trapped between the downward freezing surface, the permafrost table, and along the sides of the drained basin due to frozen ground. The trapped water expands upon freezing r[illegible]ng overlying soil upward forming a central ice-cored mound. This ice core 'grows' as more water is added slowly from the groundwater table. The pingo grows rapidly the first few years, ~1.5 m per year, then more slowly and may continue to grow for 1,000 years (Embleton and King, 1975). This type is known as a closed system pingo and often forms from a large, deep lake which is draining. The deep lakes which do not freeze to the bottom (>2m in northern Alaska) have a talik beneath their beds due to the thermal effect of the water on the permafrost. The ice cover freezes to the bottom as the lake becomes more shallow putting pressure on the talik below. A shallow lake (<2m deep in northern Alaska) will usually not have a talik below the bed and therefore is not conducive to pingo growth. Pingos range from a few meters to 70m in height and may become several hundred meters in diameter (Shumskii, 1964).

In order for a pingo to grow it must be in an area which has an abundant groundwater source (eg. a talik or a spring), overburden pressure (e.g. an ice cover) and silty, clayey material at its base to prevent infiltration (Ryckborst, 1975). There are two main types of pingos – the closed system or MacKenzie type which is the type just discussed, and the open system or East Greenland type (Price, 1972) as well as other less common types (Pissart, 1970).

An open system pingo usually occurs on a slope where artesian pressure develops in taliks. As spring water approaches the surface updoming of the surface sediments occurs (Price, 1972). Open system pingos develop with the expansion or formation of a talik (i.e. from a spring) and are associated with permafrost degradation; closed system pingos develop as a result of contraction of a talik (i.e. from beneath a formerly deep lake) and are associated with permafrost aggredation (Muller, 1959). Pingos are significant features on the Arctic Coastal Plain. Brown (1968) estimates that 1,000 pingos exist on the coastal plain of Alaska.

It should not be inferred from this discussion that these geomorphic forms cannot develop in ways other than this sequence, but moreover that the cyclicity attributed to the thaw lakes can be taken a step farther, and applied to many more of the features indigenous to the North Slope as well. As long as the climate remains cold enough to perpetuate continuous permafrost, these processes and forms will continue to evolve.

IV. STREAM AUFEIS

Aufeis fields, also known as icings or naleds, are common occurences in rivers on the North Slope of Alaska. Aufeis forms during the fall and winter seasons from water that flows to the surface from a spring, or overflows from an ice-covered river channel. Even during the coldest winters some groundwater continues to flow (Carey, 1973). The impervious layer created by the presence of permafrost restricts water flow at the base; during freeze-up the river freezes downward thus trapping any unfrozen water between the ice cover and the permafrost. Hydrostatic pressure is created when obstruction to flow occurs on the sides, for example when a river channel narrows and becomes obstructed by ice. In a stream the unfrozen water is forced up through cracks in the ice where it freezes upon exposure to the air. Layers of ice are created in this way during the winter until the groundwater supply is exhausted (Carey, 1973).

Aufeis can form on the tundra from spring water but may also form in a river channel from river and/or spring water. This is the more common and more extensive type of formation on the North Slope, in wide shallow streams (Grey and MacKay, 1977). Braided channels characteristic of the tundra in northern Alaska are good sites for aufeis formation. Aufeis is particularly common at the transition zone between the northern Brooks Range and the foothills where the river channels become more shallow, wider and more numerous upon reaching an area of less relief (Holmgren and Benson, 1974). Shallow channels often freeze to the river bed thereby trapping water between the ice cover and permafrost (Figure 8). Aufeis fields may enhance the braiding of streams by diverting stream channels in the summer when the aufeis mound is still extensive. Blocking, underflow and overflow of stream water is thus induced (Harden et al., 1977). In

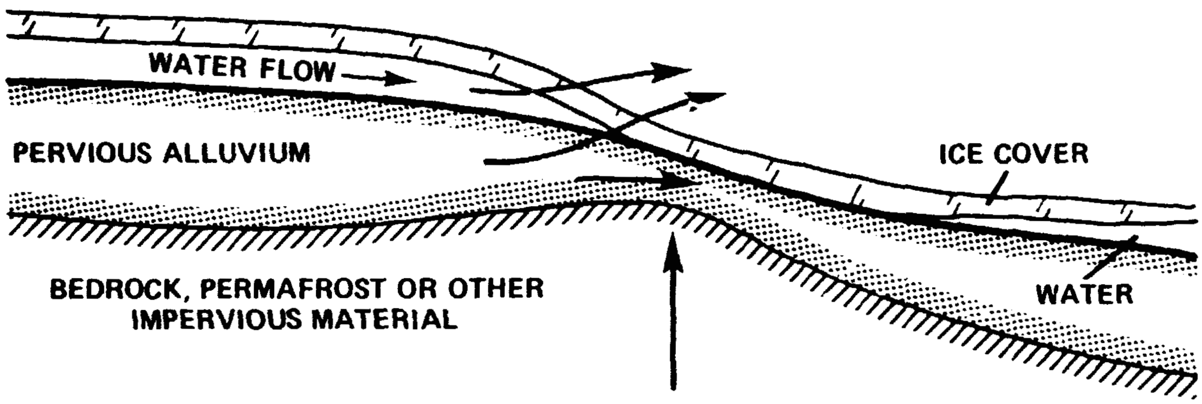

Figure 8. Profile of a Stream Partially Frozen to the Bottom and Conducive to Aufeis Formation (after Carey, 1973)

this way aufeis may contribute to the alteration of the character of the stream channel and tends to be self-perpetuating by enhancing braiding of stream channels.

The largest aufeis fields are those fed by springs discharging in a river bed rendering a continuous supply of source water available for aufeis growth throughout the winter (Grey and MacKay, 1977). For example, the Canning River aufeis field in northeastern Alaska near the coast is fed by a spring and is large aufeis field which forms in the same place each year (Childers et al., 1973). It can be seen clearly on spring and summer Landsat satellite imagery every year since 1972 and varies considerably in maximum extent between years. Measurements of Canning River aufeis using Landsat 80m resolution data reveal a $22.2 km^2$ difference in maximum extent between 1973 ($27.0 km^2$) and 1974 ($4.8 km^2$) (Hall, 1976).

The thickness of the aufeis depends upon the rate of source water seepage or flow as well as the temperature of the air. Very cold air temperatures will cause a thick layer of ice to form close to the point from whic. › water emanates due to rapid freezing upon contact with the air. Faster moving outflows occurring during warmer weather travel farther downstream before freezing and form a thinner layer of overflow ice.

The structure of the ice comprising aufeis fields is layered, resulting from successive overflowing of water. The ice is commonly white and bubbly as a result of entrapped air (Carey, 1973). Air bubbles are more numerous in aufeis layers which freeze at a rapid rate (during colder weather) because the air does not have time to escape prior to freezing (Krietner, 1969).

Pervious surficial materials are necessary for aufeis development because upward groundwater flow must occur. Topographic setting is also important. Groundwater flow rates are higher in steeper terrain creating greater hydrostatic pressure (Grey and MacKay, 1977). At the junction of mountainous and flat terrain, high rates of flow over flat areas allow braiding of channels. These braided channels are generally shallow and freeze to the bottom in the winter creating hydrostatic pressure due to the high flow rates. On south-facing slopes the permafrost table is generally at a greater depth than on north-facing slopes because greater insolation is received on the south slopes. South slopes have a thicker active layer and are capable of retaining more suprapermafrost water allowing a greater supply of water for aufeis formation. Aufeis on north-facing slopes may develop sooner because of the shallower active layer but without an adequate groundwater supply the aufeis fields will not become as extensive as aufeis forming on south-facing slopes.

It is interesting to note the similarity in the formation between aufeis and open system pingos. Both can be formed and perpetuated by subsurface springs. Strong subsurface springs produce aufeis where water flows to the surface and weak subsurface springs produce pingos where water pressure causes updoming of surface sediments (Muller, 1959).

The climatic controls over aufeis formation are more variable than the hydrologic, geologic and topographic controls. Qualitatively it can be stated that aufeis forms in areas with long, cold winters. The presence of permafrost is not a requirement but the best developed and largest aufeis fields develop in permafrost regions. The best documented area for extensive aufeis formation is in Siberia (Tolstikhin and Sokolov, 1972) where aufeis fields have been reported which are several kilometers long and several kilometers wide. One was reported to be $26 km^2$ in area and to contain $39,000,000 m^3$ of ice (Muller, 1959).

The date and rapidity of fall freeze-up is important for the development of [illegible] as is the timing of precipitation. In order for extensive aufeis formation to occur in a particular year, rainfall must occur in late summer and early fall in order to recharge the groundwater sources prior to freeze-up (Grey and MacKay, 1977). Then, if only a small amount of snowfall occurs in the first half of the winter, the rivers will freeze deeply, possibly to their beds because of a lack of insulating snowcover to retain ground heat. If substantial late winter snowfall occurs the ground will be insulated and thus retain a thermal regime conducive to continued aufeis growth (Carey, 1973). However if groundwater recharge does not occur and if an early heavy snowfall occurs, the rivers may not freeze deeply allowing little or no aufeis to form in a particular year.

As aufeis is forming in the fall and winter it acts as a heat source because heat is generated during the freezing of water (Gavrilova, 1972). Heat is then required in order to melt the aufeis in the spring. Aufeis tends to warm the underlying soil partly because it releases heat, but mainly because it insulates the ground against the cold winter temperatures in the same way that insulation from glacier ice can prevent or retard the development of permafrost.

Not only is it important to locate aufeis but also to monitor its annual variations which may be considerable. Harden et al., (1977) analyzed the distribution of aufeis in northern Alaska and found that aufeis is more abundant to the east of the Colville River than to the west. T y detected some aufeis fields on Landsat imagery that were perennial and in the next freeze-up period provided a nucleus for future aufeis development.

V. CONCLUSION

Prominent features comprising the North Slope terrain are polygonal ground, thaw lakes, pingos and aufeis fields. The existance of continuous permafrost allows these features to develop and evolve. Some of these features such as patterned ground and aufeis can form in non-permafrost zones as well but are best developed in the zone of continuous permafrost. The character of the active layer is very important to the formation and extent of development of the geomorphic features. The thickness, composition and ice content of the active layer all affect the formation of the permafrost features. In the Arctic Foothills and Brooks Range, exposure, orientation and elevation influence the thickness of the active layer. The Arctic Coastal Plain Province is very flat and the geomorphic features are characterized by more cyclic behavior. For example geomorphic evidence of the thaw lake cycle is ubiquitous with remnent lake basins, filled basins and draining basins clearly visible on the terrain of the coastal plain. Aufeis is an important feature because it is the surface expression of groundwater flow and reflects seasonal climate characteristics. It is also a geomorphic agent which may alter stream channel behavior.

The interactions between surficial processes and climate act together to initiate and perpetuate features such as thaw lakes, ice wedge polygons, pingos and aufeis deposits on the North Slope.

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