GLACIERS IN THE ARCTIC[1]

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GLACIERS are an important element of the arctic and subarctic environment. They strongly influence meteorological and climatological conditions and have local economic significance. Glaciers are delicately responsive to subtle climatological changes, and their behaviour provides a useful means of recognizing the nature and scale of these changes depending upon the degree to which the influence of the various glaciometeorological elements is understood (Wallén, 1948; Hubley, 1954; Orvig, 1954). Useful data on mean annual temperatures in remote regions can be obtained from measurement of thermal regimen in polar glaciers (Ahlmann, 1953, p. 3). Academically, study of arctic glaciers will be especially valuable if Ahlmann (1953, p. 5) is right in supposing that the great Pleistocene ice-sheets were essentially polar in behaviour and temperature regimen. Recent interest in the nature and origin of "ice islands" in the Arctic pack (Fletcher, 1950; Koenig *et al.*, 1952; Crary, Cotell, and Sexton, 1952; Montgomery, 1952; Crary, 1954; Debenham, 1954) demonstrates the need for study of arctic glaciers for practical and strategic reasons.

The size, nature, and distribution of arctic and subarctic glaciers are treated descriptively in this compilation, which is necessarily based largely on the writings and observations of others. Essentially nothing offered is new or original, and, in spite of references sprinkled liberally through the text, it is not possible to acknowledge every word or thought. Recent air photographs will certainly result in modifications of the material presented.

This compilation was prepared in 1949 for the *Encyclopedia Arctica,* a project of the Stefansson Library supported by the Office of Naval Research. It is now published in *Arctic* with the generous consent of Vilhjalmur Stefansson. Improvements have resulted from the suggestions of his staff, especially in the section on Siberia. The assistance of Hugh S. Bostock in preparation of the section on the Cordillera of Western Canada, and of Moira Dunbar in the sections on the Canadian Arctic islands is much appreciated. The manuscript has benefited by the editing of Miss A. V. Moffat.

Classification of Glaciers

Glacier classifications are diverse in philosophy and basis. Each benefits from its predecessors, so attention is directed to one of the latest. Distinctions based on morphology are of greatest use here, and the classification presented is comprehensive in this respect.

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Morphological Classification of Glaciers

(After Ahlmann, 1940a, pp. 192-193)

A. Glaciers extending in a continuous sheet, the ice moving outwards in all directions.

1. *Continental glaciers or Inland ices*: covering very large areas (e.g., Greenland and the Antarctic).
2. *Glacier caps* or *ice caps*: covering smaller areas than Continental glaciers (e.g., Vatnajökull in Iceland, Jostedalsbreen and most other large Norwegian glaciers; this group includes what by other authors are called Plateau Glaciers, Island Ices, Highland Ices, Icecaps).
3. *Highland glaciers*: covering the highest or central parts of a mountain district, from which ice streams issue through the valleys (e.g., in the interior of Spitsbergen, particularly New Friesland).

B. Glaciers confined to a more or less marked path, which directs the main movement. This group includes both independent glaciers and outlets of ice from glaciers of group A.

4. *Valley glaciers* (of Alpine type): occupy only the deeper parts of the principal valleys and obtain their supply from the heads of the valley-system (e.g., the alpine glaciers).
5. *Transection glaciers*: the whole valley system is more or less filled by ice, which overflows the passes between the valleys (e.g., the Yakutat Glacier in Alaska, most of the glaciers in the interior of Spitsbergen, and in the Alps during the last glacial period; called "Eisstromnetz" by Drygalski and by O. Nordenskjöld the Spitsbergen type; include both the Reticular and Dendritic Glaciers of Tyrrell).
6. *Circus (Cirque) glaciers*: localized to separate niches on a mountain side, or to the uppermost part of a valley (e.g., large numbers of small glaciers in the Alps and other mountain ranges, as well as in Norway; called cwm glaciers by Hobbs and others).
7. *Wall-sided glaciers*: covering the side of a valley or some part of it which is not furrowed by any marked niche or ravine (in Spitsbergen these are called Stufenvereisung by Drygalski and Flankenvereisung by Philipp).
8. *Glacier tongues afloat*: an ice stream more or less afloat at the shore of the ice-covered land.

C. Glacier ice spreading in large or small cake-like sheets over the level ground at the foot of high glaciated regions.

9. *Piedmont glaciers*: formed by a fusion of the lower parts of two or more independent glaciers of types 4, 5, or 7 (e.g., the Malaspina Glacier in Alaska; this group also includes Priestley's Confluent Ice).
10. *Foot glaciers*: from the lower and more extended parts of glaciers of types 4, 5, and 7.
11. *Shelf-ice*: connected to a glaciated inland, but receiving most of its supply from snow accumulating on it and recrystallized into a firn-like mass. It either floats on the sea or covers coastal shallows, in the latter case largely resting on the bottom (e.g., the Ross Barrier).

Refinement and further subdivision of these classes are possible on the basis of area-distribution curves (Ahlmann, 1948, p. 61). Matthes' (1942, p. 152) intermontane glacier, an ice mass occupying a spacious trough between separate mountain ranges or groups, is an additional type well represented in Alaska.

Glaciers may be distinguished on a dynamical basis as: *active*, *inactive*, or *dead* (Ahlmann, 1948, p. 63), or by their mode of flowage as *pressure-controlled* or *gravity-controlled* (Demorest, 1943, pp. 365–373; Matthes, 1942, pp. 150–153). A geophysical classification (Ahlmann, 1948, p. 66) can be made on a basis of temperature regimen, meltwater behaviour, and firn condition as follows: *temperate glaciers* attain a pressure-melting temperature throughout sometime during the ablation season, but a thin surficial layer is chilled below freezing in winter. *Polar glaciers* have a subfreezing temperature even in summer to a considerable depth, at least in the accumulation area. In *high-*

polar glaciers no meltwater forms even in summer, but in *sub-polar glaciers* some surface melting and meltwater percolation do occur.

Distribution, Area, and Volume of Arctic Glaciers

The greatest ice mass outside Antarctica is the Greenland Ice Sheet. Neighbouring islands of the Canadian Arctic west of Greenland also have sizable bodies of ice (Fig. 1). East of Greenland are the glacier-bearing islands of Iceland, Jan Mayen, Svalbard, Novaya Zemlya, and the less well known archipelagoes of Franz Josef Land and Severnaya Zemlya (Fig. 2). Of these, Franz Josef Land has the most complete ice cover. Among the continental areas considered, southern Alaska and adjacent parts of western Canada are by far the most extensively covered. Scandinavian glaciers, the largest in Europe, are small by comparison, and scattered glaciers in the Urals, Siberia, and other parts of Canada are still smaller.

The present development of glaciers in the Arctic is clearly related to sources of moisture, storm paths, and topography, more or less in that order of importance. Thus, the major glacier-bearing areas of the Arctic and Subarctic or adjacent to seas relatively free of floe ice for part or all of the year, and North Atlantic glaciers border the great re-entrant eaten into the arctic ice pack by the Gulf Stream.

The following compilation on ice-covered arctic and subarctic areas is taken principally from Hess (1933, p. 121), Thorarinsson (1940, p. 136), and Flint (1947, p. 39). It is supplemented in the Canadian Arctic by planimetric measurements made on recent maps and air photographs.

Areal Distribution of Arctic and Some Subarctic Glaciers

Region	*Area in square miles*	
Greenland		
Inland ice-sheet	666,000	
Independent ice bodies	63,000	
Total		729,000
Iceland	4,655	
Jan Mayen	45*	
Svalbard		
North East Land	4,340	
Other islands, chiefly West Spitsbergen	18,060	
Total		22,400
Franz Josef Land	6,560	
Novaya Zemlya	11,000	
Severnaya Zemlya	6,400*	
Canadian Arctic Archipelago		
Ellesmere Island	32,000*	
Axel Heiberg Island	5,170*	
Devon Island	6,270*	
Bylot Island	2,000*	
Baffin Island	14,100*	
Other small islands	300	
Total		59,840*
Scandinavia	2,400	
Continental North America	30,890	
Grand total		873,190

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*Figures marked with an asterisk throughout this paper were determined by planimetric measurements on recent maps or air photographs.

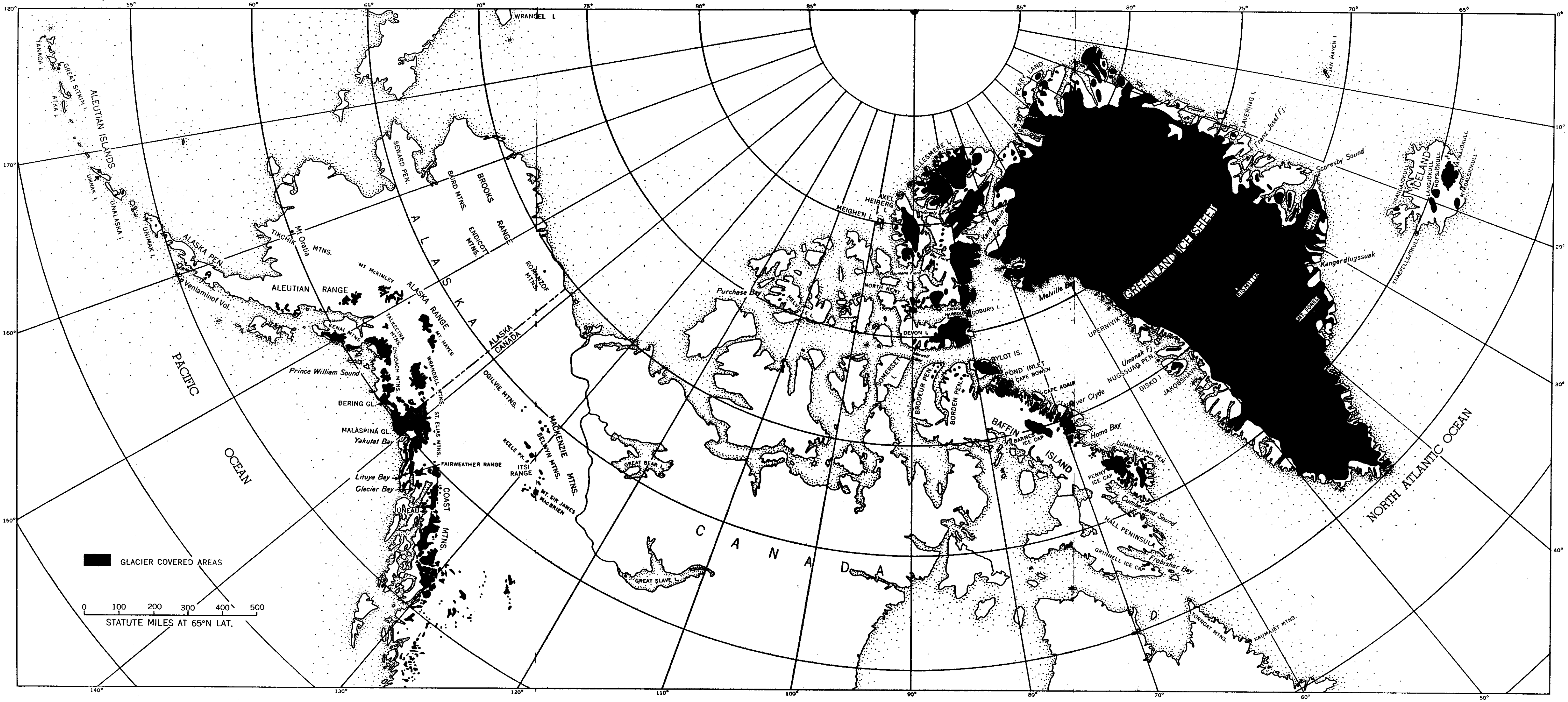

Fig. 1. Location map of arctic glaciers in the Western Hemisphere.

This 873,190 square miles constitutes approximately 15 per cent by area of the present ice cover on the earth. Thickness measurements in arctic and subarctic glaciers are limited (Brockamp, Sorge, and Wölchen, 1933; Goldthwait, 1936; Poulter, Allen, and Miller, 1949; Joset, 1952; Littlewood, 1952; Allen and Smith, 1953; Röthlisberger, 1953; Holtzscherer, 1954a), so volume figures are controlled guesses at best. Daly (1934, p. 12) assumed an average thickness for the Greenland Ice Sheet of 3,280 feet, but some 400 seismic soundings by the French Greenland expeditions give a mean of about 5,000 (Bauer, 1954, p. 46) or 5,500 feet (Holtzscherer, 1954a, p. 24) depending upon the criteria used to define the boundaries of the sheet. A thickness of 5,250 feet is probably a good figure (Holtzscherer, 1954b, p. 197). If the rest of Daly's thickness estimates are correspondingly low, the total volume of ice in arctic and subarctic glaciers is in the neighbourhood of 900,000 cubic miles. This is consistent with Cailleux's (1952, p. 9) suggestion, partly inspired by new data on the thickness of the Antarctic ice-sheet (Robin, 1954, p. 198), that the total volume of glacier ice on the globe is probably considerably greater than formerly estimated.

The Modern Regime of Arctic Glaciers

With some exceptions, especially in Alaska, glaciers in arctic and subarctic regions have been shrinking for many decades and mostly at accelerating rates (Thorarinsson, 1940). In areas bordering the North Atlantic this shrinkage, starting at some places 200 years ago and at others within the last 50 to 60 years, has been nearly universal and locally catastrophic. Many Alaskan glaciers have behaved in similar fashion, but others are now at their most advanced position in centuries (Field, 1932; 1937, pp. 69–76; Cooper, 1942). Most North Atlantic glaciers are still receding, but a number of Alaskan glaciers are advancing, have advanced recently, or have been in equilibrium for several years (Baird and Field, 1951, pp. 121–127; 1954).

The modern glacier recession in North Atlantic areas (Ahlmann, 1946, p. 24; Eythórsson, 1949b) has been attributed to the "recent climatic improvement" involving a rise in winter, spring, and autumn temperatures, and to less marked degree in summer temperatures (Liljequist, 1949). Climatic amelioration is further suggested by the decrease in average thickness of the arctic ice pack from 144 inches to 86 inches between 1893–95 and 1937–40 (Ahlmann, 1946, p. 23). The navigation season to Svalbard has become longer, birds and fish are now found farther north, frozen ground is deteriorating in many areas, and vegetative growth has increased (Hustich, 1949). This climatic improvement is attributed to increased northward circulation of warm air, brought about by changes in atmospheric pressure gradients (Petterssen, 1949), and to an accompanying secondary effect produced by the Gulf Stream System (Helland-Hansen, 1949). The North Atlantic low, and perhaps also the North Pacific low, have moved farther north to produce these conditions. This is a grossly over-simplified statement, but, regardless, a rise in temperature more than a change in precipitation has been the major result, and temperature is the factor most strongly affecting the behaviour of these glaciers (Ahlmann, 1940a, p. 190).

This climatic amelioration culminated in late 1930s or early 1940s (Kincer, 1946, p. 342; Hustich, 1949, p. 103) and was followed by more continental conditions (Ahlmann, 1953, pp. 22–27). Most glaciers of the North Atlantic region as yet display no marked reaction to this latest change, but the erratic and diverse behaviour of Alaskan glaciers may be an early response to the new conditions. Whether this is going to be a relatively minor perturbation in the history of general shrinkage or heralds a major reversal in behaviour is for the future to tell. Unfortunately, means of evaluating glacier fluctuations quantitatively are still inadequate. Data on total regime and over-all shrinkage of glaciers are more significant than fluctuations of the margins, the type of information usually available. The still scanty knowledge of the influence of various meteorological factors on accumulation and wastage makes evaluation of climatic fluctuations through glacier changes difficult and uncertain.

Greenland

The greatest mass of ice in the Northern Hemisphere and the earth's best known continental glacier is the Greenland Ice Sheet (Fig. 1). Since the first crossing by Nansen in 1888, it has been the subject of repeated exploration and investigation, especially within the last decade. Greenland is richly endowed with a variety of other ice bodies, including independent ice caps and highland, outlet, valley, cirque, wall-sided, expanded foot, tidal, and floating glaciers, and in places an ice foot. Of all glaciers listed in Ahlmann's morphological classification (1940a, pp. 192–193), only piedmont and transection glaciers were apparently lacking in Greenland. Now at least one piedmont has been reported (Flint, 1948, p. 134), and transection glaciers are described from high mountainous areas on the east coast (Drygalski and Machatschek, 1942, p. 184).

The best figure for the area of the Greenland Ice Sheet, which covers 79 per cent of the land, is probably the 665,000 square miles cited by Bauer (1955b, p. 457). The total ice on Greenland including independent glaciers and ice on islands is estimated at 700,000 to 715,000 square miles by Matthes (1942, p. 159), 733,000 square miles by Hess (1933, p. 12), and only 696,000 square miles by Bauer (1954, p. 34). The largest amount would cover close to 87 per cent of the land, the least 82.5 per cent. The sheet is about 1,570 miles long and close to 600 miles wide. Elevation of its highest point is still unsettled. Late maps show an elevation of 10,325 feet at 69°49′ N., 37°52′ W. The British-Trans Greenland Expedition (Lindsay, 1935, pp. 402, 406) reported 10,400 feet (uncorrected) on the ice 20 miles north of Mount Forel and stated that the highest point attains 10,500 feet. Flint (1947, p. 40) speculates that the highest point may exceed 11,000 feet, but Bauer (1954, p. 33) indicates that the maximum lies between 10,500 and 10,800 feet along the meridian 37°30′W. between 71°30′ and 73°00′N. The mean height is said to be about 6,900 feet by Löwe (1936, p. 317) and 7,000 feet by Bauer (1954, p. 35). The highest point of land in Greenland, so far as known, is a rocky peak of 12,139

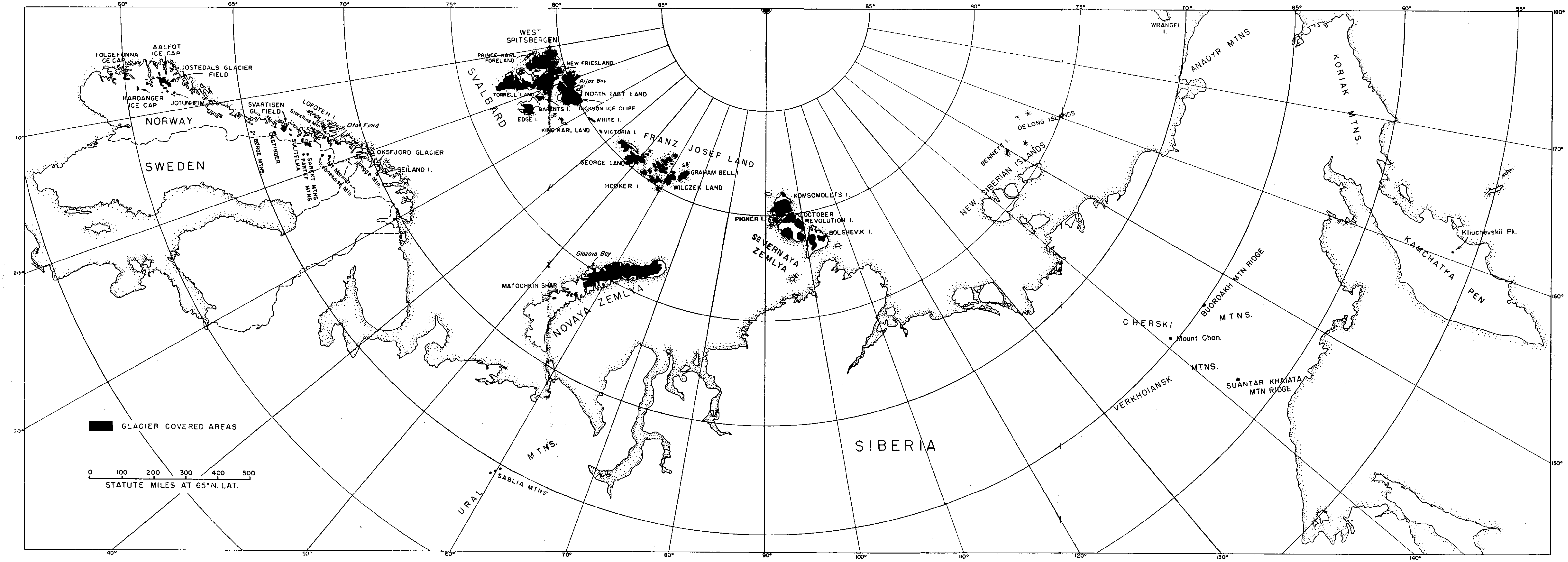

Fig. 2. Location map of arctic glaciers in the Eastern Hemisphere.

feet in the Watkins Mountains at 68°54′ N., 29°49′ W. (Courtauld, 1936, p. 202). Bauer (1955b, p. 460) estimates that the accumulation area covers 83.5 per cent of the ice-sheet and that the mean elevation of the firn limit is 4,560 feet.

The Wegener expedition made seismic soundings through the ice at various points up to 250 miles inland from the west coast (Sorge, 1933, p. 335). Thicknesses ranging from 150 to 6,000 feet, and possibly more, were presumably indicated, although the reliability of the results has been questioned (Drygalski and Machatschek, 1942, p. 29; Demorest, 1943, pp. 383–386; Ahlmann, 1941a, p. 157). An extensive program of seismic sounding in the Greenland Ice Sheet undertaken by the French Polar Expeditions provides extremely valuable information concerning thickness of the ice, configuration of the subglacial floor, and the relationship of these elements to surface form (Holtzscherer, 1954a). Ice thickness was determined at about 400 points in the southern half of Greenland yielding a median value of 5,550 feet and a maximum of 11,180 feet, the latter near the centre (38°W., 69°50′N.). Additional soundings in northern Greenland (Simpson, 1955, p. 287; Bruce and Bull, 1955, p. 892) show the ice to be at least 9,700 feet thick and indicate that Nye's (1952) calculation of greater ice thicknesses in the northern part of the sheet is probably incorrect. The French seismic data demonstrate that the subglacial floor is indeed like a great saucer with a large area in the interior extending below sea-level to a maximum of -1,300 feet (Bauer, 1954, pp. 43–44). A more accurate calculation of volume for the ice-sheet is now possible, giving 646,000 cubic miles (Holtzscherer, 1954a, p. 24; 1954b, p. 197).

The crestal ridge of the ice cap is near the centre in the north but lies much closer to the eastern margin in the southern two-thirds. It has two independent summits, a smaller one in the southernmost part and a larger one about half-way up the east coast. The cause and origin of the crestal ridge are matters for speculation. Demorest (1943, pp. 378–386) supports the view that the crest reflects a high in the underlying subglacial topography, and others maintain that it marks the site of the thickest ice (Wager, 1933, p. 154). French seismic results show that the southern summit does indeed overlie a high part of the subglacial floor. However, the rest and larger part of the crestal ridge bears no relation whatever to the subglacial topography (Holtzscherer, 1954a, p. 22). The thicker ice under the crest may be due in part to a restraint afforded to easy outflow eastward by the high eastern mountains (Holtzscherer, 1954a, p. 22), but this cannot be the whole story. The crestal ridge may mark the site of maximum accumulation, but if so, it is a relic feature for the region of maximum accumulation now lies much farther west (Hess, 1933, p. 115; Sorge, 1933, p. 335; Benson, 1954, pp. 11–12).

A broad depression extends transversely across Greenland between 66°N. and 69°N. separating the crestal summits described above (Koch, 1923b, pp. 47, 55). This depression terminates at both ends in some of the largest and most active outlet glaciers in Greenland. French seismic surveys suggest that it is in part the expression of "canals" or valleys in the bedrock floor, which extend far into the interior of Greenland and localize the rapid outflow of ice thus lowering the surface (Holtzscherer, 1954a, p. 22).

The surface of the ice-sheet slopes gently away from the crestal ridge at 5 to 50 feet per mile. Even though the surface in the interior appears relatively smooth and featureless (Fig. 3), it actually consists of a gentle but varied topography of hills, ridges, basins, and terraces with a relief measured in many tens of feet (Benson, 1954, p. 6). This micro-relief may be the product of local differences in rates of accumulation, wind drifting, wastage, and movement. If so, it is dynamic and changes from time to time. Near the edges the slope steepens appreciably and a more varied relief develops. On the west side, heads of great outlet glaciers are marked by depressed or draw-down areas termed "basins of exudation" by Peary. The basins extend at least 85 miles inland from the edge of the sheet, and crevasses indicating flowage into them are found 40 to 100 miles farther inland. On the east coast depressions at the head of the great Kangerdlugssuak Glacier (69°N.) and at the head of Waltershausen Glacier in the Franz Josef Fiord region may be of this origin. More commonly, ice along the east coast piles up behind the high coastal mountains and spills over high passes (Wager, 1933, pp. 149–150; Lindsay, 1935, p. 407). Here are great hollows 500 to 1,000 feet deep, steep sided, flat floored, 5 to 10 miles across, and separated by rounded ice ridges (Lindsay, 1935, p. 400), which may reflect the influence of underlying bedrock topography, for seismic soundings show that it is rugged (Holtzscherer, 1954a, p. 22). In places the ice-sheet exhibits a series of marginal terrace-like steps (Hess, 1933, p. 112) that have been attributed to the configuration of the bedrock floor (Hobbs, 1911, p. 129; Mecking, 1928, p. 252) or to large-scale sliding and slumping (Demorest, 1943, p. 394).

Inland ice reaches the sea at Melville Bay on a 240-mile front and in two places along the northeast coast between 78° and 80°N. for a total of 130 miles. Humboldt Glacier emptying into Kane Basin is usually described as a gigantic outlet glacier with a front 60 to 70 miles long, but it could justifiably be considered a part of the inland ice reaching the sea directly.

Greenland provides a fine display of outlet glaciers. Some are huge; Petermann Glacier debouching into Hall Basin is 15 miles wide and at least 60 miles long, Waltershausen Glacier draining to Franz Josef Fiord is about 10 miles wide and 75 miles long, and there are others nearly as large (Teichert, 1934). Most are tidal, and many are afloat. Outlet glaciers of the Disko-Umanak-Upernivik area on the west coast and of the Scoresby Sound region on the east coast produce bergs in great fortnightly cataclysms. Total annual output of bergs in west Greenland is estimated at 57 cubic miles of ice (Bauer, 1955b, p. 462). The great berg-producing glaciers are those with consistently high velocities, locally called "running glaciers". Velocities up to 124 feet per day have been recorded on Upernivik Glacier (Carlson, 1939, p. 247), and diurnal variations in flow rates are large (Sugden and Mott, 1940, p. 45). Daily movements on the inland ice by contrast amount to fractions of an inch (Hobbs, 1911, p. 135) to 3 inches at most (Bauer, 1954, pp. 49–50).

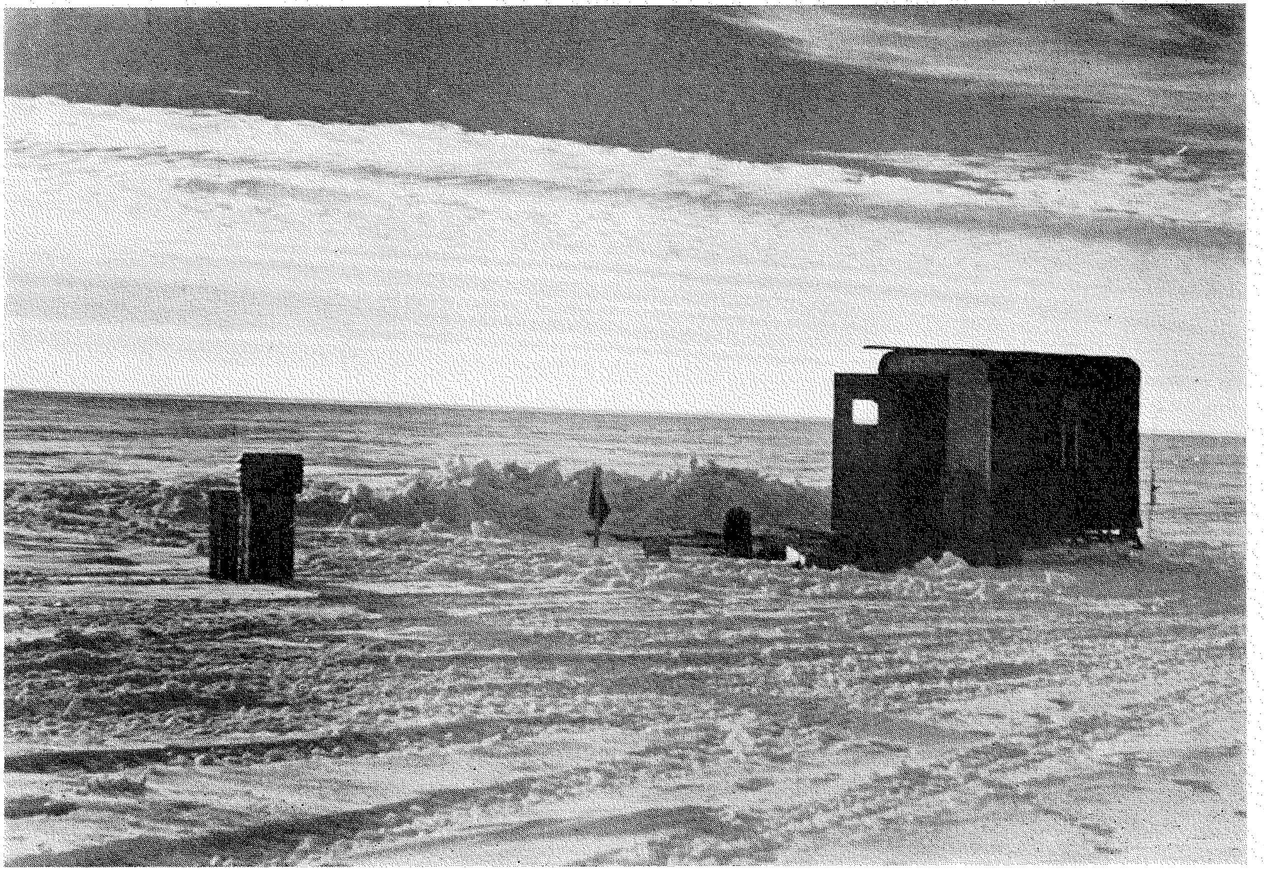

Fig. 3. Interior of Greenland Ice Sheet at 77°N. lat., 60°W. long. Photo by Carl Benson, reproduced by permission of Snow, Ice and Permafrost Research Establishment.

The largest areas of land not covered by inland ice are to the north in Peary Land (Koch, 1923a; Winther *et al.*, 1950), on the west coast between 66° and 68°N. where the sheet lies as much as 100 miles inland (Mott, 1937, p. 313), and in a strip 80 to 180 miles wide along the east coast south from 78°N. (Bretz, 1935, p. 159). Within areas not covered by inland ice are highland glaciers, independent ice caps, outlet glaciers, transection glaciers, and cirque and valley glaciers numbered in the hundreds. Some of these may not be detached remnants of the Pleistocene ice-sheet but features reborn following the shrinkage of the postglacial warm-dry period (Matthes, 1942, p. 208), for Demorest (1937, pp. 54–5) reports névé fields on Nugssuaq Peninsula that are probably of postglacial origin.

The margins of many Greenland glaciers are steep or overhanging ice cliffs (Chamberlin, 1895, p. 565). These have been attributed to differential ablation related to the low angle of the sun and to the large amount of debris in the lower ice layers (Chamberlin, 1895, p. 566; Bretz, 1935, pp. 180, 197), or to overriding by the higher ice (Odell, 1937, pp. 115–116). The sheared and fractured condition of ice in the basal parts of Greenland glaciers (Chamberlin, 1895, pp. 676–677; Bretz, 1935, pp. 180–182) is thought to be due partly to the low temperature (Löwe, 1935, p. 268; Odell, 1937, p. 124).

On a geophysical basis, Greenland glaciers are classed as polar and sub-polar. Thermal observations at Eismitte (Sorge, 1933, pp. 339, 341) and subsequent calculations (Hess, 1933, p. 113; Wegener, 1936, p. 171) indicate that the central part of the Greenland Ice Sheet is polar, but the margins are probably sub-polar. The Fröya and other low-level glaciers in northeastern Greenland are sub-polar (Ahlmann, 1946, p. 21), and Sukkertoppen Ice Cap is almost temperate (Sugden and Mott, 1940, p. 47).

Bauer (1954, p. 48) feels that the inland ice is too cold to be in thermal equilibrium with the present climate. The slow movement of the ice in spite of great thickness, the speed of seismic waves, and a 200-metre subglacial layer with the seismic velocity of frozen ground all suggest to Bauer that the ice may be at a subfreezing temperature throughout its entire thickness.

Matthes (1942, p. 156) calculates that ice now appearing at the edge of the Greenland Ice Sheet is at least 10,000 years old, and recent figures on rates of movement suggest that it may actually be much older. Bauer (1954, p. 40) calculates from the present material deficit of the inland ice that it could not have been built up in less than 100,000 years, and Wager (1933, pp. 154–155) is inclined to think that the ice-sheet originated in the Miocene.

From reconnaissance inspection, the inland ice has been said to be about in a state of balance (Löwe, 1936, pp. 317–329; Demorest, 1943, p. 398). Koch (1926, p. 105) observes that the ice-sheet in northern Greenland has varied little, and Reid (1898, p. 474) reports a description of the inland ice as it appeared in 1200 A.D., which would apply well to modern conditions. The present rate of rise of sea-level suggests that the ice-sheet is not melting as rapidly as other glaciers in the Northern Hemisphere (Ahlmann, 1948, p. 75). These generalizations do not fit with an estimated annual deficit of about 100

km^3 of water in the present regime of the inland ice (Bauer, 1954, p. 40; 1955b, p. 462). However, Bauer may have underestimated both the size of the accumulation area and the amount of accumulation, and, as he says (Bauer, 1955b, p. 462), the ice-sheet must be very sensitive to fluctuations in accumulation because the accumulation area constitutes 83.5 per cent of the total.

Some outlet glaciers and independent ice bodies have been more sensitive to climatic variation. Considerable shrinkage of ice in Greenland probably occurred during the postglacial warm-dry period (Ahlmann, 1941b, p. 198; Matthes, 1942, p. 208), and moraines a short distance beyond present glacier margins are thought to represent the greatest readvances of the postglacial period. These probably occurred between the middle eighteenth and the middle nineteenth centuries (Ahlmann, 1941b, pp. 199–202). Flint (1948, pp. 139–140, 147, 159) found evidence of two glacier advances separated by a considerable deglaciation in northeastern Greenland, the youngest possibly occurring during the nineteenth century. Advances of other Greenland glaciers in the 1850s and 1890s are recorded (Thorarinsson, 1940, p. 146; Matthes, 1942, p. 194), but since the 1890s recession has predominated. Fröya Glacier on Clavering Island was 22 per cent larger by area and 60 per cent larger by volume during its recent *Hochstand* than now (Ahlmann, 1941b, p. 203). Upernivik Glacier receded a maximum of 5,000 feet between 1887 and 1931. Jakobshavn Glacier receded 6 to 8 miles between 1851 and 1902 in spite of short-lived advances (Carlson, 1939, pp. 249–254). Glaciers in the Umanak area have been receding for 40 years (Löwe, 1934). Between 1869 and 1922 Pasterze Glacier in northeastern Greenland receded 3.9 miles (Flint, 1948, p. 125), and another glacier in this area disappeared completely. Considerable recession has occurred in other parts of eastern Greenland (Bretz, 1935, pp. 159–160, 183; Gabel-Jørgensen, 1935, p. 40; Mikkelsen, 1933, p. 388; Fristrup, 1950, p. 11), and a negative regime is reported for the Chr. Erichsen Glacier in Peary Land (Fristrup, 1950, p. 11).

A few instances of recent advance have been recorded. Taterat Glacier of west Greenland showed a lateral expansion of 100 feet in 1 month during the summer of 1938 (Sugden and Mott, 1940, p. 51), and in the Cape York (76°N.) district many glaciers advanced until 1920 (Koch, 1926, p. 107). In north Greenland in the 1920s were 15 advancing, 8 stationary, and 9 receding glaciers.

Studies of radiation and meteorological factors controlling ablation have been made in both east (Ahlmann, 1942) and west Greenland (Etienne, 1940; Sugden and Mott, 1940). The large amount of convective heat and the relatively minor role of radiation (8.2 per cent of total heat) supplied to Fröya Glacier in late summer is noteworthy (Eriksson, 1942, p. 39).

Iceland

Although Iceland is subarctic, its glaciers are treated here because they are relatively well known and are representative of a marine environment. The total area of ice approaches 4,700 square miles (Hess, 1933, p. 121), covering one-eighth of Iceland (Flint, 1947, p. 52). This is pre-eminently a land of ice caps, of which the largest, Vatnajökull (Fig. 1), covers 3,050 square miles (Hess, 1933, p. 99). The other principal caps are Hofsjökull (502 square miles), Langjökull (464 square miles), Myrdalsjökull (386 square miles), and Drangajökull (60 to 70 square miles) (Eythórsson, 1935, p. 121). Smaller caps lie in periferal positions; Langjökull has at least 4 satellites, and Myrdalsjökull and Vatnajökull have 2 or 3 apiece. Snaefellsjökull is a small isolated cap of 10* square miles in far western Iceland near Cape Önverd (Öndverdanes). Elevations on Vatnajökull range from 6,952 feet almost to sea-level. High points of other ice caps are between 3,035 and 5,581 feet, but none extends as low as Vatnajökull.

Ice tongues and outlet glaciers project from the larger caps, Vatnajökull alone having some 18, but none reaches the sea. Cirque and valley glaciers are rare except in the north between Skaga and Eyja fiords where they are favoured by rough mountainous terrain. Most of the ice caps are thin, and underlying topography exerts considerable influence on their relief and shape. Ice thicknesses in Vatnajökull as measured by seismic soundings are mostly between 1,500 and 2,500 feet with a maximum of 3,400 feet (Joset, 1952 p. 43; Joset and Holtzscherer, 1954, pp. 20–21). The subglacial topography of Vatnajökull is varied with a total relief approaching 4,000 feet (Eythórsson, 1952, p. 3). The volume of this ice cap has been calculated at 760 cubic miles (Bauer, 1955a, p. 22).

The eruption of subglacial volcanoes in Iceland attracts particular attention. Principal centres of eruption are the Grímsvötn area in western Vatnajökull, and the Katla area in south-southeastern Myrdalsjökull. In recent times, eruptions have occurred in Grímsvötn at 6- to 10-year intervals, the latest in 1954 (Thorarinsson, 1954, p. 37). The Katla area has erupted twice every century since 1580, the latest in 1918 (Thorarinsson and Sigurosson, 1947). Nielsen (1937, p. 11) distinguishes several phases in the Grímsvötn activity. The first phase is the formation of a large subglacial lake through melting by volcanic heat. Melting may precede actual volcanic outbursts (Thorarinsson and Sigurosson, 1947, p. 66). When this subglacial lake exceeds a certain level or extends to the edge of the ice cap, a great flood or *jökulhlaup* rushes forth carrying huge quantities of water, debris, and ice out over the sandy outwash areas periferal to the ice. The second phase occurs when an eruption breaks through the ice cap and ejects a huge column of ash and steam high into the air. When the eruption subsides a steaming crater lake remains, but in a year or two it is covered by ice and snow. The outbursts of meltwater sometimes create great collapse craters or "ice-caldera" (Thorarinsson and Sigurosson, 1947, p. 64). Ash erupted from Grímsvötn in 1934 served as a valuable reference horizon for subsequent studies of nourishment and wastage on Vatnajökull (Ahlmann and Thorarinsson, 1939, p. 39).

A moist maritime environment gives Icelandic glaciers, particularly in the south, an extremely high metabolic rate with large accumulation and great ablation (Ahlmann, 1939, pp. 171–188; Rist, 1952, p. 7). Some outlet glaciers are extremely active even though receding under the influence of a strong negative regime. Over a 3-year period (1936–38) Hoffellsjökull had a deficit of 332 million cubic feet of water, but nonetheless it flowed at the relatively high velocity of 2,070 feet a year (Thorarinsson, 1939, p. 202). On Hoffellsjökull and Heinabergsjökull, outlets from the southeast margin of Vatnajökull, meteorological factors (conduction-convection and condensation) outweigh radiation in causing ablation by a ratio of 60 to 40 (Ahlmann and Thorarinsson, 1938, p. 228). The regimes of these glaciers are more susceptible to variations of temperature than to changes of precipitation (Ahlmann, 1940a, pp. 188–205).

Historical fluctuations of Icelandic glaciers are well documented. From the time of colonization, about A.D. 900, to at least the fourteenth century the glaciers were less extensive than after 1700 (Thorarinsson, 1940, p. 144). An advance, starting in the early 1700s, attained a climax about 1750, to be followed by alternate periods of stagnation or recession and advance (Eythórsson, 1935, p. 134; Thorarinsson, 1943, pp. 49–50). The principal *Hochstands* recognized are in 1750/60 and 1840/50. Three lesser *Hochstands* also occurred about 1710, 1810, and 1890, and minor advances were manifested about 1870, 1910, and at the beginning of the 1920s and the 1930s. Not all glaciers reached their most advanced position during the same climax. Some were most extended in 1750, some in 1850, and still others in 1890 (Thorarinsson, 1937, p. 194). These maxima are the greatest attained within historical time and possibly within the entire period subsequent to the recession of the postglacial dry-warm period (Ahlmann, 1937, pp. 198–200).

The recession that set in during the 1890s has, with minor interruptions, been gradually accentuated until the 1930s when it became locally catastrophic. Hoffellsjökull lost at least one-third of its volume in the 45 years between 1890 and 1935 (Thorarinsson, 1937, p. 189). Glacier snouts receded 3,000 to 10,000 feet between 1890 and 1936 (Matthes, 1942, p. 193). As usual this recession was punctuated by local aberrant advances of individual glaciers, notably the outlets from Drangajökull. The glacier of Kaldalón advanced 635 feet in 1935–40, that of Reykja Fiord 2,460 feet in 1933–36, and that of Leiru Fiord 3,240 feet in 1938–42 (Eythórsson, 1949a, p. 250). At Skeidararjökull conditions are admittedly abnormal owing to frequent volcanic activities, but this ice front showed an over-all advance between 1904 and 1932. In 1929, it advanced suddenly, breaking down newly erected telephone poles. Short-lived advances have been reported for ice tongues from Vatnajökull (Wright, 1935, p. 228), and differential behaviour among ice fronts on the west rim of Hofsjökull and the southeast rim of Langjökull have been recorded (Oetting, 1930, p. 49). The erratic behaviour of Icelandic glaciers is further substantiated by the fact that in 1953, of 27 glaciers observed, 22 per cent were advancing, 4 per cent were stationary, and 74 per cent were in retreat (Eythórsson, 1954).

Jan Mayen

Jan Mayen (71°N., 8°30′W.), 375 miles north-northeast of Iceland, is a valuable glaciological reference point in the vast Norwegian Sea. Glaciers cover 45* square miles and are limited to a high volcanic peak, Mount Beerenberg (7,680 feet). Fifteen tongues protrude from a mantle of ice and snow covering the upper slopes of this cone (Jennings, 1948, pp. 168–178). Some of the tongues, particularly on the northeast, lie in valleys, but others rest on the slopes of the cone as wall-sided glaciers (Jennings, 1939, p. 128). Kjerulf, Svend Foyn, and Weyprecht glaciers, all draining the northeast slope, are the most active and reach tide water. Weyprecht Glacier, draining the inner crater of Beerenberg, is the largest and most active.

Jan Mayen, like Iceland, is an area of high glacier metabolism with large accumulation and ablation (Jennings, 1939, pp. 128–130). Above 5,000 feet, rime is an important form of nourishment owing to the maritime environment.

All glaciers except Svend Foyn and Kjerulf have moraines abandoned by recent recession, and most were retreating in 1938. The maximum advance of modern times occurred during the middle eighteenth century (Thorarinsson, 1940, p. 148). Recession from this position occurred in two phases; first a slow retreat prior to 1882–83 that left a massive outer moraine as at Kerckhoff Glacier (Jennings, 1948, p. 180), and second, a pause, recession, and readvance that produced double terminal and lateral moraines, as at the margins of South and Fotherby glaciers (Flint, 1948, p. 106). Some of the latest moraines still have cores of ice (Matthes, 1942, p. 194). South Glacier receded 2,340 feet between 1882 and 1938, and its surface was lowered 160 to 200 feet by ablation between 1883 and 1937 (Flint, 1948, p. 107).

Svalbard

The glaciers of Svalbard are relatively well known through the work of numerous exploratory expeditions, and cover about 22,000 square miles (Hess, 1933, p. 109) or 80 to 85 per cent of the land (Fig. 2). The largest island, West Spitsbergen, has three principal areas of highland ice, the outline and surface features of which are strongly influenced by the rolling relief of the dissected plateaux they rest upon at elevations between 2,000 and 3,000 feet. These are in Torell Land in the south, in the northwest, and in New Friesland to the northeast. This last mass, the largest, is a type specimen highland glacier (Ahlmann, 1940a, p. 192). Periferal to these highland ices are many small ice caps and valley glaciers. Narrow ridges and sharp peaks attain 4,000 to 5,000 feet along the west coast, giving it a strong alpine aspect. This is the area of "spitzen Berge" (Hess, 1933, p. 108). Independent valley and cirque glaciers are abundant here, and locally the heavy ice cover leads to development of transection glaciers. Outlet glaciers flow as much as 28 miles from the inland ice, and many reach the sea. Others spread out on lowlands as expanded foot glaciers or merge with their neighbours to form piedmont sheets. Two excellent piedmont glaciers occupy the east side of Prince Karl Foreland just west of West Spitsbergen (Ahlmann, 1940a, p. 199).

North East Land, the second largest island, has been the site of much glaciological research, and its relatively simple ice bodies are among the best known in this part of the Arctic. This ice is in the form of thin ice caps and highland glaciers resting on dissected plateaux. The three major bodies distinguished are: West Ice, East Ice, and South Ice. East Ice is much the largest and is separated from West Ice by an open valley joining the heads of Rijps and Wahlenberg bays. South Ice is separated from East Ice by a long, ice-filled depression, but each mass has its separate crestal dome (Ahlmann, 1933, p. 164). Among the smaller periferal caps are Glittne, Vega, Forsius, Backa, DeGeer, and Ahlmann ices (Glen, 1939, p. 6). The larger inland ices all have outlet glaciers, many reaching the sea. Laponia Peninsula harbours small cirque glaciers (Glen, 1939, p. 14), and in places along the coast is a well-developed ice foot (Ahlmann, 1933, p. 164).

West Ice covers 1,080 square miles, attains 2,000 to 2,100 feet elevation, and has a gently undulating surface with domes, hollows, and some nunataks (Glen, 1937, p. 202). West Ice has an estimated centre thickness of only 400 feet (Glen, 1939, p. 8), and the influence of underlying bedrock topography is so apparent that Glen (1937, pp. 207–208) considers it a typical highland glacier gradually disintegrating into separate domes. The inland ice ends on land in many places with a feather edge (Sandford, 1929, p. 455; Ahlmann, 1933, p. 165), and even in outlet valleys ice thickness may not exceed 1,000 feet (Glen, 1941, pp. 67–69). Lady Franklin Glacier, draining westward, is the largest outlet from West Ice, Sabine and Rijps glaciers are notable outlets to the north, and a number of glaciers pour over the south rim of the plateau into Wahlenberg Bay. Summer melting on West Ice develops through a succession of stages, ranging from a slushy firn mantle to a complex of superglacial ponds and streams on bare ice (Glen, 1941, pp. 139–140).

East Ice, the largest mass on North East Land, covers 2,150 square miles (Ahlmann, 1933, p. 168) and flows in all directions from a nearly level centre at 2,000 to 2,400 feet elevation (Glen, 1939, p. 10). The relief is more subdued than on West Ice, and this is more truly an ice cap than a highland glacier. It meets the sea in an ice cliff up to 165 feet high (Glen, 1939, pp. 10–11) and continues for 80 miles except for a brief interruption at Isis Point (Ahlmann, 1933, p. 168). This, the Dickson Ice Cliff, is said to be the largest feature of its kind outside Antarctica. Eton Glacier, draining west into Wahlenberg Bay, and Dove Glacier, draining northward, are two of the larger outlets of East Ice. Eton Glacier creates a large depression similar to the exudation basins of the Greenland Ice Sheet. The central part of East Ice has an estimated thickness of 300 to 800 feet (Glen, 1939, p. 11). The great "canals" described by Nordenskjöld (Hobbs, 1911, p. 115), extending across the southern part of East Ice, are thought to be large crevasses (Sandford, 1929, p. 463; Ahlmann, 1933, pp. 169–170).

South Ice, covering 910 square miles, with a central dome at 2,000 to 2,300 feet elevation, appears to be a true ice cap. Six outlet glaciers extend north from South Ice over the steep edge of the plateau toward Wahlenberg Bay, but only four reach tidewater. On the south, the ice cap itself reaches the sea and composes part of the Dickson Ice Cliff.

Vega Ice is a small plateau cap of 80* square miles, separated from South Ice by Erica Valley. It discharges northward through Palander Glacier into Palander Bay and southward by Rosenthal Glacier to Ulve Bay. Glittne, the largest of the small independent caps, covers 85* square miles, lies just west of Vega Ice, and discharges outlet glaciers northward into Palander Bay. The remaining caps are all smaller, and, in so far as known, none discharges to tidewater.

Other glacier-bearing islands of Svalbard are Barents and Edge, southeast of West Spitsbergen. Both have interior ice caps covering 600 to 1,200 square miles, which extend to the sea and also reach it through outlet glaciers.• Prince Karl Foreland, west of West Spitsbergen, has two highland glaciers with eastward flowing outlets reaching the sea or spreading out on coastal lowlands as piedmont sheets. The southern higher part of Great Island off the northeast corner of Northeast Land has an ice cap (Ahlmann, 1933, p. 171) that reaches the sea in a cliff along the southern margin. Much the greater part of White Island, farther east, is covered by a dome-shaped ice cap. Victoria Island, about midway between Svalbard and Franz Josef Land, is completely covered except for a narrow strip of land along its north shore (Ahlmann, 1933, p. 171). Small ice caps are shown on maps of Sven and King islands in the King Karl Land group, southeast of North East Land. Other Svalbard islands may be ice bearing, but specific information to this effect has not been found.

The ice bodies on Northeast Land have been classified as sub-polar (Ahlmann, 1933, p. 291; Glen, 1941, p. 145), although the large amount of meltwater and the isothermal condition of the firn during summer (Moss, 1938, pp. 225–227; Glen, 1941, p. 144) on West Ice suggest that it is not too far removed from a temperate condition. Thermal studies in the firn of Isachsen Plateau on West Spitsbergen (Sverdrup, 1935a, pp. 53–88) show a similar condition in some of the highland glaciers of that island. The Fourteenth of July Glacier in West Spitsbergen is temperate throughout (Ahlmann, 1935b, p. 169).

Dynamically the larger ice masses of Svalbard, and particularly of North East Land, are inactive or stagnant (Ahlmann, 1933, p. 180). The small independent ice caps are said to be mostly stagnant (Ahlmann, 1933, p. 166; Glen, 1937, p. 298), as are parts of West Ice (Glen, 1937, p. 204). Outlet glaciers and some valley glaciers are the only ice bodies showing much activity, and even these are flowing slowly. Maximum movement recorded on Fourteenth of July Glacier is 6.5 inches a day (Ahlmann, 1948, p. 57). Nordenskjöld Glacier moved 1.9 feet a day in August 1921 (Slater, 1925, p. 436), and King Fjord Glacier has attained a peak rate of 6.5 feet a day (Pillewizer, 1939).

Throughout Svalbard the glaciers are undergoing recession from a position of maximum advance attained during the first half or middle of the nineteenth century (Thorarinsson, 1940, p. 145). In West Spitsbergen (Dineley, 1954, pp. 380–383) most glaciers terminating in fiords reached this maximum position about 50 years earlier than those terminating on land (Ahlmann, 1933, p. 186). Moraines were formed by a subsequent advance in 1890 near the snout of Fourteenth of July Glacier (Ahlmann, 1935b, p. 206). Recession has been more pronounced since about 1920, and almost catastrophic in some instances since about 1930. However, it has not been uniform in place or time and has not proceeded without interruption (Ahlmann, 1933, pp. 180–185; 1935b, p. 204). Lady Franklin Glacier, for example, receded about 1.75 miles from 1899 to 1936, but part of its front showed a marked advance between 1931 and 1938.

From studies of thermal conditions on Isachsen Plateau, West Spitsbergen, it was ascertained that early summer temperatures in the firn are markedly different from place to place and that they change erratically with time (Sverdrup, 1935a). The winter's cold front penetrated to a depth of 10 metres, but by the end of July it was completely eradicated. Warming of the firn is accomplished chiefly by freezing of downward percolating meltwater, which adds to the thickness of ice bodies in the firn.

Analyses of ablation on Isachsen Plateau show that from June 26 to August 15, 1934, 56 per cent was due to radiation and 44 per cent to meteorological factors such as conduction-convection and condensation (Ahlmann, 1935a, pp. 43-52; Sverdrup, 1935b). Evaporation accounted for only 3.5 per cent. For the entire Fourteenth of July Glacier, ablation was 45 per cent by radiation and 55 per cent by meteorological factors.

Novaya Zemlya

Novaya Zemlya is bisected transversely by Matochkin Shar (Fig. 2). The southern island, low and featureless in the south, rises northward to heights of 2,800 feet where a few small glaciers lie in protected spots.

North of Matochkin Shar more rugged country attains elevations of 3,500 feet and contains many alpine glaciers. North of 74°N. the ice cover becomes heavier, taking on the characteristics of a highland glacier (Mecking, 1928, p. 157). Outlet glaciers from this inland ice reach tidewater on both coasts. One of the greatest outlets, 35 miles in length, empties into Glazova Bay on the west coast. North of 75°N. the inland ice mantles the land in the form of a cap lying on a dissected plateau 2,000 to 3,000 feet high. Several crossings of this cap have been made, and its surface is described as slightly undulating (Holtedahl, 1922, p. 374).

Large, outlet glaciers extend to the sea on both coasts, and some end in ice cliffs, 130 to 165 feet high and 8 to 9 miles long. In the northwest, outlet glaciers descend steeply to the sea from rugged country 3,500 feet high. At the northern end of the island the cap is separated from the sea by a strip of bare land several miles wide (Ellsworth and Smith, 1932, p. 74). The area of ice on Novaya Zemlya is 11,500 square miles, covering approximately 30 per cent of the land (Zubov, 1950). Average thickness is thought to be less than 800 feet, and the maximum does not exceed 1,400 feet (Flint, 1947, p. 59).

Widespread glacier recession is indicated by abandoned end moraines a mile or more from the ice (Mecking, 1928, p. 158; Lavrova, 1931, p. 131). At present the glaciers are said to be receding, thinning, stationary, and in some instances dead (Thorarinsson, 1940, p. 145). However, in 1896 the glaciers of Novaya Zemlya were expanding (Reid, 1897, p. 379). Perhaps, they were then nearing their maxima of the 1890 *Hochstand.* Lavrova (1931, p. 131) states that Novaya Zemlyan glaciers experienced intense melting and reduction within postglacial time prior to the advances of recent centuries.

Franz Josef Land

Franz Josef Land consists of at least 75 individual islands, nearly all of which have extensive, and in at least twelve instances complete, covers of ice (Ellsworth and Smith, 1932, pp. 66–67). Most islands have only a few capes or a narrow coastal strip free of ice, but some of the smallest islands are shown on recent maps as ice free. Ice covers 87 per cent of Hooker and 93 per cent of Leigh Smith, next to the east (Hess, 1933, p. 109). The largest single ice cap, 1,150* square miles, is on George Land, the next, of 775* square miles, occupies Wilczek Land, and the third, of 663* square miles, lies on Graham Bell Island. For the most part the islands of Franz Josef Land have elevations of a few hundred to 2,000 feet. Wilczek Land has the highest point (2,411 feet) and seemingly the most rugged topography.

The glaciers are principally ice caps with a total area of 6,560 square miles (Hess, 1933, p. 121) covering 85 to 93 per cent of the land, depending upon the figure accepted for the land area (Flint, 1947, p. 58; Mecking, 1928, p. 151). The ice is thin, probably not exceeding 500 to 600 feet (Flint, 1947, p. 59), and slow moving (Hobbs, 1911, p. 106). The velocity of Jury Glacier is 5 to 7 inches a day (Hess, 1933, p. 109). The glaciers of Franz Josef Land are in equilibrium (Thorarinsson, 1940, p. 145) or receding (Matthes, 1942, p. 194).

Severnaya Zemlya (North Land)

Komsomolets, Pioner, October Revolution, and Bolshevik, the four principal islands composing this archipelago, all have extensive ice caps, but no ice is shown or reported on the smaller islands. The ice forms caps, 600 to 800 feet thick (Flint, 1947, p. 60), on low, broad, somewhat dissected plateaux with elevations up to 2,300 feet. Outlet glaciers are sparse, the west coast of Bolshevik Island displaying the best examples. Caps on October Revolution and to a smaller extent on Komsomolets Island reach the sea directly along broad fronts. At some places the land is so low that glacier ice merges without sharp distinction into pack ice on the sea (Ellsworth and Smith, 1932, p. 70).

About 6,400* square miles of ice covers 42 per cent of the land (Flint, 1947, p. 59), with the following distribution: on Komsomolets Island a large cap of 2,310* square miles and a small cap of 158* square miles; on Pioner Island a single small cap of 150* square miles; on October Revolution Island four rather large caps of 740* square miles (southwest), 628* square miles (east), 625* square miles (south), and 650* square miles (north); on Bolshevik Island a western ice cap of 785* square miles and an eastern cap of 348* square miles. October Revolution Island has the greatest area of ice, but the central cap of Komsomolets Island is the largest single body. The ice coverage on the three largest islands is 21 per cent on Bolshevik, 45 per cent on October Revolution, and 65 per cent on Komsomolets. The coverage increases from southeast to northwest, owing more to open ocean on the west than to topography or latitude (Flint and Dorsey, 1945, p. 92). The glaciers of Severnaya

Zemlya were receding or stagnant in 1930-32 (Thorarinsson, 1940, p. 145; Drygalski and Machatschek, 1942, p. 179).

Other Siberian Islands

Glaciers have not been reported in the New Siberian Islands (Mecking, 1928, pp. 177–181; Flint and Dorsey, 1945, p. 92). Fossil stone ice mentioned by Drygalski and Machatschek (1942, p. 179) is probably ground ice. Lack of glaciers is due to isolation from a suitable source of moisture and to low elevation (Flint and Dorsey, 1945, p. 93).

In 1823, the small islands of Seminovski and Vasilevski in the Laptev Sea at 74° 17'N., 133° 30'E., were reported to be covered by ice, but by 1936 the ice on Seminovski had shrunk to one-eighth its former size, and it had disappeared from Vasilevski (Thorarinsson, 1940, p. 146).

Bennett Island, east-northeast of the New Siberian group at 76°40'N., 149°E., has an ice cap on a 1,000-foot basalt plateau from which several small glaciers descend to the coast (Drygalski and Machatschek, 1942, p. 179; Flint, 1947, p. 60).

Farther east, islands of the De Long group are said to have caps of ice with small tongues projecting almost to the shore (Mecking, 1928, p. 181; Drygalski and Machatschek, 1942, p. 179). Wrangell Island is said to have glacier ice (Drygalski and Machatschek, 1942, p. 179), but this is contradicted by maps and by other statements (Flint and Dorsey, 1945, p. 92).

Nothing is known concerning glacier regimes on these various islands, but the ice shrinkage on Seminovski and Vasilevski islands suggests that all the ice has diminished considerably in the past few decades.

Scandinavia

Scandinavian glaciers, even though the largest in Europe, cover only 2,416 square miles (Flint, 1947, p. 60) of which 310 square miles are in Sweden and the rest in Norway. The principal bodies are flat to dome-shaped caps resting on broad plateaux, and the largest cap, Jostedalsbrae, occupies a peneplain remnant (Strøm, 1949, p. 20). These are the type specimens for Nordenskjöld's (1928, p. 25) plateau glaciers. Outlet glaciers and tongues project to lower levels, but none reaches tidewater. Independent valley and cirque glaciers are also present.

The ice in Scandinavia is concentrated in the high coastal areas. Most heavily covered are the southern, highest part of Scandinavia and the northern uplands. The southernmost major glacier is the 108-square mile Folgefonna Icecap (Drygalski and Machatschek, 1942, p. 154) on an upland attaining 5,423 feet south of Hardanger Fiord (60°N., 6°20'E.). Two large outlet glaciers flow from this cap. About 60 miles northeast, inland from the head of Hardanger Fiord, are smaller caps including Hardanger Icecap of 52 square miles at 6,109 feet maximum elevation (60° 35'N., 7°25'E.). Small ice caps dot the upland north to Sogne Fiord, and farther north, at 61°40'N., 7°E., is the great Jostedals glacier field including Jostedalsbrae. This cap covers 365 square miles and has twenty-five outlet glaciers and tongues extending to lower elevations from a maximum height of 5,653 feet. Tungbergdalsbrae, the largest outlet, is 8.7 miles long, up to 1.25 miles wide, and descends to 1,335 feet elevation. Numerous satellitic ice caps and independent glaciers, some with elevations of 7,000 feet, cluster around Jostedalsbrae and increase the total ice-covered area to 785 square miles. To the west on the coast is the Aalfot Cap of 48 square miles, and 40 miles to the east cirque and valley glaciers are clustered around the highest peak of Scandinavia, 8,097 feet, in the famed Jotunheim district.

Between 62° and 65°N. the land is ice-free, but at 65°20′N., 13°45′E. are small ice caps and cirque and valley glaciers at about 5,500 feet in the Børge Mountains. A large glacier field at Okstinder (66°N., 14°10′E.) covers 29 square miles with three separate caps, several outlets, and many independent valley, cirque, and hanging glaciers (Drygalski and Machatschek, 1942, p. 156). Farther north is the Svartisen glacier field (66°40′N., 14°E.), at maximum elevation 5,246 feet, the most extensive ice-bearing area of northern Scandinavia. Several large outlet glaciers flow from the two principal ice caps of this field, and one, Engabrae, extends almost to sea-level. Many small satellitic caps and independent glaciers augment the total ice coverage, reported as 175 or 385 square miles (Drygalski and Machatschek, 1942, p. 156; Hess, 1933, p. 98). Planimetric measurements suggest that the smaller figure is more nearly correct.

Along the Norwegian-Swedish border, between 67° and 67°30′N., are several ice caps such as Sulitelma, Blamandsisen, and other smaller caps and independent glaciers at 5,000 to 6,000 feet elevation. Some of these extend into Sweden or lie largely in that country. About 40 miles to the east in Sweden are the Partef and Sarekt mountains, both bearing small alpine glaciers at elevations between 6,700 and 7,000 feet. Glaciers of the Sarekt Mountains, numbering more than 200, have been the subject of some glaciological study (Hess, 1933, p. 98). Glimajalos Glacier of about 9 square miles is the largest; Mikka Glacier is about 3 miles long. Farther north on Kebnekaise Mountain (7,005 feet), at 67°54′N., 18°30′E., is a cluster of small alpine glaciers including Stor Glacier, 2.2 miles long by 2,600 feet wide (Herrmann, 1931). The northernmost glaciers of Sweden appear to be small alpine glaciers on Mount Marmat (6,565 feet) and on the high ridge to the north, south of Torne-Träsk, at 68°8′ to 68°12′N. Among these is the well known Kårsa Glacier west of Abisko (Ahlmann, 1948, p. 4).

Westward in Norway are small ice caps and small alpine glaciers on Storsllins Mountain (5,722 feet) and neighbouring peaks at 5,000 to 6,000 feet. At the head of Ofot Fiord is the Frostisen Massif with 12 square miles of ice, including steep ice tongues and a reconstructed glacier. Islands of the Lofoten group have many small cirque glaciers (Drygalski and Machatschek, 1942, p. 156). Northward, the Norwegian mainland is sprinkled with small ice caps and alpine glaciers. Notable are the ice cap on Jaegge Mountain (6,286 feet), at 69°28′N., 19°50′E., and the Oksfjord Glacier on a plateau at 3,871 feet near the coast, 70°10′N., 22°E. An area of 73 square miles is cited for this glacier (Hess, 1933, p. 98), which would make it one of the largest ice caps in northern Scandinavia, but measurements on late maps indicate an area of 15 square miles, so the earlier figure may be in error. The northernmost Scandinavian glaciers are two ice caps, at 5,218 feet and 3,527 feet, covering a combined area of 15 to 20 square miles on the island of Seiland at 70°25′N., 23°5-15′E.

Scandinavian glaciers are not overly active. Average maximum movements of 3.15 inches a day on Styggedal Glacier and of 2.84 inches a day on Kårsa Glacier have been recorded (Ahlmann, 1948, p. 57). These are small glaciers, and some of the larger outlets from Jostedalsbrae undoubtedly have higher velocities. Studies in Scandinavia have established a pattern for investigation and evaluation of glacier behaviour in North Atlantic regions (Ahlmann, 1948, pp. 67–68). The so-called "climatic improvement" and its effects on glaciers was first analysed in detail here. Investigations in Scandinavia also gave birth, in 1896, to the idea that many small alpine glaciers completely disappeared during the postglacial warm-dry period and were subsequently regenerated (Matthes, 1942, p. 207).

The principal oscillations of Norwegian glaciers have been summarized by Thorarinsson (1940, p. 139). Following a minimal stand about 1700 (Faegri, 1934a), a strong advance occurred in the first half of the eighteenth century, reaching its maximum about 1750. This *Hochstand* has generally not been reached by the subsequent advances that took place in the 1810s, in the period from the late 1830s to 1850, and about 1890. The last advance is well documented by end moraines near the snout of Styggedal Glacier where the inner, unvegetated series was formed at the end of the eighteenth or beginning of the nineteenth century (Ahlmann, 1946, p. 11). Swedish glaciers seem to have had essentially the same history (Ahlmann, 1946, p. 69).

Some Scandinavian glaciers advanced slightly in 1897 (Reid, 1898, p. 473) and outlet glaciers from Jostedalsbrae interrupted the cycle of recession and shrinkage that set in after 1890 by advances in 1905–06 and 1924 (Ahlmann, 1946, p. 12). The periodic variation of these particular glaciers matches the meteorological records (Faegri, 1934a, 1934b). The late history of other Norwegian glaciers includes a slight advance in 1901–02, with subsequent slow but accelerating recession to about 1912, producing stagnation in some instances about 1906–07. Oscillations occurred between 1912 and 1932 but with recession maintaining the upper hand. Since 1932, the recession and shrinkage of Norwegian glaciers have accelerated (Ahlmann, 1940b, p. 122). Exposure of arrows buried in firn fields suggests that the present melting exceeds any since Roman times (Thorarinsson, 1940, p. 139). The modern melting is expressed by recession of glacier snouts from 100 to 2,000 feet a year (Faegri, 1934b). Styggedal Glacier on the Horung Massif in Jotunheim, lost 189.2 million cubic feet of ice between 1919 and 1935 (Ahlmann, 1940b, p. 111). This accelerated wastage is attributed largely to a rise in temperature and a lengthening of the ablation season. As usual, in spite of the prevailing recession, occasional advances are recorded, and of 20 Norwegian glaciers observed in 1953, 17 were receding and 3 were advancing (Liestol, 1954, p. 440).

Urals

The Ural Mountains were once assumed to be ice-free, but in 1929 small cirque glaciers were discovered on the eastern and northeastern slopes of Sablia Mountains (Aleschkow, 1930, p. 57). About a dozen additional cirque glaciers were subsequently found (Aleschkow, 1933), and later reports give the total as 15 (Drygalski and Machatschek, 1942, p. 179). These are located principally on slopes of the Sablia Mountains between 64° and 65°N. and on Norodnaia Peak (6,184 feet), the highest point in the Urals. Wind drifting and a northeastern exposure are important factors in the formation and preservation of these glaciers. Hofmann Glacier on Sablia Peak (5,407 feet) with a length of 0.6 mile and a terminal elevation of 2,300 feet is one of the largest. It has a typical banded structure and contains ice at least 220 years old (Aleschkow, 1930, pp. 58–61). In 1929, an abandoned moraine 25 to 35 feet high lay 50 to 60 feet beyond the ice front.

Siberia

The glaciers of continental Siberia have been described as small in number and size despite the cold climate and many highlands, but recent Russian explorations indicate they are more numerous than formerly supposed. The known glaciers, all alpine, are restricted to the following highlands (Flint and Dorsey, 1945, p. 92): Saian Mountains (54°N., 95°E.) at 10,000 feet; Verkhoiansk Mountains (63°N., 139°E.) at 7,000 to 8,000 feet; several localities in the Cherski Mountains (65°N., 142°E.) at 8,500 feet and possibly lower; Koriak Mountains (62°N., 171°E.) at 2,700 to 4,000 feet; Anadyr Mountains (68°N., 177°E.) at 3,000 feet in the eastern and 4,000 feet in the south-central part; and the high volcanic peaks of Kamchatka Peninsula (56°N., 160°E.).

Explorations during 1940 along headwaters of Indigirka River in the Cherski and Verkhoiansk ranges (Popov, 1948) located 46 valley and hanging glaciers on Buordakh Mountain Ridge (65°N., 146°E.), a part of the Cherski Range rising to 9,515 feet. The total area covered by these glaciers, omitting firn fields, is about 30 square miles (Popov, 1947, p. 41). Farther west in the Cherski Range, Mount Chon (10,215 feet), the highest peak in northeastern Siberia, harbours a group of small glaciers (Popov, 1947, p. 41).

Explorations in the Suantar Khaiata Mountain Ridge (62°N., 142°E.), a southern spur of the Verkhoiansk Range rising to 9,000 feet, have located 114 small cirque and valley glaciers (Berman, 1947a, p. 33; 1947b, p. 45). The largest, at the head of Kongor and Setania rivers, is 6 miles long, 2.5 miles wide, and estimated to be 500 to 650 feet thick. The total area of ice here has been given as 36 square miles, although one publication reports more than 250 glaciers and an ice-covered area in excess of 200 square miles (Grigoriev, 1948, p. 27). The highest peak in Siberia, Kliuchevskii (15,912 feet) on Kamchatka Peninsula, is an intermittently active volcano with glacier-clad slopes (Drygalski and Machatschek, 1942, p. 170).

Further explorations will probably add substantially to the number of known Siberian glaciers and to the ice-covered area, which is now conservatively estimated at 200 square miles. All Siberian glaciers are said to be shrinking (Popov, 1947, p. 41). The small size and number of Siberian glaciers are attributed largely to the great distance from sources of moisture (Flint and Dorsey, 1945, p. 93).

Canada

The areas treated are the mountains of northeastern Labrador and the ranges of the Canadian Cordillera north of latitude 60°. Glaciers in the Canadian Rockies and the Central Plateau and Mountain area (Bostock, 1948, pp. 43–45) south of 60°N. are not considered, and bodies of ice in the Coast Mountains and the St. Elias Range are described in conjunction with Alaskan glaciers. Like Siberia, much of continental Canada, even though cold and in part relatively high, has only small glaciers because of insufficient moisture.

The Torngat and Kaumajet mountains of northeastern Labrador contain the only glaciers of eastern continental North America. They are all small and in deep cirques on northern, eastern, or northeastern slopes. The best known is Bryant Glacier (Forbes, 1932, pp. 34–35) on the northern slope of Mount Tetragona (4,511 feet, 59°18'N., 63°54'W.). Others are reported on Mount Razorback, the Four Peaks (4,140 to 4,416 feet), and other Torngat peaks to the south and east (Odell, 1938, p. 212). The glacier on Razorback is about one-half mile long and descends to 1,200 feet (Forbes, 1932, p. 47). A small ice mass on the northeast side of Brave Mountain (4,200 to 4,400 feet) in the Kaumajet Mountains (57°45'N.) has been cited as the southernmost glacier in eastern continental North America, but if current maps are correct, the Kaumajet Mountains lie on Cod Island, and this glacier is not continental. Bryant Glacier receded 250 to 300 feet between 1908 and 1931. In 1931 it was judged to be in a state of recession and only one step removed from stagnation (Odell, 1933, pp. 205–206). The accelerated wastage of recent times has probably caused further deterioration of these glaciers and perhaps the destruction of some.

The principal glacier-bearing areas in the interior of the Cordillera of Canada, north of 60 degrees, lie near the divide between Yukon and Mackenzie rivers in the Selwyn and Mackenzie mountains. In many parts of the region groups of peaks rise to more than 7,500 feet elevation and where they stand near the divide or southwest of it small icefields and glaciers occur around their summits. In addition small patches of glacier ice, 50 to 200 yards across, are commonly found on the northerly sides of steep cirque walls at elevations above 6,700 feet in the southeast part of the Ogilvie Mountains[1] and in the higher ranges of the Yukon Plateau. Four high centres containing, or thought to contain, summits more than 9,000 feet in elevation, form areas of relatively high precipitation with relatively large bodies of glacier ice for their latitudes in this region. In each of these, one or more icefields and glaciers, seen in air

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[1] The glacier shown on maps at latitude 65 degrees 17 minutes north and longitude 140 degrees west in the Ogilvie Mountains was a large body of ice in a stream valley recorded by W. Ogilvie early in the spring of 1888. Air photographs and recent topographical mapping show that no glacier exists here but that masses of overflow stream ice commonly remain far into the summers in these mountains.

photographs or on recent topographical maps, are judged to measure 5 square miles or more in area. Comparison of these centres shows an increase in the areas of ice from north to south. In the north between Wind and Snake rivers where four peaks are reported to be about 9,500 feet high little more than 12 square miles of ice is apparent in the air photographs. To the southeast at the heads of Snake, Arctic Red, and Stewart rivers, some 25 square miles of ice can be seen around a high unmapped centre of drainage. Farther south, about the headwaters of Hess River, where Keele Peak, now determined to be more than 9,700 feet in elevation, is outstanding, 45 or more square miles of ice shows on the maps and in the air photographs. Southeast of the Canol Road in the southern Selwyn Mountains where Mount Sir James MacBrien, 9,049 feet, is the highest point and in the ranges of the Mackenzies nearest them, east of South Nahanni River, icefields and glaciers are more numerous and the main ones comprise at least 140 square miles of ice. In addition, some 50 square miles of ice in the form of many small glaciers scattered over the region, including 8 square miles or more of ice in the Itsi Range, make up a total of 270 or more square miles of ice distinguishable in air photographs or shown on maps in the Cordillera region of western Canada north of 60 degrees.

Alaska and Adjoining Parts of Canada

The glaciers of southern Alaska and contiguous parts of Canada bordering the Gulf of Alaska have been described as among the largest outside of polar regions (Gilbert, 1910, p. 9; Tarr and Martin, 1914, p. 1). Although this may be an overstatement, this region does provide an exceptionally fine display of valley, transection, intermontane, and piedmont glaciers. The ice cover is heaviest in those areas that combine elevation and copious supplies of moisture. This includes the Coast Mountains as far south as 56°30'N., the St. Elias Mountains, the Chugach Range, and the Kenai Mountains. The snowline rises rapidly inland so that interior ranges such as the Wrangell, Talkeetna, Aleutian, and Alaska, although in many instances higher in altitude and latitude than coastal ranges, have less ice.

The area of greatest glacier development centres around the Fairweather Range and the high peaks of the St. Elias Mountains (Fig. 4), the Icefield Ranges (Bostock, 1948, pp. 98–100), which rise abruptly 10,000 to 18,000 feet above the Pacific and culminate in Mount Logan at 19,850 feet. Great storms sweeping in from the Gulf of Alaska leave their heaviest snows here. This area contains long valley glaciers such as the Hubbard (Fig. 4), 80 to 90 miles, great intermontane glaciers such as the Seward and Brady (Matthes, 1942, p. 152), many transection glaciers, tidal glaciers and type examples of piedmont sheets in the Bering and Malaspina glaciers (Fig. 6) (Russell, 1891, p. 176). The ice cover tapers off southward in the Coast Mountains where cirque and valley glaciers become the rule, although even here some ice streams are fed from highland ices, like that north of Juneau, and many extend to tidewater as far south as 57°N. Westward the Chugach Mountains and high areas around Prince William Sound have large glaciers, and the ice cover in parts of the Kenai Mountains forms highland glaciers.

Fig. 4. View northward across upper Seward Glacier to head of Hubbard Glacier in heart of St. Elias Mountains, Canada. Photo by Walter A. Wood.

In the Aleutian Range the environment is more continental, and alpine glaciers and small ice caps prevail, with their best development in the higher areas between Iliamna Volcano and Mount Gerdine. Talkeetna Mountains (62°N., 149°W.) have relatively small cirque and valley glaciers. Wrangell Mountains are mantled by the most compact glacier systems in Alaska. This is partly due to elevations locally exceeding 16,000 feet, to a topography that favours an extensive cover of upland ice, and to somewhat lower coastal mountains to the south and southwest. The Alaska Range lacks glaciers in keeping with its pre-eminent elevation on the North American continent. Cirque and valley glaciers of moderate length are clustered chiefly around Mount McKinley (20,300 feet) in the western, Mount Hayes in the central, and Mount Kimball in the eastern parts of this range.

Glaciers dot high peaks scattered along the Alaska Peninsula, among which Veniaminof Volcano (8,400 feet) with a small cap and outlet tongues is notable. High peaks of the Aleutian chain are also ice-bearing, especially on the three easternmost islands, Unimak, Unalaska, and Umnak. Mount Shishaldin on Unimak (9,372 feet), the highest peak in the Aleutians, has a permanent mantle of ice and snow but has no outflowing ice streams. Isanotski Peak and Round Top just to the east have a number of small glaciers. Pogromni, a smoking volcanic cone, on the west end of Unimak harbours at least two valley glaciers. Unalaska Island has the heaviest ice cover in the

Aleutian chain owing to the extremely rugged terrain favourable to the protection and growth of cirque and valley glaciers. Small glaciers lie on the north sides of high peaks at the west end of Umnak Island, including Recheschnoi Volcano (6,920 feet) and Mount Vsevidof. In 1945 volcanic eruptions and a lava flow caused melting of a glacier on Okmok Volcano at the east end of Umnak (Robinson, 1948, p. 514).

Westward in the Aleutians, glaciers are fewer owing to lower elevations. Korovin Volcano on Atka, at about 174°10′W., is said to have glaciers (Tarr and Martin, 1914, p. 18), and five glaciers are reported (Simons and Mathewson, 1947, pp. 59, 65) on Great Sitkin Island (52°5′N., 176°8′W.). Tanaga Volcano (6,975 feet) at 51°55′N., 178°W. is certainly high enough to bear glaciers, but as yet none has been reported. Gareloi (5,334 feet) with two ice streams is the westernmost (178°54′W.) Aleutian island so far reported to be glacier-bearing (Coats, 1947, p. 98).

In the interior of Alaska, small glaciers lie among peaks of 5,000 feet elevation in the central Tikchik Mountains (Mertie, 1938, p. 14) north of Bristol Bay (60°N., 159°30′W.). Glaciers are also reported on Mount Oratia (5,400 feet) in the Kilbuck Mountains a little farther north, and on Seward Peninsula (Tarr and Martin, 1914, p. 19).

The Brooks Range of northern Alaska rises to elevations above 9,000 feet but contains only a few small glaciers owing to light precipitation. In this range small cirque glaciers are reported in the Baird and Endicott mountains on the upper Noatak and Koyukuk rivers (Smith, 1913, p. 32), but the best area of glacier development is farther east on the north slope of the Romanzof Mountains, where one valley glacier, the Okpilak, attains a length of 10 miles, dozens are up to 3 miles long, and there are scores of cliff and cirque glaciers (Leffingwell, 1919, p. 156). These ice bodies cluster around Mounts Chamberlin (9,131 feet) and Michelson (9,239 feet), and the high country at the head of Canning, Sadlerochit, Hulahula, Okpilak, Jago, and Aichillik rivers.

Alaska and adjacent parts of Canada contain thousands of individual glaciers (Fig. 5), and those of Alaska cover an estimated 20,000 square miles, or slightly more than 3 per cent of the territory (Capps, 1931, p. 1). The contiguous parts of Canada contain at least another 5,000 square miles of ice. The major part of the 30,890 square miles of ice on the North American continent lies in this area.

The coastal glaciers of southern and southeastern Alaska, being of relatively easy access, have been subjects of repeated observation, particularly in Glacier, Lituya, and Yakutat bays, the Juneau area, Prince William Sound, and lower Copper River. Fortunately, early explorations by Bering, Cook, Vancouver, La Pérouse, and others left records permitting relatively accurate recounting of coastal glacier behaviour for the past 150 years. Extensive shrinkage and recession during the postglacial warm-dry period were early recognized (Gilbert, 1910, p. 103) and later more clearly established by geobotanical studies (Cooper, 1942, pp. 18, 21; Lawrence, 1950). Glaciers in Prince William Sound, Glacier Bay, and other parts of southern Alaska

Fig. 5. Kaskawulsh Glacier, St. Elias Mountains, Canada.

(Gilbert, 1910, p. 103; Wentworth and Ray, 1936, pp. 888, 891; Field, 1937, p. 78; Cooper, 1937, pp. 38–39) are uncovering, during their current recession, stumps and remains of forests that grew far up the valleys when ice masses were much reduced during the warm-dry period (Cooper, 1931, pp. 88–93). Readvances culminating within the last 200 years have brought these glaciers to their most advanced positions in at least 6 centuries (Lawrence, 1951) and probably in many thousands of years (Matthes, 1942, pp. 210–211; Lawrence, 1953, pp. 84–85).

The climax of this last advance has occurred in various places at different times. It probably culminated in Glacier Bay 150 to 200 years ago (Cooper, 1937, p. 47), as recession of only 3 to 6 miles had occurred at the time of Vancouver's visit in 1794. In parts of Prince William Sound and on the western slopes of the Fairweather Range, the glaciers are now at their most advanced positions in centuries or have only recently receded from such maxima (Field, 1932, p. 371; 1937, pp. 69, 72; Cooper, 1942, pp. 4, 17). The advance of other glaciers has culminated at intermediate times. La Pérouse Glacier was at its maximum in 1899 (Gilbert, 1910, p. 103), having advanced since La Pérouse's visit in 1788 (Klotz, 1899, p. 526; Mertie, 1931, p. 122). After some recession this glacier readvanced in the interval between 1948 and 1952 (Baird and Field, 1954). The glaciers of Port Wells attained their maxima about 50 to 100 years ago. The Eagle, Herbert, Mendenhall, Norris, Taku, Hole-in-the-Wall, and Twin glaciers draining from the southern part of the Juneau Ice Field all attained advanced positions in the early or middle eighteenth century (Lawrence, 1950, p. 222). Thus, the coastal glaciers of Alaska appear to have attained their postglacial *Hochstands* at various times within the last 200 years, and at least some interior Alaskan glaciers have probably experienced a similar history (Péwé and others, 1953). In view of this irregular behaviour one may wonder how successful attempts will be to distinguish in Alaska the *Hochstands* of 1850 and 1890 recognized in many other parts of the world.

Alaskan glaciers are notoriously out of phase with each other in terms of current behaviour. Many have been in rapid retreat since attaining their maximum postglacial positions. A noteworthy example is the 60-mile recession of Muir Glacier in Glacier Bay since 1794 (Matthes, 1942, pp. 198–199). Field (1947, p. 369) calculates that the ice-covered area draining to Muir Inlet has been reduced about 175 square miles, or 35 per cent, during the past 65 years. Shrinkage and recession of similar proportions may have occurred in other parts of Alaska, and great shrinkage of interior Alaskan-Canadian glaciers is reported by H. B. Washburn (1941, pp. 220–222).

An early study of Alaskan coastal glaciers (Gilbert, 1910, p. 104) demonstrated that closely associated glaciers may display markedly dissimilar behaviours. Between 1794 and 1894 one glacier in Glacier Bay receded 45 miles while another only 20 miles away advanced 5 miles. In Lituya Bay, two glaciers advanced 3 miles in 108 years while a third glacier located between them receded (Mertie, 1931, p. 123). Discordant behaviours of this type

Fig. 6. Folded morainal septa in Malaspina Glacier, Alaska, looking west-southwest.

continue today at many places along the coast (Field, 1937; Cooper, 1942, p. 4), although recently the number of glaciers advancing or essentially holding their own has shown a marked increase (Field, 1948; Baird and Field, 1951, 1954). One of the most consistent long-term advances on record is that of Taku Glacier (58°26′N., 134°3′W.), the snout of which moved forward 18,200 feet since 1900 and continues to advance (Field, 1954, p. 237) even though neighbouring glaciers, some draining from the same snow fields, are receding.

An especially erratic behaviour was that of Black Rapids Glacier, draining from the southeast slope of Mount Hayes in the Alaska Range (63°32′N., 145°55′W.). This glacier had receded consistently for 2 or 3 decades when suddenly, in late September or early October 1936, it started a rapid advance (Hance, 1937, p. 778). By early March 1937, the front had moved forward nearly 4 miles (Moffit, 1942, p. 152) at an average rate of 115 feet a day. The daily rate between December 3, 1936, and March 7, 1937, approached 200 feet and possibly attained 250 feet on occasions. By September 1937, movement had ceased, and the glacier has since shrunk and receded (Péwé and Taylor, 1953).

Erratic glacier fluctuations are attributed to climatic variations, local meteorological conditions, relations between snowline and maximum snowfall (Field, 1937, p. 81); orographic factors (Cooper, 1937, p. 61; Gilbert, 1910, p. 109), reservoir lag (Tarr and Von Engeln, 1915, p. 137), threshold resistance (Moffit, 1942, pp. 155–156; Ahlmann, 1953, pp. 10–11), and various combinations of any or all of the above (Reid, 1895, pp. 278–282). Earthquakes may also cause unexpected glacier advances as demonstrated by the 1899 shock at Yakutat Bay, one of the most severe of modern times. Prior to 1899 and up to 1906, the glaciers of Yakutat Bay and vicinity were stagnant or receding. Starting in 1906 and continuing to 1913, many of these glaciers underwent rapid, short-lived rejuvenation and advance. Fortunately, this occurred during a comprehensive study, so the details are well known (Tarr and Martin, 1914, pp. 168–197). Several glaciers advanced as much as 4,000 feet in a few months, and one moved forward 10 000 feet in less than a year. The advances occurred at different times in different glaciers, being progressively later in direct proportion to the glacier's length. The rejuvenation has been attributed to great quantities of snow avalanched onto the glaciers by the earthquake. This supposedly sent a wave of accelerated movement through the glacier, eventually producing an advance of the snout. Although erratic advances of Alaskan glaciers at other times and places may have been caused by other earthquakes (Tarr and Martin, 1914, p. 193), it seems likely that the irregular behaviours of most glaciers is caused by climatic variations, influenced and complicated by the host of factors listed above.

Ellesmere Island

Ellesmere is the most northerly, the third largest, and the highest island of the Canadian Arctic Archipelago. Peaks in the northern part may reach 10,000 feet (Humphreys, Shackleton, and Moore, 1936, p. 425), but radar altimeter readings in the area suggest that the maximum height is rather less.

The east and north parts of the island are largely rugged alpine terrain, but much of the west side is lower, with gently sloping hills, and in the south is a dissected plateau. The glaciers, covering 32,000 square miles or 40 per cent of the land, are the most extensive of the Canadian archipelago.

Ellesmere Island is divided into four sections by long fiords indenting the east and west coasts. All four sections have independent caps of ice, but some are so thin and the influence of the underlying topography so strong that they are better classified as highland glaciers (Bentham, 1941, pp. 43–4). Much of the low western country is ice-free.

In the most northerly section, north of Greely Fiord and Lady Franklin Bay, there is about 9,900 square miles of ice cap and highland ice, from which long outlet glaciers extend to the sea in the north and southwest. Those on the north may contribute to an ice shelf that extends into the sea along the north coast (Koenig *et al.*, 1952, pp. 96–102). The origin and constitution of this shelf is not yet a matter of agreement. Hattersley-Smith (1955, pp. 22-26) feels that it was formed in situ largely by the freezing of sea water, but Marshall's (1955, pp. 112–113) studies of the stratigraphy and structure of the shelf suggest that much of it is composed of firn and ice accumulated directly upon a basement of glacier ice and sea ice. A number of small periferal ice caps lie to the west and northwest of the principal area, and the total ice-cover in northern Ellesmere Island is approximately 11,770 square miles.

The ice cap between Greely Fiord and Bay and Hayes fiords is the second largest, 8,500 square miles, and it probably covers a larger percentage of the land than on any other section of the island. Much of this cap lies above 5,000 feet elevation, with the highest point about 6,600 feet. Long outlet glaciers extend northwest into two arms of Greely Fiord and into Canyon Fiord, and many more reach tidewater in Kane Basin to the east and southeast. A few smaller periferal caps bring the total area to 8,760 square miles. The whole eastern side of the ice cap lies on alpine topography and is essentially a highland glacier with numerous outlets.

Between Hayes Fiord and Makinson Inlet is a single, relatively simple ice cap with outlets reaching tidewater on the north, east, and south. This ice mass attains a maximum elevation of about 6,800 feet and covers a total of 7,750 square miles. Farther west between Bay and Baumann fiords are a few small caps, covering in all about 60 square miles.

The southernmost section of the island, south of Baumann Fiord and Makinson Inlet, has two major ice caps or highland glaciers, and several smaller caps. The largest ice cap, 2,200 square miles, is at the eastern end and has numerous outlet glaciers (Bentham, 1941, pp. 43–44). Small ice caps, about 76 square miles, occupy high land on both sides of Starnes Fiord farther west. The second largest cap, 1,040 square miles, centres at about 76°55′N., 85°W., north of Heim Peninsula, and sends several outlet glaciers south to tidewater. Periferal caps amount to about 160 square miles. Farther west are much smaller caps on the upland between Muskox and Goose fiords, 140 square miles, and on Simmons Peninsula west of Goose Fiord, 50 square miles. The total area of ice in this section is about 3,670 square miles.

In addition to these ice caps, highland glaciers, and outlets, the alpine areas along the east coast contain independent valley and cirque glaciers, particularly between Princess Marie Bay and Judge Daly Promontory. Some tidal outlet glaciers terminate in spectacular cliffs (Weeks, 1927, pp. 136c–137c). Outlet glaciers and valley glaciers that fail to reach the sea form expanded-foot glaciers, and some in the Smith Bay area unite as piedmont sheets (Mecking, 1928, p. 231; Bentham, 1941, p. 44).

Little glaciological work has been reported from Ellesmere Island, and not much is known of past behaviour or present regime of the glaciers except that the signs of past recession are abundant (Koenig *et al.,* 1952, p. 96). Glaciers in the south were said to be stationary or in slow recession in the 1930s (Bentham, 1941, p. 44), but Sven Hedin Glacier just north of Princess Marie Bay appeared to be advancing in 1935 (Humphreys, Shackleton, and Moore, 1936, p. 412). The larger of the two glaciers at Craig Harbour retreated 6 feet between 1936 and 1938, and the snout is less than a quarter of a mile from an old moraine. End moraines lying short distances beyond the snouts of expanded-foot glaciers in other parts of southeast Ellesmere Island were probably formed in one of the well-known *Hochstands* between 1750 and 1890, and some glaciers north of Princess Marie Bay attained a maximum shortly before 1863 (Reid, 1899, p. 220). In 1955 many glaciers in the south half of the island were advancing (Fortier, personal communication). In the air photographs of the island, taken by the R.C.A.F. in 1950, many bergs appear in inlets on the east coast from Rawlings Bay to Princess Marie Bay, and Baird Inlet to the southeast corner.

Hattersley-Smith (1955, p. 26) is of the opinion that in north Ellesmere Island the glaciers are probably in a state of near-balance at the present time, and he found no evidence of any extensive advance or retreat within recent decades. An ice-cliff at Cape Aldrich is in much the same position today as is shown in a sketch made by Aldrich in 1876. A 20-foot advance was measured in this glacier between 1953 and 1954, and vegetation growing right up to the snouts of glaciers in the United States Range suggests that they are stationary or advancing.

Baffin and Bylot Islands

Baffin is the largest island in the Canadian Arctic Archipelago and has mountains almost as high as those of Ellesmere Island. Bylot Island is also treated here because of geographic proximity. The highest part of Baffin Island is Cumberland Peninsula, where the elevations, formerly estimated at 10,000 feet (Weeks, 1928, p. 88c) and shown on late maps as 8,200 to 8,500 feet, are probably closer to 7,000 feet (Ward and Baird, 1954, p. 345). Much of eastern Baffin Island attains elevations of 4,000 to 6,000 feet, but the terrain slopes away to a western plain. Plateaux predominate in the northwest and south.

Bylot Island, a mountainous island rising to about 6,200 feet, is nearly 50 per cent covered by an ice cap of 2,000 square miles. Large outlet glaciers descend on all sides, two on the northwest and one on the south reaching tidewater. Those on the south along Pond Inlet have spectacular ice falls (Freuchen and Mathiassen, 1925, p. 554).

Baffin Island itself has a large number of small ice caps, mostly along the east coast, of which many are better described as highland glaciers. Its total coverage is about 14,100 square miles. This figure does not include the ice cap of Bylot Island.

In northern Baffin Island, five small ice caps cover 118 square miles on Brodeur Peninsula, and Borden Peninsula has six small caps totalling 307 square miles in the north and one of 26 square miles in the south. From Pond Inlet to Home Bay there is an almost continuous succession of ice caps and highland ice, with many valley and cirque glaciers, amounting to about 6,330[1] square miles. A few outlet glaciers reach the sea in the northern part. At Home Bay the heights fall below 3,000 feet and the ice caps become smaller and more scattered, amounting to only about 385[1] square miles between 69°N. and the Penny Ice Cap. The Barnes Ice Cap, one of the two largest on Baffin Island, lies about 50 miles inland from the east coast between capes Adair and Christian. It is 90 miles long and 35 miles wide, lies on a plateau with its highest point at 3,700 feet, and covers approximately 2,340 square miles.

On the north part of Cumberland Peninsula is the other major ice cap, the 87-mile Penny Ice Cap (Baird *et al.*, 1953) attaining a maximum elevation of 6,725 feet and covering 2,300 square miles (Ward and Baird, 1954, p. 354). This cap rests upon a highly dissected plateau and has an irregular outline and an uneven surface featuring a number of separate domes. Many outlet glaciers flow from it in deep valleys and a number reach the sea at the heads of long fiords.

South of the Penny Ice Cap is rugged alpine country, with peaks to 7,000 feet, which has about 2,000[2] square miles of small ice caps, valley glaciers, and a complex of transection glaciers rivalling those of West Spitsbergen and Alaska. Few of the valley glaciers reach tidewater. An ice cap of 109 square miles lies near the east coast of Hall Peninsula, 10 to 15 miles inland from Popham Bay, and other small caps amounting to about 45 square miles lie farther south. The southernmost glaciers in Baffin Island are along the southwest shore of Frobisher Bay (Wynne-Edwards, 1937, p. 2; Buerger, 1938, pp. 4–5; Mercer, 1956) where the Grinnell Ice Cap at a maximum elevation of 2,710 feet covers 51 square miles, and Terra Nivea at 2,770 feet elevation covers 64 square miles. Outlet glaciers from the Grinnell Ice Cap reach tidewater.

Information on the physical aspects of the Barnes and Penny Ice Caps is provided by the work of two expeditions sponsored by the Arctic Institute of North America (Baird *et al.*, 1950; 1953; Baird, Ward, and Orvig, 1952; Ward and Baird, 1954). Both are subpolar ice bodies with strongly negative subsurface temperatures and limited surface melting in summer. The Barnes is essentially devoid of firn or snow by the end of the ablation season and is

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[1] These figures are approximate owing to difficulty of measuring small discontinuous areas.

[2] Approximate, owing to difficulty of measuring small discontinuous areas.

nourished largely by refreezing of meltwater on top of the cold ice. Nourishment primarily by superimposed ice identifies the so-called "Baffin Type" of glacier (Baird, 1952, p. 9), which may be fairly common within the Canadian archipelago. The Penny (Ward and Baird, 1954) and the Grinnell (Blake, 1953) ice caps are transitional, being nourished both by accumulation of firn and by superimposed ice. The thickness of the Barnes Ice Cap averages about 1,000 feet and attains a maximum of 1,350 feet according to gravity surveys (Littlewood, 1952, p. 124). Maximum thickness of the Penny is estimated at 1,000 feet in most places. The outlet glaciers of the Penny Ice Cap, although large and numerous, flow less than 1 foot a day. The Barnes Ice Cap had a slightly negative regime in 1949-50 (Baird, 1952, p. 8) and the Penny shows evidence of continuous recession from an advance culminating 150-200 years ago (Ward and Baird, 1954, p. 346). At least one outlet glacier from the Grinnell Ice Cap was advancing in 1952 (Mercer, 1956, p. 563).

Devon Island

Among the remaining glacier-bearing islands in the Canadian Arctic, Devon, midway between Ellesmere and Baffin, has 6,270 square miles of ice. Eastern Devon Island has an ice cap of 5,840[1] square miles rising to 6,100 feet. The margin of this cap is strongly digitated, and a number of small periferal caps covering about 35 square miles lie to the south and southwest. Large outlet glaciers descend steep slopes from the central plateau to the sea (Low, 1906, p. 236). Southwest of the main cap and its fringing satellites is a small independent cap of 70[1] square miles on the upland west of Maxwell Bay. In northern Devon Island three caps totalling 324 square miles occupy Colin Archer Peninsula. Two outlet glaciers extend to the sea on the north side from the eastern cap.

North Kent Island, between north Devon and the southwest corner of Ellesmere, carries a small ice cap of 74 square miles with an outlet to tidewater on the east. About 15 miles northeast of Devon Island is Coburg Island, which with 90 square miles of ice and a total area of 140 square miles is 65 per cent covered. Expanded-foot glaciers reach the sea on the northeast and southwest coasts especially.

Axel Heiberg Island

Axel Heiberg Island has an extensive ice cover. An ice cap about 145 miles long covers 3,600 square miles in the central part of the island, and a kidney-shaped cap of 1,120 square miles lies to the southwest. Smaller fringing caps on the west side of the central cap bring the total to 5,170 square miles. Large outlet glaciers lie on the west and south, only one of which reaches the sea from the central ice cap. The highest part of the central cap is somewhat over 7,000 feet, and the southern cap attains about 6,500 feet.

Meighen Island

The long-suspected existence (Stefansson, 1921, pp. 518–519) of a small ice cap on Meighen Island, west of Axel Heiberg Island, has been confirmed

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[1] Approximate, owing to difficulty of measuring small discontinuous areas.

by aerial observation and air photography. The ice cap is about 35 square miles in area and without outlet glaciers.

Melville Island

Three small ice caps, or large perennial snowfields, amounting to about 100 square miles in area, are located near the head of Purchase Bay in Melville Island. These are the only glaciers west of Meighen, Axel Heiberg, Devon, and Baffin islands, though there are many permanent snowbanks which may at times have been reported as glaciers. The absence of other glaciers in the western islands has been attributed to low elevation and low precipitation (Mecking, 1928, p. 221), a theory that is supported by the fact that the Melville caps, at about 3,500 feet, are undoubtedly on the highest land in the area.

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