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SCIENTIFIC RESULTS
OF THE POLISH GEOGRAPHICAL EXPEDITION
TO VATNAJÖKULL (ICELAND)

RAJMUND GALON, Editor

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TO VATNAJÖKULL (ICELAND), 1968

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THE POLISH GEOGRAPHICAL EXPEDITION TO VATNAJÖKULL (ICELAND) JUNE 5 TO SEPTEMBER 7, 1968

RAJMUND GALON AND JAN SZUPRYCZYŃSKI

THE PURPOSE OF THE EXPEDITION

Polish geographers and representatives of related sciences have always paid close attention to the investigations of polar and subpolar regions. To a certain extent one may speak of the specialized, almost traditional interest of Polish scientists in glaciological, geological and geomorphological studies in the Arctic. Particular preference was, and still is, given to Spitsbergen which prior to Second World War and, especially, afterwards, was destination of a number of Polish expeditions. Practically every Polish "polar" scientist has had to pass difficult Spitsbergen training, either as member of one of the Polish expeditions [1] or by taking part in scientific expeditions organized by Norway or other countries.

So far, Iceland has rather been neglected, and had not raised great interest among Polish polar students and glaciologists. No essential field studies were made in Iceland by Poles, with a few exceptions. Professor A. Kosiba, a glaciologist, made several trips to Iceland; a report describing his studies was published in 1938.[2] Professor A. Jahn also briefly visited Iceland in 1956 on his return from the United States; his main interest was the periglacial structures in Iceland.

Strictly speaking, Iceland does not belong to what is called the polar world because most of the islands lies south of the Arctic Circle. However, the glacial phenomena occurring in Iceland, although at present undergoing a distinct regression, are of exceptional interest to students of the Pleistocene glaciations which also took place on the Polish Lowland. The Iceland glaciers which originate on elevated areas and descend to the level of coastal lowlands, constitute a fairly true miniature image of the immense Scandinavian inland ice which once covered a large part of Europe and which in Poland extended as far as the Carpathians and the Sudetes. Iceland's largest glacier, Vatnajökull, is especially worth studying. It covers an area of 8400 km². The study of the Iceland glaciers, their decay and deposition of moraine and glaciofluvial deposits may be useful in explaining some stages in the decay of the Scandinavian inland ice in Europe — especially of the last glaciation which is perpetuated in landforms such as end moraines, dead-ice forms, outwash covers, etc.

[1] See an extensive recent report: *Polish Spitsbergen Expedition 1957–1960, Summary of Scientific Results*, Warszawa 1968, 165–332.

[2] A. Kosiba, Some problems in Iceland's morphotectonics and glaciology (in Polish), *Czas. Geogr.*, 16 (1938), 257–306.

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The plan of undertaking an expedition originated in 1961 in the Section of Geomorphology and Hydrography of the Institute of Geography of the Polish Academy of Sciences at Toruń. From the very beginning two scientific objectives were planned for this expedition:

(1) to obtain, as fully as possible, material on the origin of glacial and glaciofluvial landforms and deposits from the present marginal zone of glaciers, in order to correctly interpret this type of landforms and deposits of the Pleistocene glaciations in Poland,

(2) to investigate in detail those fragments of the marginal zone of the Vatnajökull glacier which so far have been least examined, or which represent the greatest variety of classical landforms.

ORGANIZATION OF THE EXPEDITION

For the research work to be carried out, the Polish Expedition first selected the northern Vatnajökull forefield, planning to concentrate on the part called Dingjujökull and Koldukvislarjokull, because no data whatsoever were available in scientific literature for these areas. Up to the present, the whole north-western and south-western part of this glacier has been scarcely investigated.

The notion of undertaking the Expedition was reported to the Institute of Geography of the Polish Academy of Sciences; the trip was originally planned for the Summer of 1962 (July-August). Participants were to be four geomorphologists from the Section of Geomorphology and Hydrography of the Lowlands of the Institute of Geography, Polish Academy of Sciences at Toruń. The Department of Geology of the Polish Academy of Sciences was also greatly interested in this proposal, because the geologists were anxious to examine volcanic landforms and rock types in Iceland. After some contacts between these two groups, a detailed research programme as well as a cost estimate were prepared.

Paralled with the steps taken in Poland, the organizers started contacting scientific centres in Iceland. From Professor Sigurdur Thorarinsson, Director of the Natural Science Museum at Reykjavik, the prominent glaciologist and geologist, comprehensive information was obtained with regard to the area of interest, and as to transport conditions in Iceland. Poland's diplomatic representation at Reykjavik was also contacted in this matter. At that time Mrs. Anna Kowalska was charge d'affaires at this representation. Further, the programme of the Expedition was forwarded to the National Research Council of Iceland whose director, Dr. Steingrimur Hermansson, gave his consent to the visit of a Polish research group. After studying the programme submitted. Professor Thorarinsson suggested that the geomorphological investigations of the Polish Expedition should be made in the forefield of smaller Alpine-type glaciers on the Öraefa massif, Iceland's highest mountain group, or in the south-western part of the Vatnajökull. Aircraft photos and maps were available for these two regions, while no satisfactory surveys existed for the northeastern parts of the glacier. Moreover, Icelandic scientists intended to concentrate their own glaciological and geological research on the latter areas.

Meanwhile, in Poland, the plans of the Iceland Expedition were put before the Polish Academy of Sciences. The Expedition was endorsed by the Committee of Geophysical Expeditions attached to the Executive Board of the Academy. However, difficulties in obtaining the necessary funds prevented the execution of this programme in 1963.

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The Iceland scheme was taken up again in 1966. On February 28, 1966 Professor Galon approached the Board of the Polish Geographical Society with the suggestion of undertaking a research expedition to Iceland under the auspices of the Society and with its financial support. The general programme of this expedition anticipated field work in Iceland in the Summer of 1967 (July and August). The voyage was to be made partly by air and, partly, on a mother-ship of the trawler fleet owned by one of the Polish oceanic fishing enterprises, while proceeding to its Newfoundland deep-sea fishing grounds. The Board of the Polish Geographical Society gave their consent to this proposal, but insisted that the Expedition was an all-Poland undertaking and that the number of research staff were increased. A new detailed programme along these lines was drawn up including a cost estimate, and in May 1966 it was submitted to the Polish Academy of Sciences. The Academy welcomed the suggestion made by the Polish Geographical Society and supported its request for a Government subsidy. A grant was assigned in June, 1967. However, for technical reasons, it was decided to postpone the Expedition until the Summer of 1968. The Polish Naval Authorities offered to take the Expedition to Iceland and back in 1968 on one of the Polish Navy ships.

Fig. 1. Research area of the Polish Geographical Expedition to Vatnajökull 1968
SK — Skeidarárjokull and its proglacial area
S — Sidujokull and its proglacial area

In August 1967 Professor Galon, the leader of the Expedition, made a brief visit to Iceland. The purpose of his trip was to solve organizational problems and to obtain the necessary maps and air photographs from the Geodetic Survey of Iceland. At that time final decisions were also reached in connection with the area to be investigated. In these negotiations Iceland was represented by Professor Thorarinsson, and with his assistance the official consent of authorities of Iceland was also obtained. It was decided to study the marginal and proglacial areas of Skeidarárjökull and Sidujökull, two outlet glaciers on the southern and south-western margin of Vatnajökull (Fig. 1).

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In the autumn of 1967 all members of the research team met for the first time to divide the tasks in preparing the Expedition. The costs of the Expedition were covered by the Governmental grant and by the contribution made by the Polish Geographical Society. The Institute of Geography of Poznań University donated medicine supplies, the Institute of Geography of the Polish Academy of Sciences and the University of Toruń jointly purchased the camp equipment which later was brought back to Poland and will be available for further expeditions.

Much valuable assistance was also given to the Expedition by the Institute of Geophysics of the Polish Academy of Sciences. The Institute offered a complete outfit of individual clothing and equipment. Some time before all this equipment had been acquired by the Polish Committee of International Geophysical Co-operation for its research expedition to Spitsbergen. The Institute of Geophysics also supplied part of the apparatus required for glaciological and meteorological investigations. Finally, the Institute of Geography of the University of Wrocław loaned some glaciological and meteorological instruments.

In view of all this, the members of the Expedition wish to express their sincere gratitude to all persons and institutions, both Polish and Foreign Citizens, for their valuable assistance.

The composition of the research staff was as follows (Fig. 2):

(1) Professor Dr. Rajmund Galon, leader of the Expedition, Institute of

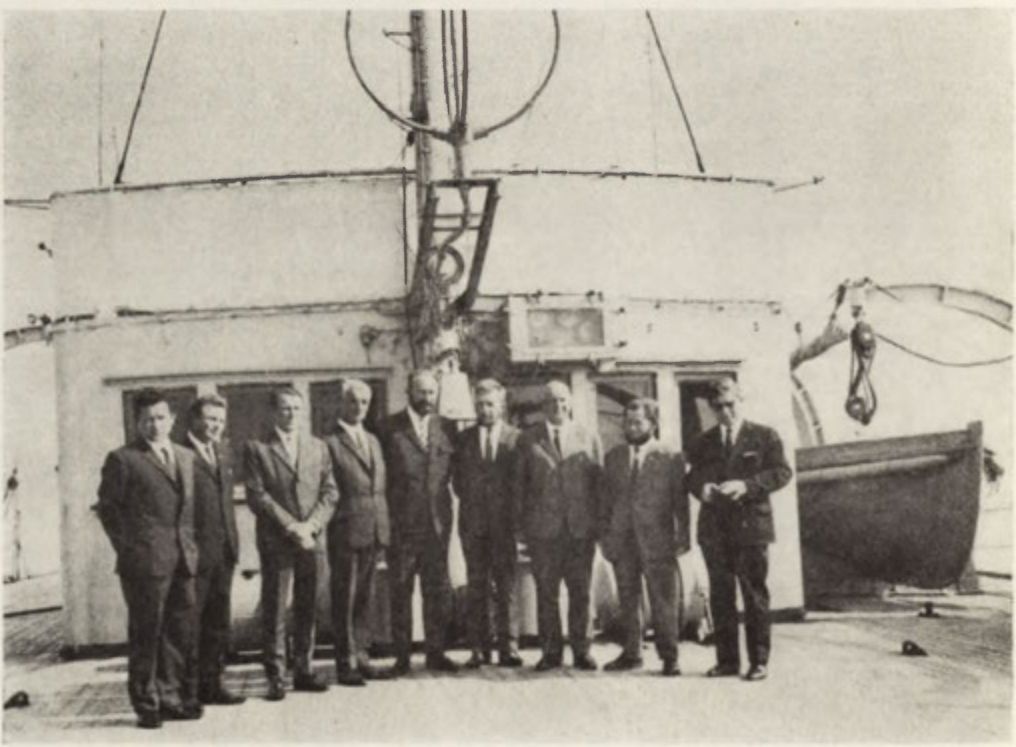

Fig. 2. Members of the Polish Geographical Expedition to Vatnajokull. From left to right: T. Konysz, Z. Churski, J. Szupryczyński (Deputy Leader), S. Jewtuchowicz, S. Kozarski, K. Klimek, R. Galon (Leader), M. Bogacki, G. Wójcik

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Geography, Polish Academy of Sciences and Department of Physical Geography, Institute of Geography, Nicholas Copernicus University, Toruń.

(2) Dr. habil. Jan Szupryczyński, deputy leader of the Expedition, Department of the Physiography of Poland, Institute of Geography, Polish Academy of Sciences, Toruń.

(3) Dr. Mirosław Bogacki, Department of Physical Geography, Institute of Geography, Warsaw University.

(4) Dr. Zygmunt Churski, Department of Hydrography, Institute of Geography, Nicholas Copernicus University, Toruń.

(5) Late Dr. habil. Stefan Jewtuchowicz, Institute of Geography, Polish Academy of Sciences, Łódź. Dr. Jewtuchowicz met a tragical death in July 1972 in Iceland. He was at that time the leader of a research expedition sponsored by the University of Łódź.

(6) Dr. habil. Kazimierz Klimek, Department of Physical Geography, Institute of Geography, Polish Academy of Sciences, Kraków.

(7) Tadeusz Konysz M.Sc., Central Office of Geodesy and Cartography, Warsaw.

(8) Dr. habil. Stefan Kozarski, Department of Geomorphology, Institute of Geography, Adam Mickiewicz University, Poznań.

(9) Dr. Gabriel Wójcik, Department of Climatology, Institute of Geography, Nicholas Copernicus University, Toruń.

It should be mentioned that much valuable assistance in organizing the Expedition was given by Dr. Stanisław Baranowski of the Institute of Geography, Wrocław University.

TRANSPORTATION

On June 3 and 4, 1968, the Polish Navy's hydrographic ship "Bałtyk" was loaded with the Expedition's luggage (camp equipment, instruments, food supply), the "Gaz 69M" automobile, and fuel. The Expedition left Gdynia on June 5. After a rather stormy passage "Bałtyk" entered the harbour of Reykjavik on June 13. On June 15 the Polish group in two hired trucks and the Expedition's car left Reykjavik, reaching the forefield of the Skeidarárjökull two days later. This trip, 331 km long, had to be made during pouring rain and over very rough narrow roads passing over areas of a basalt plateau and over coastal plains. The rate of travel often did not exceed 25 km/h.

FIELD WORK

In the late afternoon of June 16 the base camp of the Expedition was established in the forefield of the main end moraine of the Skeidarár (Fig. 3). Here it was decided, that research work on the forefield of the glacier would be done by a group of seven, including the leader of the Expedition, and that S. Kozarski and J. Szupryczyński would carry research in the forefield of the Sidu glacier (Sidujökull — Fig. 4).

The Skeidarár glacier (Skeidarárjökull) represents a typical example of a glacier lobe, which is retreating from the area of coastal plains and leaves in its forefield typical marginal deposits and landforms which resemble the assemblage of landforms left by the last Pleistocene glaciation in the Polish Lowland. The proglacial area of the Skeidarárjökull contains a wide arc of high end moraines dating back to 1890. Southward, ahead of this arcuate

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Fig. 3. Main Camp of the research group on the forefield of Skeidararjokull. Main end moraine ridge. Photo by T. Konysz

Fig. 4. Two-man camp of the research group in Djupa Valley, forefield of Sidujokull. Photo by J. Szupryczyński

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moraine belt, extends a classical widespread outwash fan; its length is 20–30 km, widening to 50 km where it reaches the ocean shore. Above the surface of the outwash sheet, 1.5 km south of the main 1890 moraine zone, rise rather low moraines of older age. In this area of the main end moraine zone comprehensive investigations were carried out by Professor Galon, Dr. Bogacki and Dr. Klimek. The youngest moraine and outwash deposits, nearest the ice margin, were the object of studies of Dr. Jewtuchowicz. Dr. Klimek also studied the contemporary sedimentation processes on the outwash plain. Mr. Konysz took several hundred photogrammetric photos which consist of the basic data for a topographical map of the forefield and the glacier margin, to be compiled in 1:10,000 scale. At the base camp and next to the glacier margin, Dr. Wójcik put up meteorological stations with recording instruments; he made observations of meteorological phenomena and actinometric measurements (Fig. 5). He also studied glaciological features on the

Fig. 5. Climatologist G. Wójcik establishing the meteorological station on the main end moraine ridge. Photo by R. Galon

glacier surface (Fig. 6). Dr. Churski concentrated on detailed hydrographical observations on the outwash plain, on studies of groundwater conditions and river dynamics (Fig. 7), and on examinations of the limnological properties of the ice-dammed lakes.

Dr. Kozarski and Dr. Szupryczyński made their independent investigations in the forefield and on the margin of the Sidujökull. The forefield of this glacier had been so far neglected by scientific expeditions because of its difficult reach. The glacier snout lies at 700 m a.s.l., flanked by mountain massifs up to 1000 m high. The wide glacier lobe is more than 30 km long. From

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Fig. 6. Measurement and camping on the glacier surface. Photo by G.Wójcik

Fig. 7. Hydrological measurement in the meltwater river

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Fig. 8. Before beginning the field work. Photo by G. Wojcik

Fig. 9. Before crossing a meltwater river on the proglacial area of Skeidararjokull. Photo by R. Galon

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a hamlet called Kålfafell, with a young farmer Snorri Björnsson assisting, the "Gaz 69M" automobile and a hired landrover car hauled camp equipment, instruments and provisions into the Djupá valley. The vehicles had covered barely 17.5 km of this route when further travel became impossible because of a steep, some 150 m high, basalt spur blocking the valley. At the base of this spur the first camp had to be established (Fig. 4). The cars, after unloading, had to return to Kålfafell. With six kilometers of difficult tracks to reach their goal, it took the two scientists eleven tours to transfer gradually in heavily loaded knapsacks all their equipment and provisions to the proglacial area of the Sidujökull glacier. Directly next to the base of one of its moraine chains they established their second tent camp, and from here they carried out their studies of the wide marginal zone of this glacier. The material collected made it possible to compile a detailed geomorphological map similar to that of the Skeidarárjökull forefield. With four end moraines as a basis, the course of recession of this glacier since 1890 up to present times has been determined; the drainage system was also reconstructed as it existed during successive stages of the glacier retreat. Dr. Kozarski and Dr. Szupryczyński also studied the geological features of the forefield, the texture and structure of the ground moraine and of the local ice-moraine hillocks. On the glacier margin they observed in detail ablation processes and, in particular, the ablation moraine and the ablation fans.

COMPLETION OF THE EXPEDITION

On August 25 both groups met at Kålfafell and next day they reached Reykjavik. Here they visited Professor Sigurdur Thorarinsson at the Museum of Natural History; on August 30 the members of the Expedition were guests of the Minister of Culture of the Iceland Government, on August 31 the leader of the Expedition was received by Dr. Steingrimur Hermansson, director of the Research Council of Iceland.

During their stay in Iceland, the Expedition's members were enjoining hospitality of the Polish chargé d'affaires in Iceland, Mr. M. Kroker and the Embassy staff.

In the late hours of August 31 the Expedition boarded the Polish Navy's training ship "Gryf" and left Reykjavik; on September 8 the "Gryf" reached its home port Gdynia.

Nicholas Copernicus University, Toruń
Institute of Geography, Polish Academy of Sciences, Toruń

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GEOMORPHOLOGICAL AND GEOLOGICAL ANALYSIS OF THE PROGLACIAL AREA OF SKEIDARÁRJOKULL

Central Section

Rajmund Galon

INTRODUCTORY REMARKS

Skeidarárjökull, a classically developed outlet glacier of the Vatnajökull large ice cap, descends to the shore plain at an elevation of c.100 m; from the west it is bounded by a basalt upland spur (Lomagnupur 668 m) and from the east by the more differentiated mountain area of Skaftafellsjökull reaching a height of 1085 m. The width of the protruding front of Skeidarárjökull is about 30 km.

For the most part the glacial meltwater is drained off by the centrally running Sandgigjukvisl River and, also, by the widespread networks of two lateral meltwater streams. In the west the glacier runoff is drained by the Sula River which starts its course from the ice-dammed Lake Graenalón at the site where the Skeidarárjökull has pushed ahead of the Vatnajökull ice cap, and which also receives part of the meltwater arriving from the western Skeidarárjökull margin. In the east the lateral stream is the Skeidará which issues partly from under the eastern Skeidarárjökull margin and, partly, from the mountain area of Skaftafellsjökull. Beyond the Skeidará river rised the high-mountain region of Oraefajökull crowned by the highest peak of Iceland, Hvannadalshnukur, 2119 m high, and containing the alpine-type valley glaciers: Skaftafellsjökull, Svinafellsjökull and Fallsjökull.

The area separating the front of Skeidarárjökull from the sea shore is some 12 km wide and is known as Skeidarársandur [1]: this area is built of glacial and glaciofluvial deposits. The principal part in the relief of this glacier forefield is played by what shall be called the main end moraine belt which extends in an arcuate chain some 3 km today's front of the Skeidarárjökull, separating the youngest marginal deposits of the glacier laid down during the past 60 years from the end moraines and outwash sheets which date back to the 18th and 19th centuries. The main end moraine zone is in close relation to a clearly defined readvance of the glacier which took place at the turn of the 20th century.

[1] *Sandur* is an Icelandic term denoting sand or a sandy plain. To limit this term to denote only glaciofluvial deposits is contrary to Icelandic conditions. This term has been coined by German scientists (who equated *Sandur* with *Sander*) and has also been accepted by other scientists. Only the English language created a suitable separate term: outwash.

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This three-zonal pattern of glacial and glaciofluvial drift spread over the Skeidarárjökull forefield, which illustrates the history of deglaciation in the Skeidarárjökull area and the processes which caused the retreat of Skeidarárjökull from the sea, is sharply split into two unequal parts by the valley of the Sandgigjukvisl River. The lateral streams, the Skeidará and, particularly, the Sula have destroyed the more lateral fragments of the end moraines and created wide outwash valleys.

Irrespective of its dissection by the Sandgigjukvisl valley, the zone of glacial and glaciofluvial drift in front of the Skeidarárjökull is sharply divided into two parts: a western and an eastern part. The middle of this zone contains a wide break in the belt of end moraines and other moraine deposits. This break is filled by an extensive outwash plain starting out from inside the marginal zone. The whole zone of marginal deposits and landforms in the Skeidarárjökull forefield is narrowest in the east, gradually widening westward. This asymmetry, caused by a higher intensity of deglaciation during the recent 60 years in the western part of Skeidarárjökull, also finds its expression in differences in the shape of the marginal forms and deposits fringing the glacier. A full series of deposits of glacial and glaciofluvial accumulation appears in the western part, that is, on both banks of the Sandgigjukvisl valley; hence this set occupies a compact section of the main end moraine zone which along its margin is terminated by wide outwash valleys. The most characteristic part of this zone adjoins the Sandgigjukvisl valley in the west (Fig. 1).

The most important element among the marginal forms in front of the Skeidarárjökull is the main end moraine zone, forming a characteristic semicircle and developed and situated in a definite classical pattern with regard to the glacier. Incidentally, the desire to study this typical end moraine was one of the incentives for arranging the Polish Iceland Expedition. Both the forefield of the main zone of marginal forms, diversified by deposits from former ice front positions of the Skeidarárjökull, and its hinterland occupied by youngest glacial and glaciofluvial deposits constitute together with the main zone mentioned a characteristic genetic assemblage. Our description of the marginal landforms of Skeidarárjökull will, first of all, deal with the main end moraine zone followed, in connection with this zone, by examination of the outwashes and of older deposits of glacial and glaciofluvial accumulation in the forefield of this zone and, finally, of the younger glacial and glaciofluvial deposits occupying the rear of the main end moraine zone. The contemporaneously forming deposits in the near vicinity to the present glacier margin will be separately considered. A detailed geomorphological study of Skeidarárjökull dwells on successive sections of the proglacial area which each time is covered by a multi-zone series of deposits and relief forms (see Geomorphological Map at the end of the volume). The first of these descriptions refers to that part of deposits and relief forms which form the west side on the Sandgigjukvisl valley (as prepared by R. Galon). Then a description of two sections of the proglacial area will be given which adjoins the marginal forms discussed on both sides (as prepared by M. Bogacki). Further parts, and fringe parts of the main end moraine zone including the valleys of the lateral streams Sula and Skeidará will be the subject of the field work undertaken by K. Klimek. Finally, the task of presenting detailed investigation of the glacial and glaciofluvial deposits which are currently developing along the retreating frontal margin of the Skeidarárjökull has been given to S. Jewtuchowicz.

The forefield of the Skeidarárjökull called Skeidarársandur has been the object of a number of partial or general examinations covering the entire Vat-

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Fig. 1. Geomorphological sketch map of Central Section of the proglacial area of Skeidararjokull

Outer zone of ice-marginal landforms (main end moraines): 1 — older end moraines, 2 — younger end moraines. Arrows designate moraine gaps, 3 — outwash fans with eroded end moraines and meltwater valleys, 4 — plain of degradation and aggradation, 5 — elongated depressions and kettles or subglacial channels, 6 — inner zone of ice-marginal landforms composed of crevasse filling ridges, eskers, kame plateaus, irregular hillocks, moraine and silt plains. Black lines designate more distinct ridges, indented lines — small escarpments, arrows — places of subglacial overflowing of meltwater, 7 — rivers and lakes. I — area of the oldest ice-dammed lake, II — area of the younger ice-dammed lake, III — area of the youngest ice-dammed lake with disappearing small lakes

najokull region or some larger parts of the proglacial area. Among more recent papers should be mentioned, primarily, the investigations continued for a number of years by Ahlmann and Thorarinsson with regard to glaciological and meteorological phenomena and to accumulation processes observed in the Vatnajokull region (1939). Thorarinsson also published a number of less extensive studies on glacier oscillations in Iceland, and a further paper on marginal deposits; the same author investigated the *jokullhlaup* (1954) process, a problem important from a hydrological and geomorphological point of view. Todtmann stayed repeatedly in the Vatnajokull region where she made glaciological and geological observations along the southern and northern Vatnajokull margin; in 1960 this author published a synthetic glaciological-geological report on Vatnajokull. Further data on Vatnajokull and on its forefield, accompanied by extensive literature, was given in a synthetic study by Okko who discussed glacial deposits in Iceland (1956). Finally there are Hjulström (1952), Price (1969, 1970) and Price and Howarth (1970), who made their re-

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markable hydrological, meteorological and geomorphological observations on the proglacial area of the relatively close Breidamerkurjökull and Fjallsjökull; these observations are of particular importance for purposes of comparison.

It should also be mentioned that a number of air photographs covering the region under discussion, taken in 1946 and again in 1965 were of marked significance in determining the spatial structure of the investigated land relief. A detailed examination of these photographs was granted to the Polish Expedition by courtesy of the Iceland's Landmaeling; the Expedition's own photogrammetric surveys were only ready for use after they had been brought back to Poland.

The authors of the present study are going to refer repeatedly to all publications previously prepared on the subject of deglaciation and of deposits and landforms, connected with this deglaciation in the forefield of Skeidarárajökull. But it should be stressed, that both the scope and the scale of accuracy of the geomorphological examinations made by the Polish geomorphologists classify this work as the first detailed investigation of the geomorphological problems of the area involved, supported by detailed geomorphological mapping.

Parallel with our geomorphological investigations and a variety of other research work, our geodesist T. Konysz made his ground-photogrammetric surveys covering a considerable part — mainly the western and central — of the proglacial area of the Skeidarárjökull. His work is added to our study in the shape of a separate topographical atlas compiled in 1 : 10,000 scale. This atlas is of high documentary value and, in conjunction with our detailed geomorphological maps and descriptions, it is well suited to throw light upon the young and youngest glacial forms which occur on Iceland, and to correlate them with the young marginal forms of the Scandinavian inland ice which mantled most of Poland in Pleistocene times.

THE CENTRAL SECTION OF THE PROGLACIAL AREA OF SKEIDARÁRJÖKULL THE MAIN MORPHOGENETIC ZONES

The area discussed here comprises the morphologically most diversified assemblage of landforms and deposits associated with the successive halting stages and decay which the Skeidarárjökull has passed through during the past 70 to 80 years. In this area one can distinguish a number of morphogenetic zones which correspond to the processes of deglaciation diversified in both time and space. Within the limits of the main end moraine zone there is noticeable a division into external, older moraines which are destroyed to a higher degree and partly converted into block-strewn regions, and into younger zone consisting of ridges, hillocks and moraine plateaus adjoining the inner side of the older moraines mentioned. Transverse depressions in both parts of the end moraine zone, the so-called moraine gates or breakes, give access to the forefield; they pass into outwash fans or they continue as valleys of meltwater streams. In the immediate rear of the end moraine zone there extends a system of elongated depressions running parallel with the moraines, and these depressions are entered by a number of channel-type valleys. Finally, there is the innermost and widest morphogenetic zone in which marginal deposits and landforms predominate, created mostly at the time of prevailing stagnation of the glacier. Hence here can be distinguished:

(1) the zone of outwash fans and valleys formed by meltwater streams,

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(2) the main end moraine zone subdivided into zones of older and of younger end moraines,

(3) the zone of elongated kettle-type depressions,

(4) the inner zone of marginal deposits and landforms.

THE MAIN END MORAINE ZONE

In order of importance, first place in our detailed discussion of the particular morphogenetic zones shall be given to the main end moraine zone and its subdivision into two parts.

The older end moraines. In the central part of the main end moraine zone the older and the younger end moraines are closely interrelated; they represent a compact system of elevations which reach relative altitudes up to 30 m and which drop in height towards both the outwash plain and the younger glacial landforms by distinct steps which show slopes at an angle of no more than 33°. Locally the boundary line between the older and the younger end moraines is indistinct. The diversity in origin and age between them is mainly indicated by the difference they show in the configuration of the end moraine deposits and in the way they have survived. In contrast with the younger end moraines which morphologically appear mostly in the form of moraine ridges, the older part of this zone has the typical features of a flat moraine plateau, slightly undulating, with the surface strewn with numerous rock blocks. The moraine plateau shows a slight northward tilt. Here and there in the zone of the older end moraines hummocks a few meters high occur; sharply defined moraine ridges are absent here. On the other hand, the moraine surface shows thin runnels parallel with the axis of the zone; these old runnels probably go back to the stage of marginal runoff and are now filled with aeolian material. The older moraines are built of uneven sized sands with many boulders; their mineral composition will be discussed later.

It is believed that the surface of these older moraines has partly been destroyed by meltwater streams during subsequent ice front positions of Skeidarárjökull. Clear proof of the destruction of these moraines is the strongly dissected southern rim of the end moraine zone (Fig. 2, 3). This sort of dissection must have been brought about by the numerous meltwater streams which were issuing under hydrostatic pressure from glacier gates and which after passing the moraine ridges escaped into their forefield. Corresponding to the glacier gates are moraine breaks (or gaps) from which both the outwash fans and outwash channels began to develop. But the older end moraines were cut apart not only by moraine gates which developed in the chain of these moraines, but also by similar though bigger breaks derived from the younger end moraines. (The author will deal with the moraine breaks in a later separate chapter). Ultimately the zone of the older end moraines became much reduced in size; locally, especially in the eastern part of the discussed moraine zone and in the west, near the last moraine gate, the older moraines were completely destroyed and, directly at the boundary between the main end moraine zone and the outwash plain, they are followed by the younger end moraines. Evidence of the original extent of the end moraine zones is the widespread stone fields encountered in many places in front of this zone, in the area of the outwash fans belonging to this zone (Fig. 3, 4).

The younger end moraines. As said before, the older end moraines of the main end moraine zone pass abruptly into the younger end moraines of this zone. However, at many places the boundary line of the younger end moraines is marked by a sharply defined moraine ridge ahead of which, from the side

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Fig. 2. The main end moraine ridge dissected by overflowing meltwater valleys (named end moraine gaps or breaks). In the foreground outwash plain with stone remnants of destroyed end moraines. Photo by R. Galon

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Fig. 3. The main end moraine ridge with traces of denudation. Photo by R. Galon

Fig. 4. Main end moraine zone. Photo by R. Galon

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of the older end moraines, lies a narrow elongated marginal-type depression. In every respect the relief of the younger end moraines is much more diversified (Fig. 5). The most characteristic of the relief forms of this zone are moraine ridges built of stones with an unusually narrow and sharply peaked ridge, predominantly built of uneven-sized rounded boulders. The height of these ridges varies from 2 to 5 m, the slope gradient is apt to be as much as 35°. These moraine ridges occur in several parallel lines, but they show many gaps in their lines. From one of these moraine ridges a crevasse filling wall forks is

Fig. 5. The main end moraine ridge with the highest culmination on the discussed area (119.3 m a.s.l.) and slope steps seen from N. In the foreground a section of the zone of elongated basins. The relative height is 46 m. Photo by R. Galon

running over the slope of the moraine zone in a NE direction. This sort of crevasse filling ridges extending at a right angle to the end moraines and apparently ingrown into these ridges, are also seen at other places in the area of the younger end moraines, but they always extend northward; this, for example, is the case near to the end moraine gaps Nos. 5 and 10 (Fig. 1).

Apart from moraine ridges, this zone contains mounds, even moraine hillocks formed recently, with large numbers of boulders on their crests and slopes. The moraine material rests in labile equilibrium. It suffices to kick one of the boulders resting on the steep slope to make it tumble down and take along further boulders. At several points mounds were observed with a depression in their centre area, that is, circular-shaped mounds, the result of melting of ice blocks buried in these elevated areas.

Besides moraine ridges, mounds and hillocks, moraine plateaus can be observed locally which, as was mentioned before, occur for the most part inside the zone of the older end moraines. In the area of the younger end

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moraines such moraine plateaus lie mostly near the Sandgigjukvisl valley, in the rear of the moraine ridge; here they form a system of plateaus separated by steps each several metres high. But moraine plateaus have also been observed in the western part of the discussed end moraine zone.

On their surfaces, at about 1 m intervals, the moraine plateaus show regular stone strips or ruts 0.3 to 1.0 m high and about 1 m wide; they are separated by flat furrows built of more fine-grained material (Fig. 6). These strips run continuously, independent of relief diversities of the moraine plateaus, in

Fig. 6. Longitudinal stone strips separated by flat furrows on the surface of the main end moraine ridge near Sandgigjukvisl valley. Photo by R. Galon

a course parallel to the end moraine ridges. On the moraine plateaus near the valley of the Sandgigjukvisl river, some 30 stone strips of this kind were discovered the last of which, counting from the lower side, is higher and rather resembles a typical moraine ridge. These minor features are undoubtedly linked with the structure of the glacier, with the occurrence of shear planes and, connected with these shear planes, with deposition in series of moraine material.

The area of the younger end moraines is also accompanied by a number of what might be called negative landforms. Usually the moraine ridges run parallel with elongated depressions with an uneven botton, now filled with aeolian sands. They bear the features of marginal drainage channels, but in the moraine plateau area many closed depressions can also be seen. The larger among them, of rather elongated shape, run parallel with the end moraines. Moreover, minor kettle holes occur on virtually the entire area of the younger end moraines. Often they are young sudden collapses of the surface, the consequence of buried dead ice blocks melting. The process of formation of this type of depressions, even of some short valleys, has been observed on the slopes of the Sandgigjukvisl valley.

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As mentioned before, the area of the younger end moraines directly adjoins the area of the older end moraines. Only where the older moraines have been destroyed does the area of the younger end moraines pass with a sharp scarp directly into the outwash plain. Conditions are different on the northern boundary of the younger end moraine zone. Here the entire area consisting of moraine ridges, hillocks and plateaus drops by distinct steps to the bottom of the elongated depressions which here constitute the next morphogenetic zone of the Skeidararsandur section discussed. Conspicuously noticeable are

Fig. 7. Weathered palagonie erratic block. Photo by R. Galon

the slope steps near the Sandigigjukvisl valley; they can be seen clearly from the valley, looking southward. Here four steps with very sharp breaks rise in succession. Near moraine gate No. 5 were found as many as six small terraces in the slope, at 5 m altitude difference or more. Similar slope steps, some 3 to 5 m high, exist near end moraine gap No. 10. It might be suggested that the sharper slope breaks mentioned here, situated in the northern slope of the area of the younger end moraines, must have originated as contact walls during successive stages of recession and stagnation of the gradually diminishing glacier (Fig. 30). This origin of these forms seems to be indicated by transverse crevasse filling ridges at some places joining at an acute angle the contact walls; this is the case, for instance, near end moraine gap No. 10 where the crevasse ridge also shows steps.

Moraine deposits. The moraine cover predominantly consists of boulders and sandy moraine material. It was found, that the boulder mantle is underlaid by uneven-grained sands deposited in different ways, locally stratified or streaked, with black silt intercalations but also containing first-size erratics. In general, this material becomes finer-grained with increasing depth (Fig. 8).

Samples of moraine deposits were collected from the end moraine ridge near the Sandgigjukvisl valley. Because of the kind co-operation of Prof. A. Ga-

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weł[2], the head of the Department of Mineralogy and Petrography at the Jagellonian University in Cracow, after the return of the Expedition to Poland these samples were analyzed with regard to their granulometric, mineralogical and chemical properties. On the basis of these examinations A. Gaweł reports that the moraine samples examined are silty-sandy deposits containing little (5.4%) material that is flushed away. In contrast to the outwash deposit this material is more fine-grained but, at the same time, it shows poorer sorting. The quantity of sideromelanite increases in fractions of less than 0.60 mm grain size; an increase is also shown in the feldspar content. Also in this

Fig. 8. Moraine deposit. Main end moraine ridge near Sandgigjukvisl valley. Photo by R. Galon

moraine deposit a large quantity of some brown glass was found which gave a light chocolate-like colouring to some of the small fractions. The larger rock fragments in the samples are for the most part grey or dark-grey basalt, light-green tuffas, or palagonites; less frequently they are sideromelanites (see Table 1).

However, at a lower altitude, near the gate above the Sandgigjukvisl River, banks of moraine material again appear. Further down, dark silts occur again, locally with 5 to 10 cm streaks and an admixture of small rock particles and gravels; they slightly dip northward. Judging from the laboratory examinations

[2] Professor A. Gaweł has prepared a separate study dealing with the granulometric, mineralogical and chemical properties of the deposits collected from the forefield of the Skeidararjökull glacier; for these examinations he used samples received from R. Galon.

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made by Prof. A. Gaweł it appears, that this deposit contains only 19.8% coarser fraction of more than 0.060 mm diameter, and that the large-size fragments are dark-grey basalt, while among the smaller-size fraction sideromelanites predominate; and that the fraction which can be flushed out consists of chips of volcanic glass.

TABLE 1. Granulometric parameters characterizing certain deposits from the Skeidararjokull forefield, in the light of A. Gawel's laboratory tests

Kind of deposit	Median Md	Quartile Q_1	Quartile Q_2	Dispersion $S = 1/2(Q_1+Q_2-2Md)$
Main end moraine	0.43 mm	0.15 mm	1.20 mm	0.245
Moraine deposit of inner moraine zone	0.49 mm	0.15 mm	1.02 mm	0.095
Outwash-type deposit in substratum of the main end moraine	1.20 mm	0.60 mm	2.00 mm	0.125

Some 20 m above Sandgigjukvisl River this silt deposit, here perhaps more distinct, reaches its bottom. A rock pavement appears here under a different, older series of sediments consisting of clearly stratified coarse-grained sands and gravels alternating with light- and dark-colours, with well-rounded boulders of various size. This series show a very much diversified composition. In the light of laboratory examinations made by Prof. A. Gaweł they represent the substratum of the moraine deposits and landforms described above and being the result of some re-advance of the glacier. In the 0.25 mm fraction fine gravels of isometric shape and rounded edges predominate. Light-coloured feldspars are also represented, especially in the 0.2 mm and the 0.6 mm fractions. The coarser material (1.2 mm and 2.0 mm) shows similar granulometric features and consists mostly of black basalts and sideromelanites (Table 1).

To make characteristic structure of the end moraine zone more explicit let us add the description of exposures into the zone where the moraine hillocks are built of stratified or streaked sands and silts and show a structure indicating settling or downflow. This deposit, consisting of alternating dark-yellow silts and loose black sands, occasionally with some gravel admixed, spread in the form of fluidal streaks dipping northward in conformity with the land surface, may be considered as ablation moraine (Fig. 9).

As seen from Prof. A. Gawel's laboratory examinations, the material resembles the moraine deposit from above the Sandgigjukvisl valley. However, this sample contains many more parts easily flushed out (34.6% as against 5.4%). The coarser fractions contain isometric grains, while in the finer fractions (of 0.43 mm diameter) elongated grains of irregular shape occur. Basaltic glass of light-brown colour is predominant here. As a new element a soiled glass of pumice-like, fibrous texture appears in this material (Table 1).

The deposits which build the main end moraine zone described up to now, can be subdivided into:

(a) stratified,
(b) streaked,
(c) unsorted deposits.

Lithologically the stratified and streaked deposits are:

(a) either dark yellow silts,
(b) or coarse-grained dark sands.

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The stratified deposits show the features of sediments found in ice-dammed lakes next to the glacier margin, or of meltwater basins in the glacier base or on the glacier surface; they also may be evidence of the runoff of some former meltwater streams or may have originated as a supraglacial ablational deposit after the melting of ice which in vertical steps had contained some rock material. The streaked deposits are mainly the product of surface ablation, especially when they show fluidal structures.

The granulometric, mineralogical and chemical examinations made by Prof. A. Gaweł on the samples supplied to him from the area of R. Galon's Icelandic research have shown, that the fine fractions (of 0.075 mm diameter) of the end moraine and of the material derived from the outwash plain and from ice-dammed deposits are all of an identical mineralogical character. All this material is simply a greyish-brown palagonite silt with a large admixture of angular volcanic glass. The coarser fractions of this material

Fig. 9. Ablation moraine on the slope of the main end moraine ridge. Photo by R. Galon

also show a lack of essential differences in their mineralogical-petrographic composition. All these deposits contain an identical worked-over material derived from basalt-type effusive rocks. However, the coarser fractions differ markedly in their granulometric composition. Moreover, A. Gawel found that in mineral and rock type the deposits of the Skeidarárjökull forefield show no traces of dust particles of labradorite, hypersthene, magnetite or apatite which might be ascribed to the 1947 eruption of the Hekla volcano. Even so, the mineralogical and petrographic features of these coarser fractions do not fail to show some effects of the Katla volcano in 1918 and, even, traces of the volcanic eruption which took place in 1783.

From the description given so far of the geological structure and the litho-

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logical features of the moraine zone it appears, that in this region no typical boulder clay has been observed. The moraine material encountered here consists of loose sands of unequal granulation and of diversified boulder material. The way how these moraine deposits appear does not show any thrusting of the material as usually accompanies glacier retreats and advances. Hence the relief forms of the marginal zone discussed seem to be caused not so much by movements of the glacier than by its structure, i.e., the effect of a segregation of the moraine material within crevasses and shear planes of the glacier. On the other hand, the formation of moraine hillocks which decrease gradual in height the nearer they come to the zone of the larger depressions, has probably taken place under conditions prevailing while the glacier was retreating and gradually decreasing in height. This problem of the way deglaciation has been proceeding in this region will be taken up again later.

The end moraine gaps or breaks. As mentioned before, the main end moraine zone, in the section containing both the older and the younger end moraines, has gaps or breaks in the moraine chains at many places. Consistent with the concept of glacier gates with which these breaks are genetically linked, the term "moraine gates" has been coined. In the section of the end moraine zone here taken into consideration, there are 10 moraine gaps, not counting the wide break in the zone of end moraines occupied by the valley of the Sandgigjukvisl River. Some of the moraine gaps are incised only in the younger end moraines at places where because of destruction of the older end moraines the rim of the younger end moraines came into direct contact with the outwash plain. However, all larger moraine gaps disrupt both the older and the younger end moraines; but in cases like this the older end moraines near a gap have been destroyed and only an agglomeration of larger rock blocks are evidence of where they once were (Figs. 1, 7).

The moraine gates differ not only in their width but also in structure. The pattern of a moraine gap as they occur in the area under discussion is illustrated by three figures (Figs. 12–14).

A concentration of meltwater takes place in the glacier gate; this causes an erosive hollow to develop from which the water is expelled by flowing over the moraine ridge. In this way a flow sill is formed, often characterized by an accumulation of very large stones; after passing this break the meltwater erodes runoff channels or regular valleys incised into the outwash fans and growing shallower in southward direction (Figs. 12, 14). It is also conceivable, that the kettle in the glacier gate is the trace of a former artesian flow gushing out from under the glacier margin.

The pattern of evolution listed above may be applied to description of particular moraine gaps, provided certain differences are taken into consideration. First of all, the majority of end moraine breaks have more than one step in their sills: two, as in gap No. 2 or, even three — and in this case it is one of these sills in which most of the meltwater was once concentrated (gap No. 9). In some gaps the sills were formed in two stages, meaning that the original sill has been gashed a second time and in it two smaller and lower sills were produced, as in gap No. 9. Generally speaking, the end moraine breaks differ as to the depth to which they are incised and to the morphological part they play within the main end moraine zone. Some of the breaks like Nos. 1, 3 and 7 are only faintly marked and do not cause any marked change in the longitudinal profile of the end moraine zone. But moraine gap No. 2 is worthy of particular interest on account of its characteristic shape (Figs. 10, 11), its connection with the zone of large elongated depressions, and of the relatively

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Fig. 10. Gap No. 2 in the main end moraine ridge (see Fig. 1). Photo by R. Galon

Fig. 11. Channel in the moraine gap No. 2 originated by overflowing meltwater (from right to left) during the standstill of the glacier margin along the end moraine
Photo by R. Galon

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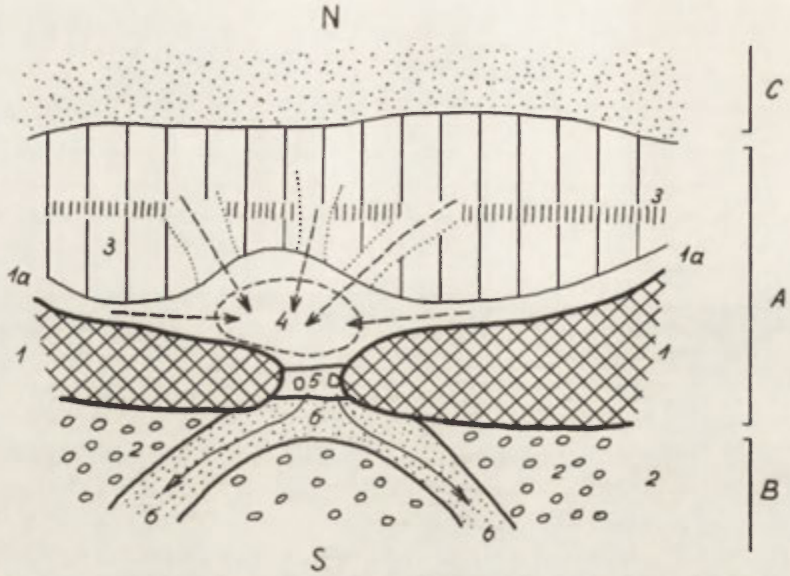

Fig. 12. Schematic diagram of a moraine gap

A — main end moraine zone, B — outwash zone, C — elongated depression: 1 — end moraine ridge, 1a — glacier margin at the time of the end moraine and when the moraine gap was formed, 2 — outwash fan associated with the above end moraine, 3 — moraine hillocks and plateaus, 4 — central erosive kettle in former glacier gate, 5 — flow sill in end moraine gap formed of large rock blocks, 6 — runoff channels incised in outwash plain

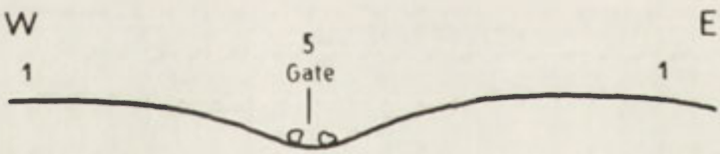

Fig. 13. Schematic diagram of a moraine gap. For legend see Fig. 12

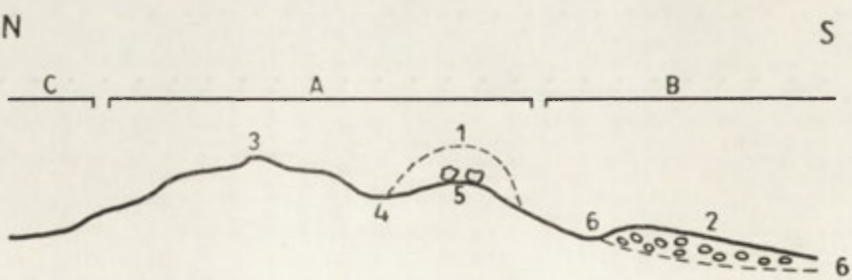

Fig. 14. Schematic diagram of a moraine gap. For legend see Fig. 12

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deep (as great as 2 m) and numerous flow channels which in the forefield of this gate have spread in radial pattern. Due to the erosive action of meltwater streams accumulating in this older gap the end moraines were destroyed and in the region of the younger end moraines a deep hollow has developed. However, it seems probable that the decrease in size of the moraine area, especially of the older moraines, near gap No. 2 might have taken place at some earlier time, when some earlier moraine break existed in the zone of the older end moraines or when the younger end moraines started developing. In this case, this might be considered as an example of how glacier gaps have persisted while the glacier was retreating.

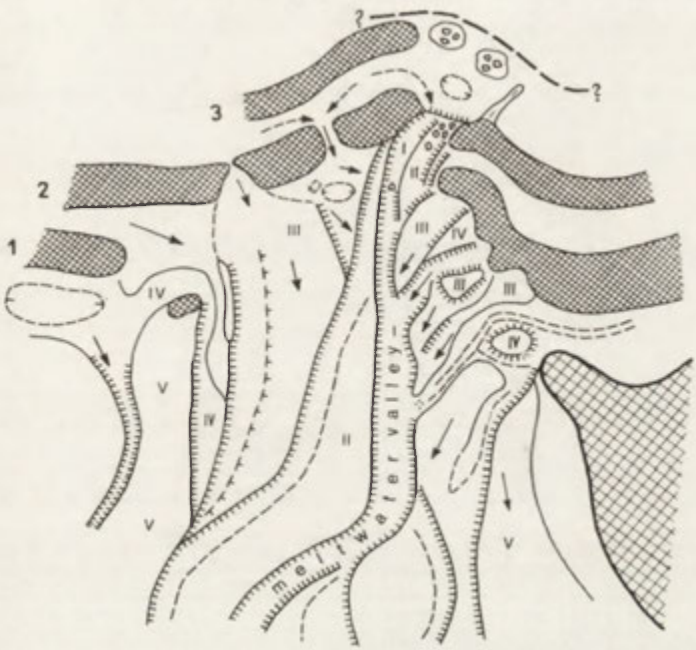

Fig. 15. Moraine gap No. 5: end moraine ridges (1, 2, 3), small dissected outwash fans (V–III) and meltwater valley with terraces (V–I)

Despite this the most thoroughly developed are moraine gaps No. 5 (Fig. 15) and No. 10 (Fig. 16) which dissect the main end moraine zone down to its base; one might call these gaps regular breaches in the moraine zone. From both gaps a number of outwash fans and tracks have started, evidence of the great power of the meltwater streams which arrived from several smaller glacier gaps. Moreover, the outlets of both these gaps pass directly into wide meltwater valleys which extend far into the moraine forefield. An important role in proglacial drainage was also played by a valley, dry today, which started its course from gap No. 5 (Figs. 15 and 17). Near this gate, farther westward, in the southern margin of the older end moraines a smaller moraine gate may be seen with its runoff track which has since been undercut by the valley starting from gate No. 5.

Gap No. 10 is a conspicuous extensive and fairly wide sill, raised on one side 18 m above a basin which was once filled with an ice mass and, from other the side, i.e., from the south, passing by a sharp 15 m escarpment into an outflow channel which today is diversified by large kettles (Fig. 16). This

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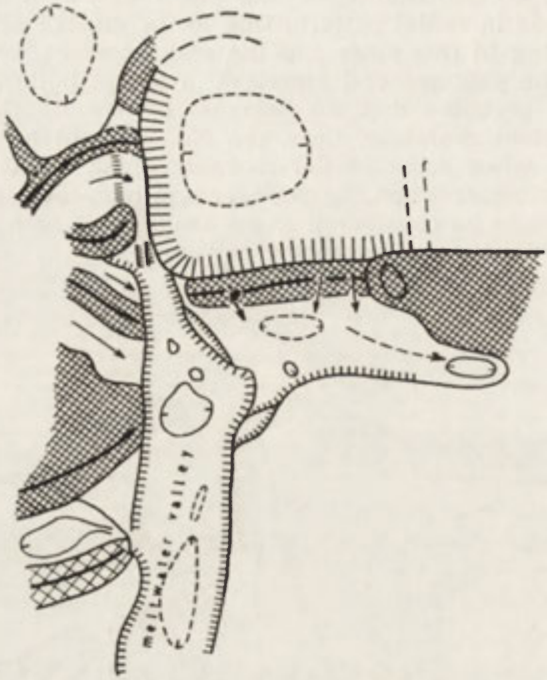

Fig. 16. Moraine gap No. 10

Fig. 17. Outwash fans in the moraine gap No. 5. Photo by R. Galon

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flow sill was later cut, and the meltwater runoff over this sill proceeded at a lower level. Finally a runoff track developed at a level that was even lower, and a meltwater valley developed which dissected the flow still down to its very base. This last-mentioned runoff continued until the glacier receded from the line of the main end moraines and until in the rear of this sill an ice-dammed lake developed.

The evolution of the moraine gaps, interlinked with ice gates in the diminishing glacier, proceeded in a complicated manner. The author will return to this topic in the following chapter which covers outwash fans and gullies. In a similar way he intends to discuss a former moraine gate, which later was transformed into the gap valley of the Sandgigjukvisl River. A further comment on moraine breaks is given by M. Bogacki.

Traces of changes due to denudation that can be seen in the discussed main end moraine zone and its numerous moraine gaps, especially in the slopes of the moraine ridges, are relatively scarce. Slope processes led to the formation of gullies, to colluvial fans and shelves, to a widening of the original moraine gaps and to landslides with sharply marked scars. However, on account of the lack of vegetation and the looseness of the material, as a rule all kinds of slope forms are soon obliterated and the slope inclination comes to assume the natural, usually rather steep angle common to sandy moraine material.

THE ZONE OF OUTWASH FANS AND MELTWATER VALLEYS LINKED WITH THE MAIN END MORAINE ZONE

The forefield of the main end moraine zone consists of two genetically differing parts: (a) — the relatively narrow zone of meltwater deposits, in the form of outwash fans and streaks dissected by flow channels and valleys incised by the erosive action of these meltwaters, and (b) — an extensive plain of degradation and aggradation, the deposits of which, as shown by the geological structure of the main end moraine zone, underlie these moraines including the outwash material connected with them (Fig. 30).

The above mentioned *plain of degradation and aggradation* which is the substratum of the series of glacial and glaciofluvial material of Skeidararjökull deposited during the recent 70 to 80 years — probably due to some more marked oscillation of the glacier — is genetically associated with the process which caused the decay of the Pleistocene ice cap during the Holocene. This process of ice retreat has been interrupted by the periods of standstill and re-advance of the glacier. Evidence of these processes of degradation are the universally occurring stratified sand and gravel beds with their southward dip, and the remnants of end moraines appearing as mere accumulations of boulders or, even, of strongly eroded end moraines shaped like flat ridges of some 10 m relative height, built for the most part of slightly rounded boulders and coarse gravels. This latter refers to the oldest end moraines, the extension of which is accompanied by accumulated piles of boulders. In the gullies one notices the way these boulder extend downward. It seems probable that the oldest end moraine in the investigated area goes back to the latter half of the 18th century. Detailed data with regard to this end moraine is given in M. Bogacki's paper.

From the above it can be seen, that the plain discussed of degradation and aggradation is older than the main end moraine zone, and that no genetic connection exists between the material building this plain and these end moraines. On the other hand, the meltwater valleys and channels which issue from the main end moraine zone and mark the period of halt of the glacier along this

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zone, dissect in many places this plain of degradation and aggradation in their southward escape. Undoubtedly along these erosive tracks, especially where they gradually vanish, processes of aggradation have also been taking place, corresponding in age to the main end moraine zone. At present these old tracks are dry, and the sands filling them are exposed to wind action. But the processes of delta-type accumulation have moved southward, towards the marine lagoon, and for these processes the material is being supplied abundantly by the powerful glacial rivers Sula, Sandgigjukvisl and Skeidarà. Hence, the area of degradation and aggradation may be definitely classified as polygenetic plain formed by the processes of accumulation and erosive denudation during the late-Pleistocene and Holocene times of deglaciation, and as a plain which gradually grew southward by the accumulation, partly in a deltaic pattern, of material brought along and deposited by contemporaneous outwash streams.

In this plain, which has originated by degradation and aggradation, boulders can be noticed striated by wind action (see M. Bogacki's paper); but, for all the aeolian processes observed and the strong winds arriving predominantly from the glacier, very few dreikanters have been found. On other hand, the deflation of sands from stationary gravel accumulations, which then remain sorted in conformity with the prevailing wind direction, is noticeable.

The zone of outwash fans and streaks which are intimately connected with the main end moraine zone is relatively narrow; it has a width of barely a few hundred meters and widens ahead of the moraine gaps. In this place it is necessary to correct an opinion, frequently stated in textbooks and often supported by suitable drawings, to the effect that the whole plain of degradation and aggradation is genetically linked with the main end moraine zone. In fact this narrow outwash zone is built of stratified boulders, gravels and sands, and the surface of the outwash sheet has a marked dip which rapidly decreases with growing distance from the end moraines.

The outwash zone is one continuous plain, meaning, that the meltwater streams have been running into the forefield not only through gaps in the moraine chains but that they were also escaping over the crests of the moraine ridges as shown by numerous minor erosive gaps in the extreme moraine walls — every one of which has been the place of water runoff from the system of elongated furrows between the ridge chains; they may be considered moraine as gates *in statu nascendi.*

However, the moraine gaps were not only the places of accumulation but also the places of erosion by meltwater flows. In the first stage, while the farest exterior end moraines were being piled up, an outwash fan was developing. In further stages of deglaciation, and always within the main end moraine zone, the outwash fans were cut. Groups of the shallow channels or valleys were being formed which spread radially into the forefield of the gates. This took place particularly intensively in front of gaps Nos. 1 and 2 (a joint feature) and of gaps Nos. 3 and 4 (a joint feature too), and similarly in front of gaps Nos. 7, 8 and 9. These small valleys are 0.3 to 1.5 m, exceptionally 2 m deep, with flat floors and steep banks. As mentioned before, southward they grow shallower and finally disappear in the degradation-aggradation plain. Some of these shallow channels dissect the zone of the oldest and moraines; others which tend toward the Sandgigjukvisl valley deepen near this valley and change into deep ravines.

In front of the moraine gaps Nos. 5 and 10, large *meltwater valleys* have developed; most important is the valley issuing from gap No. 5 (Fig. 1). While the glacier was halting along the main end moraine zone, this valley with its

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numerous terraces (Fig. 14) dry today, has been the central track of meltwater runoff. At that time, that is, at the turn of the 19th century, the Sandgigjukvisl valley did not yet exist. Figure 15 reveals the mechanism of the formation of the successive outwash fans and channels, as they were incised into the previously developed outwash plain and as they started from the base of the end moraine ridges. The highest outwash level (V) 25 associated with the older end moraines, and the next lower levels with two or three halting stages of the glacier expressed by end moraines A, B and C. The individual outwash fans or flow tracks start from moraine gates or from recesses in the moraine ridges. The outwash channel marked I represents the lowermost valley level; this is the valley floor (Fig. 15) which has been incised into higher levels during the final stage of the growth of moraine B. At that time the meltwater streams were escaping from a glacier gate in the near vicinity of end moraine B, at the place where the valley floor, lined with rounded boulders, shows an abrupt step 2 m high (an ice contact). At that same time a small crevasse filling ridge was formed, a tributary form to end moraine B.

The question arises, whether the escape of the meltwater through this moraine gate was taking place while end moraine C was already being formed. This seems possible, provided ice-dammed water had been ponded up in front of the glacier margin to the floor level (the fifth) of the outwash valley.

The result of the evolution of runoff conditions for the meltwater flow as described above was the formation of the outwash valley with the terraces mentioned before, issuing from gap No. 5 (Fig. 15). This classical meltwater valley which today is completely dry, is a useful object for studies of the way valley morphology took place under proglacial conditions. The structure and the profile of the terraces and of valley floor testify to the high transporting and eroding power of the water. At a distance of about 3 km from the moraine gate this valley becomes shallower, and as to the grain size the material today filling the old channel does not differ from the deposits of the surrounding plain of degradation and aggradation; partly it is even more fine-grained. Until recently the lowest reach of this valley was drained by a small brook; today this region shows a very shallow groundwater table. Finally the valley vanishes completely in the plain near today's lagoon and passes into the widely spread accumulation cover. The question of the now dry valley and of its relation to the Sandgigjukvisl valley will be discussed in further chapters.

Facing moraine gap No. 10 a similar but shorter and narrower valley of meltwater runoff has developed (Fig. 16) which has also been discussed by M. Bogacki. Here again the valley floor starts from an ice contact step and dissects the short outwash fans left after previous standstill stages of the glacier, as are perpetuated by chains of end moraines. Above the valley floor, here and there in the slope small terraces can be seen. From the east, this valley is penetrated by a 0.5 to 1 m higher degradation plain formed by meltwater which in three streams was escaping through the end moraine break. On this surface as well as in the valley floor several larger and a great number of small kettle holes appear, showing shores marks of former proglacial lakes. The degradation plain which here is lowered by erosion may be considered as the third terrace level, while the small step between the degradation plain and the valley floor would represent the second terrace, and the wide valley floor, the first terrace level.

The meltwater valley in front of gap No. 10 is closely linked with one definite glacier standstill, probably the same during which the outwash valley at gap No. 5 was formed, or some time later. This question is difficult to solve,

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because part of the marginal material deposited on top of the dead ice has been swept away within an extensive meltwater basin. And during the next standstill of the glacier, indicated by a younger standstill line (see Fig. 16), a wide ice-dammed lake developed here. At any rate, this valley floor from gap No. 10 was connected with this ice-dammed lake and the meltwater has been running through this gate for a relatively long time.

In a downstream direction, however, the dry meltwater valley in front of moraine gap No. 10 is rather short; relatively soon it loses its sharp contours and barely reaches the older moraines in the plain of degradation and aggradation, merely as a shallow rill filled with wind-borne fluvial materials.

Drainage of the forefield of the main end moraine zone is today fully taken care of by the Sandgigjukvisl River, which starts its run in the marginal zone of Skeidararjokull and passes the main end moraine zone by a wide gap. The valley of this central outwash river and its relation to the dry valleys described above is going to be the topic of one of the following chapters.

THE ZONE OF ELONGATED DEPRESSIONS

In the immediate rear of the main end moraine zone a system of large elongated depressions has developed, strung along the step-wide descending northern slopes of the main end moraine belt. These lengthwise extending depressions show remarkable differences in width and a great variety of their bottom features. In striking contrast are, first of all, the basin-like wide depressions and the narrow depressions between them with their characteristic sills. Next it should be noted, that northward the zone of these depressions borders on a great number of similar depressions, mostly of elongated shape, but extending in a N–S direction (Fig. 1). The deepest part of the zone of the elongated depressions, at 73 m a.s.l., contains a vanishing lake. Slightly east of this place the end moraines attain their greatest elevation (120 m a.s.l. — see Fig. 5). Hence, the relative height of the end moraines is here as much as 47 m, whereas with regard to the outwash forefield this highest part of the moraines has only 30 m.

The floors of the discussed depressions are lined with sandy material of a diversified size, mostly of wind-borne origin. Near the vanishing lake, coarse-grained dark sands were observed, in 10–15 cm intervals alternating with 1–2 cm laminae of light-coloured silt. This material represents successive processes of accumulation of a periodically widening lake and of deposition of aeolian sands brought in by frequent powerful winds from above the glacier. This sort of aeolian mantling of the floors of the depressions also extends upwards over the lower parts of the slope. But here the opposite process also takes place: erosive dissection of the slopes and a deposition of the colluvial material at the slope basis in the form of cones and loose debris.

The slopes of the depressions clearly show also traces of abrasive incisions, representing shore marks of successive stages in decreasing water levels of the ice-dammed lakes. And near the vanishing lake, next to the valley of the brook emptying to the lake, a distinctly marked littoral plateau can be observed some 3 m above the water level in the lake. Similar shore lines can also be traced in the northern slope of this depression and in one of the N–S depressions, at 2–3 m and at 5 m above the basin floor. Farther, near moraine gate No. 10 three former shore lines can be seen in the local wide and closed basins; the highest lies 5 m above the bottom of these depressions (Fig. 18). Here were also found typical ice-dammed lake deposits consisting of laminar silts; here layers of more fine-grained brown material, about 1 cm thick, alternate with

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Fig. 18. A section of the zone of elongated depressions with the 5 m shoreline of ice-dammed lake. Photo by R. Galon

Fig. 19. Ice-dammed lake deposits. Photo by R. Galon

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coarser-grained yellow material. At greater intervals, about every 10 cm, beds or intercalations occur several centimetres thick of coarse-grained black sand (Fig. 19). This series of ice-dammed deposits shows characteristic collapse structures.

Judging from laboratory examinations made by Prof. A. Gaweł, the ice-dammed deposit discussed above, which additionally contains a humus substance, consists of 93% of easily flushed out material; these parts are exclusively volcanic glass of sideromelanite type. Hence the fraction of more than 0.060 mm diameter represents only 7% of the deposit.

The zone of the elongated depressions is genetically closely linked with the formation of the younger end moraines, with concentration of meltwater in the marginal zone of the glacier, with the erosive action of this meltwater, and with melting of the stagnant bottom part of the glacier after it had retreated farther northward. This first stage of evolution of the elongated depressions might be called the *glaciofluvial stage,* in which the meltwater streams escaping from glacier gates were passing through the moraine gates into the moraine forefield.

Afterwards, when the retreat of the glacier margin and lowering of the glacier surface at the same time, prevented the meltwater from flowing through the moraine gaps, these waters were ponded up to form ice-dammed lakes. This became the second, the ice-blocked *glaciolimnical stage* in the evolution of the elongated depressions, perpetuated by shore lines in the slopes of the depressions and by laminated silt deposits in their floors.

The third stage which will be called the *glaciolimnical-fluvial stage* was associated with the stagnation of the glacier front along a line coinciding with the present-day peripheral river. Next to the new glacier margin new ice-dammed lakes were being formed (Figs. 20 and 21). At the same time a former

Fig. 20. Shore-line of the former ice-dammed lake in the N–S depression. Photo by R. Galon

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Sandgigjukvisl River issued from this ice-dammed lake in an eastward direction, towards the valley the modern river occupies; at that time the river already crossed the main end moraine zone and ran into its forefield. This sort of geomorphological-hydrographical pattern (Fig. 21) is known from a 1946 American map of the region in 1:50,000 scale, compiled from air photographs taken in 1944. The west-east reach of the river extending in the eastern part of the zone of the elongated depressions can be identified only by its lower part of the former river valley which today is used as channel only by perio-

Fig. 21. The former ice-dammed lake and the former Sandgigjukvisl on the area under discussion in 1944 on the base of field investigation and air photographs

Fig. 22. Terrace in the dry valley, the former valley of Sandgigjukvisl. Photo by R. Galon

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dical streams. The former river channel, lined with a large-block pavement, is for the most part covered by the dark debris of slope material or aeolian deposits; only near today's Sandgigjukvisl valley does the older stone-lined river channel appear from under its colluvial-aeolian mantle. Traces of a river terrace are also visible in this now dry valley, diversified by recent kettle holes (Fig. 22).

The final evolutionary stage of this deepest part of the zone of the elongated depressions brought — perhaps due to the ultimate melting of buried dead ice blocks and a renewed deepening of the bottom of this zone — an end to the eastward runoff of water from ice-dammed basins; as a result, the east-west fluvial valley lost its water flow, was gradually destroyed, and assumed the appearance described above. Farther north, another elongated depression running in N–S direction contained at that time, that is later than 1946, this principal ice-dammed lake. For some time the previous principal ice-dammed lake in the zone of the elongated depressions, extending in W–E direction, was connected by a short link valley (Fig. 1) with the new principal ice-dammed lake, relinquishing part of its water in this way. But all that has remained of this former principal lake in the deepest part of the depression is a small lake of varying extent which has now largely disappeared.

THE INNER ZONE OF MARGINAL DEPOSITS AND LANDFORMS

The zone containing the elongated depressions represents not only a separate and characteristic morphological element; it also constitutes the boundary separating the W–E pattern of landforms represented by the main end moraine zone from the predominantly N–S oriented series of relief appearing in the next morphogenetic zone of the glacier forefield. This zone extends as far as the peripheral river and, in fact, ends only with the zone in which glacial and glaciofluvial deposits and landforms at the same time are in the making (Fig. 23 and 24). Our present description refers to the area situated between the elongated depressions and the peripheral river; however, this area is in many ways associated with landforms of the immediate Skeidarárjökull forefield. The N–S oriented series of landforms mentioned before comprises every type of protruding forms such as ridges or wide mounds (Fig. 24), as well as types of concave landforms such as elongated depressions or channel-type or kettle-type basins. Apart from all these morphological features, in the direction towards the glacier can be observed a gradual downslope of these landforms, occurring by means of a number of steps in the land surface (Fig. 30).

Although the inner zone of marginal deposits shows a certain uniformity in the morphological character of its deglaciation, indicating that for the most part the deposits have originated from a stagnant glacier, a number of different types of the relief can be observed. A particular diversity in this respect shows the eastern part of the marginal zone under discussion, extending between the long N–S depression and the Sandgigjukvisl valley (Fig. 1). Here one sees a compact, relatively high moraine surface looking like a hummocky moraine plateau, with stone walls and ramparts on it running in a N–S direction and flanked by shallow depressions. The farthest eastern ridges have recently been incised by lateral erosion of the Sandgigjukvisl River. In the structure of these landforms coarse-grained material predominates containing boulders of various sizes in large quantities. These landforms must be looked upon as forms originating from crevasses.

Worthy of particular attention is the extreme western wall of the moraine plateau, adjoining the wide depression extending in N–S direction. This is

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Fig. 23. Inner zone of marginal landforms. N–S situated ridge and channel. Photo by R. Galon

Fig. 24. The young differentiated relief of the inner marginal zone. Photo by R. Galon

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a huge ridge (Fig. 23) running at right angle to the glacier margin and descending in this direction by numerous steps (Fig. 1). The southern part of this ridge is highest and longest. Its relative height with regard to the adjoining wide N–S depression is some 20 m, and the width of the perfectly flat and rock-strewn surface comes near to 60 m. The slopes of this wall carry two or three irregularly spaced terrace flattenings which can also be seen on the slopes of other walls in the western part of the examined zone. Farther northward this terraced wall lowers to a relative height of 8 to 10 m, and by the next step, to about 5 m height. From here forward the ridge is no over 3 m high, showing a smooth top surface locally as much as 20 m wide which is build by rounded boulders. After a gap passed by high-water floods in the river in exceptional cases — we reach the lowest and narrowest part of the ridge: it is 0.3–1.0 m high, at the most 15 m wide and has a flat smoothed stone surface, with isolated boulders and linear stone strips. This last section of the ridge ends at the bank of the river and is slightly undercut. It seems worth mentioning, that at present an extension of the discussed ridge is being formed in an open glacier crevasse.

Thus the ridge described above is characteristic of the discussed marginal zone; of particular interest is its N–S direction at right angles to the glacier front, its structure of stones of a variety of sizes and its flat surface, the abrupt changes in height and width it suffers several times, and the fact that it continues its path in the marginal zone of the glacier. In short, we are faced here with a powerful crevasse form which in its southern section had to same extent the nature of a median moraine which was melting out of the glacier in stages (note the above mentioned slope terraces!) at the rate of glacier retreat and decay, after its younger end moraines had been piled up. This glacier crevasse gradually decreases, but it still exists as shown by the contemporaneous process of accumulation in the recent marginal zone of Skeidarárjokull.

At the rate of gradually approaching the peripheral river, not only the huge crevasse filling ridge described above but other ridge-type landforms also decrease in size and finally disappear (Fig. 1); at the same time the moraine plateau gradually passes into an undulating plain built of dark sands and silts, covered by a great number of stone strips which sometimes cross each other (Fig. 25). On and off larger rock blocks can be seen, kettle holes and dry channels of former streams which go back to former stages in the existence of the peripheral river. Along a bend of the Sandgigjukvisl a terrace plain can be observed, some 5 m above the water level, covered with a gravel sheet. The undulating plain described above represent a very young area of glacial and glaciofluvial accumulation which extends as far as the end moraines, outwashes and ice-dammed lakes simultaneously developing in the recent marginal zone of the glacier.

Geomorphologically of great importance is the large elongated depression extending in N–S direction which form the west borders on the huge ridge described above. A fairly low sill step dissected by two dry fluvial channels running in opposite directions (Fig. 1) separates this elongated basin from the previously described zone of elongated depressions in the rear of the main end moraine zone. However, here we are facing a landform which genetically is associated with the zone of the elongated depressions. Most probably this was the track by which, at the time that the principal end moraines were being laid down, meltwater streams were escaping underneath the ice towards the basins situated behind the moraines, in some sort of subglacial channel.

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During a later stage of areal deglaciation this depressed area became the place where meltwater accumulated. This period of water storage is clearly indicated by shore lines visible in the slopes at 3 and 5 m above the floor of the depression (Fig. 20) and by the deposits filling the bottom part of this extensive landform. And it is from these deposits that winds carry off the material forming dust storms which lead to aeolian deposition within the wide depressed area.

It seems reasonable to assume, that the meltwater lakes filling the discussed N-S oriented depressions were displaced northward, together with the marginal zone of the glacier which was retreating in this direction, and contemporaneously with the formation of successive new proglacial basins in the forefield of the active glacier front.[3]

The N-S oriented elongated depression is flattening out northward; in the same direction the deposits are becoming increasingly fine-grained and more

Fig. 25. Fluted moraine-like stone strips crossing similar strips or small sandy ridges. Photo by R. Galon

silty, and in westerly direction a few open but gradually diminishing water--filled basins are visible — the remnants of former ice-dammed lakes. The peripheral river is at work incising its channel into this young plain of accumulation.

Westward from the N-S oriented wide depression discussed above extends an area highly diversified in altitude. As feature characteristic of this geomorphological zone — numerous hummocks and ridges, elongated depressions

[3] Description of the proglacial evolution of the fluvial system and of lakes given in the chapter describing hydrographic conditions in the investigated area as reported by Z. Churski.

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and smaller size, but deep kettle holes occur here. Moreover, a group of end moraine forms have been perpetuated here which in this zone are an exceptional feature and reflect the relief forms seen in the main end moraine zone. This inner marginal zone borders immediately upon the discussed chain of elongated depressions north of the moraine gaps Nos. 5 and 6. The picture of an end moraine landscape, resembling that near moraine gap No. 2 in the main end moraine zone, is created by steeply rising ridges of up to 20 m relative height which gradually pass into high hillocks. The ridges and high moraine hillocks which extend in a W–E direction, thus in conformity to the end moraines, tie in with other landforms which run in N–S direction (Fig. 1); these latter forms are either narrow walls or flat stones ridges or, even small plateaus. This is a classical example of the combination of end moraine forms with crevasse filling forms: two different types of deglaciation features. In--between the ridges and stone ridges lie deep basins, round, with shore lines marked at 5 m and 3 m above basin floors — remnants of recent ice-dammed lakes. It is worth-mentioning that all these forms show by their sharp contours that they are of recent origin; not the slightest traces of changes due to denudation are noticeable, and from the steep slopes large boulders can be seen protruding.

In addition to this group of end moraine forms closely associated with the crevasse filling ridges, the area under discussion consists either of wide stone--strewn plateaus or ridges which abruptly end high above the river plain, or

Fig. 26. Cross-section of a stratified or striped silty mound. Inner marginal zone. Photo by R. Galon

of narrow stone walls running toward the glacier. Also encountered here are hillocks and mounds of an irregular shape, some of them built of stratified silty deposits (Fig. 26). The high stone walls show terrace steps in their slopes, resembling those previously described from the eastern part of the geomorphological zone under discussion. However, alongside of all the protruding land-

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forms there are also many elongated basins running parallel to ridges, or irregularly scattered depressions. In their slopes one distinctly sees former shore lines. The reference level for both elevations and depressions is the undulate or slightly hilly plain built of ablation material. Here and there larger rock blocks occur, or hillocks, or stone stripes. Northward this plain gradually subsides ending in a sharp step, 1 m high, above the valley of the widely spread network of the peripheral river which gathers all outwash waters. In its eastern part this outwash plain passes into an area of recently formed rather extensive ice-dammed lakes which are now gradually vanishing. And in their vicinity flows the much braided peripheral river.

Surmounting the peripheral river a few erosively incised hillocks and crevasse ridges break off, together with the silt-coated ablation plain. Particularly worthy of attention here is an esker ridge of a typical shape, which continues on the northern bank of the river and extends farther as far as the vicinity of the present glacier margin (Fig. 27). This esker runs in a sinuous course, its

Fig. 27. Esker ridge. Photo by R. Galon

stone surface with a very sharp crest line of recent origin is also sinuous. The relative height of this esker varies from 3 to 8 m, its slope inclination is about 30°. At some points minor stone ridges branch off from the main esker ridge. The material building the esker is composed of a wide variety of rock fragments, starting from large boulders of some 30 cm diameter, over predominantly smaller boulders of 5 to 10 cm size, to gravel and coarse-grained poorly rounded sand. On the other hand, the large boulders are well rounded. The cross-section of the esker (Fig. 28) shows that the largest boulders predominate on the surface while the inside is built of stratified coarse-grained sands of alternating light- and dark-coloured material crossing each other (a channel facies), and that sporadically gravel lenses also occur. Toward the esker margins

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the sand and gravel strata abruptly end, contacting coarse gravel beds and boulders. Undoubtedly the esker described here is a crevasse structure or it originated, as has recently been ascertained on the glacier, in a meandering wide valley incised into the glacier surface. Everywhere on the surface, the esker shows coarse material, while lower down there are streaks of dark sands and silts with locally intercalated thin layers of light-coloured clay. Hence, this is an ablation moraine combined with local ice-dammed lake deposits.

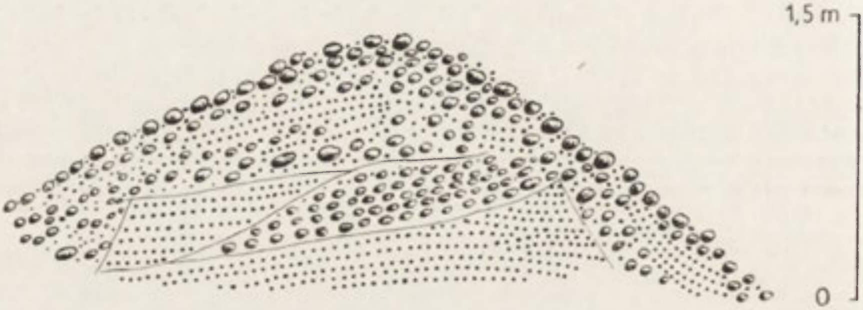

Fig. 28. Cross-section of the esker. Photo by R. Galon

Summing up our comment on the inner zone of marginal deposits and landforms as a whole, the following facts should be emphasized:

(1) The entire area shows distinctly a northward lowering. At the same time all hillocks and ridges of the end moraine type and of crevasse filling origin gradually pass in northward direction into an undulating plain built of ablation and ice-dammed lake deposits. The few crevasse filling ridges also occurring on the silt-covered undulating lain continue their course beyond the peripheral river, reaching as far as the present-day marginal zone of the glacier. This shows, that in the past 50 years a distinct change in deglaciation has set in, from frontal deglaciation (indicated by end moraines in the western part of the geomorphological zone under discussion) by way of marginal deglaciation (crevasse filling walls) to deglaciation by ablation (a silt-covered plain).

(2) The larger walls running in N–S direction show a northward tendency to grow narrower and to form sharply stepped crests — further proof of the diminishing of crevasse accumulation in a northward direction and of some sort of deglaciation by stages.

(3) The discussed marginal zone contains a great number of traces of ice--dammed lakes near the glacier margin, which in their slopes show former shore marks and tracks of former northward water runoff, in accordance with the gradual retreat of the glacier front. A number of dry valleys have also survived in which meltwater streams used to run and to feed in the nineteen forties, first, the extensive ice-dammed lake situated in the area of the elongated depressions and, next, the Sandgigjukvisl River escaping at that time from this lake (Fig. 21).

(4) At many points of this zone, in the hillocks and ridges partly running parallel with the direction of the main end moraine belt one observes large transverse incisions resembling moraine gaps leading from one basin into the next, situated farther south (Fig. 1). At these points glacier gates may once have existed through which the waters, accumulated in subglacial basins under hydrostatic pressure, were finding their way into proglacial lakes.

(5) There seems to be no doubt that both the abrupt steps in the crevasse

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filling hummocks and ridges described above, and the gaps in the moraine ridges indicate stages in which during deglaciation the glacier front has been stagnant.

(6) Many places can be seen, where deep and sharply marked kettles are either being formed at present or have developed recently; they are evidence of dead ice blocks which had been buried in the glacier substratum. They represent the last manifestation of glaciation in the area discussed.

THE COURSE OF DEGLACIATION ON THE SKEIDARÁRJÖKULL FOREFIELD: A TENTATIVE IMAGE

From the descriptions given so far it can be seen, that the retreat of the glacier from the line of the main end moraine zone across the zone of the inner marginal deposits and landforms to its present-day position has been proceeding in a diversified manner. A rough sketch shown in Fig. 29 shows the presumed path of some lines delimiting the retreating glacier margin in this area. Within the main end moraine zone on this map are indicated the chains of the more important end moraine ridges including the moraine gates incising these ridges. Also taken into account were some of the steps occurring in the slopes of the elongated depressions described above and considered as former ice-contact walls. The oldest lines marking the glacier front within this zone have been reconstructed on the basis of remnants of end moraines, today merely piles of boulders, and of the southernmost moraine plateaus, together with

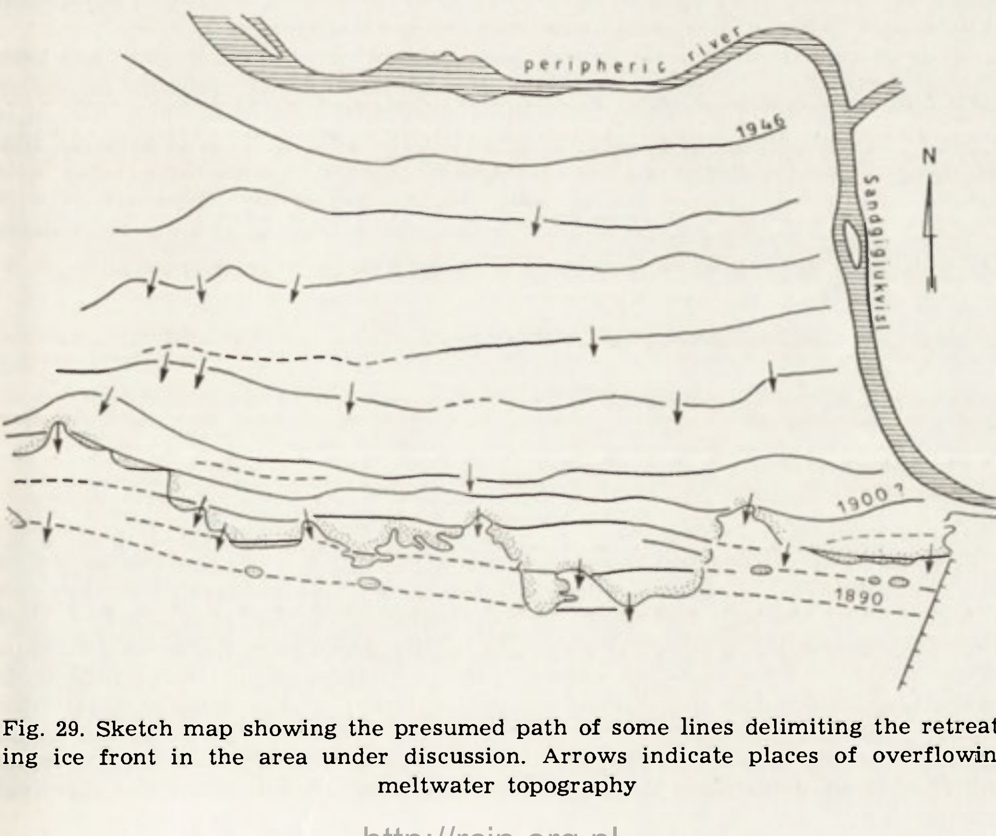

Fig. 29. Sketch map showing the presumed path of some lines delimiting the retreating ice front in the area under discussion. Arrows indicate places of overflowing meltwater topography

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their gaps. It appears, that the moraine gaps seen in the main end moraine zone belong to five separate successive lines of retreating glacier front (Fig. 29); and this is at the same time evidence of the number of more definite standstill stages of the glacier in this region.

Within the inner zone of marginal deposits and landforms has been drawn the first line of the extreme glacier margin, which runs along the marginal landforms bordering from the north upon the zone of elongated depressions, — the more so since here extend the sharply marked end moraine ridges which were mentioned. The next, that is, later lines of the ice margin have been reconstructed on the basis of steps visible in the crevasse filling forms, among them in the high N–S ridge, and on the basis of small moraine gaps seen in ridges partly running parallel with the main end moraine zone. It seems worth mentioning, that in the area where crevasse filling walls and kame forms predominate, deglaciation was not of an areal nature involving stagnation and decay of the entire glacier within the area under investigation. This assumption of areal deglaciation is belied by ridges running in W–E direction, i.e., end moraine ridges which join the crevasse filling forms or form connections between them (Fig. 1). Hence, in the light of the facts mentioned and the arguments put forward there is no doubt that here gradual deglaciation has been taking place, with the decay of the *extreme marginal zone* of the glacier and the subsequent formation of successive crevasse filling forms. It appears that some very long crevasse filling ridges have been developing during several successive standstill stages. This, on the one hand, is evidence of the longevity of a crevasse; on the other, the repeatedly seen abrupt changes in height and width of these large ridges indicate that after successive standstill stages the dimensions of the glacier must have been subject to changes.

It is more difficult to reconstruct the line in which the ice margin has been extending in the area of the undulate silty plain which is virtually devoid of stepped crevasse filling forms. Here the author made use of the map drawn in 1946, which indicates the glacier margin as it ran at that time. It appears, that at two points near the river the margin line shown on this map tallies with crest steps in the crevasse filling walls. Hence, our method of using steps in crevasse filling forms for verifying successive positions of the glacier margin has proved useful.

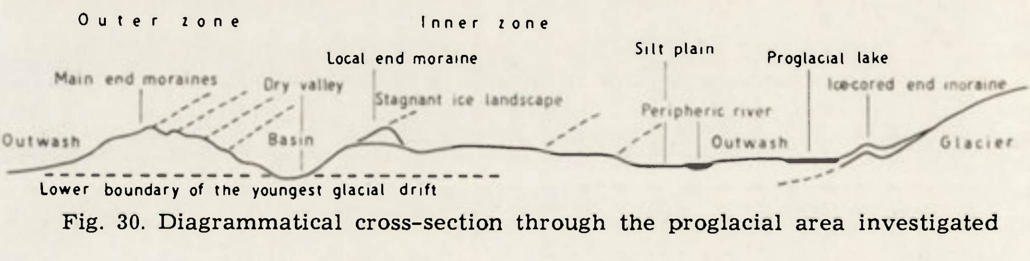

Fig. 30. Diagrammatical cross-section through the proglacial area investigated

Let us now reflect upon the transverse, that is, the N–S cross-section across the discussed area of glacial and glaciofluvial accumulation, this section line running along the gap valley of the Sandgigjukvisl River; this deeply incised channel unfolds the geological structure of the deposits and their substratum (Fig. 30). Our picture of this cross-section is diagrammatic and synthetic in character, illustrating the mutual altitude relation of the morphogenetic discussed zones and their most essential features. Generally speaking, the entire area embracing the main end moraine zone in its full arcuate extent bears the feature of a terminal depression, as is encountered both in the forefield

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of Alpine glaciers and in the area former covered by the Scandinavian glaciations. The hinterland of the main end moraines in the rear of the zone of the elongated depressions lies distinctly at a lower altitude than the end moraines, and it gradually slopes northward, reaching its lowest point in the valley of the peripheral river into which the outwash streams drain water, or in the bottom of the marginal ice-dammed lakes which drain their water into the peripheral river. To some degree the valley of the peripheral river reminds one of *pradolinas* in the Polish and German Lowland. However, under conditions prevailing in Iceland, the life of this sort of landforms is of a short duration.

The morphological and hypsometric contrasts are greatest between the ridges of the younger end moraines and the bottom of the elongated depressions which extend in the rear of the main end moraine zone. The altitude difference is here as much as 46 m. Above the outwash plain the end moraines rise only to the height of 30 m; but in opposite direction, i.e., northwards, greater altitude differences occur between the bottom of the elongated depressions and the crests of the end moraines which border on the northern rim of the region of elongated depressions (Fig. 1). Our cross-section cannot reveal the considerable differences in height exceeding 20 m which exist between the mentioned huge N–S situated ridge and the elongated depression running alongside in an identical direction; nor does it show the difference in altitude between an other big crevasse filling ridge and the plain surrounding it in the northwestern part of the investigated area. Equally invisible in our profile (Fig. 30) is the greatest of all absolute altitude differences occurring in the region under discussion: between the end moraine crests and the floor of the Sandgigjukvisl valley dissecting these moraines. This difference is about 50 m.

Our cross-section (Fig. 30) shows diagrammatically the lower boundary of the series of glacial and glaciofluvial deposits which toward the end of the past century were laid down during a re-advance of the glacier. This series

Fig. 31. Western valley slope of Sandgigjukvisl showing the superimposition of the main end moraine zone on an older sediment series (see Fig. 30). Photo by R. Galon

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comprises all the morphogenetic zones described before, consisting of (a) the main end moraine zone, including the outwash plains in its forefield and the system of elongated depressions in its rear, (b) the inner zone of end moraines, crevasse filling walls and hillocks and large kettle holes, and (c) the area containing the recent marginal landforms. Underlying the above mentioned lower boundary are older, stratified Quaternary deposits which, south of the extent of the glacial and fluvioglacial series, form the plain of degradation and aggradation. This boundary can be observed in both scarps of the Sandgigjuk-

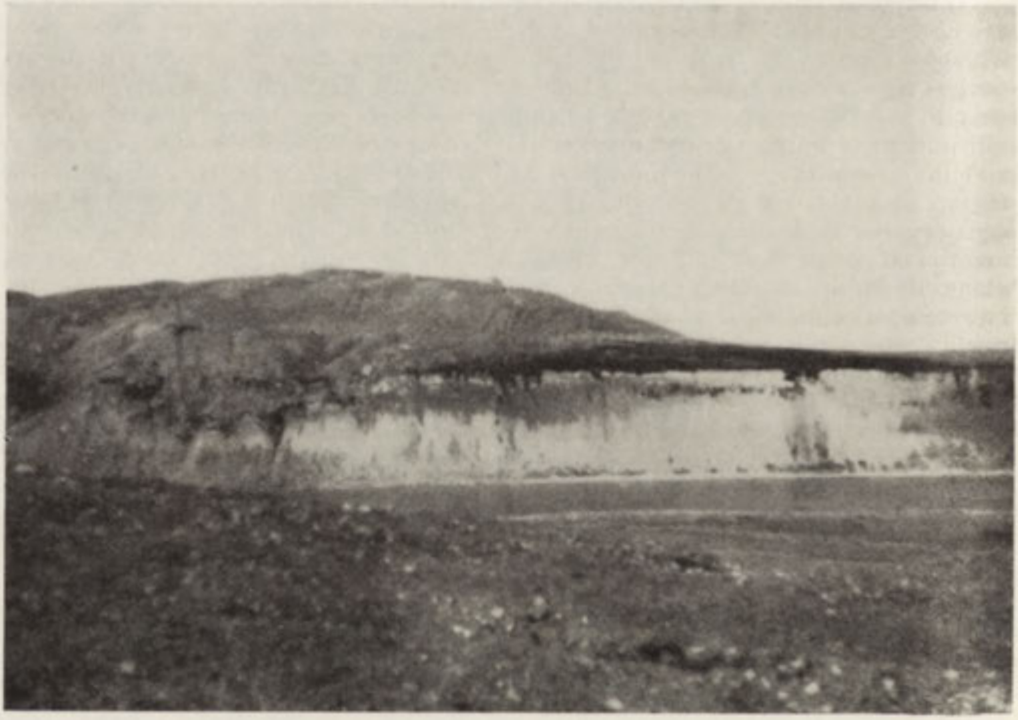

Fig. 32. Eastern valley slope of Sandgigjukvisl showing the superimposition of the main end moraine zone with small outwash fan on an older sediment series (see Fig. 30 and 31). Photo by R. Galon

visl gap valley (Figs. 31 and 32). Similarly Okko (1956, Fig. 21) has established that "the bank of the River Hanypa at Heinabergsjökull has been cut through an end moraine and underlying outwash plain. The end moraine ridge is formed on undisturbed Sandur deposits". It appears that at an altitute approximately coinciding with the level of this lower boundary line of glacier oscillation lenses and blocks of dead ice occur. When these melted a number of brooks were started which in turn produced short lateral valleys with steep gradients (Fig. 33). As was said before, melting of fossil dead ice blocks is today taking place all over the zone of the inner marginal deposits and landforms, as indicated by numerous small but deep kettle holes of recent origin. There is little doubt that these dead ice blocks represent the initial base of the glacier of the Sandgigjukvisl valley.

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Fig. 33. Dead- ice block occurring in the moraine cover and creating a small lateral valley. Photo by R. Galon

The Sandgigjukvisl is the central proglacial river which carries off from Skeidarárjökull the meltwater arriving from peripheral rivers and proglacial lakes. But in contrast to the peripheral rivers which flow at varying distance along the glacier margin, the Sandgigjukvisl River runs southward, away from the glacier; it cuts with its gap-like valley all these geomorphological zones, and on its course to the sea it passes on into the extensive plain of degradation and aggradation (Fig. 34). The present paper will deal with geomorphological development of this valley in connection with the course of deglaciation and in consideration of its relation to some neighbouring valleys which today are dry [4] (see Geomorphological Map at the end of the volume).

The Sandgigjukvisl River runs at a rapid rate, with considerable erosive power. Compared with hydrographical and geomorphological conditions as they were in 1965 and as is shown on available aerial photographs, considerable changes have meanwhile set in. In the gap section a new island has been formed. Lateral erosion at many places has led to the undercutting and steepening of the valley scarps, both in the gap section and in the moraine forefield. Undoubtedly the intensive erosive action of the river has been furthered by

[4] In our research work the Sandgigjukvisl River was treated as the line separating the areas assigned to M. Bogacki and to R. Galon, respectively. All the data on the terraces observed in the eastern slope of this valley were supplied by M. Bogacki.

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occasional *jökullhlaup* phenomena the last of which took place in 1954 (Thorarinsson 1954).

It seems certain that the Sandgigjukvisl River originated from a break in the and moraine zone and from a meltwater valley issuing from this break. Traces of this break have survived in the western part of the valley. The northward extension of the valley associated with the retreat of the glacier proceeded step-wise. During one of the evolutionary stages of the valley the river had its source in an extensive meltwater lake occupying a larger part of the zone of the elongated depressions, and ran in the eastern section of these de-

Fig. 34. Valley of Sandgigjukvisl in front of the main end moraine zone. In the background the main end moraine ridge. Photo by R. Galon

pression (Fig. 21). Gradually, by several stages, the younger northern part of the Sandgigjukvisl valley came into existence, until the present-day hydrographic and geomorphological layout had developed.

Evidence of the stage development of the Sandgigjukvisl valley are the valley terraces which follow both banks in the gap section as well as in the outwash plain (Figs. 35 and 36). The following erosive valley terraces were identified and their relative altitudes above the water level are also given:

Terrace	I	1 m
Terrace	II	2– 3 m
Terrace	III	4– 5 m
Terrace	IV	7– 9 m
Terrace	V	10–12 m
Terrace	VI	13–15 m

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Even higher above the water level lies the plain of outwash fans associated with the main end moraine zone. It will be remembered, that the surface of the outwash fans show a considerable inclination, and therefore their relative height above the water level are first as much as 20 m but rapidly decrease to about 10 m. Both the relative height and the number of the river terraces tally with the terraces observed on the shores of the ice-dammed lakes from which the river originates.

Valley terraces rarely occur near each other. Their greatest concentration can be observed at the outlet of the now dry valley mentioned above (see

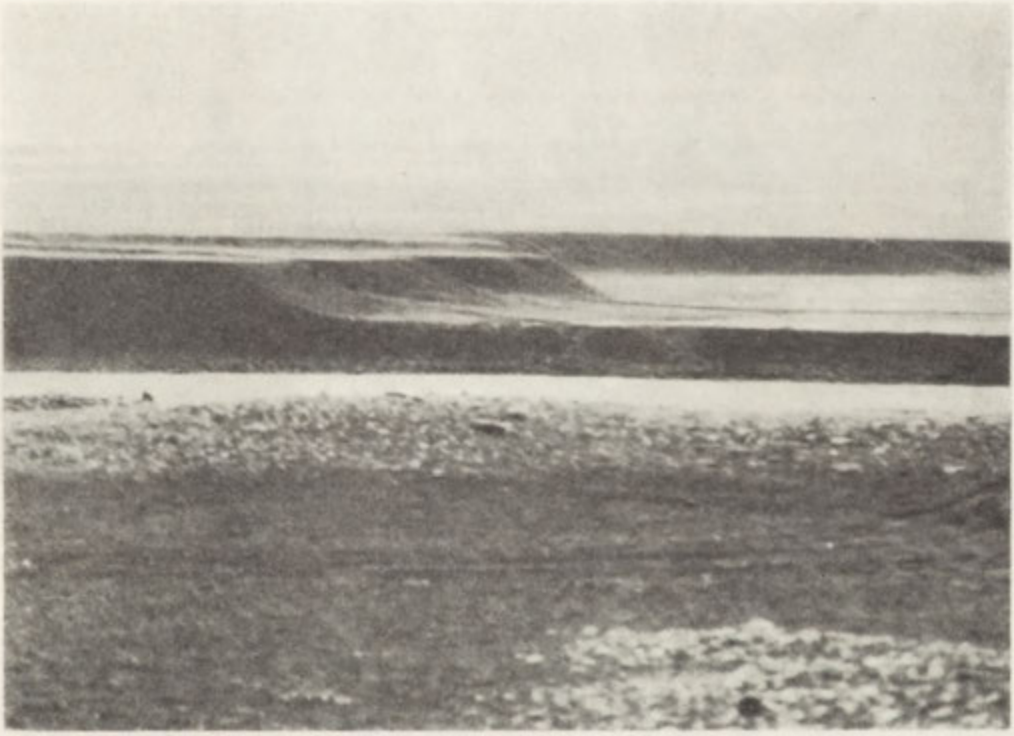

Fig. 35. Terrace of the Sandgigjukvisl valley on the area of plain of degradation and aggradation S of the main end moraine zone. Photo by R. Galon

Fig. 17), and this fact facilitates the study of the relation of this older valley to the present Sandgigjukvisl valley running in the N–S direction — a topic to be discussed later. In the lower valley reaches, crossing the area covered by the degradation–aggradation plain, the number of terraces decreases. Their relative heights also become lower and, in this respect, the Sandgigjukvisl valley gradually comes near to the appearance of its neighbour — the dry walley mentioned before. Near the outflow of the river into the oceanic lagoon the valley is incised to a depth of about 1.5 m. The lower terraces and the valley floor within the gap section of the valley, where it crosses the end moraine zone, are covered by great quantities of well rounded boulders of widely different size, meltwater kettle holes also occur here. Nearer the mouth of the river (where accumulation predominates) the material becomes more fine-grained, and even silts are being deposited.

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The origin of the Sandgigjukvisl valley representing the central proglacial river issuing from the Skeidarárjökull deserves special attention. The evolutionary stages of this valley and their periods of duration are at the same time successive stages of the course of deglaciation. However, any reflections upon the Sandgigjukvisl River and its valley must be paralleled by a consideration of the peripheral rivers and their valleys. The reason is that apart from the present-day system of peripheral valleys and ice-dammed depressions seen near the glacier margin (which are all closely connected with the Sandgigjukvisl valley) there exists, within the range of the glacial deposits and landforms

Fig. 36. Delta of Sandgigjukvisl. Photo by R. Galon

discussed above, only one older system of peripheral valleys and ice-dammed lakes which can be associated with the southward running Sandgigjukvisl valley. This older system was situated within the zone of elongated depressions in the rear of the main end moraines. Amittedly, some minor peripheral runoff tracks can be seen in the eastern part of the inner zone of marginal deposits and landforms, but they lack the character of a full valley- and - basins system which could have served as a network of peripheral drainage from a large part of glacier front.

Thus we arrive at the conclusion that only twice has the area under discussion passed through glacial conditions under which a peripheral system of concentrated water accumulation and of runoff of Skeidarárjökull meltwater might have been operating in the rear of the main end moraine zone. However, the glacier front did not have to be in a stable position during these two successive systems.

The first of these systems of drainage, situated in the immediate rear of the main end zone, existed originally as a system of proglacial ice-dammed lakes, at the time when the glacier margin was laying down the farthest southward moraine deposits of the inner zone of marginal landforms. As indicated by the 1944 aerial photographs, this system was still in operation when the glacier margin had retreated almost to the line of the present-day peripheral river (Fig. 21). While the first system of peripheral runoff lasted, the meltwa-

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ter break through the main end moraine zone had become an accomplished fact. Above the eastern rim of the gap valley, a fragment of the stone-lined channel breaks off at 17 m above the water level. And this is the highest meltwater track which has been observed in the gap valley of Sandgigjukvisl, and its counterpart south of the gap is the highest erosive step on the outwash sheet (15 m high). Within the zone of the elongated depressions, water runoff through a now dry valley took place at the level of the 11 m and 9 m terraces (called V and IV) above the level of today's Sandgigjukvisl water. Near the place where the now dry valley joins the Sandgigjukvisl gap valley, there protrudes, from under the mantle of slope debris and aeolian deposits (at the 11 m level) a fragment of the stone-lined channel of the former river channel. In the eastern bank of the gap valley an undercut oxbow basin has survived, eroded by the former river before it turned southward through the main end moraine zone into the coastal plain.

The second, that is, today's system of peripheral valleys in the nearest glacier forefield and in the rear of the crevasse filling walls is already shown in the aerial photographs taken in 1960. This means, that within the relatively short period of 15 years this fundamental change in hydrographic and geomorphological conditions has been taking place. Undoubtedly, today's silt-covered plain was first studded with proglacial lakes which retained the meltwater streams. The streams had been flowing southward into the old system of peripheral valleys and depressions. The western group of meltwater streams was carried off by the Blautakvisl, a tributary of the lateral Sula River. Gradually the evolution of a new peripheral river became possible, after the meltwater had broken through the northward part of the inner zone of marginal deposits and landforms, and after the Sandgigjukvisl walley had been formed in approximately the shape it has to this day.

Unfortunately, this basic stage in the evolution of the Sandgigjukvisl valley is least known and explained. It seems probable, that the wide gap was effected as a result of the pressure of accumulated ice-dammed masses of water, and that the new flow track was proceeded by a group of kettle holes. Nor can one exclude the possible effect of the periodical glaciofluvial floods caused by what are called *jokullhlaup* occurrences. To this evolutionary stage of the Sandgigjukvisl valley must be assigned terraces III, II and I which can be observed in the gap valley and south of it (Fig. 35) where terrace III, incised directly into the plain is widest in extent. Once more it should be stressed that the melting of dead ice blocks played a morphological role as an agency furthering the formation of the gap valley of Sandgigjukvisl.

A few remarks will be added relating to the relation between the Sandgigjukvisl valley and the dry meltwater valley running west of, and parallel with the Sandgigjukvisl. This valley has been described in the chapter dealing with the forefield of the main end moraine zone. Let us briefly consider the part this now dry valley has been playing with regard to the meltwater streams issuing from under the glacier while it was piling up its principal end moraines, and to these waters afterwards when they were accumulating in the form of ice-dammed lakes in the wide background of the main end moraines and, afterwards, in the rear of inner zone of marginal landforms. This study shows that water runoff in the two valleys took place at different periods, never simultaneously. First in operation was the western valley, now dry; it once was the principal water escape channel, though not the only one. Later on, and up to the present time, the Sandgigjukvisl valley took its place as the only meltwater track between the Sula and the Skeidara Rivers. The period

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in which one valley was in operation, was separated from the time the other valley replaced it and started operating, by a stagnation of the glacier, during which the meltwater was accumulating in the rear of the main end moraines, and part of it was escaping in lateral rivers and, east of Sandgigjukvisl, in deeply incised tracks of outwash channels. Water flow in the now dry valley was decreasing gradually; peasants of the Nupstadur village report that this flow ended some time about 1940. However, an aerial photograph taken in 1944 shows the Sandgigjukvisl river in full operation, draining a large ice--dammed water basin in the zone of the large elongated depressions. This seems to prove, that this important hydrographic change must have taken place during the rather short period from 1938 to 1944.

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REFERENCES

Ahlmann, H. W: son, Thorarinsson, S., 1943, *Vatnajokull. Scientific results of the Swedish-Icelandic Investigations 1936–37–38*, Chapter I — XI. Reprinted from Geografiska Annaler 1937–1940, 1943, Stockholm.

Eythorson, J., 1949. Variations of glaciers in Iceland 1930–1947, *J. Glaciol.*, 1.

Hjulstrom, F., 1952, The geomorphology of the alluvial outwash plain (Sandur) of Iceland and the mechanics of braided rivers, in: *Proc. XVIIth Congress IGU Washington.*

Okko, V., 1956, Glacial drift in Iceland, its origin and morphology, *Acta Geogr.*, Helsingf., 15, 1.

Price, R. J., 1969 *Moraines, sandar, kames and eskers near Breidamerkurjokull, Iceland*, Trans. Inst. Brit. Geogr., 46.

Price, R. J., 1970, Moraines at Fjallsjokull, *Arct. Alp. Res.*, 2, 1.

Price, R. J., Howarth, P. J., 1970, The evolution of the drainage system (1904–1965) in front of Breidamerkurjokull, Iceland, *Jokull*, 20.

Thorarinsson, S., 1954, Athuganir a Skeidararhlaupi of Grimsvotn 1954, *Jokull*, 4.

Thorarinsson, S., 1964, Sudden advance of Vatnajokull outlet glaciers 1930–1964, *Jokull*, 14.

Thorarinsson, S., 1967 Washboard moraines in front of Skeidararjokull (Icelandish with English Sum.), *Jokull*, 17.

Todtmann, E. M., 1960, Gletscherforschungen auf Island (Vatnajokull), *Univ. Hamburg, Abh. Auslandsk.*, Reihe C, 65, 19.

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GEOGRAPHIA POLONICA 26, 1973

GEOMORPHOLOGICAL AND GEOLOGICAL ANALYSIS OF THE PROGLACIAL AREA OF SKEIDARÁRJÖKULL. CENTRAL WESTERN AND EASTERN SECTIONS

MIROSŁAW BOGACKI

(1) CENTRAL WESTERN SECTION OF THE PROGLACIAL AREA

(A) MAIN END MORAINE ZONE

Towards the end of the 19th century, about 1890, a transgression of the Skeidarárjökull took place. The end moraines dating from this glacier advance in certain parts form a definite zone rising to 115–120 m a.s.l. Northward, towards the glacier, the land surface gradually subsides.

Up to 1968, the glacier front had again retreated and this retreat was most considerable in the western part; from the 1890 moraines the distance of the glacier margin was 3.6–3.8 km, while in the eastern part this distance was only 2.2 to 2.4 km.

The width of the end moraine zone varies: it is greatest where the innermost old moraine ridge has survived, and at these places it is in excess of 100 m. Where the old moraines have been destroyed, the width of the zone is smaller.

Very similar to the central part of the Skeidarárjökull forefield, in the western area the following zones can be distinguished: (1) a zone of outwash fans and meltwater valleys, with old eroded end moraines in the farther forefield, (2) a main end moraine zone built of older and younger moraines, (3) a zone of elongated depressions, and (4) an inner zone of marginal deposits and land forms.

The main end moraine zone extends in a western direction in a continuous belt some 1 km long, starting from the eastern boundary of the area under discussion. Further on, westward, this zone has almost completely been eroded by glaciofluvial streams (Fig. 1).

The older moraines

In the western part of Skeidarársandur, the older end moraines have been preserved only for a distance of some 0.5 km between moraine gate No. 1 in the east and the wide gap in the west. Traces remaining of these older moraines are a few piles of boulders.

The surviving older end moraines are land forms moderately high. Their relative height looked at from the outwash fans is up to 5 m; the upper part of the slope shows a gradient of up to 20°. In contrast to the central part of Skeidarársandur, in this western part the older end moraines are separated from the younger moraines by a distinct elongated depression now lined by

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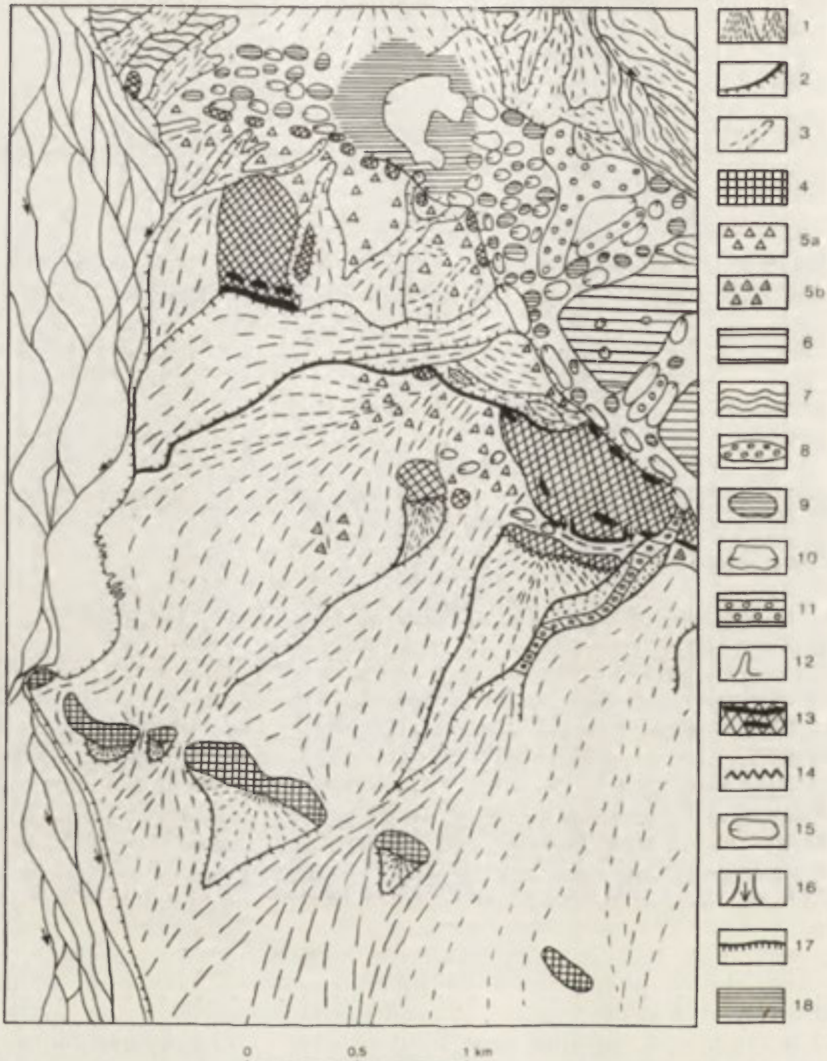

Fig. 1. Geomorphological map of the area east of the Sula River.

Forefield of main end moraine zone

1 — Outwash fans: spread of lines indicates different runoff directions on the plain of degradation and aggradation, 2 — Scarps of outwash levels and of terraces in valleys, 3 — Dry valleys, 4 — Old end moraines, 5a — Stone remnants of end moraines connected with main end moraine zone, 5b — Stone remnants of end moraines connected with older moraines

Rear area of main end moraine zone

6 — Morainic plain, with flat or undulating surface, 7 — Low stone hillocks or undulating plateaus, 8 — Large-size flat gravel-boulder ridges (crevasse forms), 9 — Isolated mounds built of silty material, 10 — Ice-dammed depressions or meltwater kettles, 11 — Pavement-type floors of former meltwater streams, 12 — Lateral valleys

Main end moraine zone

13 — High morainic hillocks and walls (more prominent ridges) from rise of 20th century, 14 — Old morainic walls from about 1890, 15 — Larger meltwater basins and elongated intermorainic depressions, 16 — Moraine-outwash gates in end moraine chains, 17 — Sharp scarp steps within main end moraine zone (ice-contact walls)

Youngest land forms presently forming

18 — Ice-dammed plains

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aeolian deposits. The difference in height between the bottom of this depression and the moraine crest reaches as much as 20 m, and the slant of the northern slope is often some 25°. In this western part of Skeidarársandur the older end moraines have been flushed away by the waters issuing from former glacier gates which now are represented by moraine gates. The moraine gates start out from the chain of the younger end moraines. Denudation has here been so powerful, that in this western part it is difficult today to trace the course of the end moraines; often not even old accumulations of boulders have survived (Fig. 2).

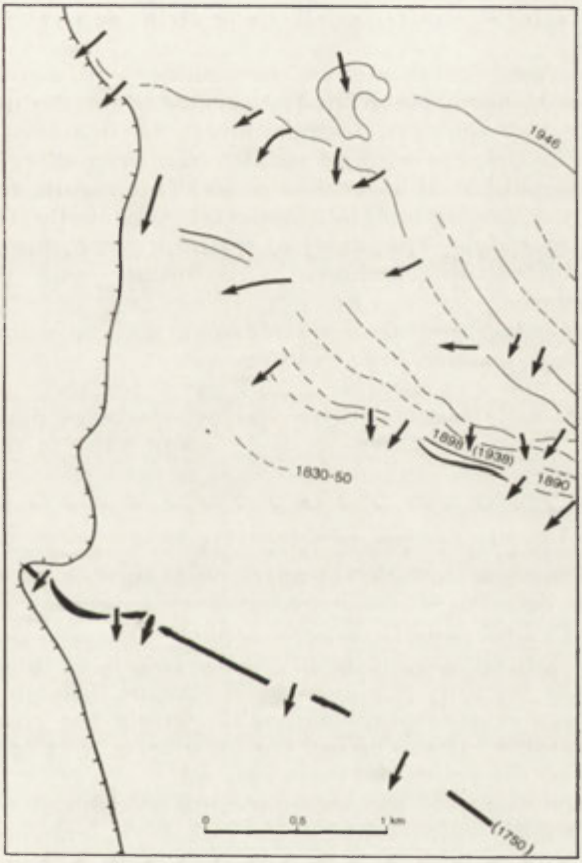

Fig. 2. Stages of glacier standstill in the area east of the Sula River from the half of the 18th century
→ run-off directions

The older end moraines are built of unequigranular sands with boulders; for the most part the boulders are embedded in the surface layers.

The younger end moraines

In the western part of Skeidarársandur the younger end moraines form a sharply marked zone. As mentioned before, they are separated from the older end moraines by a marginal depression. The younger end moraines extend from the eastern boundary of the area discussed westward for a dis-

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tance of 1 to 1.5 km. Farther on, towards the Sula River, they have left only traces in the shape of smaller or larger piles of stones.

As to relief forms the younger end moraines vary, showing sharply defined angular contours in cross-section. Here can be seen characteristic moraine walls up to 5 m high, with slope inclinactions upwards of 30°, and numerous hillocks and morainic mounds also. A further relief form diversifying zone is the great number of hollows and kettles, both older ones and those actually being formed, originating from dead ice melting.

The zone of the younger end moraines bears the features of a highly elevated, compact morainic plateau. But this plate plateau carries a variety of usually small land forms appearing as stone strips separated by furrow-like depressions.

The land forms seen in the zone of the younger end moraines are built of an extremely heterogeneous material. The surface is for the most part covered by coarse gravels and boulders; farther down the material becomes more fine-grained such as fine gravels and sands, often even silts. Frequently these deposits appear in regular strata or in streaks. This results from the way the accumulation of the muddy ablation material, repeatedly flushed down by water, has been proceeding. This mode of washing down the material is documented by fluidal structures inclined in conformity with the general subsidence of the ground.

Northward, towards the elongated depression, the end moraines descend by sharply marked steps, usually three in succession.

The deposits of the end moraines, both the older and the younger, have been laid down on an ancient, greatly destroyed erosive plain, as seen from the occurrence of a distinct pavement layer going back to the denudation of some older deposits.

The moraine gates

The main end moraine zone is cut apart by a number of valleys of different sizes. In most cases these valleys correspond to glacier gates and this is why the term "moraine gate" has been applied to them. Differences in the shape and width of these gates depend on the length of time they were in operation and on the intensity of glacial water flowing through them.

Some of these gates are barely noticeable within the chains of the older or younger end moraines. Other dissect one or several rows of morainic ridges, some even cut across the entire end moraine zone.

Moving from the east over the area under discussion one arrives at a wide and deep moraine gate which dissects the whole extent of the main end moraine zone. Southward this gate passes into a melt-water valley with terraces. The waters escaping from this gate flushed away the older end moraines situated east of this gate. Traces of this destruction are piles of big boulders.

Continuing in a westerly direction one reaches a number of smaller gates which cause gaps only in the younger end moraines. The waters arriving through these gates ran into the elongated depression which was separating the older from the younger end moraines (Fig. 1).

The western section of the end moraines has been completely flushed away. Here one meets two enormous gates or, rather, wide breaches in the moraine chain: an eastern break some 300–400 m wide and a western break with a width of about 1.5 km. These breaks start out from the youngest moraines of the main end moraine zone. A very powerful water volume must have been flowing here for a long time, probably throughout the period in which the

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main end moraine zone was formed. This seems to be indicated by well developed terraces, five of which have been observed near the end moraines.

Farther to the west of the two discussed large breaks one arrives at the area in which the River Sula maintains its vigorous destructive activity.

(B) THE FOREFIELD OF THE MAIN END MORAINE ZONE

In the western section of Skeidarársandur, which is very similar to the central part, the forefield of the main end moraines consists of two parts of different origin: a narrow zone of outwash fans directly adjoining the forefield of the moraines, and a gently sloping and widely spread plain of degradation and aggradation into which these fans coalesce. The material forming this plain constitutes the base of the main end moraines zone and the outwash fans deposited on the plain (Fig. 3).

Fig. 3. Relation of end moraines to outwash fans

The plain of degradation and aggradation and the old end moraines

It should be stressed that the material forming the plain of degradation and aggradation was not laid down at the same time as the main zone of end moraines developed, but it is older and probably should be assigned to the time when the Pleistocene glacier was retreating during the Holocene. The maximum extent of this glacier is marked in several places by old moraines. This seems to prove that the glacier retreat has been interrupted by halting stages and, perhaps, by minor transgressing readvances.

The old moraines lie some 1.5 km south of the main end zone. They are called Sandgigur and rise to 83 m a.s.l. They occupy solely the western part of Skeidarársandur, between the Sandgigjukvisl River in the east and the Sula River in the west. In the much larger area east of Sandgigjukvisl no old moraines have survived at all; they must have been flushed away by glaciofluvial streams. *Jökullhlaup* waters might also have contributed to their destruction, especially those of the 1922 *jökullhlaup* mentioned by S. Thorarinsson (1943).

The old moraines appear in the form of six separate, sharply defined mounds (Fig. 1). The mound farthest west is undercut and heavily attacked by the Sula waters arriving from the glacier; the mound situated farthest east, on the other hand, lies at a distance of 1800–1900 m from the Sandgigjukvisl River. In between all end moraine mounds, gaps called moraine gates have been eroded, and the glacial waters used to escape by these gates. The width of these gates varies, increasing from west to east, and in this same direction the mounds gradually decrease in height and become more eroded. Thus, from west to east, gates Nos. 1 and 2 are from 20 to 80 m wide; the width of No. 3 is 80 m, that of No. 4 about 240 m and of gate No. 5 some 600 m.

These old moraines are by no means large land forms; still, due to their inselberg character they are remarkable features, protruding from the Skeidarársandur area. For the most part their relative height is about 10 m; their

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slope inclination varies, the northern and southern slopes being more gentle than those towards the west and east. Moreover, all slopes are steeper in their upper than in their lower section (Fig. 4); this is due to the intensive flushing down of material from the upper slope sections and from its deposition along the slope bases. In addition, the lower slope sections are frequently overlain by aeolian material which comes to rest here in the lee of land forms (Fig. 5).

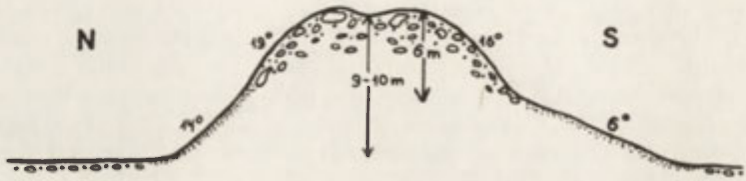

Fig. 4. N–S section across old end moraine

Fig. 5. Aeolian deposits mantling lower slope parts of end moraine slopes. Photo by M. Bogacki

The distal slopes are steeper than the proximal, and for the lower slope section, up to about 1/3 of their hight, their angle of inclination is usually 14–15°, while on the proximal slope it is 6–7°. Higher up on the slope, the distal angle is often 19–20°, and the angle of the proximal slope some 16–17°. In their western and eastern sides, that is, on their slopes towards the gates, the mound slopes are undercut by glaciofluvial streams and this causes the slopes here to become much steeper, up to 24–26°.

Usually the top surfaces of the moraine mounds are evenly flat, strewn with large boulders and pebbles. Presumably these moraines have been eroded and partly washed away by meltwater escaping from the glacier front.

The geological structure of the old moraines is highly diversified. A lot of boulder clay material is in evidence in the form of many big boulders and

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large rock blocks, mostly of palagonite, of up to 2 m diameter (Fig. 6). But a dominant constituent in the structure of these moraines are also fine and very fine deposits, volcanic dusts, or material derived from disintegration of volcanic rocks or palagonite. Very often this fine-grained moraine material is very firmly cemented. The silts and fine-grained sands are stratified; at times they lie horizontally but in most cases they are disturbed from their original position (Fig. 7), and then their beds dip at different angles and in different directions. Entire series of these deposits are thus dislocated with regard to each other. This subsidence of the beds is the result of the melting of dead ice lobes buried in the morainic deposits.

Defining the age of the old end moraines seen in the Skeidarárjökull forefield is rather difficult. From studies of end moraines and lakes, from direct field measurements made within the past 50 years or so, and from historical records it has been determined that during the past three centuries the glaciers of the Vatnajökull region have advanced several times. A smaller transgression occurred in 1710, two larger ones in the 1750s–1760s, and in the 1830–1850 period, and again a smaller one in 1890 (Thorarinsson 1943, 1963). The 1890 moraines form a clearly defined zone of hillocks and ridges. This zone can be dated fairly accurately, because the outermost ridges are marked on a Danish military map compiled in 1904. Since, on the whole, no other moraine forms have survived between the 1890 zone and the old end moraines discussed above, one might assume that these older moraines go back to the 1830–1850 transgression. Todtmann (1960), on the other hand, is of the opinion that these moraines might be even older than the 1750 transgression.

After Thorarinsson (1943) the front line of the southwestern part of the Skeidarárjökull glacier in 1756 was approximately 1700 m south of its 1904 location and in 1850 about 800 m south of this line. This would indicate, that the inselberg-type remnants of end moraines in the Skeidarárjökull forefield are older, and that they must have been formed during the transgression of the glacier in the middle of the 18th century (in 1750–1760).

In addition to this, these moraines show additional features tending to indicate that they are older than the 1830–1850 transgression. Certain species of plants are fairly well developed on them such as *Rhizocarpon geographicum* (samples of these plants brought from Iceland, have been identified by Dr. J. Zielińska of the Botanical Institute of Warsaw University). After Schytt (1959), *Rhizocarpon geographicum* may serve as an accurate index of the time a glacier retreated, because this is the first plant which takes root on bare moraines. Depending on their age, these plants have different diameters; a 10 mm diameter indicates 50 years' growth. The diameters of the plants of species *Rhizocarpon* collected in Iceland from the old end moraines indicate that the glacier must have receded from them some 200 years ago. This would mean that the moraines originated during what is called the "small ice period", that is, from 1750 to 1760.

It is worthwhile mentioning that this method of determination of the age of land forms from their vegetation is relatively new, and that it has not been verified in other regions except Sweden and on Baffin Island. Hence the data given should be treated very cautiously. It is unknown, for instance, whether the growth of plants of species *Rhizocarpon* is similar under different geographical environments. Conditions in the Swedish mountains differ somewhat from those in the Skeidarárjökull forefield in Southern Iceland. In the first place, differences in altitude a.s.l. must be considered; in Sweden the altitude is 1500 m and more, in Iceland it is less than 100 m. Also taken into account

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Fig. 6. Huge palagonite blocks resting on old end moraine. Photo by M. Bogacki

Fig. 7. Disturbed silt and sand beds partaking in structure of old moraine. Photo by M. Bogacki

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must be the difference in the substratum in which the plants are developing: in Sweden we have old crystalline and metamorphic rocks, in Iceland young volcanic rocks transformed and transported by the glacier and by glaciofluvial waters. Also, the material supplied from Iceland is very scanty, and it has proved impossible to investigate statistically. For all these discrepancies it seems that plants may be used at least as approximate indices of the age of land forms. In determining age, our agreement with the data supplied by Thorarinsson, produced by different methods, is considerable.

From the 1830–1850 transgression practically no land forms have survived. Only some rock piles in the forefield of the 1890 moraines seem to be remnants of forms dating back from this old time (Fig. 8), and our plant analysis confirms this assumption. In the western part of Skeidarársandur these types of remnants occur some 500–600 m south of the 1904 glacier margin; for the eastern part Thorarinsson (1943) believes that in the western section the glacier front was at that time some 800 m south of the 1904 glacier margin.

Fig. 8. Rock remnants of moraines from 1840–1850. Photo by M. Bogacki

The plain of degradation and aggradation is dissected by a great number of meltwater valleys and tracks which originate from the main and moraine zone; all of them are dry today. Near the end moraines these meltwater valleys and flow tracks are incised up to 2 m into their surroundings, but southward they rapidly become shallow. In the channel floors a thin layer of fine-grained dark-grey or brown sands has been deposited. Following traces of streaks in this accumulation one can track down the course of these former streams. The floor deposit mentioned is the result of the gradual cessation of

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flow at the time the mean end moraine zone was forming. The compact area of accumulation of material associated with meltwater valleys and channels starts at the distance of 1–2 km south of the old end moraines.

The zone of outwash fans and meltwater valleys associated with the main end moraine zone

In the forefield of the main end moraine zone short outwash fans have developed (Fig. 1). They belong to the older moraines and consist of stratified sands and gravels. The proximal part of these fans slants at an angle of 6–7°; this angle decreases downwards and the fans gradually pass into the plain of degradation and aggradation.

In the western section of Skeidarársandur, in the forefield of the older end moraines, the outwash fans do not issue from any moraine gates; here the waters escaping from the moraines have accumulated gravel and sand deposits over the whole extent of the end moraines. The waters arriving from the younger end moraines have incised the older outwash fans and damaged them. Some wider gates pass into well developed meltwater valleys with terraces. Today these valleys are dry.

In region under discussion the most eastern moraine gate — described in detail by Galon (p. 31–33) who calls it gate No. 10 — which forms a gap in the main end moraine zone, passes southward into a meltwater valley about 1 km long. Farther south, this valley widens and becomes shallow. In cross-section it is asymmetrical; on its right bank the highest outwash level (VI) is formed by the oldest outwash fan associated with the older end moraines of the main zone; the material of this fan overlies the plain of aggradation and degradation. On the left bank of the valley, on the other hand, both the outwash fans and the older end moraines were destroyed. Due to this the valley borders upon a slightly lowered plain of degradation and aggradation (V) which is older than the main end moraine zone and the older outwash fans associated with this zone. As to altitude, however, this plain extends at a lower level than the oldest fans, or higher than the lower erosive levels incised into the plain lie at. Events must have been as follows: in the first stage of accumulation of the main end moraine zone the meltwater deposited short fans of material of degradation and aggradation, in later stages the waters dissected these fans and continued their flow over the plain of degradation and aggradation and doing so they changed to some degree the relief of this plain.

The numbers marked on our figures refer to successive altitude levels, and here No. V, marking the plain of degradation and aggradation, denotes a level lower than that of the oldest outwash fans marked No. VI. This same way of assigning numbers to altitude levels can also refer to successive stages of water runoff during the period of accumulation of the main end moraine zone was taking place. Level No. VI of the oldest outwash fans is associated with the earliest meltwater flow, while in a later period these waters ran over the plain of degradation and aggradation marked No. V and, in the latest stages, over successively younger erosive levels incised into this plain.

Besides the valley floor a higher erosive terrace, No. IV, can be seen in the dry valley mentioned above; this terrace is associated with the younger moraines of the main end moraine zone. Northward the rough stone-lined valley floor (No. III) reaches the elongated depression extending along the rear of the main end moraine zone. The different levels observed in this dry valley might be correlated with the following levels observed and described by R. Galon in the Sandgigjukvisl valley:

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Dry valley	Sandgigjukvisl valley
VI	VI
V	V
IV	IV
III	III

In the western part of the area discussed by the author, the main end moraine zone has virtually been completely flushed away for a width of about 2 km. The meltwater streams which during successive stage of the formation of the main end moraine zone passed over this area, have denuded several erosional levels in this wide gap. The main axis of this gap section runs from NE to SW, and this is why all the water flowing here finds its way into the River Sula. Today one observes, that in its N–S flow the Sula is undercutting in this gap section erosive levels of different age and of various thicknesses.

As may also be seen in other gaps or dry valleys, erosive levels are most sharply developed only in their proximal part. In the area discussed the highest, oldest level of outwash fans (No. VI) has nowhere survived. Hence the plain of degradation and aggradation is the highest here (No. V) over which the meltwater was escaping at the time the younger end moraines were being piled up (Fig. 9). Associated with these younger moraines is also level No. IV,

Fig. 9. Outwash levels along Sula River. Photo by M. Bogacki

while level No. III belongs to the chain of end moraines bordering upon the elongated depression. Finally, level No. II, the lowest, which here and there runs in two steps, extends in its proximal part to the meltwater kettles in the elongated depression, while its distal part is undercut by Sula waters. The successive levels distinguished by the author in the wide gap of this part of the discussed region correspond in the Sandgigjukvisl valley to the following levels:

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moraine gap	Sandgigjukvisl valley
VI (lacking)	VI
V	V
IV	IV
III	III
II	II
I (lacking)	I

(C) THE ZONE OF ELONGATED DEPRESSIONS WITH MELTWATER KETTLES

In the rear of the main end moraine zone there extends, parallel to it, an elongated depressed zone. The altitude difference between the moraine crests and the bottom of the depression is as much as 20 to 30 m. Near the eastern boundary of the discussed area the elongated depression is about 100 m wide; westward it widens and is joined by a branch depression oriented in a N–S direction. In the western part it becomes rather difficult to delineate the boundaries of this depression. Its floor contains a string of recently formed, round kettles formed by the melting of dead ice blocks. In their majority these kettles are now dry; formerly they contained lakelets. Of existing lakes the largest lies in the northwestern part of the depression (Fig. 1). The relief of its surroundings shows that this lake is steadily decreasing in size. In 1968 the following levels were determined at the lake shore: taking 0.00 as the water level of this lake, there is:

0.50 m	the lake floor,
3.70 m	the ice-dammed level No. I,
6.00 m	the glaciofluvial level No. II.

Both the lake floor and level No. I were built of very fine-grained sands and brown silts brought in by winds. The structure of this material resembles varved deposits, because thin silt and sand lie in alternate sheets. This series of varved deposits was only 0.5 m thick; it was underlain by fluvioglacial sands.

In its western part, as in the central part of Skeidarársandur, several stages in the development of the elongated depression can also be observed: a fluvioglacial stage, a stage of ice-dammed basins, and a stage of basin storage and fluvial runoff. In the first-named stage the meltwater streams from the glacier were passing moraine gates in their escape into the forefield; evidence of this is a terrace 6 m high built of sands and gravels. In the stage of ice-dammed basins, part of the depression was occupied by an extensive lake basin. And in the final stage, ice-dammed basins were occupying only the deepest hollows and gradually losing their water.

(D) THE INNER ZONE OF MARGINAL DEPOSITS AND LAND FORMS

Beginning at the main end moraine zone the ground surface gradually subsides northward, toward the glacier. Between the elongated depression and the glacier margin lies a wide expanse, highly diversified in relief and dissected by the peripheral river. In this area the pattern of the land relief changes: more and more the long forms extend in a N–S direction, at right angles to the glacier front.

This area contains an abundance of mounds and hillocks and undrained basins. Also seen here are elongated, narrow esker ridges with steep scarps,

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4–5 m high, extending in a N–S direction. As to structure, the eskers are built of stratified sands and gravels (Fig. 10).

The majority of the protruding land forms encountered consist of stratified silts and of fine- and very fine-grained sands; often their structure appears disturbed. The strata are not slanting southward in conformity with the general water runoff from the glacier, but in opposite, i.e., a northward direction. The disturbances are secondary features, caused by melting dead ice blocks and by bottom subsidence resulting from these processes (Fig. 11).

Fig. 10. Esker structure. Photo by M. Bogacki

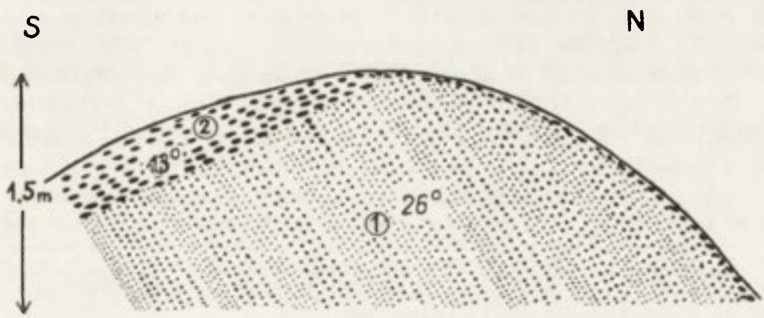

Fig. 11. Crevasse form with disturbed dip of strata:
1 — silty and very fine-grained sands, 2 — coarse-grained sands and gravels

Also seen in this area are extensive elevations built of stones and gravels. These are large forms with relative altitudes up to 10 m and with steep scarps. Their long axes frequently run in a N–S direction, at right angles to the glacier margin. Adjoining these elevations lie meltwater kettles. It seems probable that the elevated land forms have developed in large glacier crevasses. Today one sees the formation of similar forms in the zone of stagnant ice, often as extensions of older forms.

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The eastern part of the discussed area contains a widespread hilly region marked by a great number of undrained basins; for the most part this region is built of silt deposits.

(E) TENTATIVE DEFINITION OF THE COURSE OF DEGLACIATION ON THE SECTION OF THE SKEIDARARJÖKULL CONSIDERED SO FAR

The nature of the land forms encountered in this forefield and their structure indicate that at the time the relief of the inner zone of deposits was developing, the character of deglaciation has been changing from frontal and marginal to an ablation type of deglaciation, and that during this latter period widespread silt-covered plains came into existence.

The author used the occurrence of boulder accumulation, of end moraines and gaps within the N–S oriented moraine ridges and walls, as well as information gained from the 1946 map as a basis for marking the line of standstill of the glacier, tying this line in with the situation observed in the central part of Skeidarársandur. In this endeavour he distinguished: old moraines from the middle of the 18th century, the probable extent of moraines from 1840–1850, moraines from about 1900, and moraines of younger age. Within the main end moraine zone, five morainic chains can be clearly observed, among them two in the section covering the older moraines. A sixth line marking a stand still stage of the glacier extends in the nearest northern vicinity of the elongated depression.

In the middle of the 18th century, at the time deglaciation started, its character was frontal (end moraines). How deglaciation proceeded during the later stages of glacier decay cannot be stated, because for that period no land forms whatsoever have survived.

At the turn of the 19th–20th century, about 1900, the deglaciation proceeded frontally, and at that time the outer ridge of the main end moraine zone (what is called the older end moraine ridge) developed. Afterwards the glacier seems to have retreated a considerable distance, but lacking any sort of evidence it is impossible to point out where the glacier margin ran at that time. In the nineteen thirties a new transgression of Skeidarárjökull took place, and by 1938 the glacier snout had advanced close to the end moraines of the period about 1900; it is probable that at that time the younger end moraines of the main end moraine zone came into existence. Initially the deglaciation proceeded in a frontal character which later changed into a marginal character (crevasse walls) and an ablation character (areas built of silt deposits).

Fig. 12. Geomorphological map of the area east of the Sandgigjukvisl River. Forefield of main end moraine zone

1 — Outwash fans: spread of lines indicates different runoff directions on the plain of degradation and aggradation, 2 — Scarps of outwash levels and of terraces in valleys, 3a — Stone remnants of end moraines connected with main end moraine zone, 3b — Stone remnants of end moraines connected with older stages Rear area of main end moraine zone (from about 1910–1960)

4 — Morainic plain, with flat or undulating surface, 5 — Low stone hillocks or undulating plateaus, 6 — Larger-size flat gravel-boulder ridges (crevasse forms), 7 — Isolated mounds built of boulder clay, 8 — Isolated mounds built of silty material, 9 — Ice-dammed depressions or meltwater kettles, 10 — Former river channels, 11 — Stone remnants of end moraines connected with younger stages (from about 1910 to 1960)

Valleys of meltwater streams

12 — Silt-covered sheets on river banks and in river delta, 13 — Valley floor pavement

Main end moraine zone

14 — High morarninc hillocks and walls (more prominent ridges from rise of 20th century), 15 — Old morainic falls from about 1890, 16 — Larger meltwater basins and elongated intermorainic depressions, 17 — Moraine-outwash gates in end moraine chains, 18 — Sharp scarp steps within end moraine zone (ice-contact walls)

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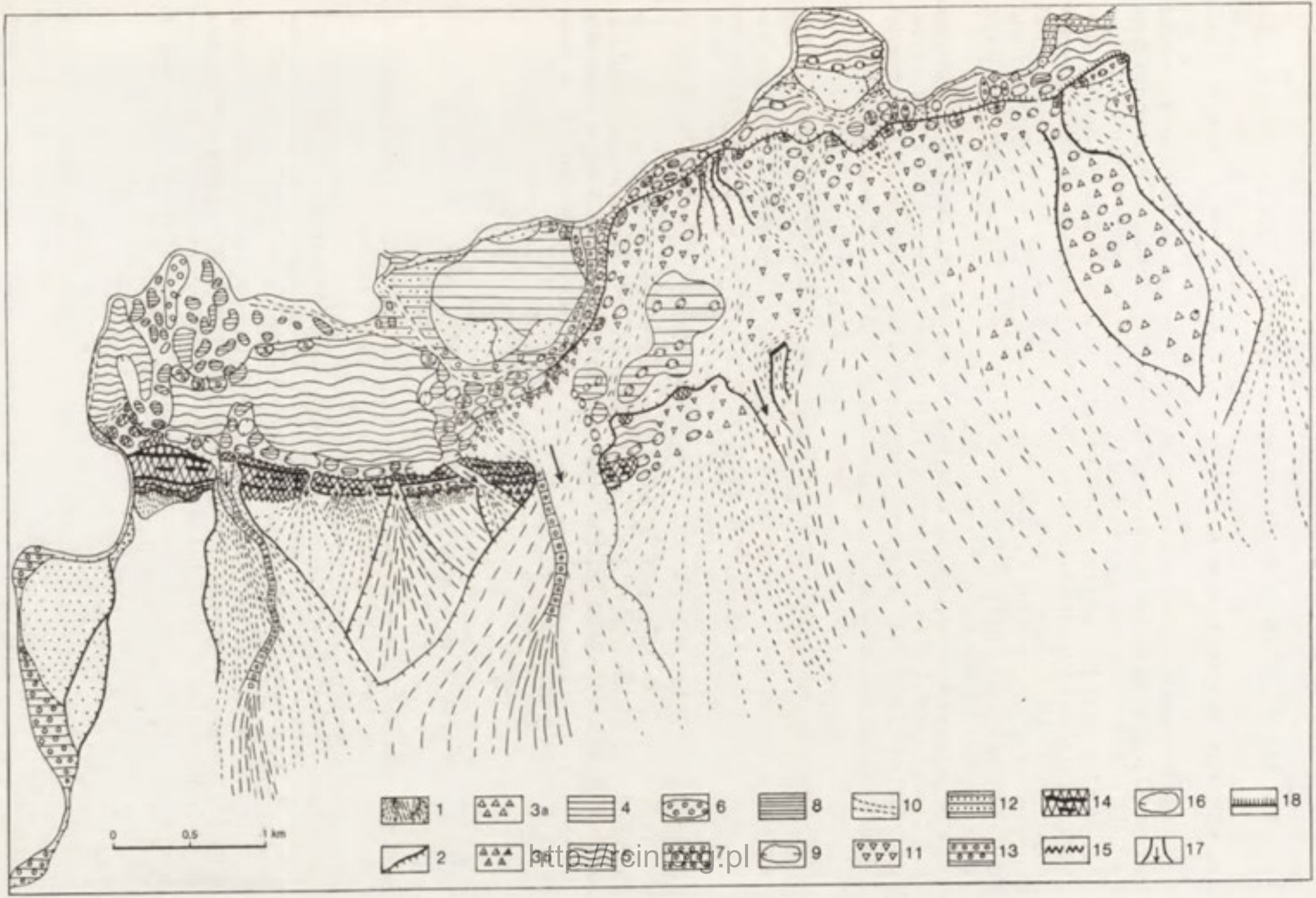

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(2) CENTRAL EASTERN SECTION OF THE PROGLACIAL AREA

(A) THE MAIN END MORAINE ZONE

This zone extends over a distance of some 8 km east of the Sandgigjukvisl River (Fig. 12). As in the central and western parts, one can distinguish here a main end moraine zone subdividing it into a zone of outer moraines of older age and a zone of younger moraine ridges, hillocks and morainic plateaus. These end moraines, both the older and the younger, are dissected by transverse depressions called moraine gates. In the forefield of the end moraines short outwash fans have developed, while in the immediate rear of the main end moraine zone an elongated depression extends and, farther north, an inner zone of marginal deposits and land forms.

The main end moraine zone runs in northeasterly direction, starting from the Sandgigjukvisl River; but in the eastern part of the section under discussion the end moraines have been completely washed away by glaciofluvial waters.

The older end moraines

In southerly direction these are the outermost land forms in the main end moraine zone. To the north they pass into younger end moraines, mostly without a boundary between them being visible. Only in the eastern part the older moraines are separated from the younger end moraines by a clearly seen, narrow elongated depression of marginal type, containing small meltwater kettles.

The older end moraines bear the features of flat plateaus with small hillocks on top. Between gates No. 6 and 7, counting from the Sandgigjukvisl River in an easterly direction, the older end moraines have been powerfully dissected by small gate-type valleys; five of them appear crowded into a distance of scarcely 0.5 km.

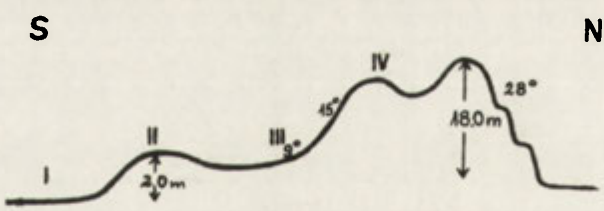

Fig. 13. Section across moraines dating from 1830–1890 period
I — outwash plain, II — moraine from 1830—1850, III — outwash fans, IV — moraine from 1890

The older end moraines are always lower than the younger end moraines (Fig. 13). They are built of unequigranular sands, gravels and larger boulders. In their distal side the moraine material is interdented with sands of the outwast fans (Fig. 13). The relation of the end moraines to the outwash fans is clearly exposed in the Sandgigjukvisl gap section.

The younger end moraines

They form a separate clearly marked zone extending in the immediate rear of the older end moraines. The relief of these younger end moraines is very highly diversified (Fig. 14). A number of morainic ridges can be observed here which run parallel to the main moraine axis. These ridges are from 1 to 3 m

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high; between them lie depressions containing ice-dammed basins. Two types of ridges can be distinguished: one with fairly sharp crests and more gentle distal slopes, the other fairly wide and flat-topped, additionally varied with strips of stone rows several centimeters, at the most 50 cm, high. Every one of

Fig. 14. End moraine zone from about 1900. Photo by M. Bogacki

Fig. 15. Rock debris and stone cover on moraines from about 1900. Photo by M. Bogacki

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these wide ridges carries up to 10 micro-strips of stone accumulation. The whole end moraine zone contains up to 12 ridges, 2 or 3 of which are wider at the top and carry the stone strips mentioned. Among these ridges, the highest are the first and eight or ninth; the remaining ridges have grown less high and are less clearly discernible.

Most probably the varied relief of the main end moraine zone goes back to the structure of the glacier, where moraine material was separated out from inside of crevasses and between slide planes of the ice.

In a southerly direction, towards the elongated depression with its kettles, the moraines grow lower by three distinct steps which may have originated from a decrease in volume of the dead ice which once filled this depression. Some of these steps may also represent contact walls of the glacier margin.

The younger end moraines are built of very much heterogeneous material: silts, sands, gravels and large boulders (Fig. 15). A certain regularity can be observed in their structure insofar as usually the cores of the morainic ridges consist of fine- and very fine-grained material. The cover layer, varying in thickness, always consists of coarse deposits.

A clearly noticeable boundary plane separates the moraine deposits from their substratum; this boundary can be observed in the Sandgigjukvisl gap as

Fig. 16. Relation of end moraine from about 1900 to older outwash plain. Photo by M. Bogacki

well as at other points where the moraines have suffered destruction. The glacier has clearly transgressed upon a surface levelled by denudation; as a trace of this degradation appears a stone pavement underlying the moraines (Fig. 16).

In many places the end moraines dating back from about 1900 have been completely flushed away — in the western part of Skeidarársandur a wide moraine area has been denuded by the Sula River, while in the eastern part the

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moraines have been mainly destroyed by waters of the Rivers Skeidara, Sigurdurfitjarall, Haaldavisl and Sugurdharfitajaralar. The three last-named rivers ran as late as 1945–1946: now they are completely dry.

Due to the partial destruction of the end moraines it is difficult to indicate definitely how far the glacier has advanced in the eastern part of Skeidararsandur. The distribution of moraine remnants indicates that the 1890 moraines ran in a SW–NE direction.

The moraine gates

The end moraines are dissected by numerous valleys (or gates) of different widths (Fig. 12). These gaps were the tracks by which the waters from the front of the glacier were escaping. In width the glacier gates vary considerably, from very small ones ending with an outwash fan to wide gaps with terraces.

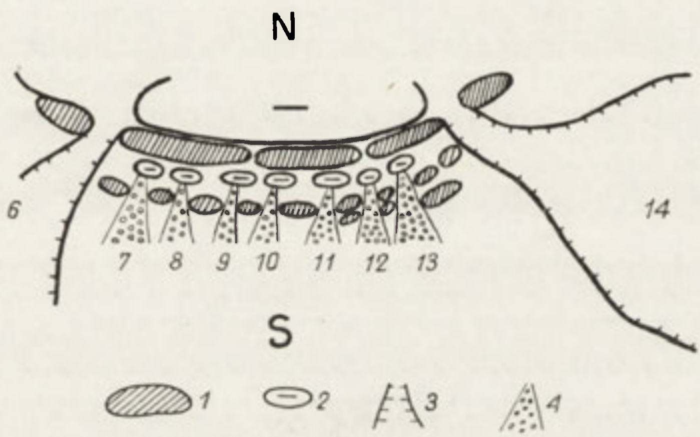

Fig. 17. Smaller gates dissecting outer moraine ridge, east of the Sandgigjukvisl River

1 — morainic ridge, 2 — depressions, 3 — scarps, 4 — small valleys (gates: 7-13); 6, 14 — large gates

Gates Nos. 7–13 may serve as examples of small valleys (Fig. 17). These are situated east of the River Sandgigjukvisl. These small gates were in operation only for a short time; usually they cut only the outermost moraine ridge and very little water used to pass thorough these gates. This moderate flow washed away the fine material leaving behind the larger pebbles and boulders (Fig. 18). This type of landform may even have originated later, at the time when the dead ice was melting. At present, this type of small gate valleys are developing on the slopes of ice-morainic ridges extending in close vicinity to the glacier snout.

Another type of small shallow valleys or gates have developed in the shape of cones. They started from different moraine chains, at times even from the youngest, such as gates Nos. 3, 4 and 5. The floors of these gates are lined with coarse residual material (Fig. 19 and 20). The gates originated at different times and were in operation for a relatively short time, probably only for the period in which one or two morainic ridges were formed.

The longitudinal profile of the small gates is characterized by steep gradients, especially in their upper part. The longitudinal profile of the cone-sha-

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Fig. 18. Gate dissecting outer moraine chain. Photo by M. Bogacki

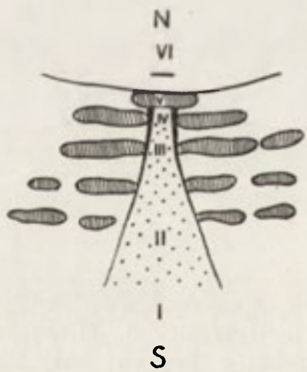

Fig. 19. Example of small gate (No. 5) starting from the youngest moraine chain (for legend see Fig. 17)
I — outwash plain, II — stone fan in gate, III — step dividing gate into two parts: the one sloping southward, the other, northward, IV — part of gate subsiding northward, V — end moraine wall separating gate from meltwater basin, VI — meltwater basin

ped gates is often disturbed. Here the upper section may slope northward, in the opposite direction to the lower part; this feature must be ascribed to dead ice melting.

The gates (or valleys) Nos. 1, 2, 6, 14 and 15 are of a different character. They have been in operation for a longer time, or throughout the period in which the end moraines dating back from about 1900 have developed. These gate features are large, cutting across the entire end moraine zone (Fig. 21 and 22). Usually their upper part abuts on one of the youngest end moraine

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ridges, or it ends with a sill step inclined towards the ice-dammed depression. However, when the buried dead ice melted, these gates ceased to function. All of them show well developed erosive levels. Mostly one sees three or four levels and, at places where the oldest outwash remained intact, even five levels (Fig. 22). Each of these levels is associated with some definite end moraine ridge (Fig. 23). The oldest and highest level (VI) corresponds to the older end moraines, level V to the first ridge of the younger end moraines, level IV to the eighth or ninth ridge, and level III to the youngest ridge bordering upon the ice-dammed depression. The lowest level (II) breaks off with a step inside

Fig. 20. Gate dissecting several moraine chains. Photo by M. Bogacki

Fig. 21. Gate dissecting entire end moraine zone. Photo by M. Bogacki

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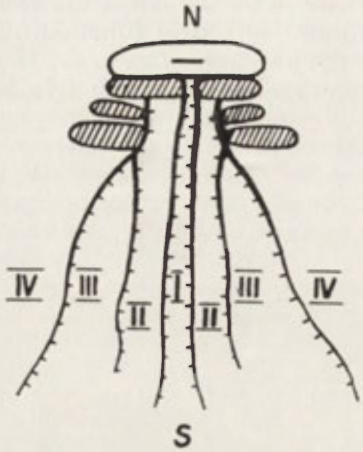

Fig. 22. Gate with successive levels of erosion (Gate No. 1)

Fig. 23. Diagrammatical profile of levels connected with end moraines from about 1900

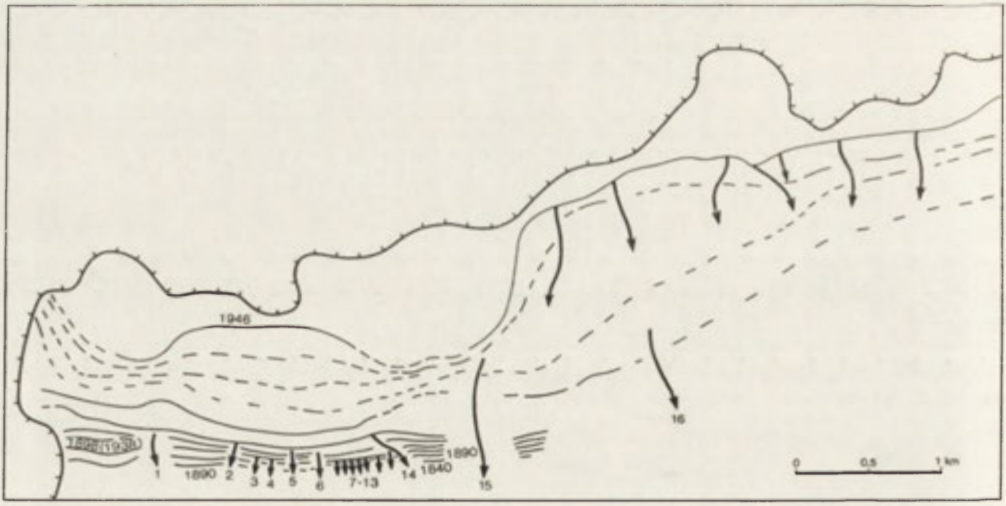

Fig. 24. Stages of glacier stillstand east of the Sandgigjukvisl River
1–16 — gates
↙ run-off directions

the depression. Over this latter level used to flow the water while the depressions were filled with dead ice. The altitude differences between successive erosive levels are usually some 2 to 3 m. These differences are greatest for the upper parts of the gates, within the end moraine zone. On the distal side of the moraines the particular levels wedge out rather early, vanishing at a distance of barely 1 to 2 km from the end moraines. At present, through one of the large gates in this area passes the River Sandgigjukvisl (cf the description of the central part of Skeidarársandur). A particular type of gate is the wide gap marked No. 16 which extends over many kilometres (Fig. 24). This enormous gap was formed due to the destruction of all protruding land forms by the water streams arriving from numerous glacier gates. Only remnants of some of these gates have survived as dry valleys lined with boulders.

(B) THE ZONE OF OUTWASH FANS AND MELTWATER VALLEYS ASSOCIATED WITH THE MAIN END MORAINE ZONE

In the forefield of the main end moraine zone a zone of short outwash fans appears, and farther south extends a wide plain of degradation and aggradation. The origin of this area has been discussed in the description of the central and western section of Skeidarársandur.

As is the case for the remaining area, the plain of degradation and aggradation constitutes the substratum of the glacial and fluvioglacial deposits dating back from about 1900 (Fig. 16).

In the eastern part of Skeidarársandur no moraines from the rise of the 18th century have survived, nor any from the 1840–1850 period. At one place alone, directly in the forefield of the main end moraine zone, a pile of boulders has been discovered with well developed plants of *Rhizocarpon geographicum*. An analysis of these plants revealed that these remnants are older than the main end moraine zone but younger than the old moraines. Hence, they may be considered as washed-down moraines from the 1840–1850 period (Figs. 13, 24).

The oldest outwash fans form a narrow zone in the forefield of the older end moraines; this means that they occur where these moraines have survived. In the region under discussion, outwash fans extend from the Sandgigjukvisl River to gate No. 1 and then again between gates Nos. 6 and 7 (counting from the river), and between the 1840–1850 end moraines and the older moraines have survived, but the outwash fan has been washed away. The fan between the Sandgigjukvisl River and gate No. 1 has developed in the forefield of moraines which have not been dissected by any gate or runoff channel; this proves that this outwash fan goes back to water runoff from the moraine zone.

The outwash fans discussed above constitute the oldest and highest level of water runoff form in front of the main end moraine zone. They are built of unequigranular sands and gravels. The short transportation distance prevented thorough sorting of the material as can be seen from the following analysis.

During younger stages of the formation of the main end moraine zone, the outwash fans were dissected by waters escaping from glacier gates. As a result the area discussed was covered by an entire generation of outwash fans which, however, accumulated relatively little material at the gate outlets (Fig. 12); these fans originated during increasingly younger moraines of the main end moraine zone. Usually the outwash fan at a gate outlet was higher the younger was the age of the water runoff which piled up that particular fan. Some of the gates must have been operating at the same place for the whole period

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TABLE 1. Mechanical composition of material building the oldest outwash fans associated with the 1890 moraines

Fraction	Percentage	Fraction	Percentage
2.5 mm	11.35	0.25 mm	8.44
2.0 mm	1.41	0.2 mm	8.67
1.0 mm	11.31	0.1 mm	12.49
0.8 mm	6.48	0.08 mm	2.42
0.5 mm	14.69	0.063 mm	2.71
0.4 mm	6.41	less than	
0.315 mm	10.51	0.063 mm	3.11

during which both the older and the younger end moraines originated. Hence, one can speak of certain longevity of some runoff directions.

At gates Nos. 1 and 8, large meltwater valleys with terraces have developed dissecting the whole main end moraine zone. The valley of gate No. 1 measures up to 200 m width in its upper part. Three erosive levels can distinctly be observed here; their heights are:

level III	1.5–2.0 m
level IV	3.0–4.0 m
level V	7.0–8.0 m

The three levels III, IV and V can be seen at practivally all gates. Level V is the principal level of outwash runoff in the forefield of the 1900 end moraines, whereas level VI forms only a short fan associated with the older end moraines. Levels V and IV are connected with the younger, and level III with the youngest moraines of the main end moraine zone, while level II, as was noted before, breaks off with a step inside the elongated depression.

In this meltwater valley, level III can only be observed over a distance of 0.3–0.4 km. Apart from the stone-lined channel floor of the former stream, at lower altitudes lie levels IV and V; but some 1 km south of the moraine zone all the levels mentioned vanish.

The valley extending downstream of gate No. 8 bears the features of a large gap about 0.5 km wide. No separate erosive levels have been preserved. Apart from the stone-lined floor, there is really only one extensive level here, corresponding to level III of the dry valley developed south of gate No. 1.

East of the last-mentioned gap one sees over a distance of about 1 km the remnants of destroyed moraines and, in their forefield, one level of degradation and aggradation. Farther east, as far as the boundary of this area, there extends the plain of degradation and aggradation (Fig. 12) which contains numerous scattered piles of boulders as remnants of the main end moraine zone and of younger moraines.

It was mentioned above that the meltwater valleys became shallow very rapidly, and that to the south they pass into a plain of degradation and aggradation in which the dry channels of former outwash streams are filled up to 0.5 m height with accumulated sand. The present-day deposition of glaciofluvial material by the three large Rivers — Skeidara, Sandgigjukvisl and Sula takes place much farther to the south, and therefore the plain of degradation passes farther southward into an area of accumulated material consisting of fairly well sorted sands (Table 2).

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TABLE 2. Accumulated outwash deposits, from 4 to 5 km south of the Sandgigjukvisl gap

Fraction	Percentage	Fraction	Percentage
2.5 mm	2.49	0.25 mm	12.09
2.0 mm	1.01	0.2 mm	11.97
1.0 mm	6.10	0.1 mm	12.09
0.8 mm	5.17	0.08 mm	0.91
0.5 mm	19.76	0.063 mm	0.34
0.4 mm	10.77	less than	
0.315 mm	16.84	0.063 mm	0.46

(C) THE ZONE OF ELONGATED DEPRESSIONS WITH KETTLES

In the rear of the main end moraine zone there extends an elongated depression. Its contours are sharply marked in the western part of Skeidarársandur, while east of the River Sandgigjukvisl the uniform depression is most often replaced by a great number of kettle holes caused by dead ice melting, which in their general outline form a depressed area parallel to the main end moraine zone.

The difference in relative height between the end moraine crests and the floors of the depressions in the rear of the moraines is from 20 to 30 m. The moraines descend by three steps in a northern direction toward the elongated depression.

Until recently the kettle holes scattered in the elongated depressions contained lakes. Some of these lakes can still be seen on the aerial photographs taken in 1945 and 1946 and on a 1:50,000 topographical map drawn in 1949. At present all these kettles are dry and filled with fine-grained silty deposits. Their top layer consists of black or brown aeolian dust.

(D) THE INNER ZONE OF MARGINAL DEPOSITS AND LAND FORMS

The elongated depression discussed with its kettle holes separates the main end moraine zone from the inner zone of marginal deposits and land forms. The character of the relief of this area differs somewhat from the character seen in the western and central parts of Skeidarársandur.

East of the Sandgigjukvisl River wall- or ridge-type forms running in a N–S direction occur less frequently. In contrast, fairly extensive moraine plateaus reach as far as the glacier margin. In the region discussed three such moraine plateaus can be distinguished: the western, the central and the eastern plateau. The western plateau is largest and highest with its 15–20 m relative altitude. Smaller and less high is the central plateau where the latitudes are of the order of 10–15 m, and smallest is the eastern plateau (Fig. 12). This entire moraine area has been dissected and partly eroded by a peripheral river.

The surfaces of the morainic plateaus are varied due to low hillocks of end moraines which rise in their northern parts and are associated with the youngest glacier standstill in this area. This was probably the line which the glacier margin occupied in or about 1946 (Fig. 24). Further, minor undrained depressions occur here, as well as a great number of low ridges running perpendicular and parallel to the glacier snout. They are up to 1.0 m high and consist of piles of stone and rubble. These forms resemble ground moraines of the fluted type, described from other areas (Schytt 1959; Todtmann 1957,

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1960; Szupryczyński 1963, 1965, 1968). They originated from the flow of a plastic moraine mass underneath the glacier.

The land forms observed in the Skeidarárjökull forefield are probably of a different origin. They may have developed in one of two ways: some grew as extensions of existing rock blocks or of ice monadnocks which shielded them. Others were piled up due to the accumulation of loose material at the outlet of gullies incised into the glacier front. The water rushing down these gullies deposited varied material close to the glacier margin. Afterwards, with the gradual retreat of the glacier, these ridges grew in length.

The rear of the main end moraine zone contains an abundance of buried dead ice lobes which are continuously melting, this creates undrained hollows which sometimes are dry and sometimes hold water. In this way the relief of this region is steadily in a state of transformation, and the structure of the land forms in a state of deformation. Some of the extensive meltwater basins were retaining ice-dammed lakes. One of the larger basins, dry at present, lies on the left bank of the Sandgigjukvisl River; it is the southward extension of an existing large lake. The floor of this basin is highly diversified, containing many sandy and silty hillocks different in shape and height.

In the meltwater basins the water table used to be much higher than it is today, as seen from abrasion-type flattened surfaces or from silt deposits on the slopes. Thus, compared with the water level observed in 1968, the highest water table in the lake was 6 m higher; fairly distinct traces of lower water levels can be recognized also at 4.5 m, 3.0 m and 1.5 m above the present water table. At the rate at which the water dropped in the meltwater basin, there emerged a steadily growing number of mounds built of sands and silts which originated in dead ice fissures. The floor of the meltwater basins contained, and still contains, blocks of dead ice. From time to time new basins of this type develop here.

Shore lines left by a high water table can also be traced in the slopes of the eastern lake where five erosive-abrasive levels are distinctly visible; in relation to the water table in the lake the altitudes of these levels were:

I	1.5– 2.0 m
II	4.5– 5.0 m
III	7.0– 7.5 m
IV	12.0–13.0 m
V	19.0–20.0 m (Fig. 25)

The water subsidence in the lakes and ice-dammed basins proceeded in conformity with the recession of the glacier, the melting of buried dead ice, and the formation of a southward runoff combined with a drop of the groundwater table.

The narrow zone of marginal deposits and land forms and of meltwater kettles extends in a thin belt between the moraine zone and the peripheral river.

Towards the end of 1946 the peripheral river ran in a channel farther south; later it moved northward, and its channel was straightened. This river forms the connection between the eastern and the western lake, and then it continues its course due south and is now called the Sandgigjukvisl. In the former channel of this river erosive terraces have survived: five levels in the eastern part, similar to those in the lake mentioned above, and four levels plus a stone-paved channel floor in the western part. Moreover, many newly formed undrained hollows occur in the former channel; today these hollows

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Fig. 25. Levels of erosion and abrasion along eastern lake. Photo by M. Bogacki

are partly filled with water. This ground ice melting took place after the river had left its former valley, and there are many depressions of this type to this very day. This is definite proof that the peripheral river (a *pradolina*) ran over a region in whose substratum dead ice blocks had been buried.

(E) TENTATIVE DEFINITION OF THE COURSE OF DEGLACIATION IN THE SECTION OF SKEIDARÁRJÖKULL FOREFIELD CONSIDERED SO FAR

The absence of any sort of land forms in the area situated south of the main end moraine zone prevents any comment on the glacier retreat from this area.

The main end moraine zone extends from SW to NE (Figs. 2, 12). The distance separating the end moraines dating from the middle of the 18th century from the younger end moraines is rather short in the eastern part; but the distance between the remnants of the 1840–1850 moraines and the 1890 is greater in the western part than it is in the eastern part.

At some points, especially in the eastern part, in 1890 the glacier might have transgressed beyond the end moraines of the 1840–1850 period. Evidence of this are ridge remnants with well developed plants situated closely ahead of the 1890 moraines. These features may have been due to the dynamics of the glacier, to its mobility which was greater in the western than in the eastern part. Even now this phenomenon can be observed: in its western part the glacier retreats fairly distinctly, while in the eastern part this retreat is negligible and, in certain sections, there is even an advance.

The author managed to draw the line of the glacier standstill for the period in which it retreated from the moraines of about 1900, up to 1946 (Fig. 12); he based this drawing on the occurrence of boulder remnants and end moraines,

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and on steps observed in the northern rim of the main end moraine zone and in the inner zone of marginal land forms. It is certain that in the eastern section of Skeidarársandur the glacier retreat proceeded in a diversified manner and in stages. At the time the main end moraine zone came into existence, the glacier retreated frontally. Its northward recession from the main end moraine zone occurred in the eastern part of Skeidarársandur in a somewhat different way to that in the central and western parts. No crevasse walls are visible which are characteristic for the remaining sections. The glacier decay must have proceeded mainly by ablation, and crevasses were of rare occurrence. Also left behind were large dead ice lobes which by melting created undrained depressions and ice-dammed basins. This pattern of Skeidarárjökull deglaciation has been described in detail by Galon (p. 147). In conclusion let it be said that the relief of the Skeidarárjökull forefield came into existence in a relatively brief time, and that the main end moraine zone has developed during two glacier oscillation stages.

(3) AEOLIAN PROCESSES IN THE SKEIDARARJÖKULL FOREFIELD

In the Skeidarárjökull forefield wind action is furthered by the complete absence of a green cover. Apart from some scarce plants of low growth and isolated weeds, the Skeidarárjökull forefield has no vegetation as far as the ocean shore.

During its stay in Iceland, from June to August 1968, the Polish Expedition found that in the forefield of the Skeidarárjökull glacier two wind directions predominate: southerly winds from the sea and northerly winds from the glacier [1]. The southerly winds are humid, bringing cloudy weather and rains; the northerly winds arriving from the glacier are dry and blow at high speed. They remove clouds of volcanic dust from the glacier surface, as well as from the outwash plains and from the ice-dammed lakes and river braids extending in front of the glacier snout (Iwan 1937; Thorarinsson, Kjartansson 1959). The intensity of these aeolian processes is highest in the western part of the Skeidarárjökull glacier next to the Sula River, because here wide areas are covered with silts and with extremely fine or normal fine-grained sand deposits.

The dust clouds thus raised are carried towards the ocean, but part of this dust is also deposited on the slopes of Lomagnupur, a basalt massif bordering upon the glacier from the west.

This type of duststorm was witnessed daily during the afternoon. When the winds were very strong which did not happen every day, not only dust was removed by them but also sands of a variety of fractions. During powerful high pressure area above the glacier strong winds were blowing from the glacier for several days. This took place, for instance, from August 11 to August 18, 1968; the violent winds caused regular dust- and sand-storms. At the meteorological station mounted on an end moraine from the 1890 period, the wind velocity was 16 m/s and, during sudden gusts, as much as 20 m/s.

At that time the winds used to remove dust as well as fairly large sand grains. Exposed in the Skeidarárjökull forefield are volcanic rocks and basalts, as well as layers of volcanic ash and cinders; the sand deposits have the character of disintegrated cinders, with a specific gravity much lower than that of quartz. As a result the wind was able to remove and whirl over the surface

[1] The meteorological observations were made by Dr. Wójcik, member of the Expedition; he is about to clarify and to report the data he has collected.

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fairly large grains. In fact, sand and cinder grains of 0.6 mm to 2.5 mm diameter were rolled along. Only somewhat smaller grains were wind-borne. For half an hour, the author caught in a plastic bag 880 mgl wind-borne sand at 1.5 m above ground surface. The grain distribution of this sand was as follows:

1.0	mm	8.1%
0.8	mm	4.5%
0.5	mm	6.8%
0.4	mm	2.3%
0.315	mm	2.3%
0.25	mm	1.1%
0.2	mm	4.5%
0.1	mm	30.7%
0.08	mm	15.9%
0.063	mm	13.6%
less than 0.063	mm	10.2%

The grain composition of the material from which the sands and dusts had been carried off by the wind, was as follows:

2.5	mm	5.8%
2.0	mm	3.2%
1.0	mm	13.0%
0.8	mm	8.5%
0.5	mm	17.4%
0.4	mm	7.3%
0.315	mm	10.8%
0.25	mm	8.2%
0.2	mm	8.3%
0.1	mm	8.7%
0.08	mm	5.5%
0.063	mm	1.5%
less than 0.063	mm	1.8%

Of this wind-borne material there have developed many of the crevasse forms, the end moraines and the outwash fans seen in the western part of the Skeidarárjökull forefield. The limited degree of sorting of these deposits is characteristic.

Finer grain fractions have also been carried off by the winds from the material coating the glacier snout. Here also, as in the glacier forefield, one sees a low degree of sorting, and the grain distribution is as follows:

2.5	mm	0.5%
2.0	mm	1.1%
1.0	mm	10.0%
0.8	mm	4.9%
0.5	mm	12.4%
0.4	mm	6.0%
0.315	mm	12.1%
0.25	mm	9.8%
0.2	mm	12.4%
0.1	mm	20.4%
0.08	mm	4.0%
0.063	mm	2.5%
less than 0.063	mm	3.4%

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The major part of the dust and sand deposits were dropped near the source from which they were derived. Hence the distance of transportation is very short, no more than 1.0 to 1.5 km; coarser grains, of course, travel a much shorter distance. All aeolian material is deposited on the slopes of the wide variety of land forms, on their lee sides. These deposits are more fine-grained and better sorted than those from which the winds had removed them. The grain size distribution of the air-borne deposits near the River Sula is as fol lows:

	0.2 mm	4.8%
	0.1 mm	39.1%
	0.08 mm	25.7%
	0.063 mm	17.0%
less than	0.063 mm	13.4%

Part of the sand and silt particles which are wind-driven over the surface of the outwash plain are retained in humid places and in depressions, and gradually they fill many old channels of former outwash streams. On the outwash sheet which in general is mantled by large rock blocks and coarse boulder and gravel material, narrow strips of fine-grained and silty deposits are built up in this manner. Some sections of shallow meltwater channels are completely filled in, and if these channels carry water, it permeates the aeolian material and often emerges as a small brook from under the range of the aeolian cover.

No further aeolian land forms are encountered in the Skeidararjokull forefield, in the proximal part of the widespread outwash plain called Skeidarar-sandur which in its proximal part has an erosive character. Some 4–5 km south of the 1890 end moraines one observes a clearly noticeable decrease in the gradient of the outwash sheet. From here, the Rivers Sula and Sandgigjukvisl at times overflow the plain and deposit the fine-grained material they have removed from in front of the glacier. The river channels are incised to barely 0.5 m depth into the outwash plain. Hence the humid ground surface and the occasional tufts of weeds tend to retain aeolian deposits in this part of the outwash plain. In the wake of every plant some of the material, rolled along or air-borne by the wind, is retained, and in this way hillocks up to 1.5 m high, 2.0 m long and 1.0 m wide are being formed. The long axes of these forms lie in a N–S direction, in conformity with the prevailing wind direction.

During winter the glacier forefield in Iceland is covered by snow which behaves in a very similar way as sandy material; this means, that large quantities of snow are carried off by winds and accumulated in the lee of northern winds. Often sand and dust collect on the snow. During winter, the glacier forefield was covered with snow while at other times there was no snow. This can be traced from sand and dust intercalations observed in the snow. When snow is covered with sand and volcanic dust it readily crystallizes changing highly into compact granular snow. The snow melts due to warm air currents arriving from the ocean. When the dust cover is thin and fails to form a continuous mantle, this usually accelerates snow melt. In the places where dust occurs in piles, snow melts more rapidly and forms depressions. But when dust and sand overlie snow in a compact layer, the snow will melt at a slower rate (Dobrowolski 1923).

In June 1968, snow covering the Skeidarárjökull forefield had survived only in the lee of northern winds underneath a mantle of sand and aeolian dust 15–20 cm thick. Underneath this mantle, the snow bed 0.5 m thick consi-

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sted of three layers: the upper and lower layers were recrystallized snow, separated by a middle layer of loose pure white snow some 15 cm thick.

In the second half of June 1968 the snow started melting at a rapid rate. Watersoaked masses of sand and aeolian dust which had covered the snow slid down the slopes. The surface of the snow cover became uneven, the thickness of aeolian deposits covering the snow also became dissimilar. This resulted in the snow melting proceeding at different rates: where the aeolian dust layer was less thick, the snow melted more rapidly. In a fairly short time the ground relief became highly diversified; a great number of hollows developed, different in dimensions and resembling land forms created by small dead ice blocks melting in the ground. Usually these hollows were round, some 2 to 3 m in diameter and no more than 70 to 80 cm deep. As noted before, snow melting proceeds at different rates, depending on the thickness of the aeolian cover. When melting occurs slowly, the hollows grow in depth and in diameter, and in this case concentrically arranged micro-levels of subsidence can be observed; least sharply marked are the outermost levels. The similar processes observed Kobendzina (1961) in the Kampinos Forest.

On the other hand, when the snow melts rapidly, aeolian material mantling the snow layer is apt to break off abruptly and to drop, and in this way deep funnel-shaped cavities are formed with steep and ragged walls. Often the bottom of this type of cavity contains remnants of unmelted snow. The walls are stratified, showing alternating bands of fine-grained or medium-grained, or extremely fine sands, or volcanic dust. Any very fine-grained or silty material dries out rapidly after the snow melts,`and in it desiccation cracks of different lengths are formed.

The depressions created by snowmelt are fairly durable forms; some of them kept their shape throughout the short Icelandic summer. The process of snowmelt and the depressed forms caused by this melting resemble what happens when buried dead ice melts. The difference lies in the scale of these phenomena. The ancient Pleistocene outwash plains in Poland are diversified by an abundance of undrained depressions of different sizes and shapes. Some of them, especially if they are sheltered from the prevailing winds, that is, if situated south of elevations and if they are of small size, may have originated from buried snow. At those times, from the glaciers, that is from the north, fairly strong winds used to blow and these winds have piled up the oldest dune forms. Wind-borne sand from such dunes may have been deposited on the snow sheet which, due to the more severe climatic conditions in Poland, must have lasted much longer during the last glaciation than they last in Iceland. And this is why in Poland the morphological results of these processes should also be of langer duration and more clearly noticeable.

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REFERENCES

Dobrowolski, A. B., 1923, *Historia naturalna lodu* (The natural history of ice), Warszawa.

Iwan, W., 1937, Über Löss und Flugsand in Island, *Z. Ges. ErdK.* Berlin.

Kobendzina, J., 1961, *Niektóre zjawiska towarzyszące procesom eolicznym na wydmach Puszczy Kampinoskiej* (Sum.: Some phenomena accompanying aeolian processes on the dunes of the Kampinos Forest), *Przegl. Geogr.*, 37,1.

http://rcin.org.pl

Schytt, V., 1959, The glaciers of the Kebnekajse-Massif, *Geogr. Ann.*, 41.

Szupryczyński, J., 1963, *Rzeźba strefy marginalnej i typy deglacjacji lodowców południowego Spitsbergenu* (Sum.: Relief of marginal zone of glaciers and types of deglaciation of southern Spitsbergen glaciers), Prace Geogr. IG PAN, 39, Warszawa.

Szupryczyński, J., 1965, Zagadnienie genezy krajobrazów wysoczyzny morenowej płaskiej i falistej (Sum.: The problem of the origin of landscapes of a flat and undulating morainic plateau), *Przegl. Geogr.*, 37, 1.

Szupryczyński, J., 1968, *Niektóre zagadnienia czwartorzędu na obszarze Spitsbergenu* (Sum.: Some problems of the Quaternary on Spitsbergen), Prace Geogr. IG PAN, 71, Warszawa.

Thorarinsson, S., 1943, Oscillations of the Iceland glaciers in the last 250 years, *Geogr. Ann.*, 25.

Thorarinsson, S., 1956, *The thousand years struggle against ice and fire*, Reykjavik.

Thorarinsson, S., Einarsson, T., Kjartansson, G., 1959, On the geology and geomorphology of Iceland, *Geogr. Ann.*, 41.

Todtman, E. M., 1957, Am östlichen Rand des Bruarjökull, Nordrand des Vatnajökull (Island, August — September 1956), *Neues Jb. Geol. Paläont.*, 7/8, Stuttgart.

Todtman, E. M., 1960, Gletscherforschungen auf Island (Vatnajökull), *Univ. Hamburg, Abh. Auslandsk.*, Reihe C, 65, 19.

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GEOMORPHOLOGICAL AND GEOLOGICAL ANALYSIS OF THE PROGLACIAL AREA OF THE SKEIDARÁRJÖKULL.

EXTREME EASTERN AND EXTREME WESTERN SECTIONS

Kazimierz Klimek

THE EXTREME EASTERN SECTION OF THE SKEIDARÁRJÖKULL FOREFIELD

The investigations covered the extreme eastern part of the Skeidararjokull foreland which for a distance of some 10 km borders the glacier snout. As in the case of the entire glacier lobe, this eastern part has also been retreating during the past 200 years, interrupted by periods of standstill and, even, of re-advances (Thorarinsson 1943, Eythorsson 1949). However, the rate of this retreat was several times slower than in the middle and western part of the lobe. This was due to the greater forces active in the eastern part of the Skeidararjokull, where a much more intensive movement of the ice mass fairly successfully balanced the wasting of the glacier caused by the ablation of its front.

During its maximum extent known in historical times — in the middle of the 18th century — the snout of the eastern part of the glacier was probably some 2 km in advance of its position in 1965. In 1890 the glacier snout was at a distance of about 500 to 700 m, and in 1955 of about 200 to 400 m from the position of the year 1965. In this case every re-advance of the glacier front, even a minor one, must have resulted in that ice margin reaching moraines previously laid down, or even overruning them. Thus in 1945, over a distance of about 1.5 km the glacier snout overran its 1904 position. The *jökullhlaup* (floods) which often affected the outwash plain of the River Skeidará, also had an influence on the relief forms of this part of the Skeidarárjökull foreland.

Owing to these circumstances the principal geomorphological units distinguished in the western and central part of the Skeidarárjökull foreland assumed a different arrangement in the eastern part of this foreland. Completely absent here is the so-called "zone of elongated depressions" appearing further westward; the "inner zone of marginal deposits and marginal land forms" (Galon 1970) here takes the shape of a deep (30–40 m) but narrow depression adjoining the steep glacier front (Fig. 1). At the bottom of this depression numerous small lakes occur. The largest of them borders on the western boundary of the investigated region. At many points this ice-marginal depression disappears completely, the glacier snout then directly joining the main end moraine zone. Here and there, particularly in the extreme eastern part of the foreland, this main end moraine zone is dissected over its whole width by large meltwater gaps, or else has been completely eroded (Fig. 2). In such places the youngest outwash fans begin directly at the glacier snout. These facts make it difficult or almost impossible to correlate individual mo-

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raine ridges and consequently to reconstruct the successive positions of the glacier front. The surviving moraine ridges lying farthest off and situated at identical distances from the glacier snout may be of different ages.

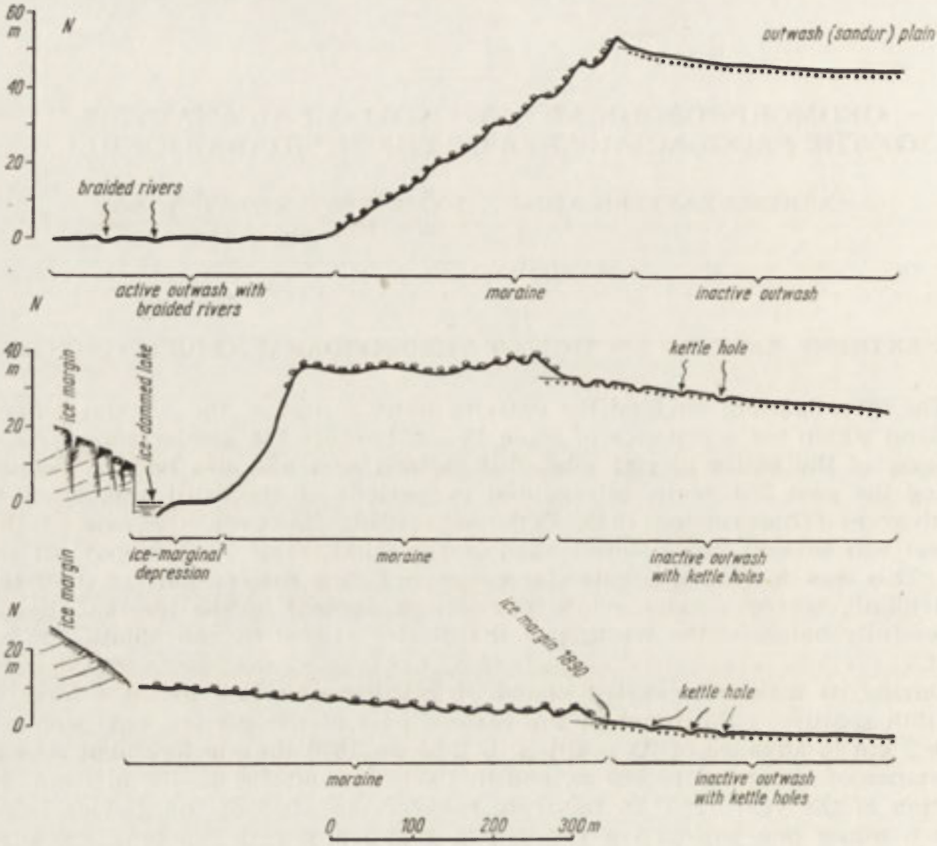

Fig. 1. Relation of the end moraine to the outwash (sandur) plain in different parts of the eastern Skeidarárjokull foreland

THE MAIN END MORAINE ZONE

To the oldest end moraine forms are assigned some low ridges, no more than 3 m high, which presumably date from the end of the 19th century (1890). They have been preserved at some places in the foreland of the younger, usually slightly higher, end moraines. These ridges with rounded tops have relatively gentle slopes, up to 20°. They are covered by gravels and boulders showing traces of weathering and block disintegration, although they are not basalts. Abutting onto these older ridges are younger moraine ridges, usually slightly higher (3–5 m) with sharp crests and steeper slopes, up to 30°. In some places it was found that younger moraines encroach upon older moraines, crossing them at low angle.

Three basic morphological types of end moraine zones were distinguished there. In the western part of the investigated section of the foreland, where the deep depression mentioned above borders on the glacier snout, the end

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morainic ridges — apart from ice-morainic hillocks developing in modern times — form the southern slope of this depression. Parallel ridges here form a step-like feature descending towards the ice margin (Fig. 1). The crest of each of the subsequently younger ridges lies at a lower altitude than that of the older ridge. The oldest of these ridges, slightly receding from the upper rim of the depression towards the foreland, rises 3 to 5 m above the outwash plain adjoining from the south and 45–50 m above the bottom of the depression in its rear. These ridges are mostly asymmetrical. Northward, towards the depression, their ice-contact slopes are steep (30–40°) up to 6–10 m high, whereas the opposite sides, much less high, have gentler gradients so that locally they even pass into a type of terrace, several metres wide. The crests of the older moraines occurring above the upper rim of the depression extend mostly in straight lines, while the younger ridges are wavy and sinuous. This would mean that at the time when these ridges were deposited, the front of the glacier, which by then had reached the northern slope of the initiating depression, had been dissected by radial crevasses in which moraine material was being accumulated.

In the extreme western part of the region investigated the younger morainic ridges are less distinct. Into their steep sides benches of abrasion terraces were carved. These originated owing to the gradual lowering of the lake water level which in 1945–1946 was occupying the entire depression in front of the glacier snout. By now it has dwindled into a miniature lake in the western part of the depression. Clearly marked benches of abrasion terraces occur at altitudes of 5–6 m, 8–9 m and 19–20 m above the water level of the above lake (Aug. 2, 1968); less distinct intermediate terrace steps can be seen at a height of 13 m and 16 m (?).

Eastward, where this lake is narrowing, the abrasion benches become less distinct or disappear altogether. Here at the base of the stepwise descending end moraine zone, fragments of a glaciofluvial terrace, some 100–150 m wide, can be observed rising 10 to 15 m above today's bottom of the depression. This terrace, built of well rounded gravels and boulders up to 0.50 m in diameter, is the remnant of the old dissected bottom of the depression.

Farther eastward the depression adjoining the ice margin narrows to 100–150 m. Its lowest parts are filled by small ice-lakes. From the side of the foreland the depression is bordered by a steep ice-contact slope, 20 to 30 m high. At many places this slope was undercut by streams flowing parallel to the glacier front. This resulted in a very steep erosion cliff (Fig. 1). The end moraine zone extending above its upper rim is an undulating plain covered by large boulders and blocks. Shallow thaw thinks 1–2 m deep occur there. They are some ten to some twenty meters in diameter and filled with sand and silt. The outer boundary of this moraine zone is marked by a moderately high (2–3 m) moraine ridge. In many places this zone passes without a distinct boundary into the outwash plain.

Where the ice-margin depression disappears, the glacier foreland has a general southward slope. Several parallel moraine ridges 0.5 to 1.0 m high occur here separated by depressions a few meters wide. Only the oldest among the ridges of this zone which in places encroach onto the remnants of the moraine from the year 1890, attain a height of 3 to 4 m (Fig. 1). Their shape is asymmetrical: the long and gentle slope of 15 to 20° turned towards the glacier shows traces of erosion, while the opposite slope is steeper, up to 30°, and bears the features of a talus heap. In its widest part this zone numbers 19 moraine ridges of this type. Thorarinsson (1967) considers these evenly shaped

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morainic ridges to be annual moraines formed between the years 1943–1963. However, this seems rather unlikely. An age of these ridges as young as this would presuppose a re-advance of the glacier front up to the line of the 1904 moraines to have taken place before the 1940s, followed beginning in 1943 by a rapid retreat at the rate of some 40 to 50 m per year. From an analysis of maps and from earlier papers by Thorarinsson (1943) and Eythorsson (1949) it appears that in the 1904–1920 period this part of the Skeidarárjokull was stagnant. In the years 1932–1947 it was gradually receding at a mean annual rate of 24 m.

Thus it seems more probable that the group of end moraines discussed has been deposited during a much longer period corresponding to the general stagnation before 1920 (higher inner morainic ridges).

East of the zone of regular moraine ridges, a five kilometer section adjoining the glacier front shows remnants of eroded end moraines. However, even with this evidence as basis it is difficult to reconstruct the successive retreat stages of the glacier front (Fig. 2; see the annex).

The transition of the end-moraines into the outwash plain shows a diversified pattern. Rarest are instances when a moraine ridge is superimposed on an older outwash and when the distal talus slope of the moraine ridge sharply abuts on the flat plain. This way of depositing a moraine ridge might have taken place at the time when, at the temporarily stagnant glacier front, the material that melted from the glacier was accumulating, and when the meltwater streams were flowing inside the glacier parallel with its margin and issuing into the foreland at some lower sites. This situation also occurs nowadays at some points of the glacier front as the result of some particular relief of the glacier substratum and crevasse pattern. At other places the moraine ridge with its gentle slopes passes into a series of small transition fans which form a subslope zone of some 50 m in width. Next to the base of the moraine this zone consists of coarse and poorly sorted material. With increasing distance from the moraine it passes into more fine-grained and uniform deposits showing a distinct fluvial structure. The small transition fans and the adjoining morainic ridges are also superimposed on the older glaciofluvial deposits. This mode of accumulation occurred at places, where small meltwater streams drained away from the ice covered by a thick mantle of the ablation moraine. Meltwater caused the moraine mud to move readily and to accumulate at the scarp foot in the form of a flattish moraine ridge. The erosion of the ridge by these rivulets led to the formation of the steeply inclined subslope zone. In both instances the outwash plain extending in the foreland of end moraines is older than the small fans.

Another way in which the end moraine passes into an outwash plain is the occurrence of a number of meltwater gaps in the moraine ridges. From these gaps the elongated outwash fans originate. Farther in the foreland they unite to form a plain (Fig. 2). These gaps varying in width from several to some twenty and more meters, show an unusually diversified relief. Thus conclusions can be drawn on the differences in the way the glacier snout was drained. While the glacier margin was stagnant at the line marked by a definite moraine ridge, the meltwater streams escaping from the foreland obstructed in some places the accumulation of moraines. Later, when the glacier had retreated and come to rest along the line of a younger end moraine, similar gaps developed in it. On their way into the foreland the meltwater streams no longer passed through all the gaps previously cut into the older moraine ridge, but they used only some of them lying at the lowest level. The con-

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centration of all the running water into a smaller number of streams increased their discharges and the energy. Hence, draining away from the end moraine zone into the foreland, the streams were doing more effective work than before. The result was a widening and deepening of the inherited gaps.

This is why in the moraine foreland these small outwash fans associated with younger stages of glacier stagnation were inserted into the older glaciofluvial covers. Under favourable conditions the dissection of the older floors of inherited gaps resulted in the formation of terrace benches on their slopes.

Successive withdrawals of the ice-margin marked by successive lines of end moraines, were accompanied by a decrease in the number of active gaps. At the same time the volume of throughflowing water was constantly growing larger. This caused the few active gaps to grow wider and deeper. As a result only one or two terrace levels have survived in the gaps, although traces are evident that meltwater streams have drained away from further younger moraine ridges into the foreland. Meltwater escaping from these longest active gaps eroded narrow valleys, 10 to 30 m wide, in the nearest foreland. Some 1 or 2 km away from the end moraines outwash fans were formed at the valley tract (Fig. 2). The continuous wasting away of the ice margin caused the gradual disappearance of direct drainage and the development of large marginal rivers, which became the connecting links between the system of ice-marginal lakes. These rivers penetrated the foreland at only a few places, carving impressive outwash (sandur) tracts. The gaps eroded by these rivers are sometimes more than 100 m wide, and their flat floors are bordered by slopes up to 15 m high. From this it appears that the deepening and widening of definite gaps in the moraine ridges, and hence the insertion of younger covers into the dissected older deposits in the foreland, has been the result of a concentration of discharge of melt-waters due to changes in the drainage pattern of the glacier front. Climatic changes which increased ablation, or local changes in the shape of the glacier substratum have been of secondary importance in this region.

OUTWASH FANS AND MELTWATER VALLEYS IN THE FORELAND OF THE MAIN END MORAINE ZONE

In the foreland of the main end moraine zone there occurs an outwash plain, in the broad usage of this term. This plain consists of several gently southward slanting erosion surfaces, erosion-accumulation or accumulation, extending along the base of moraines of different heights.

Some 1 to 2 km from the end moraines, the outwash plain constitues one level, developed by superimposition of differentiated accumulation covers on each other. In spite of its apparent monotony, the relief of this plain is varied (Fig. 3). One sees here traces of dry channels and thaw basins of various sizes. The dry channels appear as shallow grooves with erosional pavements. In places the channels are filled with fine-grained sandy material. Within the spillways or the elongated outwash fans, where the meltwater used a few channels, the dry channels are wider and deeper. Central bars have been slightly transformed by aeolian processes.

Kettle holes occurring on the outwash plain are funnel-shaped and a few meters in diameter, or else they represent large depresions of irregular shape and uneven bottom. The funnel-shaped kettle holes are mostly concentrated in the source area of the different outwash sheets, but they may also be enco-

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Fig. 3. A part of an inactive sandur in the central part of the Skeidarárjökull foreland. Photo by T. Konysz

Fig. 4. Fresh kettle formed between two stream channels. Photo by K. Klimek

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untered some 3 to 4 km away from the oldest moraines. The kettle holes can be observed in different stages of evolution, no matter how old the cover in which they occur. Some are inactive, others are still widening and deepening (Fig. 4). Here the size and the depth of the buried dead ice blocks is decisive. If the kettle holes occur at a greater distance from the end moraine zone, this means that the buried dead ice blocks must have been carried onto the outwash plain by meltwater streams. This could have happened only during the catastrophic floods which usually occur every ten years and are caused by the rapid drainage of either ice-dammed or englacial lakes. Then huge blocks are being deposited on the outwash plain (Thorarinsson 1954). The large thaw thinks occur where the moraine ridges are cut through by outwash tracks. Thaw thinks are a frequent occurrence in the gaps. They are due to the melting of dead ice blocks which became detached from the retreating main ice mass and are buried under glacial deposits. Thaw thinks developed by the time the gaps were still drained and later when the flow has ceased. They also develop nowadays. The former are characterized by rounded rims and flat floors lined with silts and small deltas extending from the sides towards the centre of the depresions (Fig. 5). The latter type shows very irregular outlines, sharp rims and an uneven floor with differences in altitude. Very often one encounters thaw thinks that originated as water still flowed. Now they are deepening under present-day subaerial conditions.

Fig. 5. Large silted up kettle. Photo by K. Klimek

INTERRELATION AND CONDITIONS OF ACCUMULATION OF OUTWASH COVERS IN THE EASTERN PART OF THE SKEIDARÁRSANDUR

In the eastern part of the Skeidarársandur five accumulation covers of different ages were distinguished (including the contemporaneously developed sandur tract of the River Skeidará). These five covers lie at several levels and withdraw towards the foreland one outwash plain. In regions where the end moraines have been completely washed out, it is very difficult to establish the

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relative ages of adjoining covers which sometimes lie at identical levels. Therefore only studies of the course of dry channels (Fig. 2) and, partly also, of the differences in the grain size of the deposits make it possible to define the boundaries between different covers. Each of these covers shows a slightly different pattern of dry channels; slight differences in the granulometric composition of gravel and boulders also occur, some noticeable even with the naked eye. The deposition of material proceeded here under somewhat different hydrodynamic conditions, and this affected both the drainage pattern and the size of the material transported.

The oldest among the preserved outwash covers is cover I. It consisted of a number of extensive fragments, the largest of which is more then two kilometers wide and several kilometers long. From the north, flat fragments of this cover begin at the base of the oldest end moraine ridge surviving in this region, or else they are bordered by scarps 8–10 m high. The scarps were eroded by waters forming the lower outwash levels. With increasing distance from the end moraines the height of scarps rapidly decreases, so that some 1 to 1.5 km off they thin out completely. The distal slopes of the moraine ridges, bordering from the north upon the these outwash plains are talus slopes. This suggests that the ridges are younger than the fragments of cover I spread at their bases. They were probably laid down towards the end of 19th century (1898?), but may also be slightly younger and correspond to a transgression which took place at the beginning of the 20th century. An abundance of funnel-shaped kettle holes can be seen within cover I; none of them appear in the adjoining younger covers, which often lie at the same altitude. This suggests, that to the deposition of these younger covers *jokullhlaups* must have contributed, these floods carried large ice blocks which were subsequently buried partially or completely in the foreland. It is difficult to define the ultimate period in which the cover originated. From a paper by Thorarinsson (1943) it appears, that in historical times the eastern part of the Skeidarárjökull lobe has reached its farthest postglacial extent in the latter half or the 18th century. At this time the glacier margin advanced some 1400 m beyond the position it had in 1904 and prior to 1965. On a map from the year 1904 two elevations are marked. These are remnants of this moraine chain. But no traces of it can be seen at the place where contemporaneously fragments of cover I appear. Hence one may conclude, that the catastrophic flood which led to the destruction of the end moraine zone marking the farthest glacier advance and contributed to depositing cover I, took place between 1750 and 1904. This may have been one of the *jökullhlaups* of the end of the 19th century, or even that of 1903.

Cover II is twofold. It borders over a considerable distance on the base of the younger end moraine ridges (IIa). It is also represented by a series of narrow outwash fans in valleys issuing from gaps that existed in the older moraine chain (IIb). After breaching the older glaciofluvial covers (at a higher altitude) these outwash fans at a distance of 1.5 to 2 km from the moraines, cover the glaciofluvial deposits of cover I. This case occurs when in the main end moraine the older ridges were destroyed, and the outwash plain begins at the base of the younger moraines. In the latter case the meltwater streams emerging from the glacier breached the older end moraines preserved ahead of them. Flow concentrated in the narrow gaps caused their deepening. This is the reason why cover II occurs at two levels. In the extreme eastern part of the area discussed, cover IIb is bordered along the ice margin by scarps which were eroded by water flowing at a lower level. Over a distance of more

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than 2 km these waters have destroyed the pre-existing end moraines. In this zone, within the surviving fragments of cover II, fairly numerous large kettle holes appear. They can be found even at 1 to 1.5 km distance from the assumed extent of the end moraines. This indicates that, in this part of the foreland of the end moraines, cover II has, in part at least, been laid down with the co-action of a *jokullhlaup*, whose waters at the same time caused the formation of the 2 km gap in the main end moraine chain.

On the 1904 map, at the place where fragments of cover II appear at present, two moraine elevation (the higher one at 101 m above sea-level) are marked, representing remnants of a moraine chain dating back to about 1750. According to S. Thorarinsson (1939), these elevations were completely destroyed by the 1922 *jokullhlaup*. From this it might be concluded that during the same period the remainders of cover II were, in part at least, laid down or transformed, and therefore originated in the years 1920–1922. Cover III does not relate to any of the preserved moraine ridges. It consists of a number of narrow outwash cones beginning in the gaps, some 50 to several hundred metres wide, which dissect both the main end moraine zone and the older glaciofluvial covers extending in their foreland. Along the ice margin the floors of these gaps are hanging above the deep ice-marginal lake basins. At many places the edges of the gaps are crooked. This shows that the throughflowing streams were meandering. Gaps hanging above the ice-marginal lake basins indicate that the lakes did exist at the time when this cover was being laid down along the eastern margin of the Skeidarárjokull. Only in the extreme eastern part of its foreland on the interfluve between the Eastern Skeidará and the Western Skeidará, an extensive fragment of cover III has survived. It is being undermined by the river from all sides. Cover III contains a great number of kettle holes several meters in diameter. This indicates that *jokullhlaup* must have largely contributed to their formation. An American map drawn from air photographs taken in 1945/46 shows in this part of the Skeidarárjokull foreland the embryonic stage of the ice-marginal lakes existing today. From only one large and elongated lake, situated more to the west, a small stream was flowing out into a wide gap formed prior to 1945. From these facts it would appear, that the various isolated fragments of cover III were laid down in the decade 1940–1950. The lowest of these fragments occurring on the interfluve between the Eastern Skeidará and the Western Skeidará was transformed by a *jokullhlaup* in 1954.

Soon after cover III came into existence the direct drainage onto the foreland ceased to function. Instead of this streams running parallel with the ice-margin developed. From the ice-marginal lakes part of the waters escaped westward to the Sandgigjukvisl system, and another part eastward into the Skeidará river. Traces of these flow directions are the youngest of the now dry outwash tracks. In the dry outflow channel from a large lake which existed in 1945–1946, a fragment of cover IVa has survived. This lies 20 m below a gap eroded by water issuing from the lake. The waters from the lakes situated farther east at that time joined the Western Skeidará. A trace of this flow can be seen in the lowermost gaps in which the narrow outwash fans begin — they represent cover IVb. On the divide between the two river systems until 1968 there was a lake from which a small brook flowed out into the foreland.

The recently formed cover V represents the sandur tract of the River Skeidará. The characteristics of this tract will be presented together with that of the Sula sandur tract when discussing fluvial processes.

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TENTATIVE DISCUSSION ON THE COURSE OF DEGLACIATION

The nature of the glacial landforms and their interrelations indicate that along the ice margin frontal deglaciation occurred. The low mobility of this part of the glacier lobe was the reason why its front retreated at a very slow rate. Higher moraine ridges correspond to periods of longer stagnation and ridges less high to periods of more rapid retreat. Soon after the marginal depression came into existence the glacier front rising above the southern sides of this depression began to waste away very slowly, depending on the rate of surface ablation. During this long period a group of stepwise arranged morainic ridges was formed.

THE EXTREME WESTERN PART OF THE SKEIDARARJÖKULL FOREFIELD

In this part of the proglacial area the investigations covered the area along the Skeidarárjökull margin from south and west. During the past 200 years the western part of the glacier retreated at 4 to 5 km, this retreat taking place at a varying rate. Periods of intensive decay and thinning alternated with periods of stagnation (1932–1939) or transgression (1840–1950, and early part of the 20th century). As compared with the eastern part, the high rate of diminution of the western part of the glacier indicates a predominance of ablation over the supply of new masses of ice — a fact probably resulting from a slower and different mode of motion of this part of the glacier. The farthest extent of the western part of the Skeidararjokull in historical times — and probably during the Holocene — was reached around the year 1750. Thorarinsson (1939) believes, basing his opinion on reports by Olafsson and Pallson who in 1756 crossed the Skeidarársandur, that at those times the ice margin lay against walls of the Lomagnupur massif. In the ice-dammed valley of the Nupsa river a lake (Nupsalon) has formed, which at its maximum extent covered a surface of 5 km^2 and held some 100 million m^3 of water.

Traces of this wide range of the glacier snout can be seen in the oldest end moraines in the western Skeidararjokull foreland. On the basis of analyses of the lichen *Rhizocarpon geographicum* they may be dated as being deposited some 200 years ago (Bogacki). This confirms Thorarinsson's opinion in this matter. The westward continuation of this moraine chain is an elongated erosional residual of moraine mound, situated in the centre of the Sula outwash plain and being undercut by branches of this river. A Danish map from 1904 shows two further moraine mounds of similar size (the upper one 108 m above sea level) lying close to the Lomagnupur rock walls, some 2 km west of the present Sula ice-spring. By 1968 these two mounds had disappeared.

THE MAIN END MORAINE ZONE

Lateral migration of the Sula branches prevented the conservation of older end moraines in this part of the foreland (Figs. 6 and 7). On the 1904 map the end moraines dating back to about 1890, have almost completely been washed away. The small erosional residual which survived on the western bank of the main Sula channel may be a remnant of this ridge. Slightly younger (after Bogacki) and deposited later than 1900 are residual moraine hillocks situated in the southern part of the investigated region, and until recently undercut by one of the Sula branches.

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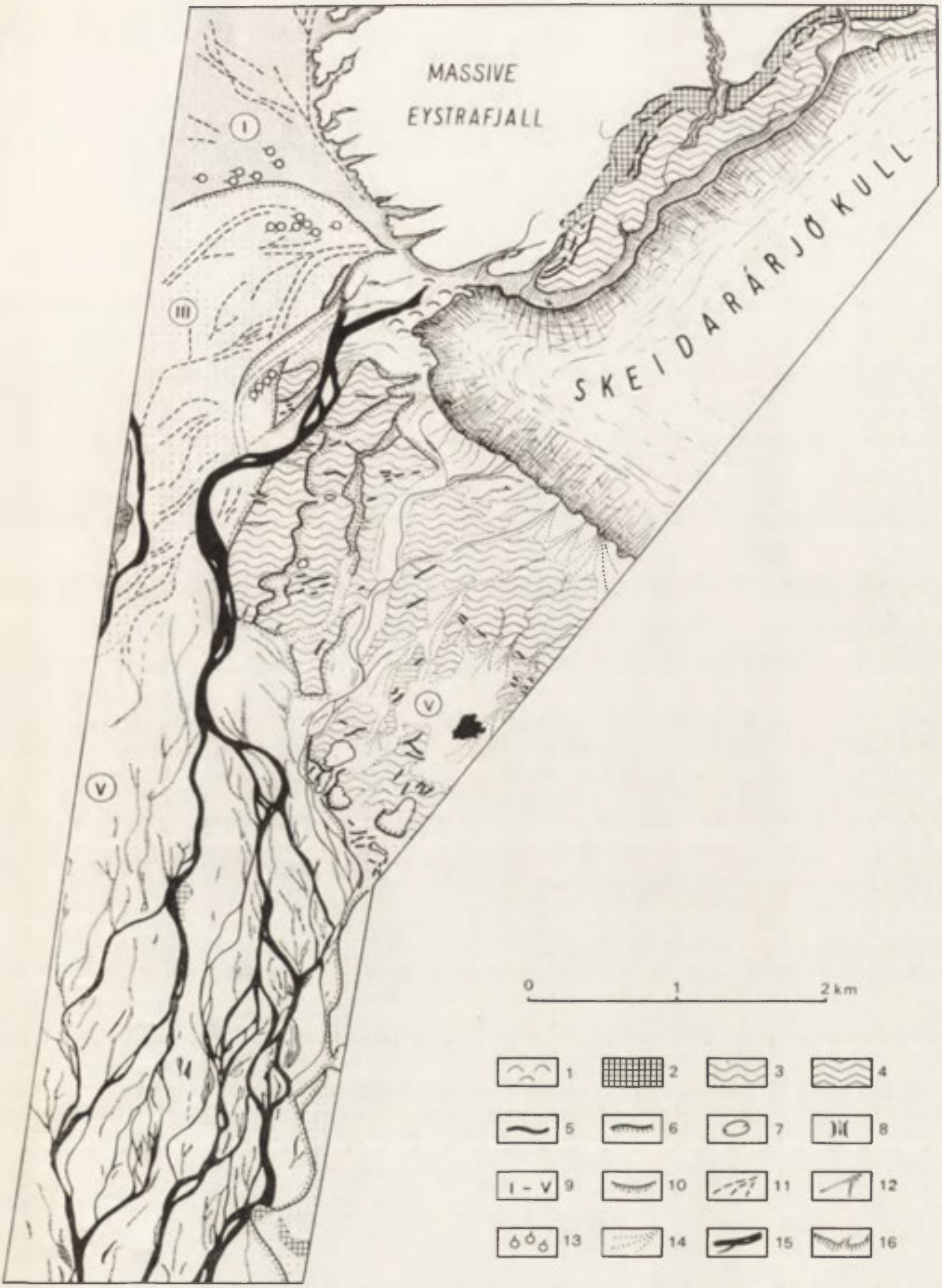

Fig. 6. Geomorphological sketch to show the northern part of the Sula outwash (sandur) valley-tract

1 — ice-moraine hillocks at the ice margin, 2 — moraine ridges (in the Eystrafjall massif) or residual hills (in the foreland of the glacier) of end moraines, 3 — undulating end moraine forming a "moraine plateau", 4 — hummocky end moraine in the Eystrafjall massif, 5 — ridges of more important moraines, 6 — ice-contact slopes, 7 — large thaw thinks within end moraines, 8 — "gates" eroded in the moraine zone by meltwaters, 9 — glaciofluvial levels of different ages, 10 — meltwater cliffs, 11 — dry channels of proglacial rivers with pavements preserved at bottom, 12 — dry channels of proglacial rivers filled with sandy-silty sediments, 13 — "gravel shadows" deposited behind ice-blocks in flowing water, 14 — small sandy-silty fans deposited at the ice margin, 15 — channels carrying water in 1960, 16 — rocky walls

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Fig. 7. Sula branches downstream the ice-spring. Photo by K. Klimek

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Fig. 8. Sula sandur in the gap. Photo by K. Klimek

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East of the main Sula channel the main end moraine zone bears the features of a rather monotonous undulating morainic plateau (Fig. 8). From the west, over a distance of more than 2 km, and from the south this plateau is bordered by scarps, 12 to 15 m high, undercut at present, or recently, by the Sula branches (Fig. 9). Eastwards the plateau is separated by a distinct ice-con-

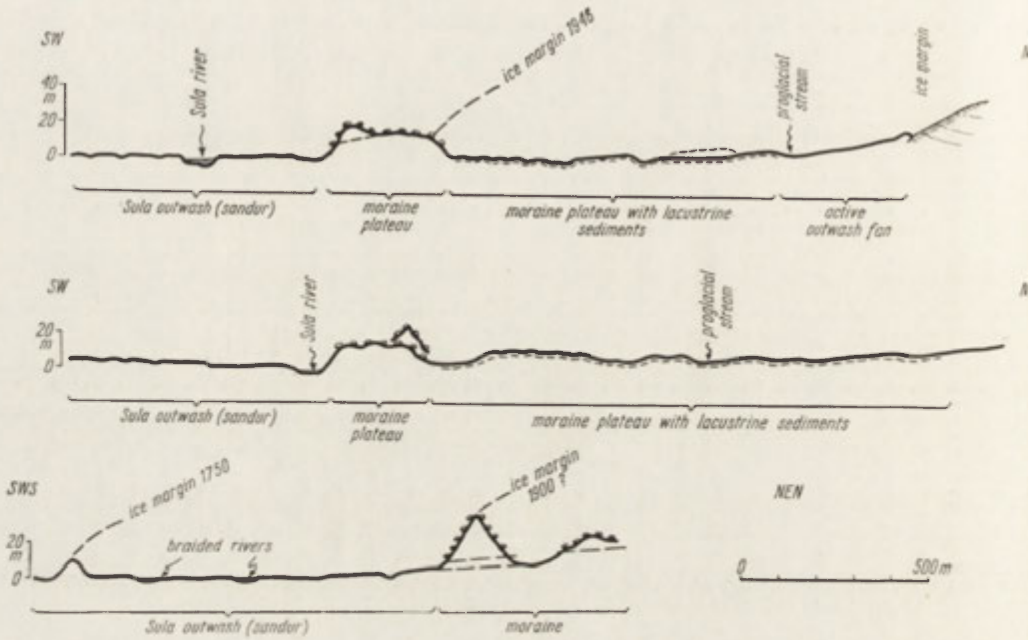

Fig. 9. Relation of the end moraine to the outwash (sandur) plain in different parts of the western Skeidarárjökull foreland

tact face, 10 m high from lower ground. The plateau surface shows undulating relief with altitude differences averaging 2 to 3 m (Fig. 9). Much space is also occupied by minor flattish-topped mounds. In the active Sula cliff it can be observed, that on the well rounded boulders and gravel (glaciofluvial deposit) rising up to 6 m above the river level lies sand and silt. In places it contains gravel and boulders. This deposit shows traces of a fine lamination, indicating that it was deposited mostly in standing water. On the other hand, the surface of higher elongated mounds is strewn with big blocks, and in hollows a sandy-silty material has accumulated.

Near the western (undermined) edge of the moraine plateau small dry valleys begin, the mouths of which are hanging above the Sula river. These are the upper reaches of small valleys carrying part of the meltwater that was escaping westward at the time when the ice margin was lying against the eastern edge of the plateau. At a few places on the eastern ice-contact slope of the plateau isolated moraine hillocks occur, 5 to 10 m high, increasing here and there the height difference to as much as 20 m. These hillocks mainly consist of stratified sand or silt capped by boulders. This suggests that these are crevasse fillings deposited in the stagnant ice mass.

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The study of the 1904 and 1946 maps leads to the conclusion, that the above described undulating moraine plateau must have been formed between 1900 and 1946 under conditions of a stagnant glacier, with a predominance of surface ablation.

THE OUTWASH FANS IN THE FORELAND OF THE MAIN END MORAINE ZONE

Between the western margin of the main end moraine zone above described and the Lomagnupur massif in the west and Eystrafjall in the north-east there lies an outwash plain due to erosion and accumulation. For the most part this plain consists of gravels and boulders up to 30 cm in diameter, in many places assuming the features of erosional pavements. Three glaciofluvial covers of different ages, clearly to be distinguished here are including the Sula active outwash tract. The radial arrangement of the dry channels indicates that while this plain was forming, the Sula River was continuously issuing from the same place, being the principal proglacial stream in this region. Near the big ice--spring of the Sula these channels are 3–5 m deep and 40–60 m wide, but with increasing distance they turn into shallow channels some 0.5 m deep, distinctly marked in the outwash plain. At the floors of the better preserved dry channels large boulders and blocks occur (moraine residuum). Even large boulders up to 50 cm in diameter show an imbrication which indicates high velocities of meltwater.

Another type of land form seen here are thaw kettles, up to several meters in diameter, containing gravel-boulder "shadows" on their downstream sides. The latter forms originated by rapid sedimentation of bed load material behind stopped ice blocks in running water. Conditions favouring this kind of processes prevailed during the *jokullhlaups* being a frequent occurrence in the Sula sandur tract. Large parts of this outwash plain are then flooded. At times like these, large ice blocks used to break off from the glacier margin, and to travel far into the foreland of the end moraines.

INTERRELATION AND CONDITIONS OF ACCUMULATION OF OUTWASH COVERS IN THE WESTERN SKEIDARARJÖKULL FORELAND

The oldest and the highest glaciofluvial cover I is squeezed into the funnel-like Nupsa valley, between the Lomagnupur and the Eystrafjall massifs. Traces of dry channels left in this cover show that the Sula river and its branches used to run west- and northwestward (Fig. 8). It is only after reaching the undercut walls of the Lomagnupur massif that the Sula and Nupsa waters joined and turned southward. An outwash fan occurring in a zone which up to 1756 was occupied by a lake indicates that this fan is not of an earlier date. On the other hand, it did exist prior to 1904 because — although in a much larger size — it is already marked on a map from this period.

Large kettles with gravel-boulder "shadows" occurring within this cover point to the considerable share of one of the *jokullhlaup* in the final stages of its fashioning. Hence this cover corresponds in age and origin to cover I distinguished in the western part of the Skeidararjökull foreland (Fig. 6).

The cover of a lower glaciofluvial level which southward adjoins cover I, has also taken a long time to be deposited. As mentioned above, where this cover now lies the 1904 map shows in front of the Lomagnupur massif two moraine mounds from the middle of the 18th century; while on the 1945–1946 map they are no longer marked. The arrangement of the dry channels indicates that at that time the Sula branches trended in western and south-wes-

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tern directions. Within this lower cover thaw thinks with gravel-boulder "shadows" occur. Hence in the formation of this cover *jökullhlaups* have also played a large part, and this may refer to any one of the floods which occurred in 1913, 1922 and in 1934. Thus the latter cover must be considered as a counterpart to covers II and III distinguished in the eastern part of the Skeidarárjökull foreland.

This youngest Sula outwash plain formed at present day will be discussed later in the chapter dealing with present-day fluvial processes.

THE INNER ZONE OF MORAINE DEPOSITS AND LANDFORMS

The moraine plateau terminates eastward above an ice-contact slope some 10 m high (Fig. 9). From the west the plateau is bordered by a lower undulating plain extending as far as the ice margin in the north. Above this undulating plain elongated mounds rise 1 to 2 m high, covered with gravels or boulders. These are separated by extensive somewhat lower grounds covered with sandy-silty deposits. The stony elongated mounds, running at right angles to the ice front, have the appearance of crevasse fillings and resemble small eskers. Northwards the materials building this plain merge beneath deposits of small, sandy-silty outwash fans laid down by minor meltwater streams flowing from the glacier snout. At many places these fans have fully covered the underlying glacial deposits, and in this way changed the originally undulating plain into a flat one. This entire area has been freed from under the ice some time after 1946 and, judging from preserved landforms and deposits, wide areas of stagnant ice remained here. Glacial deposits to large extent have been laid down in an environment of standing ice-dammed water.

THE COURSE OF DEGLACIATION

The shape and structure of residuals of the main end moraine zone and the adjacent "inner zone of marginal deposits and landforms" indicates that deglaciation in this area consisted almost entirely of a gradual lowering of the surface of the stagnant ice dissected by crevasses. None of the new masses of ice were supplied to the glacier snout, so that conditions were unfavourable for the formation of larger end moraines. In many places ice-dammed subglacial basins must have existed where moraine material was accumulating in a water environment. Such a mode of deglaciation occurred because in its western part the Skeidarárjökull was less dynamic.

GLACIAL LANDFORMS IN THE EYSTRAFJALL MASSIF

The Eystrafjall massif borders the Skeidarárjökull to the west. It is built of palagonites, tuffs, breccias and basalts of the Quaternary Móberg formation (Thorarinsson 1959, Einarsson 1967). The dome-like field surfaces lie at an altitude of 550–850 m, and from there they fall in precipices to the glacier surface lying some 100 m lower (Fig. 6). On the slopes of these fields, above the rock walls, occurs a group of well preserved glacial landforms, which cover and area 400–700 m wide lying against the present ice margin (Fig. 6). Parallel sharp-crested and steep-sided (20–30°) lateral moraine ridges, 5 to 10 m high, are found here. These ridges, having the appearance of accumulation moraines, consist of angular blocks up to 3 m in diameter. In the moraine foreland where these ridges cross depressions, short and steeply inclined outwash fans have

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developed. In the dammed valleys the fan covers pass into the deposits of small ice-dammed lakes, thus forming peat plains. Where the substratum is steeply inclined and the meltwaters were unable to escape into the foreland, the outer side of the oldest moraine ridge has the features of a talus heap on bedrock. In the rear of the main moraine zone the bedrock is covered with a thin layer of moraine debris. In niche-like depressions at the base of precipes small kame terraces were observed. These are composed of sand-intercalcated gravels which were laid down in ice-dammed lakes, similiar to those which nowadays appear next to the glacier margin at a somewhat lower level. Along the glacier margin large blocks are piling up. These fall off from the rapidly weathering rock faces and build low pseudo-morainic hillocks.

The oldest moraine ridge rising 130 to 150 m above today's glacier level corresponds in age to remnants of the oldest end moraine from the middle of the 18th century occurring in the western foreland of the Skeidarárjokull. The three successively younger end moraine ridges, confined to a few hundred meters, were formed before 1904; among them are likely to be found moraines dating from the years 1840–1850 and 1890. The hummocky moraine landscape extending in the rear of these morainic ridges goes back to the years 1904–1946. Since 1946 the glacier margin lying against the steep rock walls has been lowered some tens of meters as a result of ablation, but in the horizontal direction it has moved only slightly.

From the fields towards the glacier, streams flow down fed by rainwater or meltwater from perennial snow patches. Crossing the zone of the postglacial landforms these streams dissected the glacial, glaciofluvial and glaciolimnic deposits, together with the bedrock. Canyons several scores of meters deep, with numerous waterfalls at the outcrops of basalt rocks occur. These incisions illustrate the intensity of the depth of erosion which has been operating here during the past 200 years.

FLUVIAL PROCESSES AND RELIEF OF THE ACTIVE OUTWASH PLAINS (SKEIDARÁ AND SULA)

In the western Skeidarárjokull foreland occurs the active Sula outwash plain. The Sula river emerges from a powerful ice-spring on the ice-margin and, breaching the remnants of the main end moraines, penetrates the foreland where it branches into a number of channels forming an outwash tract several kilometres wide. In the eastern foreland of the glacier the Skeidará sandur is active. The Skeidará river issues, like the Sula, from powerful ice-springs situated near the glacier margin. It carved its way through the remnants of end moraines and older outwash covers and flows into the foreland. Here, together with the Morsa waters it forms an outwash tract 4 to 5 km wide and many kilometres long. The active sandur tract with a typically braided-river pattern have in their upper courses steep and ungraded longitudinal profiles (for Sula it is 6‰) which gradually flatten downstream. Morphological processes taking place here, and the deposits and landforms due to the action of these rivers, are determined by the regime of proglacial rivers and the structure of the substratum. Rivers, whose drainage basins are at least 20% covered by glaciers, are characterized by a distinct glacial regime (Pardé 1954) being very similar for drainage basins of different geographical latitudes (Arnborg 1955; Axelsson 1967, Davidov and Pronin 1967; Bachurin 1960; Fahnestock 1963; Østrem *et al.* 1967). During the four summer months representing 33% of the year, the run-off from these drainage basis is 85–90% of the annual

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total. During the two months of maximum ablation (July, August), i.e., 17% of the year, the run-off amounts to some 40 to 70% of the yearly total. Such a river regime clearly affects the rate, the quantity and the size of the material transported. The amount of materials transported by rivers depends on their energy and this, in turn, upon the velocity, depth, gradient, and friction along the channel floor and sides. Every increase in discharge increases the energy of the river and the quantity of material transported. This quantity increases proportionally with the square of velocity of the streamwater (Hjulström 1935, Nevin 1946). Hence the increase in river energy caused by discharge concentrated in short periods makes it possible to transport much more material than could be carried out by a river with similar but uniform discharge throughout the year. The glacial regime also increases the competence of the river. The ability of rivers to transport a defined gravel size increases 2.6 power of velocity (Hjulström 1935; Nevin 1946). The concentration of discharge in a short period and the resultant increased competence of the river in a defined section enables the river to transport gravel of much larger sizes than those transported by rivers with identical annual discharges but a different regime.

Rivers in the Skeidararjökull foreland transport mineral materials either in suspension or as a bed load. The daily quantity of suspended material is usually very large and changes depending on the way a river is fed (outflows from ice-springs, ice-marginal lakes, glacier snout) (Fig. 10). On the average, large rivers issuing from ice-springs carry 1500–3000 g/m^3 of suspended matter, small ephemeral streams due to the ablation of the glacier front transport from 500 g/m^3 in the morning to 37,000 g/m^3 in the afternoon. Finally, rivers issuing from proglacial lakes carry less than 100 g/m^3. The turbidity of stream water increases with increased flow, as has been observed on all the proglacial rivers (Davidov and Pronin 1967; Fahnestock 1963, Østrem *et al.* 1967; Walker 1953), although in some cases the maximum turbidity may precede the maximum flow by a few hours.

Bed-load in a river is transported in different ways. Boulders of various sizes and gravel are rolled on the bottom during high water stages. Smaller gravels are subject to saltation, whereas fine gravel and sand is moved in the form of a bottom layer even at low water level. Where the current is slow, sandy materials are being moved in the form of ripple-marks. As a result of the diversity of fraction in the channel all forms of transportation may occur simultaneously in the given section.

In an active sandur system one observes a network of braided channels. Each of their sections may be in a different stage of development. All channels are incised into loose glaciofluvial deposits. Their characteristic features are straight courses of the particular sections, great width and small depth, and a marked symmetry in cross-section. The widths of channels varies widely: the small ones may be some 50 cm wide and a few centimetres deep, while the large channels are some tens of metres wide and several metres deep. In the longitudinal profile longer sections with gentle gradients alternate with short sections containing rapids. Along the channels it is easy to distinguish sections shaped either by erosion or aggradation, and sections of equilibrium where transport prevails.

For the most part, channel sections shaped by erosion are incised to a greater depth. On the floors of abandoned channels erosional pavements occur showing a more or less well developed imbrication (Fig. 11). In these sections bottom erosion starts in periods of dominant removal of materials

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when the supply of materials from the upstream section is less. This usually occurs when gradients are low or when the flow has been stabilized. The finer material is carried off, the coarser one being left *in situ* and reoriented in position. In this way erosional pavements are developing, which are not even washed away during short periods of increased velocity (Fig. 11). Over the

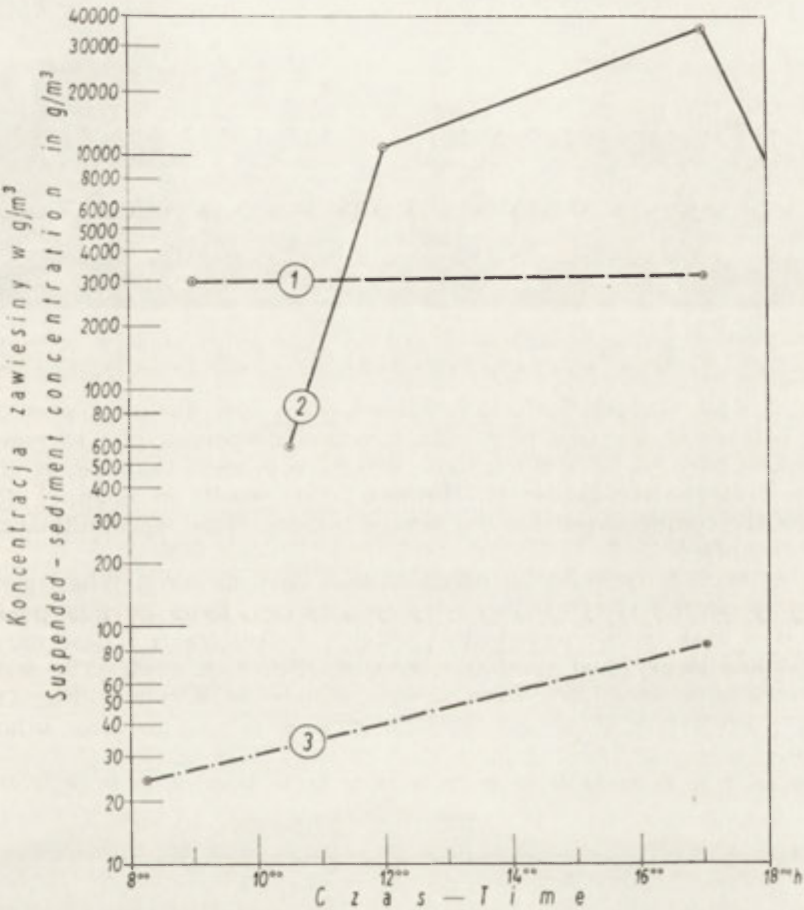

Fig. 10. Changes in the suspended load carried by proglacial rivers in the Skeidarárjökull foreland
1 — rivers fed by large ice-springs, 2 — rivers fed by the ablation of ice front, 3 — rivers draining from ice-marginal lakes

surface of erosional pavement the material from upstream sections may be moved, as was described by Arnborg (1957). The gravels in this pavement dip upstream and parallel to the flow, but along the river banks they may deviate up to 45° from this direction. This is caused by the repeated gravel reorientation when the water level is falling. In zones of rapids the shooting flow of water and the great kinetic energy cause a powerful erosion of the substratum, and hence the rapids are subject to parallel retreat. Downstream them sections of deeply incised channels occur, just like the gorges eroded in compact bedrock downstream the waterfalls. In dry sections of these deeply

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Fig. 11. Sula sandur. Boulder paved dry channel. Photo by K. Klimek

incised channels elongated erosion hollows occur, for the most part produced by the vortexes at the foot of rapids. Channel deepening due to recession of rapids seems here to be a much more effective process than depth erosion in channels with gentler gradients. However, the results of channel deepening are promptly compensated by the successive increased accumulation of material in this place.

In river sections shaped by accumulation bars develop. When growing in size they cause the rivers to branch and to stray. Even during peak floods interchannel bars can be seen in the active outwash tracts. These bars occupy a many times larger surface than the water-filled channels. The size of the bars varies in diameter from a few to some hundreds of meters (Fig. 12). These forms are in different stages of development — some are growing, others being degraded.

Fig. 12. Central bars on the outwash plain of the Skeidará. Photo by K. Klimek

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Field observations and experiments (Leopold, Wolman 1957) indicate that in straight river sections central bars are being formed in the channel axis. Here the water is overloaden with the bed-load material (Hjulstrom 1952). Furthermore, it may be the result of the reduced flow velocity caused by local shallowness (Krigström 1962). However, it is hitherto unknown how these forms develop. In the dry channels occurring within the Skeidarársandur plain bars in different stages of development were observed. Thus it was possible to study one of the ways of their formation. Bars are being formed when as a result of a marked increase in flow (velocity) the material occurring in the channel or eroded from its sides is transported in very large quantities along the channel bottom. The impulse to deposit the material at a defined place of the channel floor proceeds from vortex flow, most often caused by the presence of boulders which move at a much slower rate than the smaller gravel and sand. Downstream of these boulders where both horizontal and vertical backward vortexes occur, the transported materials are being deposited. Under favourable conditions this may cause an increase both in the size of the obstacle and in the vortex. When the flow velocities and the quantities of transported material are steadily growing, first a spindle-shaped bar is formed with its sharp end pointing upstream. It rapidly grows in width and downstream (Fig. 13). If bars start to develop near one of the river banks,

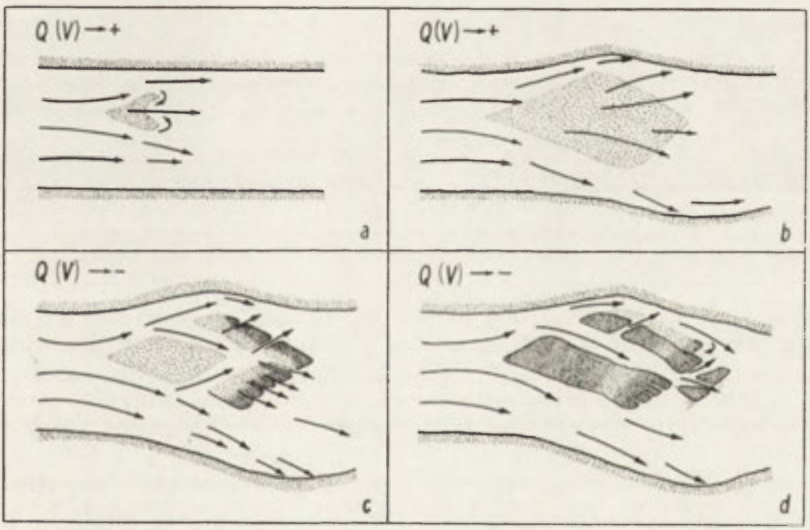

Fig. 13. Development of central bars in proglacial river channels (partly after A. Krigström, 1962)

they grow asymmetrically, producing protrusions oblique to the channel axis. The growth of bars in the channel axis drives the current towards the banks. The squeezed water body, flowing at an increased velocity, erodes the floor of the channel and undermines its sides. Usually this leads to a widening of the channel and to changes in its cross-section. When the water level is falling, the flow tends to concentrate in the lateral channels and this causes the bar to emerge from under the water surface. With falling water levels the bar's surface is subject to erosion. Then erosional pavements or linear incisions are formed as the result of water concentration in the individual streams

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(Fig. 14). Those channels which carry less water soon cease to function while others carrying more water cut up the bar into several parts. Materials derived from these incisions form small deltas on the downstream side of the bar which, when the water level continues to fall, merge into the older part of the bar thus increasing its size (Fig. 13).

Channel deepening due to current sliding over the bar surfaces is the cause of bank recession and channel migration. The glaciofluvial and glacial deposits forming the banks of proglacial rivers are loose and porous. When dry, they have a tendency to stand up in vertical walls, but after the liquefaction they are liable to turn into a flowing mass, especially when fine-grained fractions predominate. At low water level or when the dynamical axis of the channel

Fig. 14. Bars edge linearly dissected. Photo by K. Klimek

has moved near to the opposite bank, the retreat of the channel sides proceeds at a relatively slow rate. From vertical sides the first to be wind-eroded or washed out by rainwater is the fine-grained material; whereas the coarser material is falling off to form talus heaps at their bases (Fig. 15). Since during low water level only part of the fallen material can be swept away the talus heap continues to grow at the expense of a retreat and a shortening of the upper part of the cliff (Fig. 16). When the water level rises slightly and the lower part of the talus heap becomes flooded, the natural angle of repose of the liquefacted mass changes causing the material to slide off and to uncover the previously hidden part of the cliff. During high level the water rapidly infiltrates into the lower part of the wall and this causes part of the cliff-forming sediments to move readily and to disturb the equilibrium of the bank. There follows a breaking off of large packs of materials, and in this way the line of the river bank is pushed back. The material which slid into the water undergoes selective erosion. The fine fraction is carried away in suspension, the coarser part is removed as bed-load, and only the larger boulders remain *in situ* (Fig. 17). Removal of the materials contributes to the uncovering of the lower parts of the cliff, so that further layers of deposits become liquefacted and the bank settles again. Thus the principal cause of rapid proglacial river cliff retreat is the falling off and the slumping of loose cliff forming deposits which readily change into a liquid mass in their lower parts. This is a result of

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Fig. 15. Periodically inactive cliff of proglacial river. Photo by K. Klimek

Fig. 17. Residual boulders washed out from the end moraine. Photo by K. Klimek

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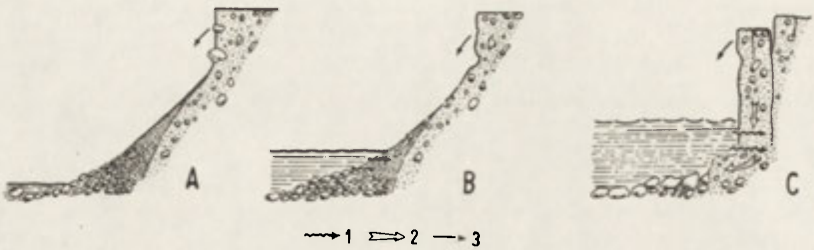

Fig. 16. The mode of retreat of channel-sides of proglacial rivers cut into sand gravel-boulders, depending on the water level
1 — infiltration of water from the channel into the sediments, 2 — slumping, 3 — falling off

the repeated and considerable changes in the water level. The formation of central bars combined with bank migration leads to rapid changes in the river network. A current channel which has started braiding and in which usually one of the branches carries a larger amount of water, causes continuous bank migration until the channel divides are destroyed and the branches join again. Periodical deepening of some sections is balanced by increased aggradation in the other ones. Thus it appears that the rivers draining into the Skeidarárjokull foreland are in a state of equilibrium in the sense used by Mackin (1948).

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REFERENCES

Arnborg, L., 1955, Hydrology of the glacial river Austurfljet, in: Hofellsandur a glacial outwash plain, *Geogr. Ann.*, 37.

Arnborg, L., 1957, Erosion forms and processes on the bottom of the river Angermanalven, *Geogr. Ann.*, 39.

Axelsson, V., 1967, The Laitaure delta, *Geogr. Ann.*, 49A.

Bachurin, G. V., 1960, Riyeka Kaindy (Kaindy River), in: *Miezhdunarodnyy Geofizicheskiy God*, Izd. Leningr. Univ.

Davidov, L. K., Pronin, A. G., 1967, Vodnyy ryezhim riyek vytyekayushchikh iz lednikov Fedchenko i Zerevshanskogo i ikh verkhnikh pritokov (Water regime of the rivers flowing out of the Fedchenko and Zerevshan glaciers and of their upper tributaries), in: *Krupneyshye ledniki Srednyey Aziy*, Leningrad.

Einarsson, T., 1967, The extent of the tertiary basalt formation and the structure of Iceland. Iceland and Mid-Ocean ridges, *Rit*, 38.

Eythorsson, J., 1949, Variations of glaciers in Iceland 1930–1947, *J. Glaciol.*, 1.

Fahnestock, R. K., 1963, Morphology and hydrology of a glacial stream, *U. S. Geol. Surv. Prof. Paper* 422–A.

Galon, R., 1970, Mechanism and stages of retreating of the Skeidararjokull (Vatnajokull, Iceland), *Bull. Acad. Pol. Sci., Ser. geol. geogr.*, 18.

Hjulstrom, F., 1935, Studies of the morphological activity of rivers as illustrated by the River Fyris, *Bull. Geol. Inst.* Uppsala, 25.

Hjulstrom, F., 1952, The geomorphology of the alluvial outwash plain (sandur) of Iceland and the mechanics of braided rivers, in: *Proc. XVIIth Congress IGU Washington.*

http://rcin.org.pl

Krigström, A., 1962, Geomorphological studies of sandur plains and their braided rivers in Iceland, *Geogr. Ann.*, 44.

Leopold, L. B., Wolman, M. G., 1957, River channel pattern: braided, meandring and straight, *U. S. Geol. Surv. Prof. Paper* 282–B.

Mackin, J. H., 1948, Concept of the graded stream, *Bull. Geol. Soc. Amer.*, 59.

Nevin, C., 1946, Competency of moving water to transport of debris, *Bull. Geol. Soc. Amer.*, 57.

Østrem, G., Brodge, C. W., Rannie, W. F., 1967, Glacio-hydrology, discharge and sediment transport in the Decade Glacier area, Baffin Island N.W.T., *Geogr. Ann.*, 49A.

Pardé, M., 1955, *Fleuves et rivieres*, Paris.

Thorarinsson, S., 1939, The ice dammed lakes of Iceland with particular reference to their values as indicators of glaciers oscillation, *Geogr. Ann.*, 21.

Thorarinsson, S., 1943, Oscillations of the Iceland glaciers in the last 250 years, *Geogr. Ann.*, 25.

Thorarinsson, S., 1954, Athuganir á Skeidarárhlaupi og Grimsvotn 1954, *Jökull*, 4.

Thorarinsson, S., 1967, Forvitnilegir jokullgardar vid jadar Skeidarárjokulls, *Jökull*, 17.

Thorarinsson, S., Einarsson, T., Kjartansson, G., 1959, On the geology and geomorphology of Iceland, *Geogr. Ann.*, 41A.

Walker, R., 1953, Raport on the silt concentration in glacial rivers, *Acta Nat. Islandica*, 1.

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GEOGRAPHIA POLONICA 26, 1973

THE PRESENT-DAY MARGINAL ZONE OF SKEIDARÁRJÖKULL

STEFAN JEWTUCHOWICZ

Morphology of the marginal zone was formed by numerous agents, and therefore the programme of investigations took several directions into consideration, such as the classification and distribution of glacial landforms as well as the dynamics of the phenomena caused by deglaciation. The border part of Skeidarárjökull, the ice-moraine belt stretched along the glacier edge and the adjacent moraines were thoroughly examined.

The morphological sketches of the individual segments of the terrain, which were made in the course of the research work, together with aerial photographs, enabled a morphological map of the Skeidarárjökull marginal zone to be drawn (Fig. 1A—see Annex).

DYNAMICS OF THE PROCESSES CONTROLLED BY GLACIER RECESSION

DISINTEGRATION OF SKEIDARARJOKULL

Thorarinsson (1943) who investigated oscillations of the Iceland glaciers, proved that Skeidarárjökull's recession had been increasing since 1933. This is shown by the glacier disintegration which controls glacial relief. Observations show that the mode and course of disintegration depend to some extent on the glacier structure and the configuration of its surface. Crevasses, deep river beds and hollows on the glacier surface, ice pyramids and englacial channels intensify ablation and facilitate water circulation at various levels of the ice-sheet. Therefore the tenacity of the glacier becomes weaker and this promotes quicker melting and disintegration.

In the literature two hypotheses explain this process: areal and frontal deglaciation (Klimaszewski 1960). During areal deglaciation the glaciers fall into dead-ice blocks, while the frontal one is characterised by recession of the glacier front.

Investigations carried out in the present-day glaciated areas show that deglaciation takes place in many ways. Klimaszewski (1960) states that areal deglaciation is a common phenomenon in the western part of Spitsbergen. According to Szupryczyński (1963) the Spitsbergen glaciers display areal as well as frontal retreat. Thorarinsson (1943), Okko (1955), Todtmann (1960), Price (1969) maintain that glaciers retreat due to shrinkage caused by surface melting. This opinion is in agreement with the notion of areal deglaciation in view of the fact that vertical wastage of glaciers leads to the formation of dead-ice

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blocks which during advancing ablation are detached from the active ice mass. Gravenor and Kupsch (1959) call this process "ice disintegration".

As the ice melts, the glacier becomes overlaid by deposits whose thickness increases. In the part where huge surface moraine exerts pressure on ice the glacier becomes inactive. According to Ahlmann's investigations (1938) it may be called the transition of the active glacier into dead-ice blocks.

One way of disintegration observed in the area of Skeidarárjokull is the detachment of dead-ice fields from the bulk of the glacier. This process occurs chiefly in places where the surface is diversified and intraglacial channels are numerous. Scattered and unevenly distributed rock debris over the surface of the dead-ice field cause more intensive melting than on the active glacier. A great number of streams cut into the ice give rise to the individual ice-pyramids. Due to thinning of the ice the tops of englacial channels, which formerly stretched in the upper part of the glacier, collapse. The advancing ablation produces a strongly differentiated relief, therefore, the morphology of dead-ice fields differs from that of the glacier surface. Here, rock material is transported only at small distances and remains within the area of dead ice whose continuous melting favours the formation of a cover of deposits. The increase of thickness marks the transformation of the dead-ice field into an ice-moraine field.

In the eastern part of Skeidarárjokull, along the River Skeidara, a dead-ice field, 3 km long and 1 km wide, is being formed. The border between the glacier and dead-ice zone constitutes a depression, eroded by meltwaters, and rock material accumulated on the ice. Unlike the glacier surface on which only scattered rock debris occur, the surface of dead-ice blocks is covered with rock material, whose thickness could exceed 10, 20 and even more centimetres (Fig. 2).

Detachment of a dead-ice field may be observed in the vicinity of lake I. The border of the dead-ice is marked by the differentiation of the ice surface and the cover of ablation deposits (Fig. 3). The dead-ice fields are also formed in other segments of the glacier front.

Another common way of disintegration of Skeidarárjokull is the detachment of dead-ice ridges parallel to the glacier edge. The mechanics of this is simple. The edge of the glacier undergoing ablation more intensive than ice in other parts, becomes overlaid with a layer of rock debris. Melting of the ice is impaired by deposits, while it increases in the places uncovered. This process gives rise to the formation of a ridge which hampers the meltwater outflow transversal to the edge of the glacier and directs it along the ridge. Due to the mechanical and thermal water erosion the ridge covered with deposits is gradually detached from the bulk of the glacier and an ice-moraine ridge is formed (Fig. 4). Sometimes several ridges occur, one next to another. For the formation of ice-moraine ridges the outcrops of rock material from the glacier as well as slide planes are of great importance (Fig. 5); it may also be facilitated by crevasses in which water circulates.

In the area of marginal lakes, the calving of the glacier into lakes does not favour the formation of dead-ice fields or ridges. Ice blocks, detached from the glacier, which have fallen into a lake melt and the remaining rock material which is deposited on the bottom makes the lake shallower.

As a result of disintegration the position of the Skeidarárjokull edge undergoes some alterations which may be called a recession of the glacier front. In the course of disintegration shown by the detachment of dead-ice fields the glacier retreats rapidly. Huge fields break off from the bulk of the glacier

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Fig. 2. Dead-ice field in the eastern part of Skeidararjokull. Photo by S. Jewtuchowicz
1 — glacier, 2 — border-line between dead- and active ice, 3 — field of dead ice

Fig. 3. Field of dead-ice near Lake I. Photo by S. Jewtuchowicz
1 — dead-ice field, 2 — glacier

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Fig. 4. Ice-moraine ridge. Photo by S. Jewtuchowicz
1 — ice-moraine ridge, 2 — depression caused by ablation of ice, 3 — glacier

Fig. 5. Ice-moraine ridge along the shear plane. Photo by S. Jewtuchowicz

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which in the eastern part of Skeidarárjokull retreats about 1 km after a single detachment. When the ice-moraine ridge separates from the glacier the latter retreats some distance, while after calving in the lake the retreat does not exceed several metres. Data obtained during investigation showed that the most common way of disintegration of this glacier is the detachment of dead-ice fields and ridges. Transformation of the active glacier into dead-ice blocks produces the dead-ice zone of various width along the Skeidarárjokull edge. The zone displays marginal glacial relief.

In the process of deglaciation some stages may be distinguished. Because of the structural and morphological properties of the glacier, melting of ice is particularly intensive in some segments; its surface is soon covered with rock debris due to which the segment itself acquires the character of a dead-ice field. Transformation of the active glacier into dead-ice blocks is an important stage in the formation of glacial relief. The next stage produces further melting of the ice which leads to the formation of ice-moraine areas. The slow decay of the dead-ice blocks deeply buried under deposits forms the final stage of deglaciation during which the full formation of surface relief takes place.

Observations made on Skeidarárjökul show that the recession of glacier can be manifested in a different way. On the same glacier occurs the detachment of dead-ice fields and dead-ice ridges.

INFLUENCE OF DEAD ICE ON THE DEVELOPMENT OF LANDFORMS

The formation of a zone of dead ice on the edge of Skeidarárjokull plays an important role not only in the creation of marginal relief, but also in its further development. An increase in the thickness of deposits within the dead--ice area produces the ice-moraine landforms whose upper parts are built of

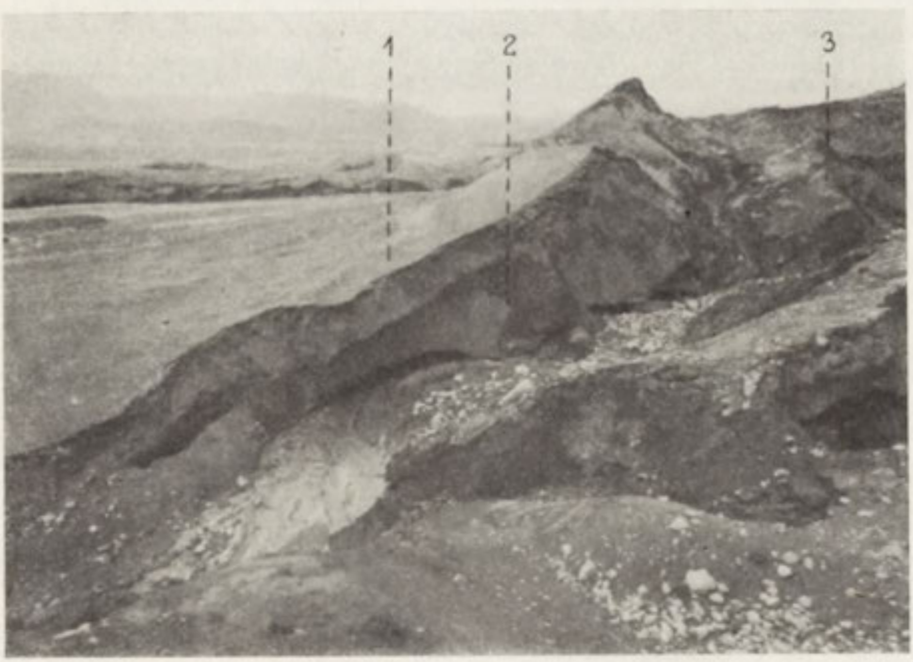

Fig. 6. Transition of the glacier into dead-ice. Photo by S. Jewtuchowicz
1 — mineral deposits, 2 — dead ice, 3 — glacier.

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Fig. 7. Kettle-holes and trenches due to melting of dead-ice. Photo by S. Jewtuchowicz
1 — glacier, 2 — dead-ice field, 3 — kettle-holes and trenches formed due to melting of dead--ice, 4 — Lake III

Fig. 8. Subsidence of the slope of an ice-moraine ridge and formation of kettles due to melting of dead-ice. Photo by S. Jewtuchowicz
1 — ice moraine ridge, 2 — kettle-holes, 3 — meltwater erosional channel, 4 — ridge of the erosional origin

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mineral material, whereas dead-ice blocks still remain in their lower parts. While the dead ice covered by glacial formation melts and becomes thinner during further deglaciation, the released rock material enriches the thickness of mineral deposits. Melting of buried ice blocks is one of the agents causing changes and development of landforms in the marginal zone. The greater the thickness of the dead ice the more extensive changes in the surface relief are caused by the ice-melting.

In the marginal zone there is a continuous mass of dead ice under deposits (Fig. 6). Its occurrence on the glacier foreland is evidenced by the active land--slides caused by melting of ice within the ridges of the 1890 front moraine, which at present is situated 3 km from the glacier edge. Conspicuous traces of dead-ice melting are to be seen on the sandur between the glacier and the old moraine mentioned above.

The changes of landforms produced by the melting of dead-ice blocks are different in character. They were investigated by Jewtuchowicz (1962) in Spitsbergen. Price (1969) describes the influence exerted by the dead ice on the surface relief of Breidamerkurjökull in Iceland. In the foreland of Skeidarárjokull kettle holes are very frequent, the shallow ones are dry, but deeper holes contain water. Melting of the dead-ice blocks makes the individual kettle holes join into small lakes (Fig. 7). The kettle holes stretching between moraine ridges cause rapid slides of debris. On the bare tops of ridges the melting of ice, being more intensive, changes their shape and size. Their ice-moraine ridges initially high with pointed tops and very steep slopes become lower, wider, with rounded tops, the angle of sloping depending on the value of the rest angle of the material they are constructed of. During slides terrace-like steps are formed on the slopes. In places, the steps are broad and contain kettle holes (Fig. 8). The inversion of the initial relief takes place as a result of the melting of ice.

ROLE OF MELTWATER IN RELIEF DEVELOPMENT

The recession of Skeidarárjökull produces huge quantities of meltwater, which flows in streams down the glacier. Between 15 June and 20 August, 1968, Skeidarárjökull was drained by 250 fluvioglacial streams of various sizes. Of these 180 drained the glacier surface.

Large quantities of meltwater penetrate the fissures inside the glacier draining channels, through which it pours out at the edge of the glacier. In the course of the investigations 50 active englacial channels were discovered from which many streams issued. The englacial channels occur on different levels and, as a result of it, meltwater also outflows on various levels of the glacier edge. Most of the meltwater is drained off by channels in lower parts of the glacier, because they accumulate here coming from areas lying far from the Skeidarárjökull edge. The water is discharged under hydrostatic pressure up to 1.5 m. These deep channels give rise to large rivers.

The influence of meltwater action on the relief of the Skeidarárjökull marginal zone is evidenced by the washing out of accumulational landforms, especially in places lying within the reach of the meltwater outflowing from englacial channels. In these segments of the glacier foreland the meltwater produces almost flat erosional surfaces (Fig. 9).

In the marginal zone the drainage net is frequently changed and superimposed valley system can be observed. The meltwater action is not confined to the smoothing of the relief but owing to erosion it forms areas displaying fairly diversified surfaces in some places of the foreland of Skeidarárjökull.

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Fig. 9. Surface smoothed by water from the englacial channel in the marginal zone. Photo by S. Jewtuchowicz

1 — glacier, 2 — outlet of englacial stream

EROSIONAL AND ACCUMULATIONAL LANDFORMS IN THE MARGINAL ZONE

MARGINAL LAKES AND PONDED LAKES

In a number of places within the marginal zone the edge of the glacier runs across the lakes which are very frequent in this area. The origin of the lakes situated at the edge of Skeidararjökull is closely associated with the glacier structure and presence of dead-ice blocks. Numerous englacial channels facilitate the ice melting and the formation of kettle holes. If the dead-ice blocks melt beneath the buttom of a lake, the depression becomes deeper and the outflow of meltwater is dammed by the glacier and moraine. The marginal lakes do not possess names, so the Polish Expedition gave them numbers for better guidance. Near the edge of Skeidararjökull, 10 lakes were distinguished, but due to continuous melting of ice the new kettle holes and trenches containing water are formed; later they widen into lakes.

Numerous streams flowing directly from the glacier discharge into marginal lakes and deposit the debris material in them. The lakes become shallower and within the area of old moraines many lakes dry up.

Along the glacier edge there are also some depressions filled temporarily with water. They do not form lakes but shallow swamps. A very extensive swamp is situated west of lake I; it has no fixed shores, its size being dependent on the quantity of meltwater supplied by the glacier.

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RIVER VALLEYS

In the marginal zone, among the remarkable erosional forms are the valley segments where large rivers start their course from the Skeidararjökull meltwater. The Sula and the Blautakvisl in the west and the Skeidará in the east belong to these valleys. The main source-segment of the Sula is formed by several englacial channels at the south-east edge of the glacier. The water is discharged here under hydrostatic pressure. The Sula is also supplied by streams flowing from the glacier surface but with smaller quantities of water than the englacial channels. The quantity of water pouring out from the channels into the Sula is too large to be contained in its bed and it outflows over a wide area destroying the ice-moraine belt and the adjacent old moraine. Due to the action of the water an extensive erosional plain is formed.

In the western part of Skeidararjökull the Blautakvisl collects water mainly from streams draining the glacier surface and thus, at the time of our observations, it had received less meltwater than the Sula.

The source of the Skeidará is situated at the eastern edge of the glacier. This river begins from the englacial channels but it is also fed by streams flowing from the glacier surface as well as by its right-hand tributary draining the south-east edge of Skeidararjökull (Fig. 10). The Skeidará and its tributary considerably levelled the ice-moraine zone and a part of the old moraine, thus, forming an extensive erosional plain.

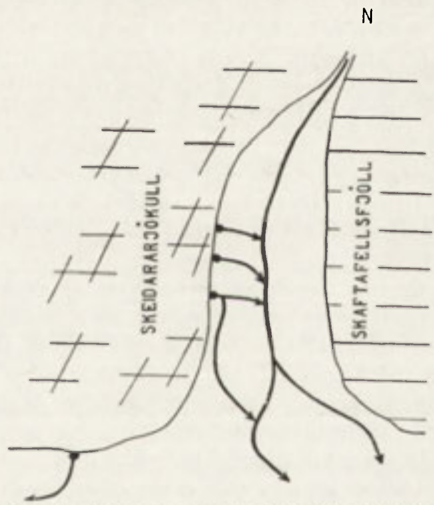

Fig. 10. Streams of the Skeidará River

The valleys of the three rivers are transverse to the edge of the glacier. There are also, in the marginal zone, valleys parallel to the glacier edge. They are largely valleys of rivers joining lakes. Between lake V and the western ponded-lake at the Skeidararjökull edge, a long meltwater channel has been formed by lakes and rivers joined together. This channel plays an important role in the drainage system of Skeidararjökull. It collects and drains off the water from two thirds of the whole investigated marginal zone.

The distance between the channel and the glacier edge varies: between the ponded-lake and lake I it is about 3 km, while towards lake II it becomes smaller. Morphological observations have revealed that the meltwater channel consists of segments of different age. The older one, situated farthest from

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the glacier, between the western ponded-lake and lake II, belongs to the previous stage of deglaciation. The estern part of the older channel near lake II is now inactive, and at the edge of the glacier a new marginal channel is being formed. It comprises the valley east of lake II, a line of sinks west of lake III — still in the course of formation, lake III, a trench formed by water flowing in the pass between lakes III and IV, lake IV and a channel running along the glacier to the intervalley edge in the neighbourhood of lake V (Fig. 1B).

Between lakes V and VIII few englacial channels coming out at the Skeidararjokull edge are poor in water and fail to form either rivers or marginal channels parallel to the glacier edge as it is the case between lakes I and V. In this part of the area investigated, water flows outside the present marginal zone and it accumulates in small depressions of old valleys or in ponded lakes.

East of lake VIII the ice-moraine zone borders a large valley of the right tributary of the Skeidará, which has been mentioned above. The tributary flows first westwards parallel to the glacier edge and then, near lake X, it turns to the south and discharges into the main river. This tributary's valley is an old marginal channel. Lake X, which originated from the subsidence trench after dead ice, forms an indirect link in the rising marginal system of drainage in the eastern part of Skeidararjökull.

These facts witness the existence of a drainage system in the eastern and western parts of Skeidarárjokull which depends on the pattern of valleys and lakes that was formed in the former stage of deglaciation, whereas in the middle part of the glacier front either a new channel is in the course of formation between lakes II and V, or the lack of sufficient meltwater hampers the formation of an outflow system, as is the case between lakes V and VIII.

MORAINE RIDGES AND HILLS

Among accumulational landforms within the present marginal zone are ridges, hills, sandar and eskers.

Ridges are formed in the dead ice area. Being detached from the glacier they are being cut by meltwater streams and this produces the hilly topography. Initially the ridges and hills possess steep slopes and pointed tops. Their height depends on the thickness of ice in particular segments of the glacier edge; some of them are about 35 m high. Because of the melting of ice and according to the content of debris material in the ice, their height becomes reduced and they rarely exceed 4 m. The subsidence and sliding of material caused by the melting of ice makes the countours of these landforms smooth, their bottom larger and tops rounded.

According to Flint (1942) the relief of dead-ice areas is characterized by the frequent occurrence of kettle holes, low hills, ridges and closed depressions.

In some segments of the marginal zone the moraine ridges rise under the ice edge. The stream flowing along the ice edge undercuts the ice and produces an overhanging wall, under which ablation material is accumulated (Fig. 11). After the hanging ice face has melted there remains a moraine ridge.

Where the glacier front is high and solid, streams flowing on its surface erode bays at the ice margin. Moraines that develop in the bays are flattened by water (Fig. 12). In some bays, in the places left by the retreating glacier edge, debris accumulates forming a ridge transverse to the glacier front (Fig. 13).

On the eastern side of Skeidarárjokull, on the glacier surface emerges a me-

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dial moraine ridge, about 8 m high, transverse to the end moraine (Fig. 14). The proximal part of the medial moraine gradually enters into the ice and disappears. After the ice melts, the moraine is unconspicuous among others glacial forms of the marginal zone.

This clearly shows that the ice melting and water activity play the main

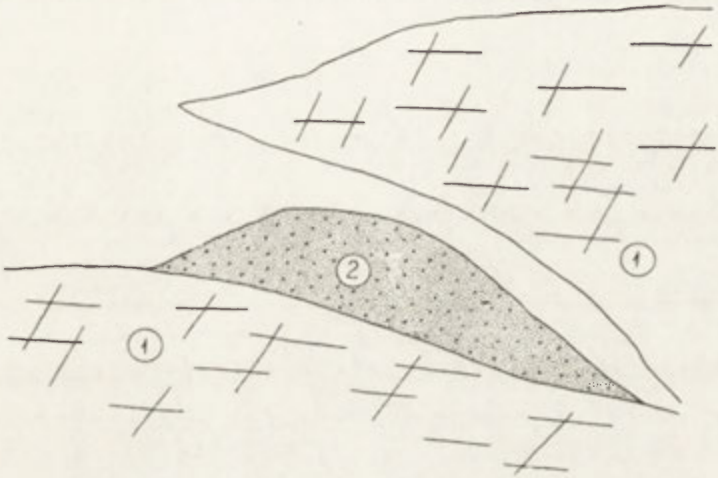

Fig. 11. Accumulational moraine ridge under the ice edge
1 — ice, 2 — moraine

Fig. 12. Accumulation in bays of the glacier edge near Lake IV. Photo by S. Jewtuchowicz
1 — moraine, 2 — glacier, 3 — present-day marginal channel draining the foreland of the glacier, 4 — older moraine

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role in the accumulation of rock material. The structure of the deposits depends on ice and water.

Surficial deposits, accumulated by water are stratified. In general they are slightly inclined and often indistinct. In the upper series of deposits the melting of the underlying dead-ice causes dislocations. They are formed by the shift of material either vertically or down the slope line. Melting of ice is also

Fig. 13. Accumulational moraine ridge, transverse to the glacier edge
1 — glacier, 2 — moraine ridge

Fig. 14. Medial moraine in the eastern part of Skeidarárjokull. Photo by S. Jewtuchowicz
1 — end moraine, 2 — medial moraine, 3 — glacier surface

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the cause of bending layers which adapt themselves to the underlying ice outline (Fig. 15).

The layers of rock material in the glacier are of various sizes. Lenses from 20 cm to several metres in length are very frequent. In some places material is not stratified.

In the marginal zone where water activity is hampered, material is accumulated directly from the ice, and the deposits remain in unchanged positions. Their structure reflects the distribution of rock particles within the ice. It is then a structure controlled by that of the glacier. Where the ice was interbedded with rock material, the bedding of layers is preserved in deposits, whereas in parts of material chaotically distributed within the ice body, there are structureless deposits (Fig. 16).

It is well known in glaciology that due to the ice movement the layers are bent upwards, rising to the margin of the glacier (Fig. 17). In case of direct accumulation from the ice the deposits show the same bending of layers as

Fig. 15. Bending of the layers caused by dead-ice melting
1 — dead-ice, 2 — sand layers

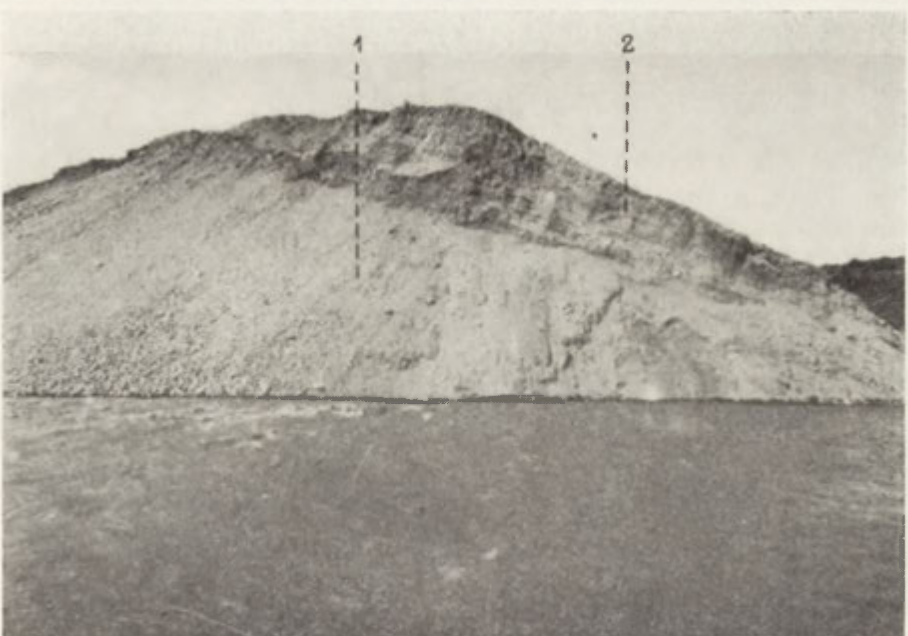

Fig. 16. Structure of deposits melted out of dead-ice. Photo by S. Jewtuchowicz
1 — dead-ice, 2 — sand layers

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Fig. 17. Structure of ice. Photo by S. Jewtuchowicz
1 — ice encrusted with rock material, 2 — ice slide plane

Fig. 18. Structure of deposits accumulated directly from ice. Photo by S. Jewtuchowicz
1 — unstratified deposit, 2 — layer of sand

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exists in the ice stratification (Fig. 18). Thus, in the hills the deposits are inclined at about 90° (Fig. 19).

Describing the deformations in the Greenland glaciers Drygalski (1897) stated that some deformations of ice layers were due to the load exerted by debris. Such phenomena may also be observed in the Skeidararjokull area (Fig. 20). In the course of direct accumulation from the ice the deformations become well preserved in the moraine (Fig. 21).

The above described moraine ridges and hills being formed as the glacier retreats may be called accumulational recessional moraines.

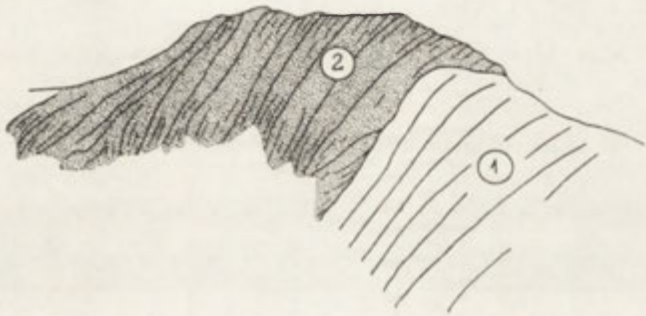

Fig. 19. Inclination of the layers controlled by the glacier structure
1 — ice, 2 — deposit melted out of the ice

Fig. 20. Deformation of the glacier layers. Photo by S. Jewtuchowicz

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Fig. 21. Deformations of the glacier layers reflected in the structure of deposit of direct glacial accumulation. Photo by S. Jewtuchowicz

SANDAR (OUTWASHES)

Sandar occur chiefly in the western part of the marginal zone. They are built either by deposits transported by streams directly from the glacier or by material washed out by water from the ice-moraine belt. Meltwater overflowing sandar during intensive ablation does not exceed 10–30 cm in depth, but its speed is sufficient to carry great quantities of material suspended in water, while coarser particles are rolled along the botton.

Investigations on the present formation of the sandur of the Gas glacier in Spitsbergen showed that the sheet outflow of meltwater is very important as a morphogenetic agent (Jewtuchowicz 1962). These observations are enriched by those carried out in Iceland.

In the course of the investigations carried out in the marginal zone of Skeidararjokull the intensity of meltwater outflow over the surface of the sandur varied. The increased volume of water supplied from the glacier was marked by waves 10 cm high and 20–30 cm long. During intensive ablation the waves were coming at 10–20 minute intervals. Decrease of waves gave rise to the accumulation of the suspended material of the coarser particles which were dragged along the bottom. This process formed the stratification of a sandur.

The frequent occurrence of sandar as glaciofluvial landforms testifies to the important role of water in the relief forming processes.

ESKERS

In the marginal zone of Skeidarárjokull the eskers occur individually and in groups. One esker ridge, 600 m in length, stretches along the western shore of lake I (Fig. 22). At present it is cut by rivers in a few places. Its proximal

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Fig. 22. Marginal zone near Lake I. Photo by S. Jewtuchowicz
1 — esker, 2— ice-moraine ridge, 3 — Lake I, 4 — older moraine, washed

Fig. 23. Formation of an esker in the englacial channel. Photo by S. Jewtuchowicz
1 — esker, 2 — glacier

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part at the glacier edge is hidden in the ice, while its distal segment lies within the older moraine and passes into a narrow sandy delta, about 50 m long, which is covered in part by lake I.

At present, in the region of lake II the eskers are in the state of formation in abandoned englacial channels. When water was flowing through the channels transportation prevailed over accumulation. After the flow of water had ceased, material melted out of the upper sections and walls of the channels was deposited. The released meltwater segregated the rock particles but failed to transport them.

Due to the thinning of the glacier the upper parts of some channels collapse. Rock material from the walls of channels covers ice-blocks derived from the collapsed upper parts of channels. In the place of the former channel a trench is formed and gradually it is filled with debris, then an esker develops. This formation of esker can be seen in various parts of the marginal zone (Fig. 23).

In the foreland of Skeidarárjökull also appear groups of eskers in which the arrangement of individual ridges resembles a river system. The best developed landforms are situated at the western ponded-lake. Some of them are about 500 m long and 3 m high. The main N–S ridges are joined by oblique short ridges. Between eskers there are lakes.

Another group of eskers lies east of lake I. They are 600 m long and stretch from the ice-moraine zone to the marginal channel formerly mentioned. Some of them are long and straight, others are winding or even ring-like.

Investigations carried out in the present-day glacial areas revealed various origins of the eskers. Klimaszewski (1960) distinguished subaquatic formation of eskers between Kongsfiord and Eidem-Bukta in Spitsbergen. Szupryczyński (1963) describes supraglacial and englacial origin of eskers of the Warenskjold glacier. Jewtuchowicz (1962) investigated eskers of subglacial origin in Spitsbergen. Price (1969) who also observed the Iceland eskers on the Breidamekurjökul glacier believes that they were formed in the subglacial, englacial and supraglacial channels. In the present marginal zone of Skeidarárjökull the eskers are formed in englacial channels.

RELIEF OF THE MARGINAL ZONE

The closest foreland of Skeidarárjökull is an ice-moraine belt and an area with deposits in which dead ice has been mainly melted away. In the ice-moraine belt the melting of dead ice blocks which constitute the core of ridges and hills produces changes in the outlines of the landforms, but in places where dead-ice blocks are deeply burried the relief shows greater stability.

Both areas are often separated by a meltwater channel parallel to the ice edge, by a sandur just being formed, or by a moraine flattened by the erosional activity of water. Morphology of these forms as well as the thickness of dead ice underlying the deposits show that the two areas arose in different phases of deglaciation.

The ice-moraine belt is the result of the present-day recession of Skeidarárjökull which started in 1933 (Thorarinsson 1943). The adjoining older moraines were formed during the deglaciation which gave rise to the recession of Skeidarárjökull in the twenties. Thorarinsson (1943) distinguishes periods of stagnation and advance of the southern glacier Vatnajökull and Skeidarárjökull between the two recessions. The morphology of the foreland of Skeidarárjökull shows that stagnation of this glacier was not uniform. Therefore,

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three different segments can be distinguished in the marginal zone. The first segment is situated between Eystrafjall Mts. and lake II, the second segment lies between lakes II and VIII, and the third one — between lake VIII and the Skaftafelsjöll Mts. (Fig. 1A).

The first segment is an ice-moraine belt, about 100 m wide. Streams flowing from the glacier have cut the belt into ice-moraine fields. The surface relief of these fields is conditioned by the configuration of the dead ice buried under deposits. The most frequent landforms here are ice-moraine ridges up to 35 m in height, but also very common — especially in the region of source-segments of the Sula and the Blautakvisl — are ice-moraine fields whose almost flat surfaces are marked by traces of former dead-ice ridges running in various directions. Here, the deposits accumulate directly from the melted dead-ice blocks. Because of the steep sloping surfaces the glaciofluvial streams do not accumulate the carried rock material, but are intensively eroded. However, in places where the ice-moraine ridges were destroyed by water, begins the formation of sandar. Among rounded landforms appear eskers, which unlike the moraine ridges, are transverse to the glacier edge.

The valleys in the ice-moraine belt are either transverse or oblique to the glacier edge. Because of great height differences the front of the glacier and its foreland, short narrow valleys with steep slopes have been formed; their inclination increases in dead ice to 90°. A dense and deep network of valleys causes frequent slides and solifluxion of debris that build upper parts of the ice-moraine convex forms. Together with relief development, the valleys gradually fill with debris, while high hills become lower. Thus contrasts characteristic of the former relief become obliterated. The surface of the ice-moraine zone is diversified by the funnel- or trench-like depressions caused by ice melting. The trenches are often filled with water and form marginal lakes.

In the south, the ice-moraine zone adjoins an area in which ice had been melted away during the previous recession of Skeidararjökull. Here, erosion has been hampered because of the thicker mineral mantle covering the dead-ice blocks and the greater distance from the ice edge, and therefore the morphological transformations of the land surface are slower.

Instead, along the contact line, between the ice-moraine zone and older deposits the activity of meltwater and land forming processes are very intensive. Only larger rivers like the Blautakvisl and the Sula break through the older deposits to the more distant parts of the foreland. Small meltwater streams cannot penetrate through the compact morphological forms and they flow along the ice-moraine zone, flatten their surface, flow into lakes giving rise to ponded-lakes, and facilitate the formation of sandar. The water activity is revealed in the relief of the area by the predominance of sandar and moraine surfaces flattened by erosion. The older moraines adjoining the ice-moraine zone form islands usually separated by sandar. Hence, the older and new relief-forms are interrelated.

Some landforms stretch over the area of the older moraines and of the ice-moraine belt. A good example is the esker on the western side of lake I. Its distal part lies within the area of the older moraine, while the proximal one is in the ice-moraine belt. The location of eskers on zones of various ages provides a reliable testimony that the marginal structure of Skeidararjökull, favourable to the formation of eskers, remained unchanged during the two recessional stages recognized by Thorarinsson (1943). If there had been a transgression of the glacier, the arrangement of englacial channels in the ice body would have been altered and the esker structure interrupted. The continuity

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of the esker examined proves that during its formation there had not been any oscillation of the Skeidarárjökull front.

Also the eskers situated between lakes I and II run continuously from the ice-moraine belt across the older moraine to the marginal channel, which drains this segment of the Skeidarárjökull foreland. The esker lying west of lake I as well as the group of eskers between lakes I and II bear witness to the continuity of the surface forms, which were distributed only by the activity of the present-day meltwater streams.

The continuity of eskers shows that since the recession in 1920 (Thorarinsson 1943), the ice had not advanced on this segment, which explains a remarkable outwashing of older moraines adjoining the dead-ice belt. Long meltwater streams flow into lakes, ponded-lakes or marginal rivers lying farther from the glacier, and whose origin belongs to the period of a previous recessional phase.

Segment 2 of the Skeidarárjokull marginal zone, situated between lakes II and VIII, displays the same relief development of the ice-moraine belt in the vicinity of lake II as in segment 1. Numerous eroding streams divide this area into ice-moraine fields. Here, water and dead-ice are the dominant agents in shaping the surface from which ice has melted away. The older moraine adjoining the ice-moraine belt is eroded by water flowing to lake II and forms islands.

An ice-moraine ridge, about 3 m high, separates the ice-moraine belt from the older moraine. In many places the ridge is cut by streams flowing from the glacier into the lakes.

The two areas of segment 2 present different modes of surface relief. In the ice-moraine area water is the dominant agent in shaping the relief; numerous eroding streams form stripes transverse to the glacier edge, while on the older moraine there are ridges parallel to the glacier front according to ice slide-planes. The surface relief of the older moraine shows the prevalence of the glacier structure.

East of lake II the ice-moraine belt becomes narrower and in places is no more than 10 m wide. In the vicinity of lake VII it is a long narrow strip, developing along the debris outcroping from ice (Fig. 24). The older moraine zone is also diversified with ridges parallel to the glacier edge. Their origin is associated with ice slide-planes. Both the older and present-day relief in this part of segment 2 is mainly due to the structure of the glacier. Unlike segment 1, this one has neither sandar nor erosional moraine plains. Near lake VII there is no marginal valley draining the foreland.

Morphological data show that the whole of segment 2 is poor in meltwater, the glacier forefield is slightly eroded and the glacier front is more stable than elsewhere. As the result of the present glacier recession, which started about 1933 (Thorarinsson 1943) the glacier edge has retreated about 100 m in the neighbourhood of lake II, whereas between lakes VII and VIII — 10 m on average.

In segment 3 of Skeidarárjökull, lying between lake VIII and the Skaftafellsfjoll Mts., the ice-moraine belt is of varying width, and it occurs on the lake shore only in the glacier-edge bays where small ice-moraine fields are formed. At the eastern border of the glacier the belt is as wide as 1 km. Its surface is differentiated, and in many places the ice-moraine elevations are 20 m and more in height. Owing to the diversity of the terrain the glaciofluvial streams erode intensively and cutting into the ice form deep vertical walls. A dense net of fissures, transverse and oblique to the glacier edge, dissect the ice-mo-

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raine belt into blocks of different sizes. Because of the surface irregularities the accumulation processes and, as a result of it, the glacial relief are differentiated. The older moraine forms the southern shores of lakes VIII and IX, as well as the southern valley-side of the right tributary of the Skeidará. The ice-moraine belt, unlike segment 2, is separated here from the older moraine by lakes and by the above mentioned river valley.

In morphology of the whole marginal zone three segments may be distinguished. The eastern and western parts are similar in their drainage pattern system, strongly eroded moraine and mode of the glacier disintegration. The

Fig. 24. Marginal zone in the vicinity of Lake VII
Photo by S. Jewtuchowicz
1 — glacier, 2 — ice-moraine ridge, 3 — older moraine, 4 — accumulational ridge of the older moraine

middle part of the segment is to some extent different. Here, small quantities of meltwater and its weak activity on the glacier forefield, lack of well-developed drainage system as well as narrow ice-moraine belt are proof of a greater stability of the glacier edge than in the two other segments.

According to Thorarinsson (1943) and on the basis of analysis of the surface morphology it may be assumed that during the recession which started in 1920 there was a short stagnation in the middle part of Skeidarárjökull, but at the same time the glacier retreated in its western and eastern part. Only in this way may the morphological differences in the area investigated be explained.

The present recession of the glacier, which started about 1933 (Thorarinsson 1943) also involved the middle part of Skeidarárjökull. As a result a narrow ice-moraine belt and the marginal channel between lakes II and V were formed. The relief-forming processes in the middle parts of the marginal zone are distinctly less intensive than in its other parts.

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CHARACTERISTIC FEATURES OF THE PRESENT-DAY PHASE OF DEGLACIATION

The Skeidarárjökull edge is bordered by the dead-ice zone, where glacial landforms are formed. The deglaciation is therefore manifested by the transition of the active glacier into dead-ice and by the formation of the marginal-zone relief.

The present stage of deglaciation results in the general detachment of wide fields and ridges of dead-ice. Three phases of land-relief formation may be distinguished within the dead-ice zone.

In the first phase intensive surficial ablation promotes the accumulation of a thick layer of debris on the ice which protects it against radiation and atmospheric factors. The structure of ice and activity of meltwater also play an important role in the relief modeling. Depending on the structure, ice-moraine ridges and hills, massive moraines, eskers and kames develop. The accumulational and erosional activity of meltwater within dead-ice fields gives rise to small sandur surfaces and flat areas due to the outwashing of moraines. There are also some landforms whose origin Flint (1948) regards as ice-contact forms, i.e., deposits accumulated through the support of an ice-face.

The second morphogenetic phase within the dead-ice fields begins when the thick cover of deposits hampers surficial ablation. Then the ice melts under rock debris and the deglaciation processes operate under the surface. This causes further changes in the relief. Melting of the ice masses buried inside the moraine ridges and hills makes these forms lower and wider. Kettle holes and small basins, chaotically arranged, are also formed.

The sub-surface deglaciation is often the cause of marginal lakes that play an important part in draining the glacier and its foreland. The drainage system during the glacier edge disintegration is not necessarily associated with the sequence: end moraine — *sandur* — *pradolina,* as Keilhack (1898) claims to be always present in the Pleistocene glaciated areas. The drainage system of Skeidarárjökull confirms the views of Maas (1904). Marginal lakes at the glacier edge, that are the result of thermokarst processes — i.e., melting of dead-ice under deposits — collect water which, flowing down the sloping surface from one lake to another, forms a marginal channel.

The disintegration of Skeidarárjökull into dead-ice fields causes rapid and irregular retreat of the glacier edge and a loss of large masses of ice, which in turn testifies to the acceleration of its decay. The decay of the glacier ice, manifested by the detachment of large fields of dead-ice, is the most important characteristic of the present stage of deglaciation.

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REFERENCES

Ahlmann, H. W., 1939, Über das Entstehen von Toteis, *Geol. Foren. Stockh. Förh.*, 60, 3.

Drygalski, E., 1897, *Grönland-Expedition des Gesellschaft für Erdkunde zu Berlin 1891–1893*, Berlin.

Flint, R. F., 1948, *Glacial Geology and Pleistocene Epoch.*

Gravenor, C. P., Kupsch, W. O., 1959, Ice disintegration features in Western Canada, *J. Geol.*, 67.

http://rcin.org.pl

Jewtuchowicz, S., 1962, Studia z geomorfologii glacjalnej północnej części Sörkappu (Glacial morphologic studies in northern Sörkapp), *Acta Geogr. Lodz.*, 11.

Jewtuchowicz, S., 1971, Współczesna strefa marginalna lodowca Skeiderárjökull na Islandii (Sum.: The present-day marginal zone of Skeidarárjökull, Iceland), *Acta Geogr.* Łódź, 27.

Keilhack, K., 1898, Die Stillstandslagen des letzten Inlandeises und die hydrographische Entwicklung des pommerschen Küstengebietes, *Jb. Preuss. Geol. Landesanst.*, 19.

Klimaszewski, M., 1960, *Studia geomorfologiczne w zachodniej części Spitsbergenu między Kongs-Fjordem a Eidem-Bukta* (Geomorphological studies of the western part of Spitsbergen between Kongsfjord and Eidembukta), Zesz. Nauk. UJ, 32, Prace Geogr., Ser. Nowa, 1.

Maas, G., 1904, Zur Entwicklungsgeschichte des sog. Thorn-Eberswalder Haupttales, *Z. Dtsch. Geol. Ges.*

Okko, V., 1955, Glacial drift in Iceland, its origin and morphology, *Bull. Comm. Géol. Finl.*, 170.

Price, R. J., 1969, *Moraines, sandar, kames, and eskers near Breidamerkurjökull, Iceland*, Trans. Inst. Brit. Geogr., 46.

Szupryczyński, J., 1963, *Rzeźba strefy marginalnej i typy deglacjacji lodowców południowego Spitsbergenu* (Sum.: Relief of marginal zone of glaciers, and types of deglaciation of southern Spitsbergen glaciers), Prace Geogr. IG PAN, 39, Warszawa.

Thorarinsson, S., 1939, The ice dammed lakes of Iceland with particular reference to their value as indicators of glacier oscillations, *Geogr. Ann.*, 21

Thorarinsson, S., 1943, Oscillations of the Iceland Glaciers in the last 250 Years, *Geogr. Ann.*, 25.

Todtmann, E. M., 1960, Gletscherforschungen auf Island (Vatnajökull), *Univ. Hamburg, Abh. Auslandsk.*, Reihe C, 65, 19.

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A SYNTHETIC DESCRIPTION OF DEPOSITS AND LANDFORMS OBSERVED ON THE PROGLACIAL AREA OF SKEIDARÁRJÖKULL. CONCLUSIONS WITH REGARD TO THE AGE OF THE DEPOSITS AND THE WAY IN WHICH DEGLACIATION IS PROCEEDING

RAJMUND GALON

(A) PRINCIPAL MORPHOGENETIC ZONES

Omitting from consideration its dissection by the Sandgigjukvisl valley, the zone of glacial and glaciofluvial deposits constituting the Skeidarárjökull forefield consists of two clearly perceptible parts: a western and an eastern part. The middle of this zone is marked by a wide break in the zone of end moraines and other moraine deposits. This break is filled by a large outwash sheet which starts out from inside the marginal zone. In the Skeidarárjökull forefield the width of the whole zone of marginal deposits and landforms is smallest in the east and gradually increases westward. This asymmetry, caused by the more rapid rate of deglaciation during the last sixty years in the western part of Skeidarárjökull (Fig. 1), is also illustrated by differences in the appearance of the marginal deposits and landforms along the glacier frontal margin. A full series of deposits of glacial and glaciofluvial accumulation can be observed in the western part of the discussed zone, i.e., on both banks of the Sandgigjukvisl valley, and here this series comprises a compact section of the main end moraine zone bordered by wider outwash valleys (Fig. 1).

The most characteristic section of this zone is that adjoining the Sandgigjukvisl valley from the west. After R. Galon's investigations the following morphogenetic zones may be considered to occur here:

(1) the forefield of the main end moraine zone, with remnants of older eroded end moraines,

(2) the main end moraine zone (Fig. 2),

(3) the zone of elongated depressions (Fig. 3),

(4) the inner zone of marginal deposits and landforms, passing into an undulating plain of ablation and ice-dammed basins, all drained by a peripheral river.

Near the glacier margin (Fig. 4) can be seen:

(5) ice-moraine ridges, morainic plains, outwash deposits and ice-dammed lakes, either in the process of formation or recently formed. This youngest marginal zone has been described by S. Jewtuchowicz.

The morphogenetic zones (1) to (4) enumerated above extend on both sides of the discussed section of marginal deposits and landforms, that is, eastward from the Sandgigjukvisl River, as well as westward from the central part for which R. Galon has prepared his report. This is the area reserved for M. Bogacki's research. Westward, in the hinterland of the main end moraine zone

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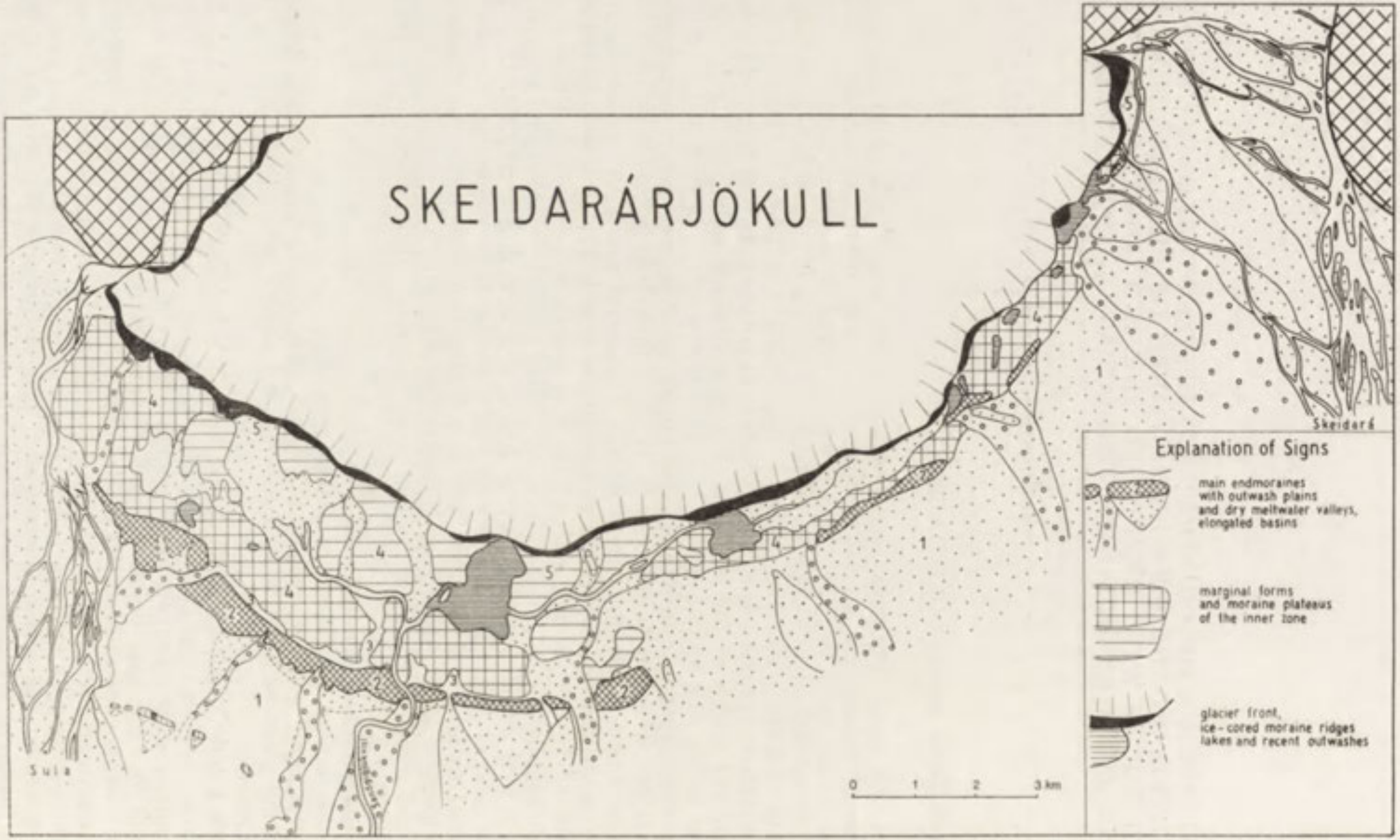

Fig. 1. General geomorphological map of the proglacial area of Skeidarárjökull

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Fig. 2. Main end moraine zone cut by the valley of Sandgigjukvisl. Photo by R. Galon

Fig. 3. Zone of elongated depressions with diminishing lake and inner moraine zone seen from the main end moraine ridge. In the background the marginal river, youngest moraine deposits and margin of Skeidararjokull. Photo by R. Galon

Fig. 4. Front of Skeidararjokull. Glacier gate and meltwater river. Photo by S. Jewtuchowicz

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which is widely spread but split up into sections and dissected by outwash valleys, there continues a chain of elongated depressions against which the moraine plain with its kames and crevasse ridges impinges. However, the more the main end moraine zone approaches the Sula River, the western lateral stream, it gradually shrinks in width and becomes undercut by the Sula River (Fig. 5). Farther on, the marginal zone turns northward and, according to K. Klimek's observations, changes into an assemblage of hillocks which at the base of the basalt plateau pass into a lateral moraine. In that same direction the chain of elongated depressions also vanishes, assuming the shape of a se-

Fig. 5. Sula River undercutting the western section of the main end moraine zone. In the background margin of Skeidarárjökull. Photo by R. Galon

ries of isolated shallow hollows. On the other hand, the inner zone of marginal deposits and landforms continues to extend westwards, first in its previous geomorphological appearance but, farther on, as a hummocky area which northward joins deposits of a younger glacial accumulation and is dissected by valleys of outwash streams or, elsewhere, interrupted by an extensive area of kettle holes.

East of the Sandgigjukvisl valley the zonal pattern (1) to (4) described above continues. The main end moraine zone again shows its typical appearance. Also clearly noticeable is the zone of elongated depressions. But within the inner zone of marginal deposits and landforms the huge crevasse ridges so characteristic of the area west of the Sandgigjukvisl valley are replaced by a great number of groups of moraine hillocks. Moreover, this zone is limited by the greatest of the ice-dammed lakes fronting the Skeidarárjökull; this lake is part of the peripheral drainage system. Surrounding this lake lie moraine plateaus which extend up to the ice-moraine ridge and are dissected by valleys formed by meltwater streams. Among them lie quite a number of dry valleys.

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Next to be seen is the wide break in the zone of the moraine deposits mentioned before. Apart from some fragments of a moraine plain, moraine deposits are absent here. However, the initial part of the widespread outwash sheet contains great piles of stones, remnants of former end moraines. These old end moraines were still entered on a topographical map dated from 1904 showing, among other forms, what is called the Håalda mound[1] some 40 m high; these forms were destroyed during a later stage of glacier retreat, probably at the time when the glacier front was converging upon the contact wall separating the outwash mentioned from the moraine hillocks. These hillocks are counterparts of the inner zone of marginal deposits and landforms, and stop short at the wide peripheral valley with its proglacial lakes and meltwater streams; this valley ends nothward and is limited by an ice-cored moraine ridge extending along the margin of the glacier. After Helgason and Thorarinsson (1943), in 1904 and especially in 1922 *jokullhlaup* waters have greatly contributed to producing the wide break in the end moraine zone.

The extreme eastern part of the zone of moraine deposits described by K. Klimek is relatively narrow, because here deglaciation has been less intensive. This is why the main end moraines which here are moderately high pass directly into the inner zone of marginal deposits and landforms; here and there the boundary between the one and the other type of formation is marked by a small escarpment. It seems probable that the end moraines of the eastern part of the zone of moraine deposits originated somewhat later than the end moraines of the western part of this zone. This assumption is also supported by the difference in distance visible in the west and the east between the glacier margin and the end moraines. The two joined marginal zones end eastward sharply at the peripheral valley mentioned. The ice-cored moraine ridge becomes steadily narrower and locally completely vanishes in an eastward direction. Along the glacier frontal margin extends a range of ice-dammed lakes, and the marginal zone consists alternately of end moraines and moraine hillocks which appear mutually interdented. Only at one place orderly arranged small end moraine ridges have survived occupying the whole width of the marginal zone. This occurrence has been described by K. Klimek, but it was mentioned for the first time by Thorarinsson (1967). Eastwards has developed the area covered by extensive outwash plains; these plains divide the narrowing marginal zone which ultimately ends at the bank of the Skeidarå, the eastern lateral river.

Parallel with the diversity of the zone of moraine deposits are great differences in the way in which meltwater streams find their way to the sea. The previously mentioned asymmetry observed in the main end moraine zone as the result of a farther retreat of the glacier along its western margin, corresponds in front of the glacier to a more extensive spread of the fluvial pattern all over the western part of the area under investigation. The meltwater streams, initially accumulated in ice-dammed lakes and peripheral rivers, ultimately combine to escape in the Sandgigjukvisl River which crosses the moraine zone by a deeply incised valley. On the other hand, in the east one sees in front of the glacier a range of unconnected ice-dammed lakes, most of them lacking surface run-off.

A remarkable feature in the forefield of the main end moraine zone and of the outwash sheets which start well inside the marginal zone are the great number of dry valleys encountered. They are evidence of the changes which

[1] The moraine mound called Håalda is supposed to date back to the middle of the 19th century (Thorarinsson 1943).

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have successively been taking place in the manner of water runoff since the time the glacier halted along the line of the main end moraines, through successive stages of retreat in which the inner zone of marginal deposits and landforms developed, until the present-day position of the Skeidarárjökull margin.

During the main period of standstill of the glacier marked by the line of the main end moraines, a number of less important runoff tracks of meltwater streams existed. During the time the inner zone of marginal deposits and landforms was developing, the meltwater runoff was for the most part concentrated in the central part of the glacier forefield, and this led to the formation of a wide outwash track and of a wide break in the arc-shaped moraine chain. In fact, at that time the moraine zone became divided into a western and an eastern part. Consideration should also be given to the part which *jökullhlaup*-type maltwater floods must have been playing in the formation of the enormous outwash stream, especially of the masses of water which ran here in 1904 and 1922. At the rate of the further retreat of the active ice margin — with this deglaciation proceeding more intensively in the western part of the glacier — both the concentration of the meltwater flow and its escape shifted to the western part of the marginal zone, and by several stages the present-day pattern of valleys and of proglacial runoff developed, with the Sandgigjukvisl as the central flow track. The two lateral Rivers Sula and Skeidará were in operation throughout this period of deglaciation described above; thus the valleys of these rivers bear a polygenetic character.

(B) AGE OF MARGINAL DEPOSITS AND LANDFORMS

In the forefield of the main end moraine zone and the outwash fans connected with this zone, within a plain of degradation and aggradation, some 1.5 km from the glacier lies an end moraine belt sharply marked through seriously damaged by *jökullhlaup* floods; this belt is formed by a series of isolated flat-topped stone elevations, followed in their extension by accumulations of rocks of all sizes. Judging by the information so far available these end moraines, the oldest found in the forefield of the Skeidarárjökull which in the framework of our research were described by M. Bogacki, go back to the middle of the 18th century. This estimate seems to be confirmed by analyses of rock lichen (*Rhizocarpon geographicum*) made by J. Zielińska of the Institute of Botany of Warsaw University. In the outwash sheet, nearer the glacier, i.e. some 500–600 m from the main end moraine zone or in the near forefield of this zone, accumulations of large boulders or low stone mounds have been discovered at many sites; they undoubtedly represent some destroyed recession-type end moraines whose origin Icelandic researchers assign to the 1830–1850 period.

As has been determined by the 1968 Polish Expedition, the main end moraine zone is the result of a recent advance of the Skeidarárjökull. The deposits laid down during this glacier oscillation overlie older moraine-glaciofluvial sediments bearing the character of a plain of degradation and aggradation, and the boundary between the lower and the higher deposits is clearly marked (Fig. 6). Icelandic investigations assign the formation of the main end moraine zone to the last decade of the 19th century, i.e., to about 1890. But from a geomorphological point of view this claim can only be referred to the "older" moraines inside the main end moraine zone, because the "younger" end moraines of this zone have developed in later years, during subsequent successive stages of glacier retreat, probably in the way illustrated on page 47. The 1904

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Islandic map drawn on 1:50,000 scale still does not show the ice-dammed lakes situated in the rear of the main end moraine zone and in front of the glacier. Evidently at that time the glacier was still piling up what we call the "younger" end moraines. This slow retreat of Skeidarárjökull — this phenomenon was determined in an identical way by Price (1969) for Breidamerkurjökull — lasted in both cases well into the thirtieths, and perhaps into the forties, of the 20th century. Presumably the Skeidarárjökull margin coincided at those times (1937?) with the line of end moraine ridges bordering in the northward direction the zone of elongated depressions which at that time bore the cha-

Fig. 6. Boundary between the younger drift and older glaciofluvial deposits. Photo by R. Galon

racter of ice-dammed lakes. From these lakes the waters escaped southward by crossing the main end moraine zone, and in this way they took advantage of the then existing moraine gaps, especially of gaps No. 5 and 10, situated in the research field assigned by R. Galon. At that time the now dry valley starting out from gate No. 5 must have been the main runoff track of the meltwater streams which were accumulating in the zone of the elongated depressions in the rear of the main end moraine zone, but facing the very slowly retreating Skeidarárjökull front. Inhabitants of Nupstadur assert that water flow in this valley continued until 1940.

In the further course of the 20th century the Skeidarárjökull glacier — very similar to the Breidamerkurjökull (Price 1969) — has been retreating at a more rapid rate; as shown on an aerial photograph (in 1944) the glacier margin reached the repeatedly mentioned line (see also the paper of Z. Churski) approaching the then actual peripheral river. This was the time when the initial phase of the new system of main drainage by the Sandgigjukvisl River was developing, still linked with the remnants of ice-dammed lakes and

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with the peripheral river in the immediate rear of the main end moraine zone. The further rapid retreat of the glacier observed in 1968 led to the drainage system as it is now, with ice-dammed lakes situated farther north and new peripheral streams. And at the same time the Sandgigjukvisl River continues her role as the central artery carrying the meltwater streams, arriving from the Skeidarárjökull terminal depression, through the semicircular system of end moraines and farther off towards the sea.

(C) THE PROCESS OF DEGLACIATION IN THE LIGHT OF STUDIES OF THE MARGINAL DEPOSITS AND LANDFORMS OF THE SKEIDARÁRJÖKULL

(1) Examinations made on the Skeidarárjokull by the geomorphological group of the Expedition revealed that, after its oscillating motions some 80 years ago, this glacier has been down-wasting in a diversified way. In the stage during which the main zone of end moraines was developing, including a great number of outwash cones and of meltwater valleys incised into the plain of degradation and aggradation, the glacier was waning and frontally receding, and at that same time its height was decreasing. Afterwards, while the inner zone of marginal deposits and landforms was developing, the glacier still ended in a prominent front along which locally large end moraines were piling up. At the same time, in the rear of these developing end moraines, near the steadily diminishing glacier, a system of fissures formed in which the glaciofluvial material was melted out. During protracted frontal retreat periods this material had been accumulated in the bottom part of the glacier, or it had been washed into these fissures or into other depressions in the glacier surface by supraglacial meltwater streams. As a result it occurred that in the marginal zone of the glacier ending in a prominent front deposits were accumulated characteristic of marginal, or surface deglaciation. Under favourable conditions this might have caused the formation of isolated blocks of dead ice, separated from each other by the previously, developed crevasse ridges. Thus here developed a type of "deglaciation by dissipation" in which the range of the stagnant marginal zone was limited to the nearest transverse crevasse or to some larger shear plane where next moraine ridge of a frontal character could develop; and where in the rear of this ridge a successive process of marginal stagnation and glacier disintegration might take place, combined with the formation of a corresponding new series of deposits and landforms. This shows that the occurrence of marginal deposits and landforms at certain intervals in the inner zone of marginal deposits and landforms has been dependent not only on climatic oscillations, but also on the texture of the glacier, on its crevasse system, and on the structure of its shear planes.

The next stage of deglaciation expressed by the formation of an undulating silty plain, with rare crevasse ridges and numerous slight stone streaks, might be assigned to a type of "deglaciation with prevailing ablation" in which few crevasse forms developed and shallow ice-dammed basins were formed in the forefield. However, in future it is probable that by melting dead ice blocks buried underneath the silt cover the flat or undulate surface of this sheet is going to be diversified in its relief.

The way deglaciation proceeds at present, as described in detail by S. Jewtuchowicz from a geomorphological point of view and by G. Wójcik with regard to its glaciological aspect, can be considered as a new stage in the process of glacier retreat. Evidence of this is the development of a clearly marked end moraine ridge, the piling up of numerous outwash fans, and the

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formation of a system of peripheral rivers closely linked with meltwater streams escaping at present from a great number of glacier gates or from artesian wells. However, the moraine ridge mentioned which — with minor erosive breaks in its chain — follows the glacier margin, is in fact a glacier wall covered by the relatively thin mantle of an ablation moraine (ice-cored moraine ridge). Similar forms have been encountered on Spitsbergen and described by Szupryczyński (1963). Thus as shown here, the ice-moraine ridge in front of the Skeidarárjökull represents the extreme part of the glacier, broken off by transverse crevasses or separated from it by ablation waters (see G. Wójcik). Therefore it seems that in the present stage of deglaciation the structure of the glacier might affect the course of its retreat. When the glacier core is melted out, the ice-cored moraine wall shrinks and becomes lower, while the moraine material is being compacted and assumes the form of relatively low ridge or sometimes, based on the principle of inversion of relief, of two moraine walls. Jewtuchowicz called attention to the important relief-forming role played by meltwater flowing over the glacier surface or escaping from englacial tunnels; by incising the ice the meltwater streams cause a compacting of the coarse moraine material in the glacier crevasses — the prospective crevasse walls. Where meltwater streams cross these ice-moraine walls, there accumulates in the stream valleys a mass of rock material extending at right angles to the future crevasse ridges. At any rate, it is evident that at present times glaciation proceeds in a very diversified manner, undoubtedly different from what it used to be in the preceding phase of deglaciation which left behind an undulating silt plain with few crevasse ridges. Maybe this new type of deglaciation goes back to the fact that the glacier has now a more mountainous character, with its surface inclined more steeply and dissected by transverse crevasses.

It is a well-known fact that deglaciation proceeds in two directions, upwards from the glacier surface and into the interior of the glacier. The former may be called frontal deglaciation, the latter deglaciation by ablation. The relation of frontal deglaciation to ablation depends on how active the given glacier happens to be. When it stagnates or breaks up into blocks of dead ice, deglaciation by ablation predominates. On the other hand, when the glacier is kept in motion, apart from ablation frontal deglaciation additionally occurs to a greater or minor degree. It must be kept in mind, however, that of whatever type the retreat of an active glacier may be, in a manner so far not defined its cause can always be traced to a reduction in the supply of new ice masses from the area of glacier alimentation.

Our investigations of the proglacial area of the retreat of Skeidarárjökull revealed that the entire forefield area including the main end moraine zone contains fossil blocks of dead ice. Incidentally it might be mentioned that dead ice blocks have contributed to the formation of the valley of the Sandgigjukvisl River. Evidence is available that the nearer the glacier, the more the buried dead ice blocks tend to coalesce. At any rate, close to the glacier margin compact ice underlies the outwash plain, the moraine ridges and the floors of meltwater channels. Thus the conclusion seems justified that under present climatic conditions a glacier decays only in its upper part and that this deglaciation is due to ablation (Similar observations were made by Price (1969) on the nearby Breidamerkurjökull glacier). Underneath a protective mantle of gradually increasing glacial, limnoglacial and glaciofluvial deposits the lower part of the glacier, of a volume and shape so far unknown, remains intact; its gradual melting sets in during later stages of deglaciation. With this aspect

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in mind we may distinguish, for a given glacier, surface deglaciation and underground deglaciation. Each of these processes creates landforms of its own. Surface deglaciation leaves marginal forms, underground deglaciation produces kettle forms. Figure 7 shows this diagrammatically. The contact line between the glacier margin and the deposits in its forefield lacks sharply marked boundaries; one rather observes some mutual interdentation which at times moves gradually up the glacier surface. But the glacier margin assumes a sharper, even a steep front at places where it is in contrast with an ice--dammed lake, or where it is undercut by meltwater streams escaping from glacier gates.

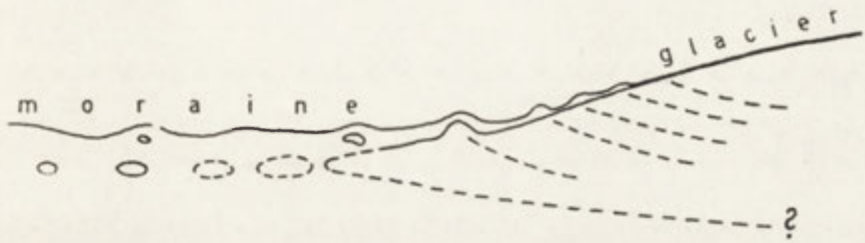

Fig. 7. Ratio of the vanishing ice mass to the increasing moraine cover

Closely linked with the process of deglaciation discussed above is the problem of moraine accumulation. If one has to assume that the bottom of the glacier below the ground surface, the moraine and other deposits left on the ground have in fact been developed supraglacially, not subglacially; thus obviously they cannot be considered as ground moraine in the true sense of the term. These landforms may rather be called an ablation moraine, here and there overlain by outwash or ice-dammed lake deposits. Further, a variety of kettle forms appear here, and often "collapse features" can be observed in the structure of these deposits.

Here a certain comparison with the area one covered by Scandinavian glaciations comes to mind. The enormous number of kettle-type depressions scattered over the area of the last glaciation seem also to indicate that the bottom of the inland ice has once extended at least to the bottom of these basins, and that the moraine forms lying higher must have been produced by surface ablation. This interpretation explains many structures observed in moraine forms which would be difficult to explain were this a ground moraine. And, what is more, this is by no means the only instance that experience gained in Iceland in the matter of glacial geomorphology can be transferred to areas of Scandinavian and, also, to Alpine deglaciations. However, this topic exceeds the scope of our present paper and is to be subject of separate later reflections.

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REFERENCES

Ahlmann, H. W.: son, Thorarinsson, S., 1943, *Vatnajökull. Scientific results of the Swedish-Icelandic Investigations 1936–37–38*, Chapter I–XI. Reprinted from Geografiska Annaler 1937–1940, 1943, Stockholm.

http://rcin.org.pl

Price, R. J., 1969, *Moraines, sandar, kames and eskers near Breidamerkurjökull, Iceland,* Trans. Inst. Brit. Geogr., 46.

Szupryczyński, J., 1963, *Rzeźba strefy marginalnej i typy deglacjacji lodowców południowego Spitsbergenu* (Sum.: Relief of marginal zone of glaciers and types of deglaciation of southern Spitsbergen glaciers), Prace Geogr. IG PAN, 39, Warszawa.

Thorarinsson, S., 1967, Forvitnilegir jökulgardar vid jadar Skeidarárjökulls, *Jökull,* 17.

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REPORT ON THE PHOTOGRAMMETRIC SURVEYS MADE BY THE POLISH EXPEDITION IN 1968 ON THE FOREFIELD OF THE SKEIDARÁRJÖKULL

Tadeusz Konysz

In the Polish Vatnajökull Expedition, as in a number of preceding Polish undertakings of this kind, apart from other scientists geodesists also took part; it was to be their job to take care of the geodetic and photogrammetric surveying, which constitutes an indispensable part of this type of research work.

The programme assigned to the geodesists in the forefield of the Skeidarárjökull was to accomplish a condensation of the local geodetic network, to supplement the geographical research by field measurements, and to carry out photogrammetric surveys on the end moraine of the glacier.

The scope of the geodetic programme had been stipulated at a preliminary meeting held in Toruń in 1968 with Professor R. Galon, the leader of the expedition.

After the expedition reached the region assigned to it, and following a preliminary field reconnaissance, a number of unforeseen difficulties came to light which made it practically impossible to carry into effect the programme of the anticipated geodetic work.

The basic difficulty appeared to be the lack of geodetic datum points within a reasonable distance. The possibility of tying in with reference points farther away was extremely difficult because no suitable means of transportation were available. The possession of a land rover failed to solve this problem, because crossing the Sula River into the area where the survey would have to be made was extremely difficult.

The geodetic network of Iceland, especially in the part covered by glaciers, is thinly scattered. The lack of building material, especially in the areas mantled by ice, made the Icelandic geodesists install their trigonometric signals on elevated remote points, many of them being almost inaccessible. The only transport in general use in Iceland which would enable a group of surveyors to move rapidly from point to point is a helicopter.

Unfortunately, the Polish Expedition did not have this means of moving from place to place. The opportunity to hire a helicopter was also limited, because in Iceland they are practically in constant use for all sorts of purposes. Under these conditions the extent of photogrammetric work and the indispensable field surveying had to be reduced to a minimum. In compensation it was decided to lay out a separate triangulation network, by means of which photogrammetric measurements could be made. As a result, 12 reference points were laid out on elevated ridges of the end moraine, more or less evenly spaced between the glacier margin and the outwash plain.

The triangulation sides of the network polygon were measured indepen-

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dently twice, by the parallax method, using separate basic grids for each side. The altitude of the 12 reference points was found by trigonometric levelling, starting and closing the survey from point Sandgigur for which the 1:50,000 Icelandic map reports the altitude. The datum points were built of stone, by mounting small pillars 70 to 80 cm high, which had a steel pipe painted white on top of them.

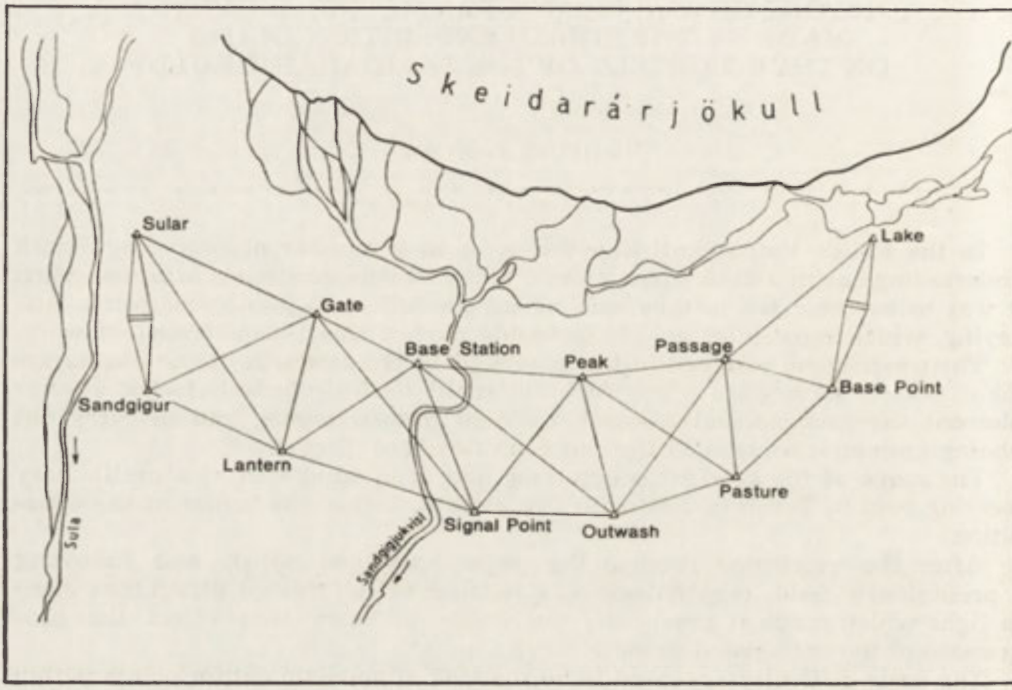

Fig. 1.

After the geodetic work was done, photo-theodolite pictures were taken and the base stations and control points were tied in. A total of 430 photo--theodolite pictures were taken from 76 base stations, and 25 control points were joined in. Areal photography was for the most part done from the crest of the end moraine, towards both the glacier and the outwash plain. Where the moraine shape appeared to be particularly intricate, photographs were taken from combination points in both eastern and western directions. In addition, two kinds of photo-theodolite pictures were taken from the glacier surface towards the end moraine. The pictures taken from longer base lines were used to increase the density of the survey grid; those from shorter base lines served to determine relief forms. Thus the pictures taken in this way covered twice the total area comprising the moraine ridge, the outwash plain and the marginal part of the glacier — all of them were extremely diversified in their relief forms. On the base stations and control points, rock piles averaging 60 cm in height were placed as marks.

All geodetic and photogrammetric surveys were made using a Zeiss photo--theodolite Theo-10 and a Zeiss photocamera 1318.

While performing the geodetic work discussed above, quite a number of

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unexpected difficulties were encountered which had not been taken into consideration in the original planning.

In the first place the climate and the field conditions prevailing in the forefield of the Skeidararjokull which made work often very difficult and, at times, completely impossible. A specific microclimate, not at all comparable with Poland's climatic conditions, here is the rule. Heavy rainfall, strong and changeable winds, dense fogs — all this greatly affected all the surveying processes. Depending on the meteorological conditions at a given time, the geodesist establishes his working programme; but any minute he must be prepared to change this plan. Fair days with little wind, had to be used for photography, because such days are rare in these regions. More often days occurred in which the fair weather lasted for barely a few hours, mostly in the morning when the wind comes from the glacier. The afternoon usually brings a change, from north to south winds; the wind force rises and clouds arrive from the sea, accompanied by rain. Precipitation does not last long, usually a day or two; then the sky clears and winds from the glacier disperse the clouds and the weather turns fair again. Then after a few days of relatively fair weather, the same cycle repeats itself.

The geodesist working in this region has to become used to these erratic meteorological conditions, and has to be prepared to use every type of weather to accomplish the work assigned to him.

Any failure to take advantage of conditions favourable to photography, delays or even prevents from performing the programmed work. Strong wind obstructs the use of even of fair days, because the camera will vibrate and the picture will be blurred. Instrument work is also thwarted: like the camera, the theodolite also vibrates and, in addition, on flat land surfaces the air assumes a wavy motion called atmospheric refraction which impairs the accuracy of instrument readings. Hence, of necessity, times like these must be used to attend to other tasks such as visual field examinations, erection of base stations and control points — or for all types of domestic work at the field station. As to geodetic measurements, they are best performed in the early morning hours then usually the atmospheric conditions are fairly stable. Even on cloudy days the morning air is clear, the cloud ceiling is mostly high, and objects in the field are clearly visible. This enables the operator to make accurate angle measurements for far distances.

But more often cloudy days prevent photography due to occasional showers and to the fog-like atmosphere they leave behind. On and off, geodetic surveying can be done on this type of day; but for the most part weather of this type is used for examining the topography, for completing records, sorting supplies and putting them in order, and for a variety of other jobs.

Under polar and subpolar conditions there should be two geodesists to perform geodetic surveying — men fully trained in the wide range of photogrammetric and geodetic work involved.

As a rule, sets of surveying equipment are standardized as to size and weight. However, one might frequently reduce weights by a suitable selection of equipment belonging to the theodolite. Under the conditions encountered on Iceland, atmospheric refraction often made it unsatisfactory to determine the lengths of base lines by the use of the measuring rod. The use of modern range-finders appeared to be more suitable.

Before leaving Poland, the whole surveying equipment to be taken should be carefully selected or specially built. Any case of unsuitable equipment may expose the geodetic group to wasted physical effort and to loss of time.

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When it comes to photography in the field, 40 or more films may be necessary when the weather is favourable. This makes it necessary to carry a suitable number of film holders which must be charged and emptied every day. Of course, this requires the use of a fixed or portable dark-room, because the bright polar night precludes doing this in the base tent. Thus it is essential to have a light-proof bag for any sort of photogrammetric work to enable film holders to be emptied and refilled in daylight. Otherwise one would have to take a full set of film holders every time, and after every day one would have to empty and refill them after returning to the dark-room at the base station — a waste of time and physical effort.

A further essential element in a surveyor's equipment are geodetic poles. For field work in Poland where suitable transport facilities are easily available, the size and weight of these poles is of little importance. But under expedition conditions, where the equipment must be carried on one's back, every kilogramme dispensable is highly important. That is why in this case it proved advisable to have the poles made of light aluminium tubes, with a steel pin at their base. Signal flags which are also essential are used rarely in Poland. But in Iceland where the landscape is a monotonous grey, coloured flags fluttering in the wind are easily seen and facilitate the marking of trigonometric signals and control points. Without this sort of flags, fairly high rock piles would have to be put up which, resembling the surface in colour are difficult to distinguish. On sunny days with considerable atmospheric refraction, mistakes can easily occur.

Office work starts with the processing of the negatives of all photo-theodolite pictures and the calculation of the geodetic network. The next operation involves composing a primary topographic map prepared from autogrammetric instrument readings. As the final stage of office work, the map is compiled cartographically, including symbols and markings.

After processing, our negatives were verified as to their protographic and photogrammetric perfection; this was done by comparing the films with a suitably chosen standard negative. In our case it was found that the negatives were properly contrasted; on most films the photographic picture was clear, with a medium-grade optical density and only slight haze. The number of spoiled pictures did not exceed 2%.

The photogrammetric quality for our photos was verified on the basis of imperfections determined in the interior and exterior orientation of the pictures. In surveys made under polar climatic conditions it was found, that due to vibrations of the camera the films were often not in a tight position against the base frame of the camera, and this fact disturbs elements in the interior orientation of the pictures. On top of this, strong winds do not only lead to vibrations of the camera, but they also affect elements of the exterior orientation causing the camera to be twisted and tilted. In order to eliminate the effect of these faults upon the accuracy of the map, every negative was first compared with a standard negative placed against the base frame of the camera. All negatives, on which the discrepancies observed in dimensions of the base frame exceeded the admissible ± 0.05 mm value, were subjected to detailed investigations. For this purpose the distances were measured on the stereocomparator between the collimating marks of the negatives to be checked and those of the standard negative. From the differences found, the necessary correction of the focal point was calculated from the respective collimating marks of the camera. Next, a graph was drawn, showing the increase in corrections to be made in focal distance due to unsatisfactory clamping of the

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negative. In our case, investigations classified the photogrammetric properties of the photo-theodolite pictures as merely adequate. The reason was, that a fairly large section of the films had faults due to unsatisfactory clamping against the base frame, and a certain number of pictures revealed distorted elements of interior orientation; these latter pictures were qualified as requiring treatment on the autograph.

The geodetic network was calculated by method of average. The mean error in the position of the level and the elevated central point of the network was ± 25 cm.

Before instrument compilation of the map was started, the geodetic and photogrammetric points were identified on the negatives of our pictures. For this purpose every single point marked in the field was identified on the photointerpretoscope from the portrayal of its topographical position. In doubtful cases the position of the point on the film was determined by graphic intersection.

The original map was compiled in 1:5000 scale by the autogrammetric method, by means of the Zeiss autograph A-7 and the stereoautograph. The land relief was entered by contour intersection at 1.25 m intervals. The actual cartography was done in the classical way, sketching the relief forms and the situation on blue cyanotype plates.

The method of ground photogrammetry used to compile the map of the forefield of the Skeidarárjökull had its good and its bad points. Among the good points should be mentioned the low cost of field work done by photographing the ground surface. The drawback of the method is the high cost of preparing the original map, and of using for this task accurate photogrammetric instruments. More economical results could have been obtained by applying the aircraft method of photogrammetry. However, during the first stages of work, i.e., for field surveying, this method requires numerous personnel, as well as money to cover the cost of aircraft services and the purchase of very expensive photographing apparatus. Nevertheless, in the long run when summing up the cost of open-air and office work, it appears that aircraft photogrammetry reduces the time needed for office work a number of times. In our case the lack of the airplane service for making photogrammetric and geodetic pictures made it impossible for the Polish Iceland Expedition to apply the most modern method of preparing a topographical map of the area under investigation.

The experience gained within recent years has resulted in a wide application of aircraft for geodetic and photogrammetric work. If our expedition had had the use of a small helicopter for the transport of surveying instruments and research apparatus, this aircraft also equipped for taking photogrammetric pictures from the air, a great many difficult problems would have been avoided and the cost of the topographic map would have been much lower.

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THE RESULTS OF THE METEOROLOGICAL INVESTIGATIONS ON THE FOREFIELD OF SKEIDARÁRJÖKULL

Gabriel Wojcik

(1) GENERAL REMARKS

Meteorological investigations were carried out in the forefield of Skeidarárjökull at the Meteorological Station (Fig. 1), situated on the end moraine just above the expedition base, by the gate of the River Sandgigjukvisl through the end moraines. The station was 114.5 m a.s.l. at a distance of 20 km from the sea and 3 km from the glacier's front. The surface of the moraine, built of basalt stone, gravel and sand, ended towards the North with gentle slope towards the glacier, and to the South with a steep scarp of 14 m towards the outwash plain stretching between the end moraine and the sea. Such a lo-

Fig. 1. Meteorological station of the Polish Glaciological Expedition to Vatnajokull, 1968, situated on the culmination of the end moraine of Skeidararjokull (h = 114.5 m a.s.l.). Photo by G. Wojcik

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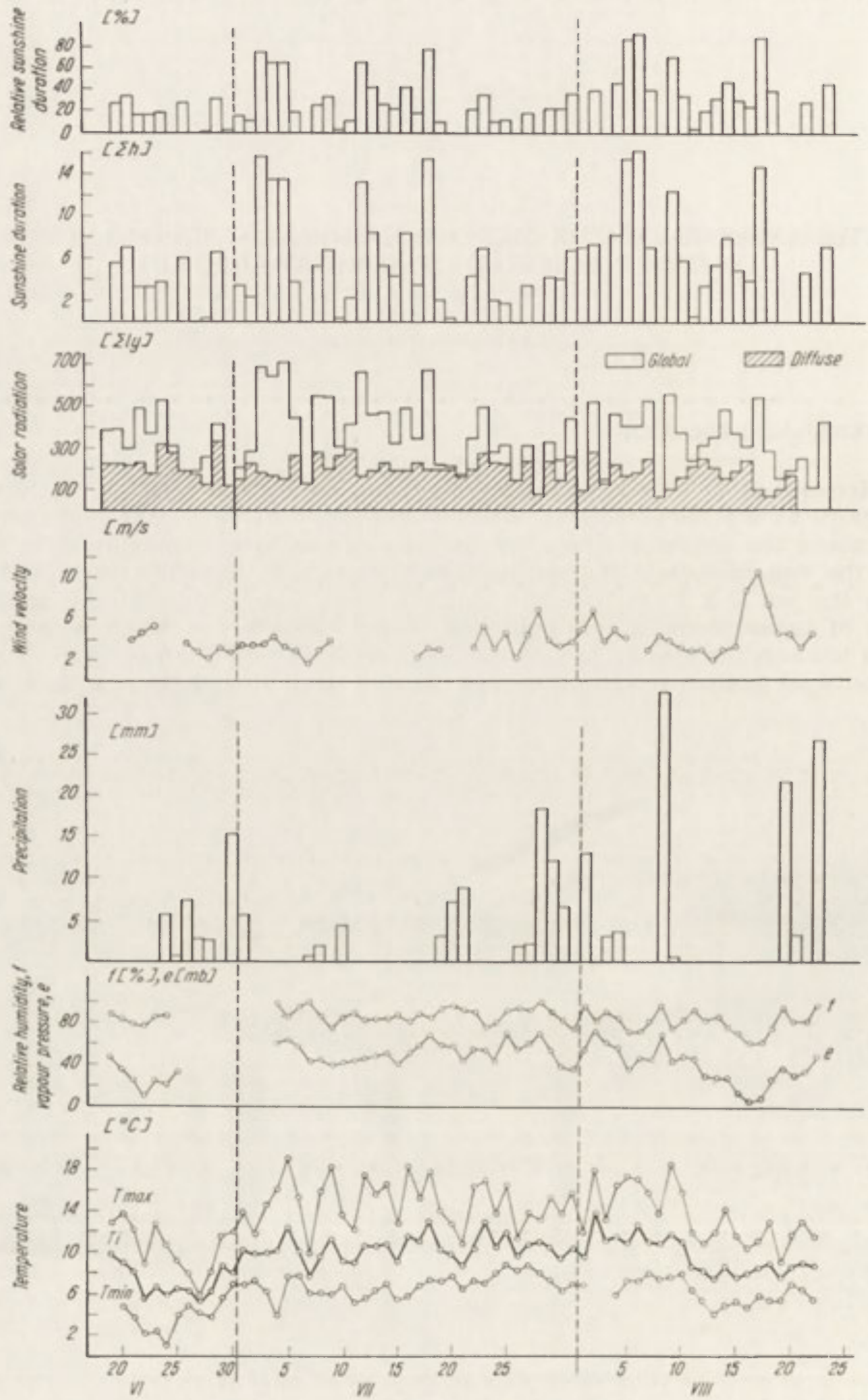

Fig. 2. Variations of the meteorological elements for the period 19th June–23rd August, 1968 in the meteorological station on the forefield of Skeidarárjokull

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cation of the station on the culmination of end moraines made it possible to record the influence of the maritime environment on the one hand and of the glacier on the other, and the materials collected permit knowledge, if only in a general way, of some effects and climatic processes taking place over the area between the sea and the glacier. The close vicinity of the station to the glacier enabled us to understand the influence of the climate in the forefield upon the deglaciation processes developing in the frontal zone of the glacier and, conversely, the influence of the glacier upon the climate of the forefield.

Meteorological investigations were carried out for a relatively short period of less than three months (20th June to 23rd August 1968). Even so, the choice of meteorological elements for investigation and the use of the automatic method of observation that helped daily variations in the investigated meteorological elements to be recorded, made the materials gathered a valuable and interesting contribution to the knowledge of the climate in the zone near the glacier.

Regular meteorological observations were not carried out owing to the small number of workers on the expedition staff. Nevertheless, observations were made every morning and every evening to control the authomatic recorders and to introduce corrections. The following recordings were carried out at the station: air temperature and air humidity, wind velocity, sunshine duration, intensity of the solar radiation, total and diffused. Measurements were also taken of precipitation and of the direct solar radiation on the surface perpendicular to the course of the sun's rays within the range of the whole spectrum and with the red RG_2 filter for the long-wave part of the spectrum. Observations of wind directions and cloudiness were carried out during field work. At a subsidiary station on the glacier air temperatures were recorded.

In the present report the amount of material expressed in figures is small, it is deliberately limited to the mean and extreme values. It is presented in Tables 1 and 2 and in diagrams (Figs. 2–16).

(2) FEATURES OF THE METEOROLOGICAL ELEMENTS

SUNSHINE DURATION

Systematic records of sunshine duration, using a Campbell-Stokes heliograph were made throughout the period from June 20 to Aug. 23, 1968. The recording cards were changed every day near midnight. Of essential importance for heliometric and solarimetric measurements is the horizontal panoramas seen from the position of the instruments; we delimited the outline of the horizon by theodolite readings (Fig. 3). The altitude of the screens of the horizon in the zone of sun setting and rising amounts as seen in Fig. 3, to about 3°, and because it is at about this solar altitude that the heliograph starts recording, the location of this instrument must be considered favourable.

The daily values of sunshine duration by hours and tenths of an hour and those of the relative sunshine duration expressed as percentage of possibility, are presented in Table 1 and Fig. 2. On the other hand, the diurnal variations of the sunshine duration are shown in Figs. 4 and 16.

During the period of our observations there were 11 sunless days (2 in June, 4 in July and 5 in August), 18 days with relative sunshine duration of $< 20\%$ and 2 days with sunshine duration of $> 80\%$ (Aug. 5 and 6). The longest daily

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TABLE 1. Daily values of the meteorological elements in the forefield of the Skeidararjokull in Iceland for the period from 19th June to 23rd August, 1968

Date	Sunshine duration		Solar radiation ly/d		Temperature °C				Humidity			Wind velocity m/s	Precipitation mm
	Number of hours	% of possible	Global I_T	Diffuse I_D	T_i	T_{max}	T_{min}	Range	Relative %	Vapour pressure mb	Humidity deficit		
VI 1968													
19	—	—	385.0	225.5	—	—	—	—	88	10.8	1.4	—	.
20	5.6	27	390.0	225.5	9.1	13.8	4.9	8.9	84	9.7	1.9	—	.
21	7.1	34	295.0	215.0	8.2	12.2	3.7	8.5	79	8.6	2.3	3.9	.
22	3.3	16	492.5	229.0	5.6	8.9	2.2	6.7	77	7.0	2.1	—	.
23	3.3	16	371.5	181.5	6.8	12.8	2.3	10.5	85	8.4	1.5	—	0.0
24	3.8	18	531.0	321.0	5.9	10.8	1.2	9.6	87	8.1	1.2	—	5.4
25	.	.	308.0	279.5	6.6	9.3	4.0	5.3	—	—	—	—	0.4
26	6.1	29	188.0	188.0	6.4	8.0	4.9	3.1	—	—	—	—	6.9
27	.	.	188.0	179.5	5.3	6.0	4.3	1.7	—	—	—	—	2.7
28	0.3	1	262.5	127.0	6.3	8.2	3.8	4.4	—	—	—	2.1	2.6
29	6.6	32	414.0	336.5	8.8	11.7	5.7	6.0	—	—	—	3.2	.
30	0.1	1	122.5	117.0	8.0	12.0	6.9	5.1	—	—	—	—	15.2
VII 1968													
1	3.4	16	210.5	152.0	10.3	13.9	7.0	6.9	—	—	—	—	5.6
2	2.3	11	284.5	227.0	9.9	11.8	7.3	4.5	—	—	—	3.4	0.0
3	15.7	76	692.5	187.5	10.0	14.6	6.3	8.3	—	—	—	3.6	.
4	13.5	66	646.0	170.5	10.2	16.1	4.0	12.1	—	—	—	—	0.0
5	13.5	66	707.0	157.0	12.5	19.1	7.7	11.4	86	12.4	2.1	3.1	0.0
6	3.8	19	447.0	263.5	10.2	15.3	7.8	7.5	95	11.8	0.6	—	.
7	.	.	135.5	129.5	7.6	9.8	6.1	3.7	100	10.4	0.0	1.6	0.4
8	5.3	26	550.0	281.0	9.7	15.8	6.1	9.7	87	10.4	1.6	2.9	1.8
9	6.9	34	543.0	201.0	11.3	18.3	6.1	12.2	75	10.0	3.4	3.7	.
10	0.3	2	306.5	263.0	9.2	13.6	6.7	6.9	88	10.2	1.4	—	4.2
11	2.2	11	414.5	294.5	9.1	12.2	5.3	6.9	91	10.5	1.1	—	.
12	13.3	67	659.5	165.0	10.7	17.4	5.6	11.8	83	10.6	2.3	—	.
13	8.5	43	460.0	187.0	10.7	15.7	6.3	9.4	85	10.9	2.0	—	.
14	5.3	27	468.5	229.5	10.9	16.5	7.0	9.5	85	11.1	1.9	—	.
15	4.3	22	321.0	189.5	9.2	12.8	5.5	7.3	88	10.2	1.4	—	.
16	8.4	43	493.0	191.5	11.7	18.3	5.9	12.4	82	11.2	2.5	—	.
17	3.4	18	346.5	231.5	11.1	15.2	6.9	8.3	91	11.9	1.3	2.3	0.0
18	15.4	79	673.5	196.0	13.0	17.9	7.4	10.5	86	12.9	2.1	3.2	.
19	2.0	10	218.0	194.0	10.3	14.0	7.3	6.7	95	11.9	0.6	3.0	2.9

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20	0.2	1	216.0	197.5	9.8	12.9	7.8	5.1	97	11.8	0.3	—	6.9
21	.	.	170.0	165.5	8.8	10.8	6.6	4.2	93	10.5	0.8	—	8.7
22	4.3	22	346.5	196.5	10.5	16.4	7.4	9.0	91	11.6	1.1	3.2	.
23	6.9	36	496.0	276.5	13.0	17.0	7.1	9.9	76	11.4	3.6	5.3	.
24	1.9	10	277.0	223.5	10.6	13.7	8.0	5.7	81	10.4	2.4	3.1	.
25	1.7	11	307.0	220.5	12.0	16.5	8.9	7.6	93	13.0	1.0	4.6	.
26	.	.	145.0	145.0	9.7	11.4	8.3	3.1	96	11.6	0.4	2.1	1.7
27	3.4	18	309.5	233.0	10.9	13.8	8.9	4.9	93	12.1	0.9	4.3	2.0
28	.	.	75.5	71.5	11.0	13.2	8.1	5.1	100	13.1	0.0	—	18.3
29	4.1	22	332.0	231.5	10.6	15.4	7.3	8.1	91	11.6	1.2	3.9	12.1
30	4.0	22	239.5	142.5	9.5	13.8	6.4	7.4	84	10.0	1.9	3.2	6.4
31	6.7	37	439.0	254.5	10.4	15.8	6.9	8.9	76	9.6	3.0	—	.
VIII 1968													
1	.	.	88.0	88.0	9.6	12.0	7.0	5.0	98	11.7	0.3	4.9	13.1
2	7.2	40	516.0	275.0	—	—	—	—	84	13.3	2.5	6.6	.
3	0.1	1	144.0	126.5	—	—	—	—	92	12.3	1.0	3.7	3.1
4	8.2	46	461.0	213.0	11.6	16.4	5.9	10.5	86	11.7	1.9	4.9	3.5
5	15.4	88	401.5	164.5	10.9	17.3	7.3	10.0	74	9.6	3.4	4.0	.
6	16.2	93	401.5	181.5	12.6	17.0	7.3	9.7	74	10.9	3.7	—	.
7	7.1	40	521.5	244.5	11.0	15.8	8.0	7.8	82	10.7	2.4	2.8	.
8	.	.	58.5	58.5	10.8	13.8	7.5	6.3	99	12.8	0.2	4.3	31.3
9	12.4	72	549.0	96.0	11.8	18.4	7.6	10.8	76	10.5	3.3	3.8	0.5
10	5.8	34	366.0	153.5	11.0	15.8	7.9	7.9	84	11.0	2.1	3.0	.
11	0.4	2	235.5	206.5	8.6	11.9	6.4	5.5	94	10.6	0.6	2.8	0.0
12	3.4	20	309.5	246.0	8.1	10.9	5.3	5.6	85	9.1	1.7	2.8	.
13	5.4	32	348.5	200.5	7.4	11.9	4.0	7.9	87	9.0	1.3	1.9	.
14	7.9	48	479.0	153.5	8.6	14.1	4.9	9.2	78	8.8	2.4	2.8	.
15	4.9	30	366.5	185.5	7.6	11.6	5.2	6.4	71	7.4	3.0	3.2	.
16	3.9	24	303.5	233.5	7.9	10.5	4.6	5.9	62	6.6	4.0	8.7	.
17	14.7	90	536.0	87.5	8.5	11.0	5.8	5.2	62	6.9	4.2	10.3	.
18	7.0	39	277.0	66.0	8.9	12.8	5.4	7.4	76	8.6	2.8	7.2	.
19	.	.	143.5	94.0	7.5	9.0	5.3	3.7	95	9.8	0.6	4.4	21.6
20	.	.	186.5	173.5	8.5	11.6	6.8	4.8	83	9.2	1.9	4.3	3.2
21	4.6	29	246.5	—	8.8	12.3	6.3	6.0	83	9.4	1.9	3.1	.
22	.	.	98.0	—	—	—	—	—	—	—	—	4.4	26.6
23	7.1	46	421.0										

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sunshine duration occurred on Aug. 5 lasting 16.2 hours, i.e., 95% of the possible sunshine duration.

The sunshine duration was highest in the first 10-days of August when the sum of sunshine hours was as great as 72.4 hours, the equivalent of 41% of the possible sunshine duration. The lowest sunshine duration was in the last 10-days of June — only 30.6 hours, i.e., 15% of that possible. Only for July was it possible to obtain a full monthly cycle of observations, and therefore the values for this month may be used in comparison with relevant data cited in literature. Thus we found that in Skeidarárjokull forefield the total of sun-

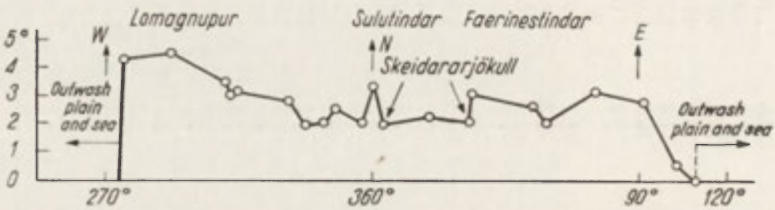

Fig. 3. Physical horizon of the meteorological station on the forefield of Skeidarárjokull, measured with the theodolite

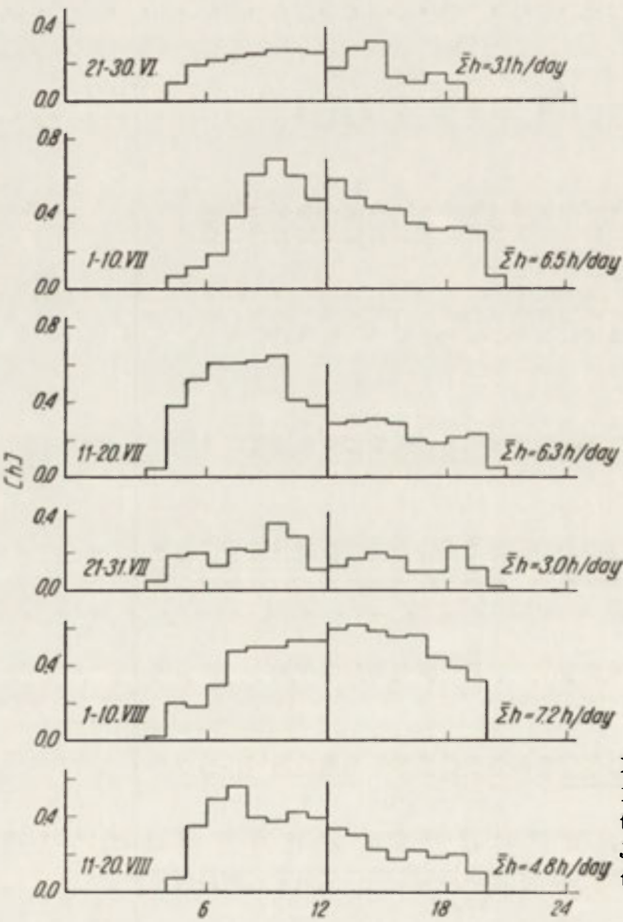

Fig. 4. Mean daily variations (for 10-day periods) in sunshine duration (in hours) for the period 21st June–20th August, 1968 in the meteorological station on the forefield of Skeidarárjokull

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shine hours for July 1968 was 160.7, representing 27.2% of the possible total for this month. From material published by Einarsson (1966) it appears that at Reykjavik the mean July sunshine duration for the 1957–1960 period was 215.3 hours. For six full 10-day periods, from June 21 to Aug. 20, our observations revealed a total sunshine duration of 311.3 hours which equals 28% of possible (Table 2).

TABLE 2. Ten day sums or mean values of meteorological elements in the forefield of the Skeidararjokull in Iceland for the period from 21st June to 20th August, 1968

Element		VI	VII			VIII		21 VI–
		21–30	1–10	11–20	21–31	1–10	10–20	20 VIII
Sunshine duration	Number of hours $\sum_n$	30.6	64.7	63.0	33.0	72.4	47.6	311.3
	% of possible	15	32	32	16	41	29	28
Solar radiation $\sum$ ly	Global	3173	4523	4270	3137	3507	3185	21,796
	Diffuse	2175	2034	2074	2160	1600	1646	11,689
Temperature °C	T_i	6.8	10.1	10.6	10.6	11.3	8.2	9.6
	T_{max}	9.9	14.8	15.3	14.3	15.8	11.5	13.6
	T_{min}	3.9	6.5	6.5	8.4	7.6	5.4	6.4
	Range	6.1	8.3	8.8	6.7	8.1	6.0	7.3
Humidity	Relative %	83	89	88	89	85	79	86
	Vapour pressure mb	8.2	10.9	11.3	11.4	11.4	8.6	10.3
	Humidity deficit	1.8	1.8	1.6	1.5	2.1	2.2	1.8
Wind velocity m/s		—	3.1	2.8	3.8	4.2	4.8	3.7
Precipitation mm		33.2	12.0	9.8	49.2	51.5	51.4	207.1

The mean ten full day diurnal variations of sunshine duration are presented in Fig. 4, while the mean diurnal variation for the whole period of our observations is shown in Fig. 16. It can be seen from these curves, especially those representing periods of limited frequency of cyclonic weather conditions and a predominance of sunny weather which periods promoted the formation of convective clouds, that the diurnal course of sunshine duration is irregular and asymmetrical; maximum diurnal sunshine duration occurs at different intervals before noon. At an average, the maximum for the whole period falls from the 9 h to 10 h interval and lasts 0.47 hours. Some time about noon a distinct decrease in sunshine duration used to set in, caused by the formation of convective clouds and by the movement of a chain of convective clouds above the meteorological station, travelling from the shore line towards the glacier, driven by a sea breeze (Fig. 17). The latter part of the day was characterized by increased clouding and reduced sunshine duration compared with the first half of the day. This phenomenon causes the diurnal pattern of sunshine duration to be asymmetrical: the average for the whole period of observations indicates that the first half of the day receives 55% and the second half 45% of the sunshine duration.

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DIRECT SOLAR RADIATION

Our measurements of direct solar radiation upon a perpendicular surface to the solar rays were made by means of the Janiszewski thermoelectric actinometer No. 106660 coupled with a Kipp millivoltmeter No. 39480. This system was calibrated by means of the Linke-Feussner actinometer No. 80 owned by the Climatology Department of the Copernicus University at Toruń. The above measurements were made under a clear sky or with only slight clouds, but only at times when the angular distance of the cloud from the sun was at least 30°. Our measurements of the intensity of the total solar spectrum and of its "red", longwave section were made using a Schott RG_2 filter 2 mm thick.

All our measurements were made in series, at half-hour intervals of GMT time, and afterwards converted to true local solar time. Each series comprised four readings with regard to total solar radiation and two referring to the long-wave part of the spectrum. From these readings there have been calculated the mean values attached to the moment found in the first and the last readings of the given series; and the values thus calculated were used as a basis for calculating solar radiation intensities. Following each actinometric measurement, additional measurements were made for determining the intensity of global radiation upon a horizontal surface, and of the radiation reflected from the moraine surface, using the Janiszewski pyranometer for this purpose. Parallel with this were observations of clouds (the degree to which the sky was covered with clouds and the kind of clouds observed), of wind direction and of visibility.

Measurements of direct solar radiation were made during 10 days, but only from 4 days there have been registered longer stretches of observations. A total of 131 measurement series has been registered and they became the base for the following calculations:

(1) the intensity of direct radiation penetrating the atmosphere:

(a) for the whole range of the spectrum which covers the range from 0.3 to 3.0 μ (7.14), or the range from 0.2 to 4.0 μ (13),

(b) for the "blue" short-wave part of the spectrum, with wave lengths $\lambda < 0.625\,\mu$,

(c) for the "red" long-wave part of the spectrum, with wave lengths $\lambda > 0.625\,\mu$.

(2) the extinction of solar radiation; as standard of extinction we adopted F. Linke's "new" turbidity factors (2, 3, 7, 11, 13, 14, 17) which were calculated for:

(a) the whole radiation spectrum (Θ_c),

(b) short-waves (Θ_k),

(c) long-waves (Θ_d).

The spectral composition of direct solar radiation is presented in Fig. 5. The minimum long-wave radiation, $\lambda < 0.625\,\mu$, coincides with the period from 10 h a.m. to 1 h p.m., amounting to 64% with regard to the radiation of the full spectrum. With decreasing solar altitude the percentage share of long-wave radiation increases due to the more rapidly growing losses in the short-wave part of the spectrum. By 8 h p.m. the share of long-wave radiation is as much as 82%. On the average, for the period from 6 h a.m. to 8 h p.m. the share of long-wave radiation is 69% while the share of short-wave radiation is 31% with regard to the radiation of the whole spectrum.

The diurnal course in the pattern of spectral radiation shows a certain asymmetry: in the first half of the day the part of long-wave radiation is slightly greater than in the second half. Thus, at 6 h a.m. in the morning it amounts

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to 74%, and at the same solar altitude in the afternoon, that is at 6 h p.m., it is 72%. This radiation pattern is due to the corresponding formation of water vapour during the day. The vapour pressure increases gradually, from its morning minimum to its maximum which is attained during the afternoon hours. Because the absorption of solar radiation caused by water vapour affects more intensively rays of greater wave lengths, the percentage share of this type of radiation decreases during the afternoon.

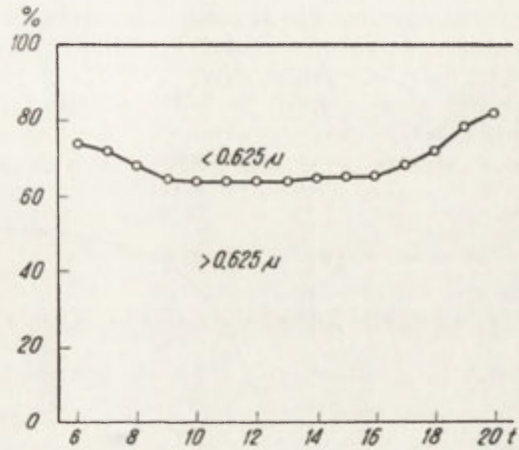

Fig. 5. Mean daily variation of the spectral composition of the solar direct radiation at the front of Skeidararjokull according to measurements from 5, 6, 16, 18 July and 5, 6, 10 August, 1968

Examples of the diurnal courses of direct radiation under different conditions of air transparencies are presented in Fig. 6, while Fig. 7 shows the mean diurnal course computed from all measurements. As a rule, the sections of the curves representing morning periods are more regular and steeper than the sections indicating afternoon periods. The latter show smaller or larger irregularities and disturbances, caused by decreasing transparency of the air due to an increase of water vapour and aerosol in the atmosphere.

When the sky is clear or slightly clouded, a local air circulation described by the author (Wójcik 1970) is forming above the Skeidarárjökull forefield,

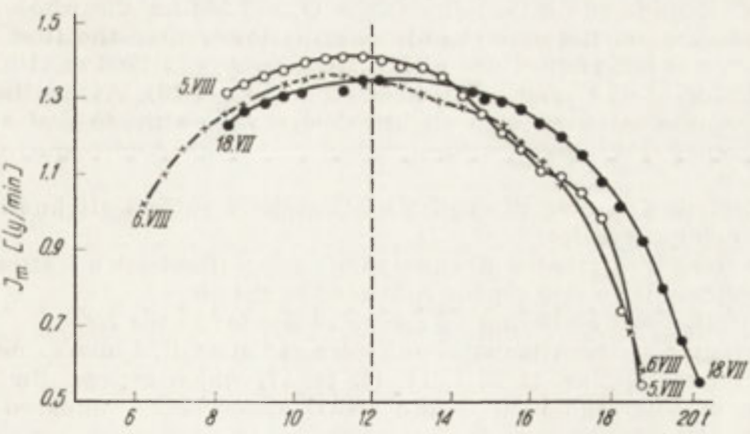

Fig. 6. Daily variations of the direct solar radiation I_m at the front of Skeidararjokull for 18 July, 5 and 6 August, 1968

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and this circulation renders practically impossible any stagnation in the atmosphere and any development of calms. During the day the wind velocity increases, very similar to the increase in water vapour pressure, from its morning minimum to its maximum in the afternoon hours. The wind velocity adds considerably to the air turbulence which evolves at the rate the surface of the forefield becomes warmer. Sunny weather promotes drying of the soil surface which contains a large share of the finest fractions. Thus it happens on days without or with little clouding that all over the Skeidararjokull forefield dust storms develop which lead to a marked increase in air turbidity and a corresponding decrease in solar radiation. A good example illustrating these conditions may be seen on Aug. 5, 1968 (Fig. 6) when in the time from 1^{45} to 4^{15} p.m. an abnormal drop in radiation occurred, just due to a strong increase of turbidity clearly seen — even with the naked eye.

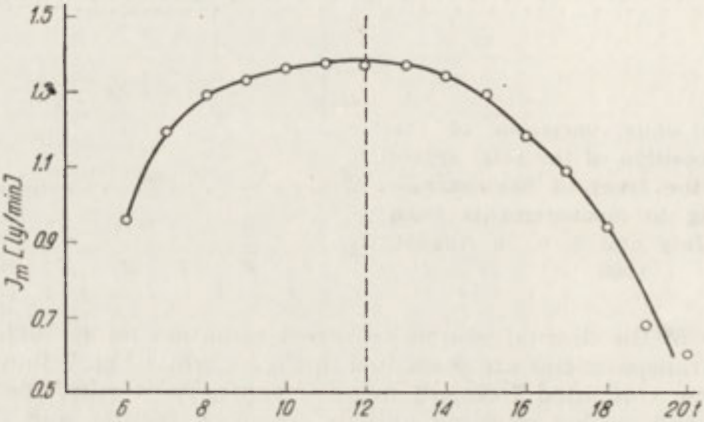

Fig. 7. Mean daily variation of the direct solar radiation I_m at the front of Skeidararjokull according to measurements from 5, 6, 16, 18 July and 5, 6, 10 August, 1968

The highest values determined for direct radiation are: 1.425 and 1.411 ly/min. The former figure was obtained on July 5, 1968 at 1^{46} p.m., with the sun at 45.2° altitude and a turbidity factor $\Theta_c = 1.243$ for the whole spectrum. The latter figure, on the other hand, which is lower than the first by barely 0.024 ly/min, was determined one month later, on Aug. 5 1968 at 11^{15} a.m., with a solar altitude of 42.4° and a turbidity factor $\Theta = 1.261$. As can be seen the second maximum occurred at a slightly lower solar altitude and at a worse air transparency. Both these values control each other and seem to be trustworthy.

Solar radiation passing through the atmosphere suffers attenuation caused by the following precesses:

(a) scattering by particles of atmospheric gases (Rayleigh's scattering),

(b) absorption by water vapour contained in the air,

(c) absorption and scattering by matter suspended in the air.

As a measure of the attenuation of solar radiation F. Linke's "new" turbidity factors were applied (2, 3, 7, 11, 13, 14, 17) which express the degree of attenuation of solar radiation in the actual atmosphere, compared with the attenuation occurring in an ideal atmosphere, i.e., an atmosphere devoid of suspended matter and containing a definite amount of water vapour equalling 1 g/cm².

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The turbidity factor can be expressed by the equation:

$$\Theta = \frac{\log I_0 - \log I_m}{\log I_0 - \log I_{mid}}$$

where:

Θ — turbidity factor,

I_o — solar constant,

I_n = intensity of direct radiation measured for a given mass,

I_{mid} = intensity of direct radiation in an ideal atmosphere.

The optical masses of the atmosphere were calculated for given solar altitudes with due consideration given to Bomporad's corrections for atmospheric pressure, and the measured intensity of radiation I_m was reduced to the mean distance of the earth from the sun. On the other hand, the turbidity factors were calculated by the use of suitable tables prepared by Słomka (Słomka 1959).

As mentioned before, the turbidity factors were calculated for the whole solar spectrum (Θ_c), for its short-wave part (Θ_k) and its long-wave part (Θ_d). The most important data referring to these factors are presented in Table 3.

TABLE 3. Extreme and mean values of turbidity factors in total range of spectrum (Θ_c), in its short-wave (Θ_k) and long-wave (Θ_d) parts, as determined at the snout margin of Skeidarárjökull by measurements made on July 5, 6, 16 and 18 and on August 5, 6 and 10, 1968

	Θ_c	Θ_k	Θ_d
Maximum	2.239	2.134	2.438
	Aug. 5, 6^{43} p.m.	Aug. 5, 6^{43} p.m.	Aug. 5, 6^{43} p.m.
Minimum	1.227	1.233	1.028
	July 5, 8^{44} a.m.	July 18, 8^{13} p.m.	Aug. 5, 8^{17} a.m.
Mean	1.447	1.556	1.362

The mean diurnal course of the turbidity factors, calculated from all measurements made, is presented in Fig. 8. As can be seen, these factors are rather low-evidence of a high transparency of the atmosphere and of a relatively small extinction of solar radiation in Iceland.

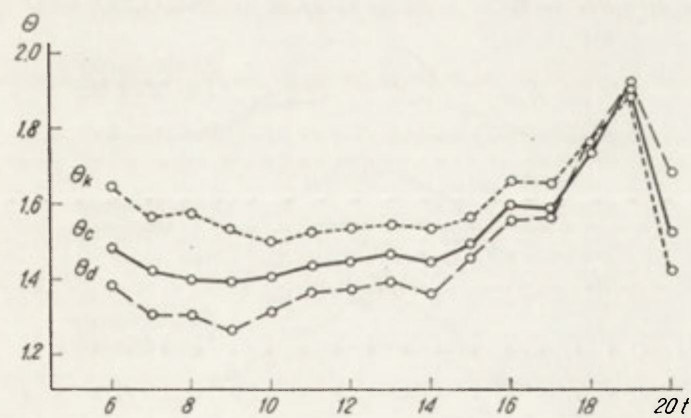

Fig. 8. Mean daily variation of the turbidity factor Θ at the front of Skeidararjökull according to measurements from 5, 6, 16, 18 July and 5, 6, 10 August, 1968

Θ_c— for the whole solar spectrum

Θ_k— for the short-wave part of solar spectrum ($\lambda < 0.625\ \mu$)

Θ_d— for the long-wave part of solar spectrum ($\lambda > 0.625\ \mu$)

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GLOBAL AND DIFFUSE SOLAR RADIATION

The recording of global and diffuse solar radiation upon a horizontal surface was made by the use of Kipp thermopiles coupled with a clockdriven recording millivoltmeter. This system was calibrated by comparing it with the indications of a Kipp pyrheliometer Linke-Feussner No. 80, borrowed from the Climatology Department of the Copernicus University at Toruń.

The recording of global radiation was carried out between 19th June and 23rd Aug., that of diffuse radiation between 19th and 20th Aug. 1968.

The day-by-day course of the daily sums of global radiation (I_T) and diffuse radiation (I_D) is presented in Table 1 and Fig. 2. The daily sums of I_T vary, from 58.5 on Aug. 8 to 707.0 ly on July 5, with a mean value derived from six full 10-day periods amounting to 357.3 ly/day.

In literature of this topic there exist only one detailed paper on the global solar radiation in Iceland, — in Reykjavik where this element was recorded in the years of 1956–1960 (Einarsson 1966). According to this source, the mean 4-year sum of global solar radiation for July was 13,020 ly/month. The analogous value calculated by us for July 1968 for the Skeidararjokull was 11,970 ly.

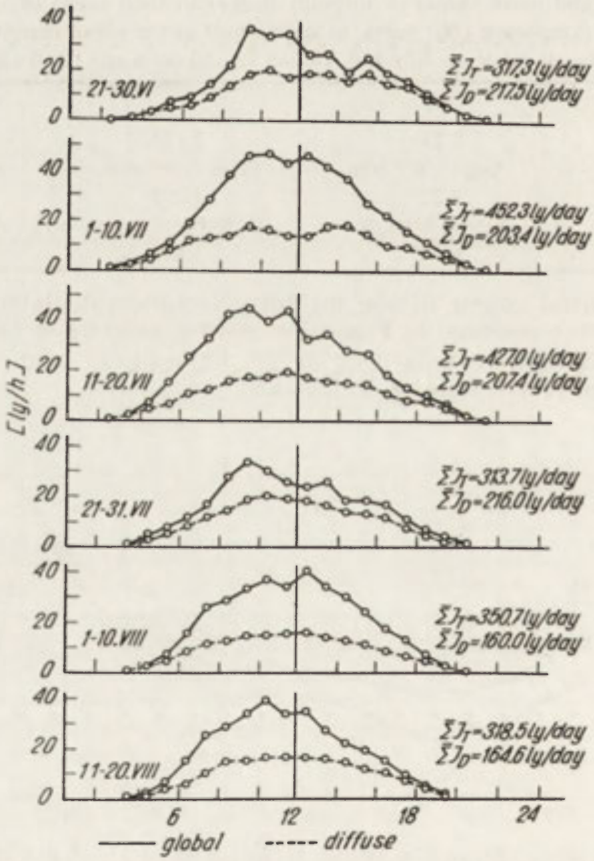

Fig. 9. Mean daily variations (for 10-day periods) in global solar radiation I_T and diffuse solar radiation I_D for the period 21st June–20th August, 1968 in the meteorological station on the forefield of Skeidararjokull (Σ ly/h)

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With regard to the amount of heat sums received, the first and second 10-day period of July mere particularly favourable, whereas the last 10-day period of June (Table 2) was rather handicapped. In the diurnal course of the majority of 10-day period and, on the average, of the whole period of observations (Figs. 9 and 16), the maxima of global solar radiation occurred during the late morning hours, mainly in the interval from 9 h and 10 h, and they averaged 38.3 ly/h. It should be remembered, that the maximum of sunshine duration also takes place at this same hourly interval. Around noon the greater part of the 10-day periods show in their diurnal course a decrease in radiation values, so that the morning period averages 55% and the afternoon period 45% of the daily sum of heat received.

Our recording of diffuse radiation (I_D) was maintained for the period from June 19 to August 20. The daily sum of diffuse solar radiation (Fig. 2) appeared in the range from 58.5 (Aug. 8) to 336.5 ly/day (June 29), with a mean daily value of 191.6 ly for the six full 10-day periods. August 8 was a rainy day on which the sum of diurnal precipitation reached its maximum, and thick stratiform clouds were covering the sky. Throughout this day there prevailed only diffuse radiation, considerably reduced. On the other hand, the

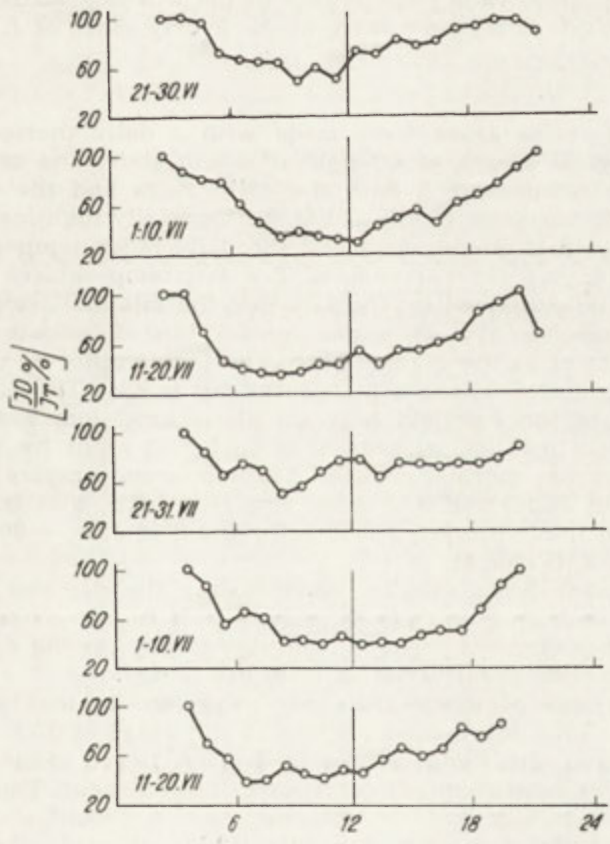

Fig. 10. Mean daily variations (for 10-day periods) in the ratio of diffuse solar radiation to global solar radiation $\frac{I_D}{I_T} \times 100\%$ for the period 21st June–20th August, 1968 in the meteorological station on the forefield of Skeidarárjökull

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highest diurnal sum of solar radiation in June resulted from the longer duration of this day and from moderate clouding, with clouds *Cu* and high clouds *Ci* which contributed to the process of diffusion of solar radiation.

The 10-day sums of diffuse radiation I_D are presented in Table 2. The highest values were observed in the last 10-day period of June and the third 10-day period of July — thus at the time when the sums of global solar radiation were lower. The diurnal courses of I_D are shown in Figs. 9 and 16. The diurnal curves of the course of I_D are enough regular, and the mean course for the whole period of observations shows some asymmetry: the maximum hourly value occurs in the morning from 10 h to 11 h, and its mean value is 17.8 ly/hour. Of the sums of diffuse radiation I_D, 51% arrive before noon and 49% in the afternoon. For the six full 10-day periods the sum of global radiation I_T is 21,795.0 ly, that of diffuse radiation I_D 11,706.0 ly. For the whole period of observations the ratio of I_D to I_T is 54%. This means that the share of the second component, i.e., direct solar radiation, has a mean value of 46%.

The value of the quotient I_D/I_T, expressed as a percentage, shows a diurnal course (Figs. 10 and 16) which depends on the altitude of the sun and weather conditions. We note highest values approaching 100% at the time of sunrise and sunset, while these values are lowest in the morning hours, between 9 h and 11 h. During this latter time interval the hourly share of I_D is some 45%.

TEMPERATURE

Recordings of temperature were made with a daily thermograph placed in the meteorological screen at a height of 1.5 m above the stone surface of the moraine. Air temperatures, both the daily mean and the extreme, have a great day-to-day variation (Table 1, Fig. 2). Especially high leaps are characteristic of the maximal temperature and the daily mean temperature. On the other hand, the day-to-day variation of minimal temperatures is more smoothed. This may be explained by the fact that the minimal temperatures usually establish themselves at night under the influence of cool air blowing down from over the glacier as the glacier wind. The temperature of the air moving down the melting glacier changes little from day to day. The air passing over the narrow belt of the forefield between glacier and meteorological station (about 3 km broad) that has already been cooled at night by radiation, does not undergo any great thermal changes. The minimum temperatures fluctuate between 1.2° (24th June) and 8.9° (25th and 27th July), that is, in the range of 7.7°. The mean minimum temperature for the 21st June — 20th August period amounts to 6.3° (Table 2).

Maximum temperatures usually occur during the daytime. They result from the solar conditions and cloudiness, as well as air circulation. As the combination of these factors changes from day to day, so the daily maximum temperatures also show great variation from day to day.

The daily maxima of temperature are comprised in the interval of 6.0° (7th June) — 19.1° (5th July), thus varying in the range of 13.1°. Climate descriptions stress that the temperatures in Iceland never exceed 20° (O'Dell, Serebryannyy). Our measurements confirmed this statement. The highest maximal temperatures usually occurred in sunny weather about noon, most often between 12 a.m. and 1 p.m., e.g., 5th July (19.1°), 9th July (18.3°), 9th Aug. (18.4°). On the other hand, the low maxima were found in advectional, cloudy and rainy weather, e.g., 27th June (6.0°), 7th July (9.8°), 19th Aug. (9.0°). The mean maximum temperature was 13.6°.

The daily amplitudes of temperature fluctuate between 1.7° (27th June)

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and 12.4° (16th July). High amplitudes, in the range of 10° and over, however, do not occur frequently. During our investigations they occurred ten times.

The daily means of temperature, as regards day-to-day variation, take a middle position between the maximal and the minimal temperature (Fig. 2). They fluctuate between 5.3° (27th June) and 13.8° (2nd July), that is in the range of 8.5°. The mean temperature for the period 21 VI–20 VIII is 9.6°. The mean temperature for July is 10.4°. According to the distribution of July temperature in Iceland (Serebryannyy), it is 10° in the region of our investigations. It may be worth while to add for the sake of comparison that the mean July temperature, calculated for the years 1964–1967 was 11.2° according to the data of Vedurstofa Islands at the Kirkjubaejarklaustur (situated about 45 km to the West of the Skeidarárjökull), and the mean for the summer time (June, July, August) was 10.2°. The corresponding data at the Hofn, situated ca. 100 km to the East of the Skeidarárjökull were 9.6° and 9.2° respectively.

The daily variation in temperature for 10-day periods is presented on Fig. 11 and 16. Some of the periods show a better exposed daily variation, with

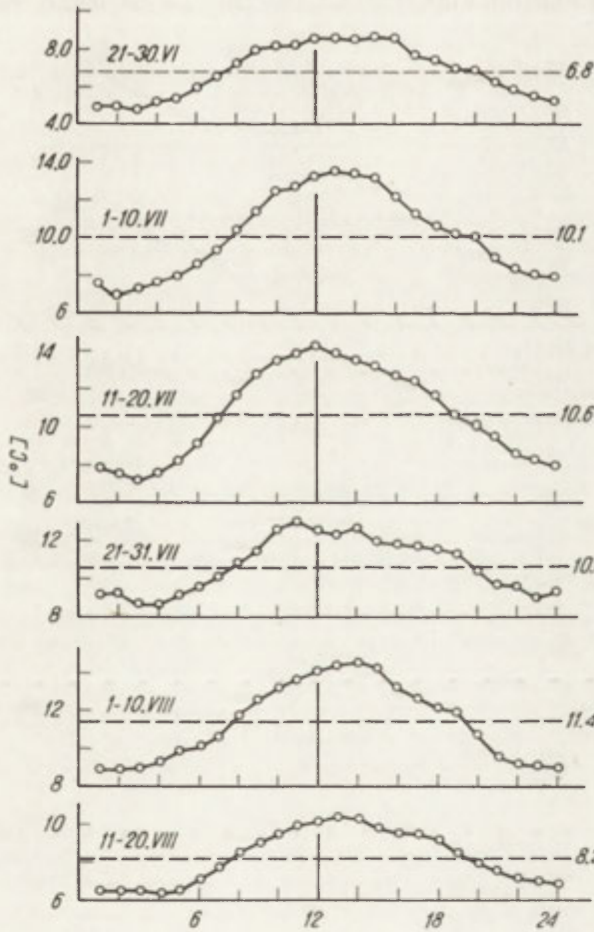

Fig. 11. Mean daily variations (for 10-day periods) in air temperature for the period 21st June–20th August, 1968 in the meteorological station on the forefield of Skeidarárjökull

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a greater differentiation between day and night temperatures. Such variations belong to the first and the second 10-day periods of July. A comparison between the corresponding curves and diagrams in Fig. 2 shows that these periods were characterized by high insolation, large amounts of radiation and little precipitation. The prevailing weather at that time was of the solar type. On the other hand, some of the other 10-day periods, e.g., the third ones of June and July display a less varied daily variation in temperature. A comparison with other elements proves that there was frequent and heavy precipitation, reduced sunshine duration, and less solar radiation. The prevailing type of weather then was advectional.

AIR HUMIDITY

The mean daily values of relative humidity are spread in the high interval, namely, from 62% to nearly 100% (Table 1, Fig. 2). The mean humidity for the whole period equals 86% (Table 2). The daily variation of the relative humidity is presented on Figs. 12 and 16. The lowest mean value of relative

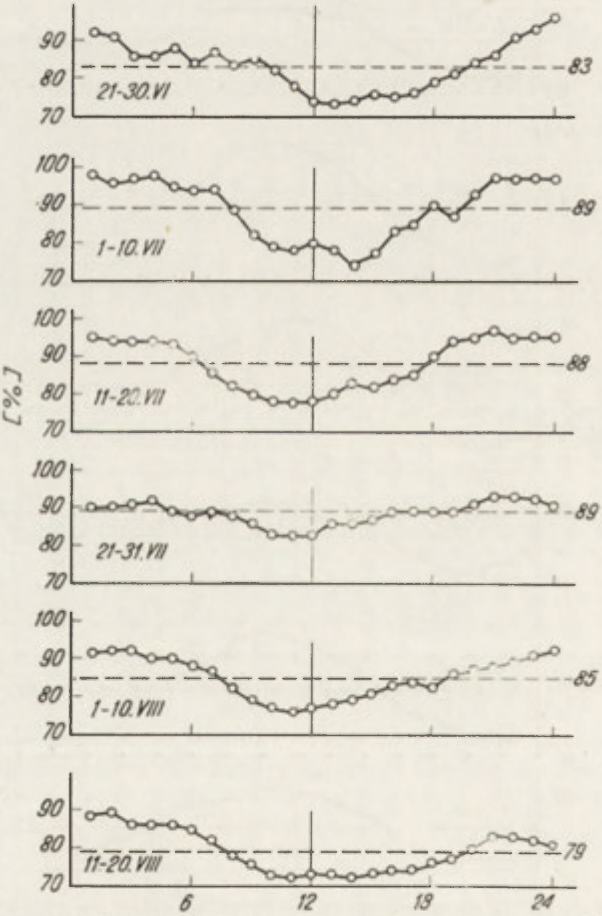

Fig. 12. Mean daily variations (for 10-day periods) in relative humidity for the period 21st June–20th August, 1968 in the meteorological station on the forefield of Skeidararjokull

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humidity amounting to 78% is to found in the daily variation between 11 a.m. and 2 p.m., the highest value, on the other hand, amounting to 92% between 11 p.m. and 2 a.m. (Fig. 16).

The water-vapour pressure was calculated from air temperature and relative humidity by means of psychrometric tables. The daily mean values of water-vapour pressure fluctuate between 6.6 mb (16th Aug.) and 13.3 mb on 2nd Aug. (Table 1, Fig. 2), the mean value for the whole period being 10.3 mb (Table 2). The 10-day mean values (Table 2) fluctuate between 8.2 mb and

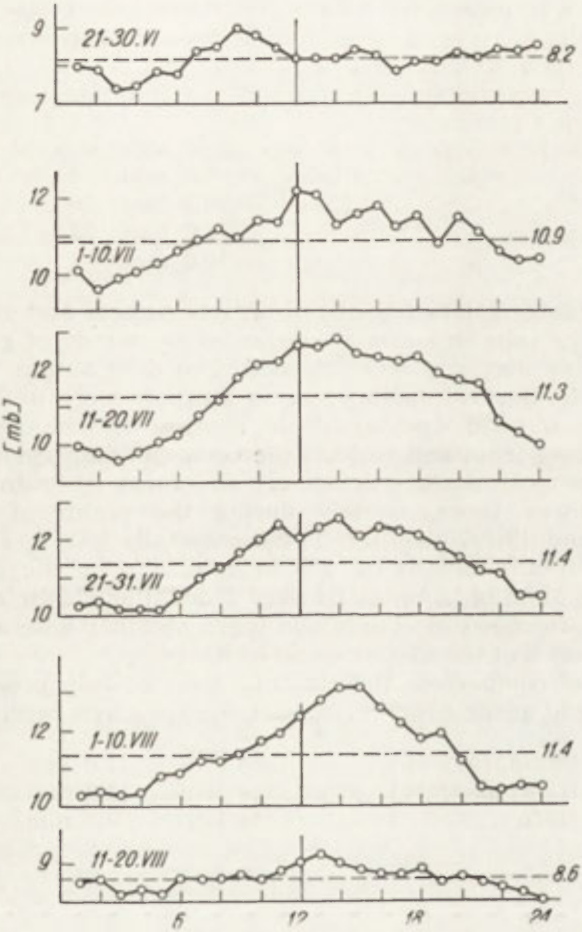

Fig. 13. Mean daily variations (for 10-day periods) in water vapour pressure for the period 21st June–20th August, 1968 in the meteorological station on the forefield of Skeidarárjokull

11.4 mb. In the daily variation (Fig. 13), the highest values occurred in the afternoon (in accordance with the temperature distribution) between 1 and 2 p.m. The mean values calculated for the same time interval during the whole period equalled 11.2 mb. The lowest values occur in the morning hours, that is at 3 a.m. (9.3 mb). In the daily variation of vapour pressure assymetry is visible: the greater values occur in the afternoon, the lower values, however, before noon, in accordance with an analogical daily distribution of air temperature.

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The water-vapour pressure reaches high values often close to the saturated vapour pressure. This fact is proved by a humidity deficite. Its daily values fall within the interval of 0.0 mb (7th and 28th July) — 4.2 mb (17th Aug.), the mean value for the whole period being 1.7 mb. The small humidity deficit determined the limited water evaporation.

PRECIPITATION

Precipitation was measured with a Hellmann rain gauge placed on the ground. During the 63 days of observations, precipitation occurred on 32 days, giving 207.1 mm water (Tabl. 1, Fig. 2).

The frequency distribution of classes of precipitation accepted in climatological literature is as follows:

daily precipitation	=	0.0 mm	6
„ „	⩾	0.1 mm	26
„ „	⩾	1.0 mm	23
„ „	⩾	10.0 mm	7

The heaviest daily rainfall occurred on 8th August and reached 31.3 mm. It was very heavy rain and was accompanied by winds of great velocity, so that the tents gave way and were flooded. The distribution of rainfall was very characteristic. Several spells of rainy weather were divided by spells of moderately good or good weather. This rhythm in the weather sequence undoubtedly follows from and reflects the corresponding cyclogenetic rhythm. The precipitation took almost exclusively the form of rainfall. Snow was recorded only three times, namely during the nights of 20th/21st June, 23rd/24th June and 19th/20th July. These snowfalls left a short-lived snow cover in the mountain regions ca. 200 m a.s.l., and on the glacier the snow covered even the ablation zone at its very front, though for a short time. On 10th July in the afternoon in cloudy and foggy weather with a drizzle a single thunder roll in part E of the glacier could be heard.

For the sake of comparison the monthly sums of July precipitations at the Skeidarárjökull and at the neigbouring stations are given here:

Skeidarárjokull	1968	17.0 mm
Kirkjubaejarklaustur	1964–1967	112.9 mm
Hofn	1964–1967	39.3 mm

WIND VELOCITY

The wind velocity was recorded with an electric contact anemograph giving signs every 500 m of the wind flow. Owing to a temporary break-down of electric contact in the head, there were gaps in the observations, particularly in June. Therefore, in the present work anemograms for July and August were taken into account.

The mean daily values of wind-velocity are presented in Table 1 and on Fig. 2. The region investigated was characterized by great air movement. Periods of still air occurred rather seldom, usually at night and lasted at most a few hours. The mean daily values of wind velocity fluctuated between 1.6 m/s (7th July) and 10.3 m/s (17th Aug.). Also on 16th and 18th August high wind velocities were recorded. The daily means amounted to 8.7 and 7.2 m/s respectively.

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The daily variation in wind velocity is presented on Fig. 14 and 16. On Fig. 14 against each of the 10-day curves the number of observational days is given in brackets. These variations are typical and interesting, being the reflection of local air circulation in this region. The highest mean wind velocity in the daily variation (Fig. 16) occurs in the afternoon hours, namely between 3 p.m. (5.7 m/s) and 4 and 5 p.m. (5.8 m/s), the lowest was recorded at 7 a.m. (2.2 m/s).

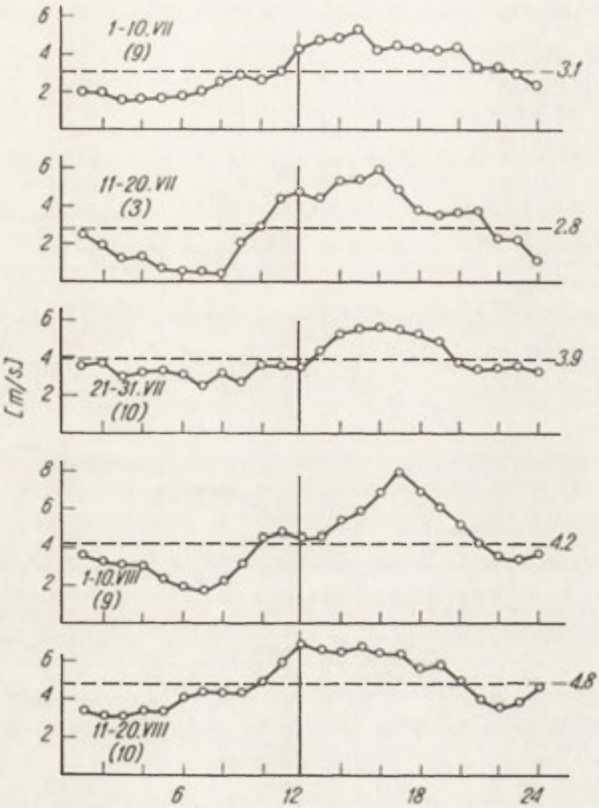

Fig. 14. Mean daily variations (for 10-day periods) in wind velocity for the period 21st June–20th August, 1968 in the meteorological station on the forefield of Skeidararjokull

From the daily maximum in the afternoon on, the wind velocity gradually diminishes till it reaches the minimum in the morning hours. Since then the velocity increases up to its afternoon maximum (Fig. 16). Thus, the mean daily variation in wind valocity for the whole period shows much regularity, with one maximum in the afternoon and one minimum in the morning. During the night the wind at the station blows from the glacier and has the component N; it extends a further over the forefield and joins the on-shore breeze. During the daytime, however, there is a sea breeze with the direction component S (mainly SSW). The mechanism of this circulation is discussed on page 178.

The curves of daily variation in wind velocity for some of the 10-day periods, having predominantly solar weather, show a reduction of values about

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the noon. It occurs when the see breeze (general direction S) meets the glacier wind (general direction N). At this time, moreover, an increase of cloud is observed, which cancels the warming up of the dark ground, and mitigates the convection and turbulence. As a result wind velocity slightly diminishes. The instance of wind on 16th and 17th August mentioned before was a typical fohn blowing from the Vatnajokull down the valley of the Skeidararjokull and onto the far forefield. The character of the wind is proved by its velocity, thermo-hygrometric changes and other weather effects, as for instance cloudiness. The wind velocity steadily increased from 16th August on until it reached

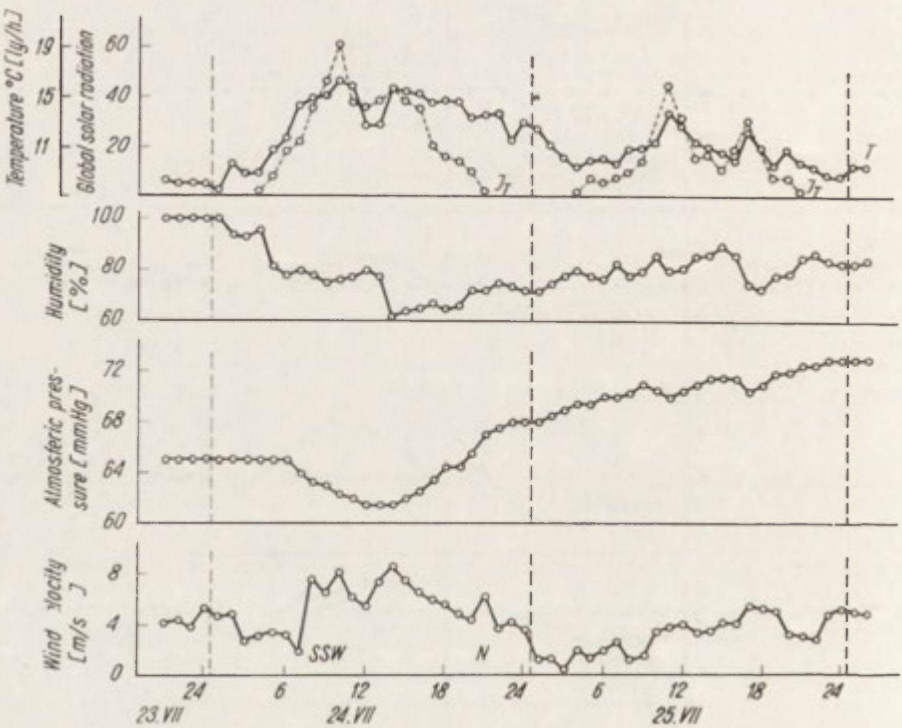

Fig. 15. Daily variation in the global solar radiation (I_T), air temperature (T), relative humidity (f), atmospheric pressure (P) and wind velocity (v) on 24th and 25th July 1968 in the meteorological station on the forefield of Skeidararjokull

its maximum on 17th August between 11 and 12 a.m. (mean hourly velocity was equal to 16.6 m/s), in gusts the wind velocity reached much higher values. On 16th and 17th August an increase in the mean daily temperature and the extreme temperature was clearly marked. The mean daily temperature on those days was ca. 8.5° and it may be regarded as a result of adiabatic processes taking place in the air violently descending in the direction of the seashore plain from the glacial area over 1500 m high. Relative humidity and vapour pressure reached the lowest daily values on those days: relative humidity 62%, water-vapour pressure 6.6 and 6.9 mb, the deficit of water vapour pressure however, had the highest values 4.0 and 4.2 mb (Table 1).

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Other weather effects were also typical of the fohn. On 16th August in the upper zone of the Skeidararjokull there appeared *Altocumulus lenticularis* clouds. It is highly probable that on the windward side of the Vatnajokull, that is N, precipitation occurred at that time. Next day the sky was clear and the wind velocity at its highest. On the 18th August advectional weather established itself with general overcast sky and precipitation characteristic of cyclone fronts.

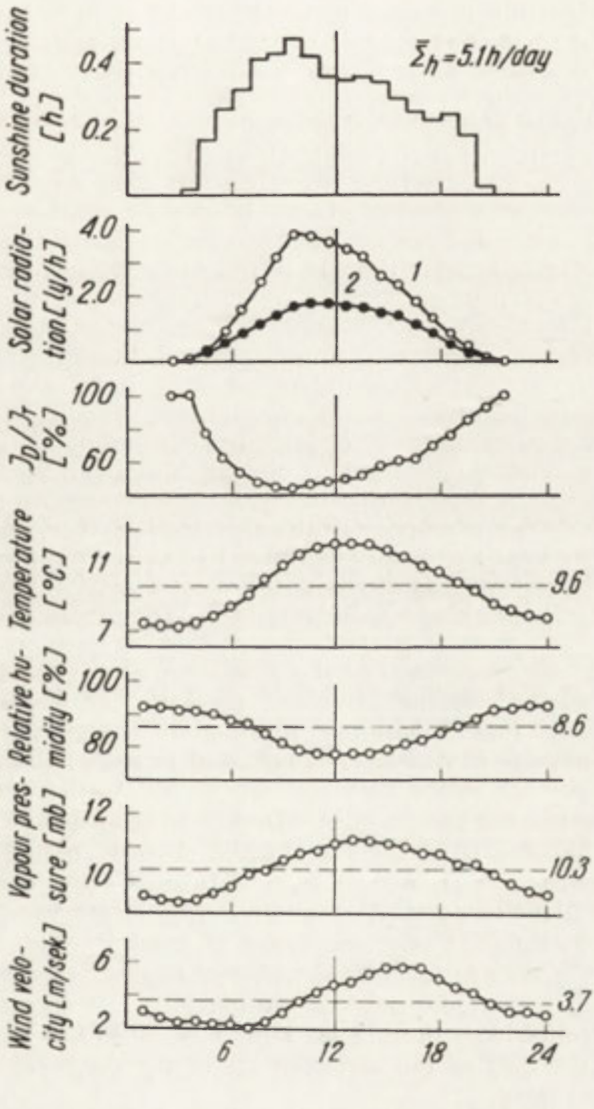

Fig. 16. Mean daily variations in meteorological elements for the whole period from 21th June to 20th August 1968 in the meteorological station on the forefield of Skeidararjokull:

1 — global (I_T), 2 — diffuse (I_D), $\Sigma I_T = 367$ ly/day, $\Sigma I_D = 192$ ly/day

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(3) THE QUESTION OF LOCAL CIRCULATION

In the sunny weather in the forefield of Skeidarárjokull the system of air exchange was set up under the influence of local factors, namely, the glacier, the ocean and the stretch of outwash plains and moraines about 23 km broad, lying in between. The glacier constitutes an environment that is "cool" day and night, and as a result high-pressure is formed over it. So air currents flow down over its immediate forefield in the form of a glacier breeze, blowing day and night but at a varying extent over the forefield.

During the night the moraines in the forefield of the glacier and the outwash plains lying further away cool under radiation, which makes the higher pressure over the glacier extend over the further parts of the forefield, and air currents increase their range. The sea region at night is warmer than the neighbouring outwash plains. Thus, lower pressure forms over the sea as compared with the pressure over the land. Air currents move from the land out to sea as a land breeze. The air flow seawards extends over a broader and broader belt of the shore. Since the glacier breeze gradually extends its range, there is a moment when it joins the land breeze, and since then over the whole area between the glacier and the sea the N wind blows, or one which a N component. The wind is characterized by moderate and regular velocities in the range of 3–4 m/s.

In the day, even in moderately sunny weather, the air circulation becomes more complicated. After sunrise the dark outwash plains and moraines gradually warm up, which results in lower atmospheric pressure (Fig. 15), development of convective currents together with convective clouds of initially slight vertical structure. The clouds used to appear first over the shore. It should be supposed that this is mainly due to the wetter ground on the shore, which makes the amount of water vapour in the air over it greater. As soon as the atmospheric pressure over the sandrs becomes less than over the sea, in the zone over the shore the pressure gradient changes its direction for the opposite to that it had during the night, and currents of air flow from the sea to the land as a sea breeze.

In accord with the extension and the deepening of the local depression above the outwash plains and the moraines, the sea breeze gradually increases its range towards the inland, whereas the range of the glacier wind undergoes a systematic reduction. A visible symptom and proof of the widening of the range of the sea breeze is the displacement of the wall of convective clouds from the sea towards the glacier (Fig. 17). At the peak phase of this situation the air currents blowing from the sea and the outwash plains as well as from the moraines towards the glacier are reaching directly the head of the glacier itself. This takes place at noon. At that time the glacier wind has its shortest range. Evidence of thus far reaching range of currents having S components blowing from the moraines towards the glacier is the immense inconstancy of the wind direction just at the head of the glacier. This shows the convergency of the air currents from the glacier with those blowing from the sea. Further evidence is the halt of the arrangement of the convective clouds exactly before the glacier's head.

At the meteorological station the change from the northern to the southern circulation is observed before noon. For instance, on 24th July (Fig. 15) it occurred suddenly between 7 and 8 a.m., together with a remarkable heightening of velocity. The evening change of direction to the opposite direction, i.e.,

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the return to the night system was recorded as late as 10 p.m., it being likely to have occurred a little earlier.

The daily wind is characterized by much greater velocities than the night wind. Undoubtedly, it is connected with the development of large scale turbulence during a fine day, which heightens the wind velocity.

Fig. 17. Wall of the convective clouds moving under the seebreeze from the shore towards the front of Skeidarárjökull. Its passage over the meteorological station in the about-noon hours was the cause of a break-down of the diurnal variation in all meteorological elements. Photo by G. Wójcik

The behaviour of all meteorological elements towards noon is very interesting at the station as seen in Fig. 15. Thus, between 11 a.m. and 1 p.m. there occurs a reduction of solar radiation, temperature and wind velocity. This is a result of the passing over the station of the convergence zone with the wall of convective clouds due to the extending range of on-shore winds. The clouds reduce radiation, on the other hand, the temperatures of the ground and the air become lower, turbulence is reduced and all this results in reduced wind velocity. The relative humidity also reacted regularly, that is, by a certain increase in values.

(4) CLIMATIC CONDITIONS AND THE RECEDING OF THE FRONTAL ZONE OF THE SKEIDARÁRJÖKULL

During advectional weather the climatic conditions along the whole front of the Skeiderárjökull are almost identical, and they determine a more or less identical rate of the ablation processes. During clear, sunny weather, however,

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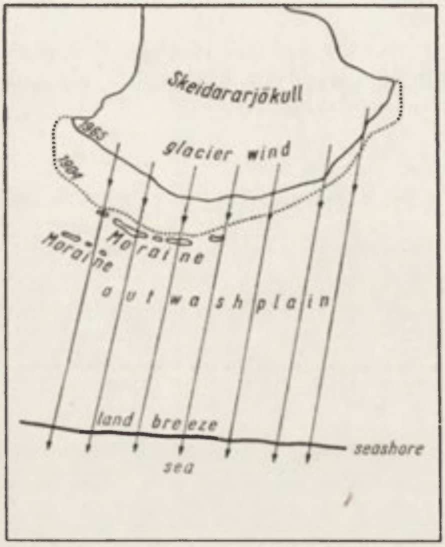

Fig. 18. The scheme of the development of the local air circulation on the forefield of Skeidararjokull at night

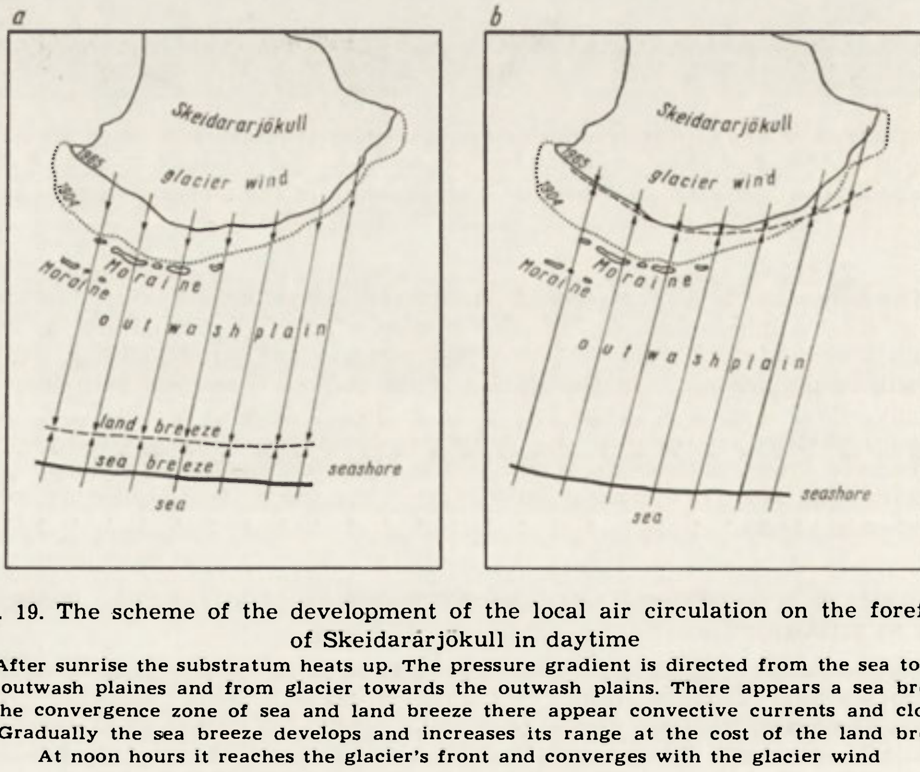

Fig. 19. The scheme of the development of the local air circulation on the forefield of Skeidarárjökull in daytime

(a) After sunrise the substratum heats up. The pressure gradient is directed from the sea towars the outwash plaines and from glacier towards the outwash plains. There appears a sea breeze. In the convergence zone of sea and land breeze there appear convective currents and clouds. (b) Gradually the sea breeze develops and increases its range at the cost of the land breeze. At noon hours it reaches the glacier's front and converges with the glacier wind

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the climatic conditions in the eastern and the western part of the frontal zone of the glacier are largely differential. The phenomenon is connected with a slight differentiation of the local circulation mentioned above in the western and the eastern parts of the glacier.

Observations of the position of the wall of convective clouds (Fig. 17) surrounding the front prove that in the western part it lies almost exactly over the front of the glacier, but the system gradually increases its horizontal distance from the front, the more eastwards it is. This is a significant fact that can be explained as follows: the Skeidarárjökull is bounded on the east by a fairly high mountain range of the Skaftafells (height range 1000 m), which further eastward passes into the highest orographic part of the Oraefa. Along the axes of the numerous valleys and glaciers to be found there, a much stronger air flow takes place than along the axis of the Skeidarárjökull. These neighbouring mountain and glacier winds of wide range join and strengthen the glacier breeze in the eastern part of the Skeidarárjökull and constitute a bar-

Fig. 20. The early stage of the advective fog in the middle part of the ablation zone of Skeidarárjökull. Photo by G. Wójcik

rier in its farther forefield against the warm southern currents blowing from the sea, the moraines and the outwash plains (Fig. 19). Therefore, the system of convective clouds is being stopped over the eastern part at a larger distance from the front than over the western part. It follows from the above remarks that the western part, as well as the middle part of the frontal zone have (due to warm currents of air heated over the moraines and outwash plains) a much warmer climate than the eastern part where it is formed under the more intensive influence of cool currents of the glacier and mountain breezes.

In Fig. 18 and 19 the positions of the glacier's front in the years 1904 and 1965 are presented; it can be seen that the glacier was not receding in the same

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way: during almost exactly 60 years it receded about 3 km in the western part, whereas in the eastern part by only about 500 m, and in parts even less. In my opinion, this unequal rate of recession of the glacier's front is mainly caused and determined by the differential climatic conditions in the zone at the glacier, which have been discussed above. But one cannot categorically exclude the action of still another agent, namely, a differentiation in the movement of the ice masses towards the front of the glacier. If the glacier in the eastern part were more dynamic as compared with the western part, the faster flow of ice masses would constitute an agent balancing ablation and so would be another cause of the slower recession of the front. In order to solve this problem, it would be necessary to carry out simultaneous measurements of the movement of the glacier in both the parts in question.

Indirect evidence proving differential climatic conditions along the front of the glacier is provided by observations of the flora of the zone in the glacier's vicinity. So, in the western part of the glacier the flora appears just after the receding glacier, this being true not only of less demanding plants such as moss, but also of flowering plants, which are more demanding. In this part of the glacier algae and dense patches of moss were observed at a distance not exceeding 50 to 100 m from the front of the glacier. At the same distance in places sheltered from the glacier breeze and exposed to the south saw tufts of flowering plants were seen. In the eastern part of the immediate forefield of the glacier, however, plants were hardly to be found, only at a distance in the range of 1 to 2 km there was a cover of moss and lungwort, that is harder plants. The discribed succession of plants to the places abandoned by the glacier under the same edaphic conditions (the same type of ground) may only be explained by differential climatic conditions.

In the future, it would be worth while carring out additional studies for the purpose of more exact investigation of the system of local air circulation presented in this paper and of the thesis posed here concerning highly differential climatic conditions along the front of the Skeidarárjökull. In connection with this it would be necessary to carry out meteorological field investigations along two profiles, on the west and east side, spreading from the front of the glacier towards its further forefield. At several points of each of the profiles, recordings of air temperature, wind velocity and wind direction should be made. For the purpose of an exact definition of the thickness of the air region involved in the local air circulation, however, it would be advisable and desirable to make at least a dozen pilot-balloon observations.

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REFERENCES

Averkiyev, M. S., 1956, Vozmozhnye mesiachnye i godovye sumy pryamoy sol'nechnoy radiatsi na gorizontalnuyu poverkhnost pri razlichnoy prozrachnosti atmosfery dla shirot 40–70° (The possible monthly and yearly sums of direct solar radiation on the horizontal surface under different transparence of the atmosphere for the latitudes 40–70°), *Vestn. Moskovsk. Univer.*, 2.

Averkiyev, M. S., Ryazanova (Beryukova), L. A., 1963, Sol'nechnaya radiatsya v idealnoy atmosfere i mutnost atmosfery realnoy (Sum.: Solar radiation in the ideal atmosphere and the turbidity of the atmosphere of reality), *Vestn. Moskovsk. Univer., Ser. V, Geografia*, 5.

http://rcin.org.pl

Barashkova, E. P., 1959, Mutnost atmosfery v Karadage (Turbidity of the atmosphere in Karadaga), *Trudy Glav. Geof. Observat. im. A. J. Voyeykova*, 80.

Dziewulska-Łosiowa, A., 1962, Próba oceny zakłócenia przezroczystości atmosfery w Warszawie (Sum.: Tentative estimate of disturbances in transparency of the atmosphere in Warsaw), *Przegl. Geof.*, 2.

Einarsson, M., 1966, Global radiation in Reykjavik and its relation to some meteorological elements, *Met. Ann., 4,20.*

Eythorsson, J., 1949, Temperature variation in Iceland. Glaciers and climate, *Geogr. Ann.*, 31.

Gjessing, Y. T., 1968, Local climates and growth climates of the Sognefjord region, the radiation climates, *Met. Ann.*, 5, 10.

Kluge, M., 1969, Osłabienie bezpośredniego promieniowania słonecznego w Górnośląskim Okręgu Przemysłowym (Sum.: Attenuation of direct solar radiation in the Upper Silesian industrial region), *Biul. Zakł. Bad. Nauk. Górnośl. Okr. Przem.*, 11.

Kosiba, A., 1958, Badania glacjologiczne na Spitsbergenie w lecie 1957 r. (Sum.: Glaciological investigations of the Polish IGY Spitsbergen-Expedition in 1957), *Przegl. Geof.*, 2.

Markin, V. A., 1962, Osnovnye cherty radiatsyonnovo rezhima lednikovykh pokrovov Zemli Frantsa Yosifa (Sum.: The main features of radiation regime of Franz Josef land glacier covers), *Met. Gidrol.*, 2.

O'Dell, A. G., *Kraje Skandynawskie* (The Scandinavian World), Warszawa 1961.

Paszyński, J., 1959, Wstępne wyniki badań przezroczystości atmosfery w Bydgoszczy (Preliminary results of the investigations of the atmospheric transparency in Bydgoszcz), *Przegl. Geof.*, 2.

Serebryannyy, L. R., 1969, *Islandia* (Iceland), Moskva.

Sivkov, S. J., 1968, *Metody rascheta kharakteristik sol'nechnoy radiatsii* (Methods of computation of the solar radiation), Leningrad.

Słomka, J., 1959, Ekstyncja atmosferyczna we Wrocławiu w latach 1950–1955 (Sum.: The atmospheric extinction in Wrocław in the years 1950–1955), *Acta Geoph. Pol.*, 7, and the manuscript of a Doctoral Thesis.

Słomka, J., 1966, Badania dopływu promieniowania słonecznego w zakresach widma ważnych dla procesów energetycznych i biologicznych (The investigations of solar radiation in the ranges of spectrum valid for the energetic and biologic processes), *Mater. Prace Zakł. Geof.*, 11.

Spinnagr, G., 1968, Global radiation and duration of sunshine in Northern Norway and Spitsbergen, *Met. Ann.*, 5, 3.

Voloshina, A. P., 1961, Aktinometricheskiye i obshchemeteorol'ogicheskiye nablyudeniya na sedl'ovine Elbrusa (5300 m) (Sum.: Actinometric and general meteorological observations on the Elbrus saddle — 5300 m a.s.l). *Vestn. Moskovsk. Univer., Ser. V, Geografia*, 1.

Wójcikowa, A., 1966, Niektóre aspekty stosunków solarnych Górnego Śląska na przykładzie Chorzowa (Some aspects of solar radiation conditions in the Upper Silesia using the example of Chorzów), *Wiad. Uzdrow.*, 1/2.

Wójcikowa, A., 1971, O wartościach bezpośredniego promieniowania słonecznego w warunkach atmosfery silnie zanieczyszczonej na przykładzie Górnośląskiego Okręgu Przemysłowego (Sum.: On the value of direct solar radiation under conditions of heavy air pollution with the Upper Silesian industrial district as an example), *Zesz. Nauk. UMK*, 26, *Geografia.*

Wójcik, G., 1972. The problem of asymmetrical recession of the Skeidararjokull's head (Iceland). Bull. Acad. Pol. Sci. Ser. sci. geol. geogr. 20, 2.

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GLACIOLOGICAL STUDIES ON THE SKEIDARÁRJÖKULL

Gabriel Wójcik

GENERAL FEATURES OF THE GLACIER

Skeidarárjökull is one of the numerous glacier lobes issuing from the great Vatnajökull ice cap; by these lobes the ice masses originating in the accumulation zone are being removed. The Skeidarárjökull lies in the southern part of Vatnajökull and is an example of a glacier combining features of a glacier lobe (an ice front with a snout widening in the lowland) and of a valley glacier (a markedly extended ice tongue).

The area from which Skeidarárjökull is fed constitutes a fragment of the Vatnajökull ice cap, and in its elementary outlines it has the shape of a triangle whose acute apex lies far to the north, near the Bardarbunga peak. Its western boundary runs along a summit line of cryptovolcanoes, volcanoes and nunataks. From south to north these peaks are: Geirvortur (1441 m), Thordarhyrna (1659 m), Haabunga (1700 m) and the Grimsvötn volcano; in the massif of the latter the loftiest peak is 1789 m high while the centre drops to 1300 m. Finally this chain of peaks ties in with the highest western part of the glacier called Kverkfallahryggur, some 1800 m high; as has already been pointed out by Kosiba (1938), is in its general position in conformity with the tectonic direction of these regions.

As seen in the map of 1904 the dimensions of the glacier are: length 72.5 km, of which 40.0 km is the length of the accumulation section situated above the 1100 line which, after Ahlmann (1937) is considered the firn contour-line; and 32.5 km is the length of the ablation section. The length of the arcuate snout margin is 27.5 km, the width in its narrowest part, in between the steep scarps of Sulutindar in the west and Fearines in the east, is. 8.6 km. After Thorarinsson (1939), the total glacier surface is 1722 km^2: 1211.5 km^2 the accumulation zone and 510.5 km^2 the ablation zone. Thus the ratio of the zone of ablation to that of accumulation is 1:2.4:

The long profile of the ablation zone shows certain local differences in slope. The slope of the glacier surface in the eastern part, below the southern flank of the Fearinestindar rock spur is the greatest. This part of the glacier shows the greatest intensity of crevassing; here all the cracks are very wide and deep (Fig. 1).

The glacier surface is irregular as the result of altitude differences caused by ablation (ice-moraine pyramids and medial moraine ridges) and by stresses (crevasses).

The whole glacier surface is dissected by a system of transverse crevasses of different length and width (Figs. 1 and 2) which locally may cross each other. On the average the length of the transverse crevasses reaches some tens

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Fig. 1. Crevasses with rims rounded by ablation in strongly sloping part of glacier. Eastern part of Skeidararjokull, below Faerneste rock spur. Photo by G. Wójcik

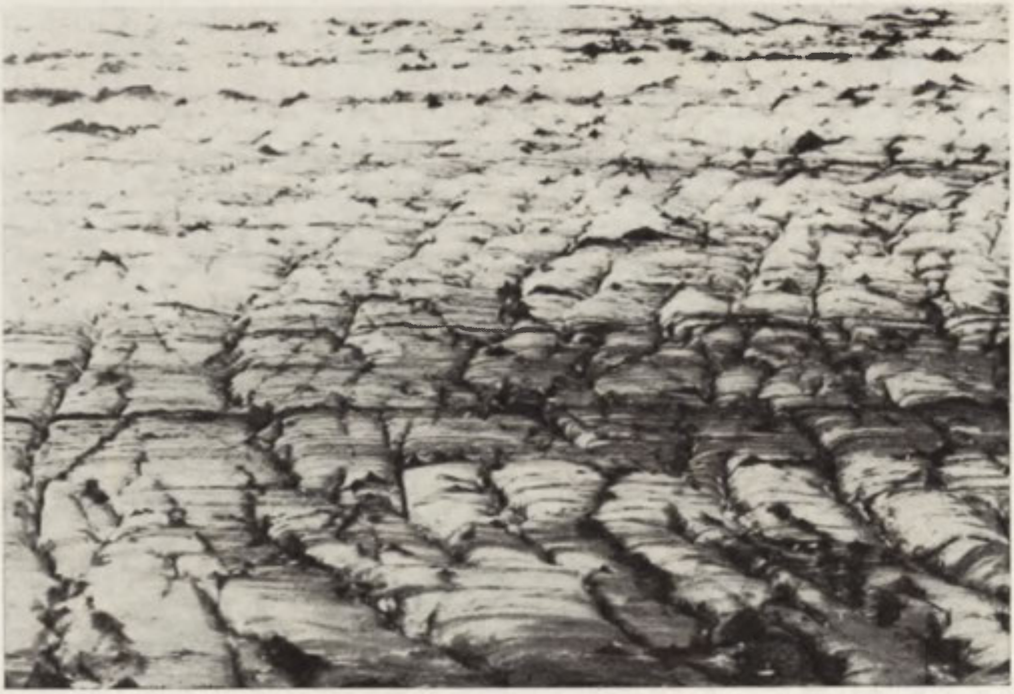

Fig. 2. System of transverse crevasses seen from Sulutinder scarp. Ablation zone of Skeidararjokull. Photo by G. Wojcik

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of meters while for the most part their width is about 1 m. The rims of these crevasses are rounded by ablation and, due to this, they appear much wider at the top. The glacier surface in its middle part is most intensively and in a particular way diversified in height. This surface shows the greatest amount of crevasses and the highest rate of movement. Here ablation of the walls of the widely gaping crevasses reaches far down, forming ravines which often are several meters deep and end at the bottom in fissures, sometimes bridged over, sometimes open. There also occur features shaped like hollows containing temporary glacial lakes. Longitudinal cracks occur almost exclusively in the snout part of the glacier. They originate from stresses produced by the spreading of the ice masses after they passed the mountain spurs which had constricted them. These cracks are also subject to transformation by processes of ablation (Fig. 3).

Where the ice-moraine layer is exposed in the marginal zones and the glacier front, the rock material held in this layer melts out. This released material is carried off by meltwater and by rainwater. The material is accumulated in

Fig. 3. Longitudinal cracks in snout part of Skeidarárjökull. Photo by G. Wójcik

gullies, in cracks and crevasses, in supraglacial and englacial tunnels and round the margin of the glacier snout. In this way the covers on the top of ablation-accumulation moraines originated, sometimes of the remarkable thicknesses of a dozen, or in exceptional cases, of some 50 or more cm. It is here that ablation creates convex forms on the glacier surface: ice-moraine cones (all over the surface) and bands of medial moraines. These convex forms are the result of locally diversified ablation due to a differentiation of the thicknesses of moraine cover and the albedo.

Ice-moraine cones occur: (a) in rows following the structure of the glacier, (b) in rows conforming to the slope of the glacier surface, and (c) in rows con-

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sistent with the direction of cracks. The structural rows commonly occur in the zone of the glacier front (Fig. 4) which is about 2 km wide, as seen in aerial photos taken in 1960 and 1969 and as observed in 1968 by the author. This zone, steeply sheared off by ablation, shows outcrops of ice-moraine layers accumulated in successive seasons, appearing on the surface as dark and light-coloured bands. These are what are known as the "dirt bands" distinctly visible on aerial photographs, and along these bands the pyramid rows appear.

Cone rows conforming to the slope of the glacier surface are seen on the glacier snout and on its marginal zones, especially where the glacier margin is inclined abutting upon the mountain scarps. They usually develop along

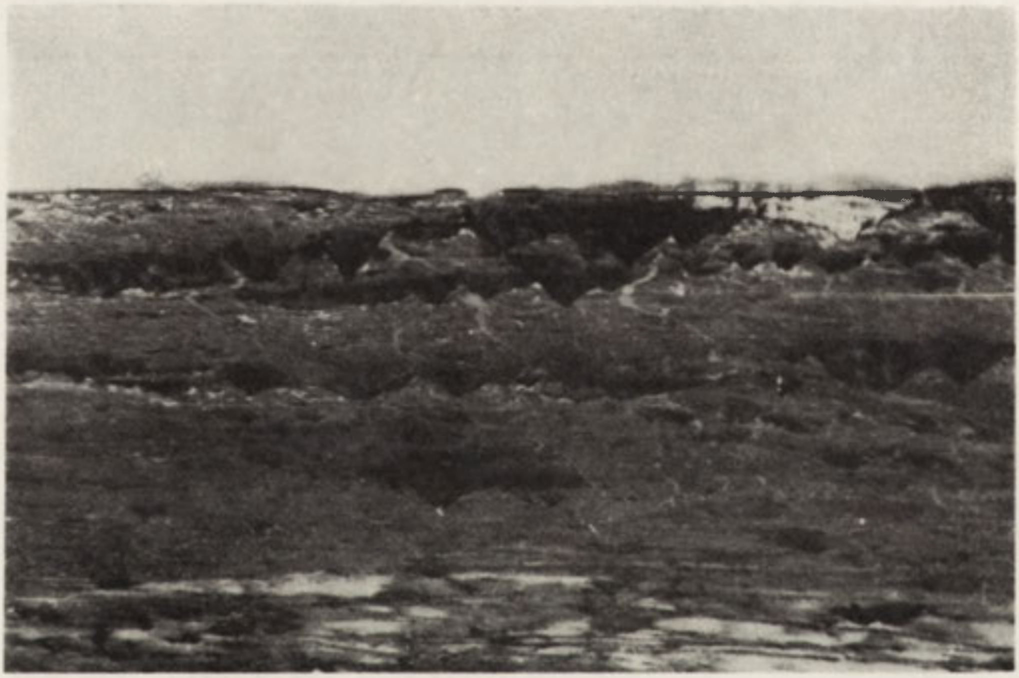

Fig. 4. Rows of ice-moraine cone running parallel to snout margin of Skeidararjokull. Photo by G. Wojcik

abandoned flow channels in which ablation water has accumulated moraine material. Pyramid rows consistent with longitudinal cracks are encountered on the glacier snout, and rows running parallel to transverse crevasses on the marginal zones of the glacier.

Medial moraine bands extend in the western part of the glacier, in line with the Grimsvotn volcano; this peak is an abundant source of moraine material in this part of the glacier. Numerous bands originate here built of fine volcanic sand; the largest is the band of a medial moraine which consists of basalt debris (Figs. 5 and 6). The eastern part shows a medial moraine of large dimensions, probably supplied with material derived from the Faerneste group. Only in some sections and, predominantly during the initial period of ablation, do medial moraines show an unbroken course and a fairly smooth surface; later on, when they grow higher, the moraine cover assumes different

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Fig. 5. Bands of medial moraines built of volcanic sand, in western part of Skeidararjokull near Profile No. 2. Photo by G. Wojcik

Fig. 6. Medial moraine built of basalt rocks. Western part of Skeidararjokull near Profile No. 2. Photo by G. Wójcik

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thicknesses and is gradually broken up into disrupted pieces. What are called "glacier tables" of various sizes also develop.

As far as the author was able to observe, as a rule all convex forms produced by ablation develop and grow in size within one season of ablation. Their height grows as long as the rock material covers and protects them from intensive melting (Fig. 7). On Skeidarárjökull this type of form grows to be 3 to 4 m high. The protective cover deposit is demolished by a variety of agencies: by lively gravity downflow of the water-soaked material — especially on days with strong solar radiation — by slopewash due to atmospheric precipitation, and by wind action which is enhanced on a large-scale when a few fair-weather days have quickened surface desiccation (Fig. 8). The intense destruction of such forms always takes place on slopes facing the sun, because the surface of these slopes, set almost perpendicular to solar radiation, receives the maximum heat per surface unit, so that slopes situated in this way melt rapidly, and it takes only about half the time of normal ablation to make them concave.

An illustration of the intensity of these processes is the observation of July 25, 1968. At about 11 a.m. on this day, the cloud cover above the higher part of the ablation zone abruptly vanished and a spell of intensive insolation set in. After just a few minutes all the cones started evaporating into the cool air

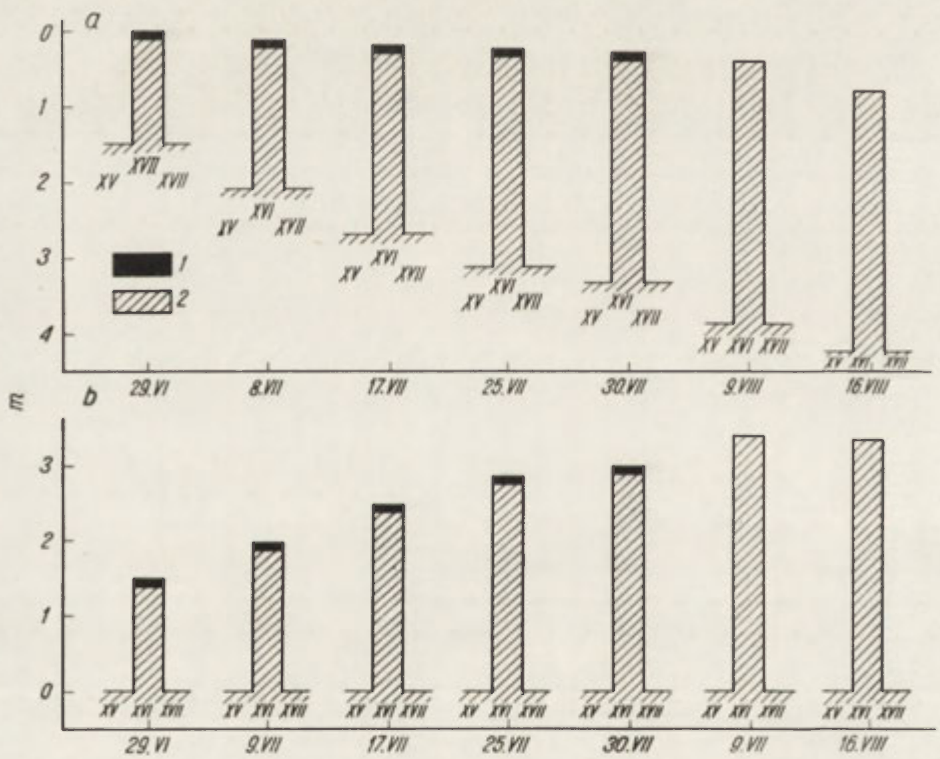

Fig. 7. Development in size of the ice-moraine cone in the period 29 June–16 August 1968, according to measurements of ablation in points T_{XV}, T_{XVI} and T_{XVII}

(a) Decrease of the top surface of the ice-moraine cone sheltered by a 10 cm layer of volcanic sand (point T_{XVI}) and decrease of the general surface of the glacier (points T_{XV} and T_{XVII}). The top surface decreases less than the general surface of the glacier. (b) Increase of the cone's size when it is protected by the moraine cover and decrease when protective cover is lost. 1 — sand cover, 2 — ice

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Fig. 8. Example of aeolian erosion on Skeidarárjokull. Photo by G. Wójcik

with such rapidity that in a short while a fairly dense fog developed. Having no thermometer on hand, the author was unable to measure the temperature of the evaporating surfaces. But on the very next occasion met in this same region, and under identical conditions of insolation but a somewhat stronger wind, he managed to make measurements. Thus, on August 9, at 2.55 p.m., the glacier cover, barely 2 cm thick, consisting of dark volcanic sands and resting on the slope of a cone exposed to the sun, had a temperature of 34°C. For all its slight thickness, the outer surface was drying and appeared distinctly affected by the destructive wind action.

From the moment the protective cover is lost, the feature passes into the stage of vertical degradation and suffers complete destruction during that same season of ablation (Fig. 9); or it is left in a mutilated shape until the next

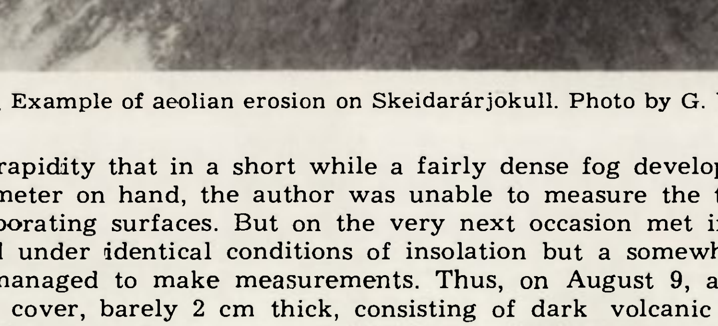

Fig. 9. Supposed development in time of the ice-moraine cone under the climatic conditions of the ablation zone of Skeidarárjokull, according to measurements of the ablation in the points T_{XV}, T_{XVI} and T_{XVII} in the period 29 June–16 August 1968. Extrapolation is marked by broken line

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season, but without any chance of regeneration. Only a certain number of ablation features on which a moraine cover thick enough to last until the successive season of ablation has been preserved, maintain the possibility of further development, and it is only these which finally attain the highest relative heights. This, however, refers only to isolated specimens, because autumn and winter rains and all-year foul weather continue the process of removing all remnants of moraine covers surviving on isolated protruding glacier forms.

The following data are evidence of the protective capacity of a moraine cover. They concern the rate of ablation, observed at sheltered (T_{XVI}) and an unsheltered points (T_{XV} and T_{XVII}) of the moraine cover (Fig. 7 and 16). In the time from June 29 to August 9, 1968 the author observed that at Point T_{XVI}, situated on the surface of a medial moraine which was covered by a 10 cm layer of volcanic sand, the ice melted to a thickness of barely 49 cm, while

Fig. 10. Pure firn ice. Middle part of ablation zone of Skeidararjokull. Photo by G. Wójcik

nearby, on the surface of "dirty" ice, at Point T_{XVII} the thickness of the ice melted away during the same period of time was 238 cm. These figures show, that in the profile illustrated in Fig. 14 it took 41 days for the medial moraine to increase 189 cm in height. It should be added, that in the time from August 9 to 16 the moraine cover at Point T_{XVI} disappeared completely — a process which unfortunately was not observed in detail — and that here an ice layer 38 cm thick melted, whereas at Point T_{XVII} the ice cover was reduced by only 30 cm. All this shows that a definite process of vertical degradation had set in.

With protective covers of adequate thickness the height of glacier forms depends on the rate at which ablation takes place; this rate is highest in the zone of the glacier snout and decreases the farther up the glacier one moves.

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Observations revealed that the same rule applies to all convex forms seen on a glacier; they are highest in the snout part and become lower with increasing distance from the snout. This vertical degradation is most conspicuously visible at the medial moraines whose relative height increases towards the glacier snout — a feature observed not only on Icelandic glaciers but also elsewhere, for example on Spitsbergen.

Apart from convex forms of ablation, the Skeidarárjökull surface also shows concave forms in the shape of depressions such as cryoconitic hollows. This type of form commonly occurs in the zone covered by pure firn ice (Fig. 10), overlying the ice-moraine layer which appears on the surface in the snout face and in the marginal zones. These hollows develop at places where dark rock material dispersed over the glacier surface by winds and ablation waters has melted into the ice; on Skeidarárjökull these hollows have attained remarkable dimensions and are up to some 50 or more cm.

THE SKEIDARÁRJÖKULL DRAINAGE SYSTEM

On the Skeidarárjökull the waters derived from ablation and rainfall rapidly penetrate the glacier ice; for the most part they are carried off by a sub--surface drainage system so that the glacier surface appears to be "dry". In the ablation zone a well developed surface runoff was visible only on the snout surface and on some parts of the glacier where the surface gradient was gentle and smooth, for instance on the western half of the central part of the ablation zone. On July 17, 1968, over a distance of barely 2 km between Points T_{XII} and T_{XVII} (Fig. 14, Profile No. 2), no less than seven larger meltwater streams — apart from some minor ones — run down the glacier surface; their average width was 1 m, a few were even wider than 1.5 m (Fig. 11). This well developed drainage system completely disappeared below the ice surface at some lower part of the glacier where the surface ran almost horizontally. From the snout of the glacier the ablation waters escape mostly on the ice surface by ablation gullies or in open cracks (Fig. 4).

In part of the marginal zones the glacier surface is sloping at times rather steeply, toward the confining mountain scarps. The groove-like depressions thus formed at the contact point between glacier margin and rock face contain ice-dammed lakes of perennial or temporary nature, varying in size. Graenalon, the largest of these lakes has a size of 18 km^2 (Thorarinsson 1953, p. 272). Into this lake a cliff-like fragment of Skeidarárjökull descends, and icebergs originate here in large numbers (Fig. 12).

From the major part of the glacier, that is, from the zone of accumulation as well as from the upper and middle part of the zone of ablation, the ablation waters escape into two peripheral subglacial flow arteries which near the margin of the glacier snout issue under hydrostatic pressure and in this way originate two powerful streams, the Sulukvisl in the west and the Skeidará in the east. In contrast, the waters from the lower part of the ablation zone pour into the forefield in a wide sheet all over the ice surface, partly also by numerous small subglacial channels; they all accumulate their flow in several proglacial lakes which are drained by the River Sandgigjukvisl which in its run dissects the end moraine zone by a splendid gap (Fig. 13).

The rivers forming what was called "peripheral arteries" carry at certain intervals abrupt floods of water which intermittently had been stored in the depression of the Grimsvötn volcano and in the large Lake Graenalon. The

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Fig. 11. Glacier stream in western part of Skeidararjokull, near Profile No. 2. Photo by G. Wojcik

Fig. 12. Lake Graenalon, into which Skeidararjokull fragment descends. Icebergs are produced here. Photo by G. Wojcik

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Fig. 13. The frontal part of Skeidararjökull and its forefield with the patterns of rivers draining off meltwaters. In foreground fragment of englacial tunnel exposed by ablation. Photo by G. Wójcik

mechanics of this most characteristic event, known by its distinctive name *jökullhlaup,* has been convincingly explained and described by Thorarinsson (1953, 1953a). Thus, a *jökullhlaup* which last ten days, on average starts with the bursting of an ice barrier; this takes place after the water basins mentioned above are full to overflowing. Up to the 1940s this event took place in surprisingly equal intervals: the Lake Graenalon floods took place every four years, for example in 1935 and 1939, while for the Grimsvötn depression they occurred approximately every ten years, as shown by the known series: 1903, 1913, 1922, 1934. During these tremendous floods one and a half km^3 of water escape from Lake Graenalon, while from Grimsvötn the huge mass of seven km^3 is released; this huge inundation submerges enormous areas of the glacier forefield. To give an example: during the 1934 *jökullhlaup* from Grimsvötn, the flood covered an area of some 1000 km^2 of the outwash plain, and the torrents rushing seaward carried icebergs of the size of three-storey buildings (Thorarinsson 1953, p. 268).

During a certain period of a Grimsvötn *jökullhlaup,* a volcanic eruption usually occurs; after an explanation given by Thorarinsson (1953), these eruptions occur as a result of pressure reduction which occurs in the Grimsvötn depression due to the escape of the huge mass of water.

ABLATION

In his glaciological investigations the author made large-scale measurements of ablation, using wooden stakes 2.5 m long inserted into holes drilled

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in the ice; with ablation these holes were successively deepened. The wooden stakes were arranged in two profiles (Fig. 14): one longitudinal (No. 1) the other transverse (No. 2). The longitudinal profile which in an upward direction started at 50 m from the rim of the snout, was 12 km long, and in it nine control points were provided. Points T_{I} to T_{VI} were set up on June 21, 1968, the remaining three some time later, on July 17. With rare exceptions, control readings were made every day, sometimes even twice a day, in the morning and evening in order to determine variations in ablation between daytime and night. The results of these control measurements are recorded in Tables 1 and

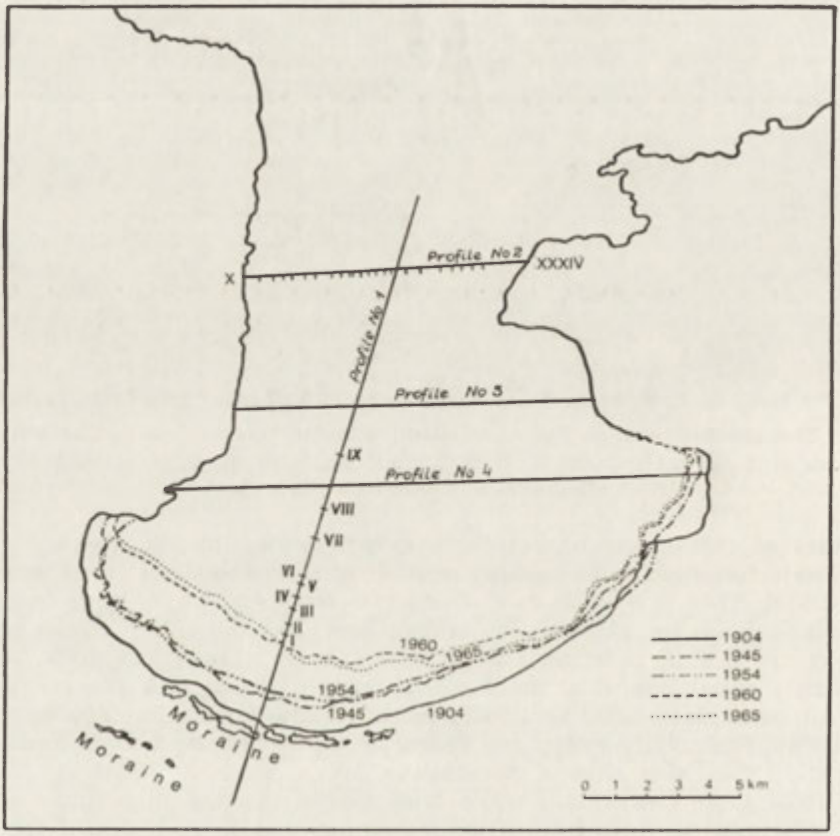

Fig. 14. Area of glaciological investigations in summer 1968. Arrangement of profiles (Nos. 1–4), stakes (T_{I}–T_{XXXIV}) and indication of changes in range of Skeidarárjökull snout

2 and in Fig. 15. The transverse profile (No. 2) was set up at a distance of some 12 km from the snout rim, following the width of the glacier where it is narrowest between the Sulutindar and Faerneste mountain spurs. This profile — in which measurements of ice movement were also made — had a length of 8.6 km and here 23 control stakes on the ice were placed, from T_{XI} to T_{XXXIII}; this profile was set up on June 29, 1968. For this profile control readings were made less often, because of the considerable distance from the camp (on the average once every 7 or 8 days). The way ablation proceeded

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TABLE 1. Ablation on the Skeidarárjökull along longitudinal profile in the water equivalent, Summer 1968

Number of measurement	Period	Ablation, cm — Number of point I	II	III	IV	V	VI	VII	VIII	IX	Intensity of ablation, mm/h — Number of point I	II	III	IV	V	VI	VII	VIII	IX
	VI																		
1	21/22	3.32	3.32	5.81	3.32	2.92	3.32				1.66	1.66	1.93	1.44	1.21	1.33			
2	22/24	10.80	12.04	8.30	9.96	8.30	8.30				2.15	2.41	1.97	1.95	1.66	1.62			
3	24/25	2.49	2.95	2.49	1.66	0.83	1.66				1.24	1.45	1.31	0.87	0.42	0.92			
4	25/28	13.27	11.20	9.96	9.96	8.72	8.30				1.86	1.57	1.39	1.40	1.23	1.17			
5	28/29	3.32	3.32	4.15	4.98	3.32	3.32				1.38	1.38	1.13	2.07	1.38	1.37			
6	29/1	12.44	14.10	10.80	10.80	9.96	11.62				2.48	2.82	2.12	2.16	1.99	2.32			
	VII																		
7	1/5	32.40	32.37	29.88	30.70	27.40	28.22				3.20	3.20	3.02	3.61	3.04	3.10			
8	5/8	16.60	14.11	13.27	14.10	12.44	12.44				2.57	2.17	2.04	2.20	1.86	1.86			
9	8/17	49.80	53.90	56.40	48.20	45.65	46.50				2.86	3.10	3.28	2.85	2.70	2.75			
10	15/17	10.80	10.79	10.80	10.80	10.80	10.80				2.56	2.57	2.49	2.39	2.35	2.35			
11	17/19	15.77	16.60	18.26	18.26	14.94	14.10	10.80	12.45	11.63	3.09	3.25	4.05	3.59	2.83	2.76	2.12	2.49	2.27
12	19/22	19.72	17.43	17.43	22.41	14.94	14.94	14.94	14.10		2.85	2.49	2.49	3.16	2.13	2.13	2.10	1.96	
13	22/23	8.30	6.64	8.30	7.47	7.47	7.47	5.82	5.81		3.45	2.76	3.45	3.11	3.11	3.11	2.42	2.54	
14	23/23	5.82		4.98	2.49	2.49	3.32	3.32			5.28		4.49	2.76	2.76	3.68	4.15		
15	23/24	2.49	6.64	3.32	2.49	4.15	5.81	8.30	11.62		1.91	2.66	2.37	1.66	2.59	3.62	6.82	5.53	
16	24/25	4.98	4.98	4.98	5.81	2.49	3.32			27.40	2.37	2.49	2.49	2.76	1.24	1.66			1.97
17	25/25	6.64		3.32	3.32	3.32	3.33				6.64		3.32	3.68	3.68	3.68			
18	25/29	20.75	23.25	24.90	21.60	18.25	23.25	23.25	19.91		2.26	2.28	2.74	2.34	1.98	2.50	1.84	1.59	
19	29/30	3.32	4.98	4.98	3.32	2.49	1.66	2.49	3.32	17.49	1.84	2.62	2.71	1.75	1.24	0.87	1.31	1.84	1.43
20	30/30				4.15		5.81	4.98	4.15	1.66				3.19		4.84	4.52	3.78	1.66
21	30/31	7.47	8.30	9.14	4.98	10.80	8.30	5.81	4.15		2.30	2.68	2.95	2.76	3.27	4.15	2.90	2.08	
22	31/31	2.07	1.24	0.83	1.66						5.20	3.20	2.74	5.52					
23	31/5	33.20	28.21	34.96	34.87	29.04	31.55				2.75	2.33	2.86	2.86	2.32	2.54			
	VIII																		
24	5/6	4.15	2.49	1.66	2.49	2.49	0.83				3.18	1.92	1.28	1.77	1.91	0.66			
25	6/6	5.81	4.57	6.64	7.47	6.22	3.73				4.81	4.15	6.03	7.47	6.22	4.15			
26	6/7	2.49	2.07	3.32	2.49	2.49	3.73	38.20	39.85		1.91	1.38	2.22	1.66	1.66	2.07	2.32	2.41	
27	7/7	4.15	4.15	3.32	3.32	1.66					4.61	5.92	5.53	5.53	3.32				
28	7/9	6.64	5.82	7.47	8.30	8.30	14.20	9.13	13.27	18.20	1.73	1.49	1.87	2.07	2.02	3.14	2.02	2.89	2.09
29	9/9	6.81		6.64	7.47						4.47		5.10	6.23					
30	9/12	14.94	18.25	16.60	15.76	18.25	16.60				2.20	2.26	2.41	2.25	2.26	2.05	2.77		
31	12/13	0.83	0.83	0.83	0.83	1.66	0.83	24.08	21.60		0.58	0.58	0.58	0.58	1.10	0.58	2.17	2.25	
32	13/13	3.32	3.32	1.66	2.49	2.49	1.66	0.41			4.73	4.73	2.36	4.15	4.98	3.32	1.04		
33	13/22	39.85	36.51	37.33	38.20	29.90	28.22				1.89	1.73	1.78	1.80	1.40	1.33			

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TABLE 2. Ablation on the Skeidarárjokull along the longitudinal profile

Period	Abla-tion	Number of point I	II	III	IV	V	VI	VII	VIII	IX	XXVI
June 21–	(a)	450.5	426.0	449.0	441.0	384.0	396.0				
August 22,	(b)	373.96	354.38	372.73	366.13	315.01	327.13				
1968	(c)	2.51	2.37	2.51	2.47	2.12	2.22				
July 17–	(a)	185.5	165.5	190.0	189.0	158.5	175.0	153.0	155.0	128.00	100.0
August 9,	(b)	153.97	137.37	157.81	156.90	131.54	145.34	127.04	128.63	106.32	83.0
1968	(c)	2.79	2.49	2.86	2.85	2.38	2.63	2.30	2.33	1.92	1.5

(a) thickness of layer of melted ice, cm
(b) this same in water equivalent, cm
(c) average intensity of ablation in water equivalent, mm/h

in this profile is illustrated in Table 3 and Fig. 16. The longitudinal and transverse profiles intersected each other at Point T_{XXVI}.

In its upslope extent the longitudinal profile dissected first the steeper part of the glacier near its snout; this surface was covered by an ablation moraine of diversified, but moderate, thickness. This zone is some 2 km wide and in it were Points T_I to T_{IV}. Point T_V was set up on the boundary between

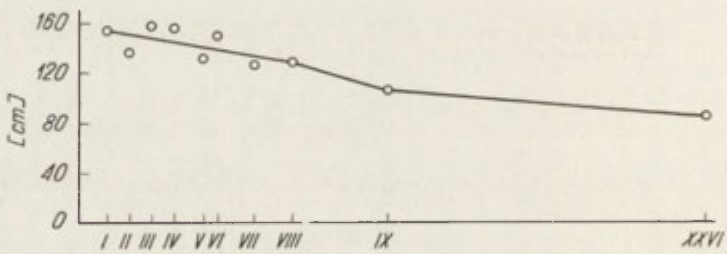

Fig. 15. Ablation on Skeidarárjokull along longitudinal profile (No. 1) in period from July 17 to August 9, 1968, expressed in cm of the water equivalent

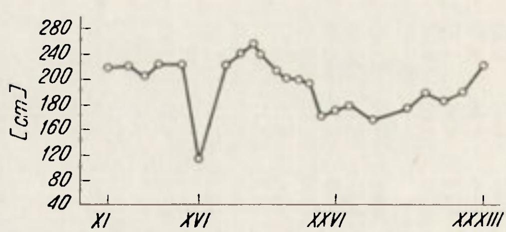

Fig. 16. Ablation on Skeidarárjokull along transverse profile (No. 2) in period from June 29 to August 16, 1968, expressed in cm of water equivalent

"dirty" and pure ice, while Points T_{VI} to T_{IX} were situated fully in the pure ice zone. In Table 1 the effect of ablation of the ice is given in cm of water equivalent, and the intensity of ablation in mm of water equivalent per hour, in values averaged from two successive readings; in an analogous way the data for profile No. 2 shown in Table 3 are computed. In Table 2 the author added ablation values in cm of the height of melted ice. In this table, the first part contains data for the period from June 21 to August 22 with regard to con-

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TABLE 3. Ablation on the Skeidararjokull along the transversal profile in the water equivalent, Summer 1968

Number of point	Ablation, cm							Intensity of ablation, mm/h						
	29.VI–8.VII	8–17. VII	17–25. VII	25–30. VII	30.VII. 9.VIII	9–16. VIII	Total 29.VI–16.VIII	29.VI–8.VII	8–17. VII	17–25. VII	25–30. VII	30.VII–9.VIII	9–16. VIII	Mean 29.VI–16.VIII
XI	45.65	45.65	37.35	16.60	44.00	27.40	216.65	2.15	2.12	1.96	1.37	1.83	1.58	1.87
XII	45.65	44.00	37.35	16.60	46.50	29.90	220.00	2.12	2.05	1.91	1.37	1.93	1.72	1.91
XIII	45.65	34.88	34.05	15.77	46.50	27.40	204.25	2.12	1.68	1.79	1.30	1.92	1.60	1.77
XIV	44.80	48.20	36.52	15.77	44.00	33.21	222.50	2.10	2.24	1.91	1.31	1.82	1.96	1.93
XV	49.00	49.00	35.70	15.77	45.65	27.40	222.52	2.30	2.27	1.88	1.31	1.89	1.62	1.94
XVI	9.96	9.14	4.15	4.15	13.28	31.55	72.23	0.46	0.42	0.21	0.35	0.55	1.88	0.63
XVII	49.75	49.75	35.70	17.43	44.80	24.90	222.33	2.34	2.32	1.87	1.45	1.86	1.49	1.94
XVIII	51.50	51.50	39.86	18.27	51.50	27.40	240.00	2.42	2.40	2.08	1.52	2.13	1.64	2.09
XIX	48.20	52.30	47.30	21.60	55.65	29.08	254.13	2.27	2.43	2.47	1.80	2.30	1.76	2.22
XX	44.00	49.00	39.86	22.42	53.20	29.08	237.56	2.09	2.27	2.10	1.87	2.20	1.76	2.07
XXI	37.35	47.30	39.86	19.93	47.30	19.10	210.84	1.76	2.20	2.07	1.66	1.96	1.16	1.84
XXII	34.90	38.20	41.60	24.90	26.58	34.05	200.23	1.64	1.78	2.17	2.09	1.10	2.07	1.75
XXIII	43.20	46.50		57.26	51.50	15.77	198.46	2.03	2.16		1.85	2.12	0.96	1.74
XXIV	33.22	35.70		62.25	46.50	13.28	190.95	1.57	1.66		2.00	1.92	0.82	1.67
XXV	25.75	29.70		34.05	35.70	14.94	140.14	1.21	1.39		1.10	1.48	0.92	1.24
XXVI	28.25	27.40		40.70	42.30	10.80	149.45	1.33	1.28		1.31	1.75	0.67	1.31
XXVII	21.60	25.75		50.65	42.35	16.60	156.95	1.02	1.20		1.64	1.75	1.03	1.38
XXIX	28.25	31.55		39.85	37.35	14.12	151.12	1.33	1.46		1.29	1.54	0.95	1.34
XXX	31.55	30.70		49.80	44.80	18.25	175.10	1.49	1.42		1.60	1.85	1.24	1.56
XXXI	25.75	29.06		57.30	38.02	10.80	161.11	1.22	1.35		1.85	1.57	0.75	1.44
XXXII	34.90	34.05		46.50	43.20	16.60	175.25	1.66	1.58		1.49	1.79	1.15	1.56
XXXIII	44.85	45.70		57.28	50.65	19.93	218.41	1.12	2.13		1.84	2.10	1.37	1.95

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trol points T_I to T_{VI}; the second part refers to a shorter period, from July 17 to August 9, but also considered here are further control points set up at a later date. In his computations the author calculated water equivalent values, assuming the specific gravity of ice to be 0.83 g/cm^3; he determined this figure by testing several samples collected from the top layer of the glacier ice. The altitudes of particular control points were determined by altimeter readings referred to a master point marked on a moraine near the glacier margin, for which point the altitude a.s.l. was established by the geodetic method.

As mentioned before, Point T_I was situated 50 m from the rim of the glacier snout, at 95 m a.s.l., while the distance from the rim of Point T_{VI} was 2030 m and its altitude 330 m a.s.l. In the time from June 21 to August 22, hence within 62 days, an ice layer 450.5 cm thick melted at Point T_I, corresponding to a water equivalent of 373.96 cm. At Point T_{VI}, placed on pure ice, an ice layer 396.0 cm thick melted, corresponding to 327.13 cm water. The rate of ablation is highest in the zone of dirty ice because of its high absorption of solar radiation, and amounts to 2.51 mm/h. The figures given in Table 2 indicate that ablation decreases with growing distance from the glacier snout. This feature must be ascribed to:

(1) changed climatic conditions, which grow more severe the farther one moves upward from the glacier snout and the more the altitude above sea level rises,

(2) changes in the albedo — a feature particularly in evidence at Point T_{VI}.

The reduced ablation observed at Points T_{II} and T_V is worthy of attention. These two points were set up at places where accidental accumulation of moraine material caused a decrease in ablation of the underlying ice. The ablation, occurring during a shorter period of time — from July 17 to August 9 — but observed at the higher control points added later, can be recognized in the data given in the second part of Table 2 and in Fig. 15. Here one notices a practically linear decrease of ablation with increasing distance from the glacier snout.

For purposes of comparison the author adds some figures illustrating the effect of ablation upon some neighbouring glaciers. He found records for some glaciers descending from the Oraefa massif, called Morsarjökull, Skaftafellsjökull and Svinafellsjökull showing that in the summer of 1954 the mean value of ablation was respectively 6.5, 5.6 and 5.85 cm/day (Ives, King 1954; King, Ives 1955). In the Summer of 1966 this mean value measured on Breidamerkurjökull near the glacier snout was 8.0 cm/day (Price 1969). It is probable that in all these earlier reports the authors had in mind the thickness of melted ice, not its equivalent expressed in water. The present author found from observations at Points T_I to T_{IV} near the snout of Skeidarárjökull, that ablation removed an average of 7.1 cm ice, equivalent to 5.9 cm water. Ablation was greatest at Point T_I where the mean daily rate was 7.2 cm ice, or 6.0 cm water. According to tests previously cited, made by Ahlmann and Thorarinsson (1938) on Hoffellsjökull, during July 1936, 1937 and 1938 the mean ablation amounted to 212 cm water. The figures obtained by the author for July 1968 on Skeidarárjökull were: at Point T_I 249 cm ice = 206.7 cm water, and for the snout section (T_I to T_{IV}) the mean value was 247 cm ice = 205.0 cm water.

In higher parts of the glacier, along the transverse profile (Fig. 14, Profile No. 2), measurements of ablation were made from June 29 to August 16. The results obtained from successive measurements are given in Table 3, and Fig. 16 shows the ablation sums for particular points expressed in cm of water equivalent. Along this profile the rate of ablation is distintly diversified:

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in the marginal zones ablation processes proceed at a higher rate than in the middle zone. The reason seems to be twofold: in the first place, in the marginal zones the microclimate is warmer than in the middle zone due to the warming effect of the nearby mountain scarps. Further, in the marginal zones the ice contains a greater amount of darker rock material which increases the absorption of solar radiation and in this way hastens the ablation. The wooden stakes set in this part showed the highest figures of ice wastage. In the time from June 26 to August 16, the ablation at Point T_{XVIII} was 240.0 cm, at Point T_{XIX} 254.13 cm — equalling 2.09 and 2.22 mm water per hour respectively.

In the middle part of the profile, built of pure firn ice, where admittedly the surface layer was porous but at the same time brillantly white, the higher albedo value caused ablation to proceed at a less intensive rate. Hence during

Fig. 17. Fragment of former glacier surface exposed by ablation, showing traces of probable volcanic eruption. Middle part of ablation zone of Skeidarárjökull. Photo by G. Wójcik

the same period of time as mentioned before, ablation at Point T_{XXV} was 140.14 cm and at Point T_{XXVIII} 132.91 cm water; this gave values of 1.24 and 1.17 mm water per hour (Table 3). At some places on this part of the surface were exposed — or still existed-fragments of the ice surface plentifully strewn with ashes and volcanic sand (Fig. 17) — probably traces of former volcanic eruptions.

The way ablation was proceeded at Point T_{XVI} situated in the medial moraine, where the ice surface was sheltered by a 10 cm layer of volcanic sand is worthy of attention. In the time period mentioned before ablation here was barely 72.23 cm while at the nearby Points T_{XV} and T_{XVII} it was 222.53 and 222.33 cm water (Table 3, Fig. 7). After the protective moraine cover had slid off or had been washed away, the rate of ablation changed rapidly. This is in-

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dicated by the figures given in Table 3 for the time from August 9 to 16: at T_{XV} ablation rose in these few days to 27.40 cm, at Point T_{XVI} to 31.55 cm and at Point T_{XVII} to 24.90 cm water.

Plainly, the intensity of ablation depends on weather conditions. Among all meteorological elements, ablation goes most distinctly with solar radiation, especially with direct radiation. It appears that hourly values of ablation are highest for cloudless weather, less high when the sky is clouded, and least high when it rains. Thus during sunshine on August 6, causing a relative sunshine duration of 93%, measurements made at Points T_I to T_{VI} disclosed a mean ablation per hour on the glacier snout of 5.44 mm water. On August 13 clouds covered the sky, but most of the clouds were cirrus clouds and clouding was 80% and relative sunshine duration 32%; under these conditions the mean ablation was 4.04 mm water per hour. By constrast, in rainy weather on June 28 and the night following this day, the mean ablation was only 1.55 mm water per hour.

It is a fact, however, that indirectly rainfall affects ablation to a fairly marked degree. Rain water flushes away the rock material melted out from the ice, and in this way prevents the formation of the protective cover of an ablation moraine on the ice surface. After heavy rainfall, the ablation always takes place with increased intensity, especially on the steeper slope of the snout.

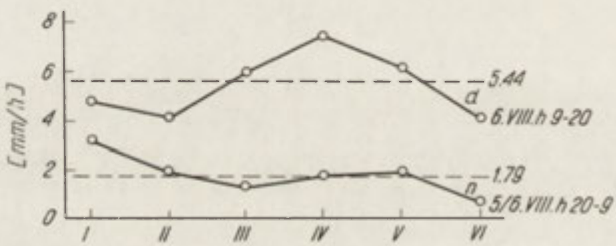

Fig. 18. Ablation on the snout of Skeidararjökull along longitudinal profile No. 2 (T_I–T_{VI}) in the daytime (d) and night (n) under the fair weather, in mm/h of water equivalent

It should be added that the character of the weather also affects the diurnal variation of ablation. When the weather is fair, the difference in ablation between daytime and night is considerable; the difference diminishes when clouds cover the sky, and almost vanishes during rainfall. These differences can be seen from the following data: on August 6 the mean daytime ablation was 5.44 mm water per hour; in the night from August 5 to 6 it was only 1.79 mm (Fig. 18) and almost the same, 1.77 mm, during the following night. Unfortunately, no data are available indicating ablation during cloudy or rainy nights.

GLACIER MOVEMENT

This movement was measured on Profile No. 2 (Fig. 14) which, as mentioned before, was established June 29, 1968 at the narrowest cross-dimension of the glacier between Sulutinder and Faerneste, 8.6 km wide. In this profile 25 stakes were set up in a straight line: 23 on the ice surface (T_{XI} to T_{XXXIII}) and

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one stake each in the mountain scarps (T_{X} in the west, T_{XXXIV} in the east). The stakes set in the ice surface were also used for determining ablation.

Shortly before leaving Iceland, on August 16, 1968, the author had the position of all stakes checked geodetically; this field work including the calculations involved was done by the geodesist of the Expedition, Tadeusz Konysz, M. Sc. (Figure 19) shows position of the stakes as they were on June 29 and, again, on August 16. As can be seen, all stakes set in the ice had moved downward. This movement was greatest for the middle part of the glacier: here Point T_{XXIV} had moved farthest, i.e., 56.5 m during a period of 48 days, equal to 1.18 m travel per day. Had this rate of movement persisted the year round irrespective of seasons, the glacier would have moved at this point a distance of 430.7 m/year. It is worth noting that the zone of maximum glacier "flow" coincides fairly accurately with the zone containing the maximum number of crevasses, between Points T_{XIX} and T_{XXV}. The rate of glacier movement decreases systematically from the middle part toward the margins. The extreme stakes moved least: Point T_{XI} set at 510 m from the mountain scarp moved a bare 25 m, that is 0.52 m/day, while Point T_{XXXIII} situated 1300 m from the eastern rock wall moved 32.5 m or 0.68 m/day.

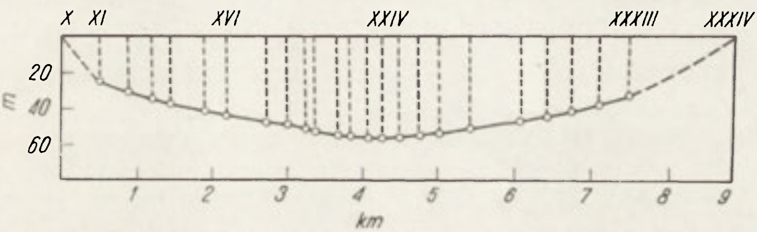

Fig. 19. The movement of Skeidararjokull along transverse profile (No. 2) in the period from June 29 to August 16, 1968

For comparison's sake the author cites figures illustrating the movement of neighbouring glacier lobes. Thorarinsson (1939) reports that investigations, made in 1936–1938 for a full year by a Swedish-Icelandic Expedition, determined in 1937 for Hoffellsjökull a mean rate of movement of 110 cm/day and for 1938 a rate of 80 cm/day, while for the summer seasons the daily figures were, respectively, 211 and 137 cm/day. On nearby glaciers descending from the Oraefa massif, King and Ives (1955) made measurements of the glacier movement; according to these authors, in the time from June 30 to August 13, 1954, the Morsarjökull moved at an average rate of 30 cm/day equalling 108.5 m/year. The change observed in the position of a mark on the ice surface showed that from August 14, 1953 to July 14, 1954 the glacier had moved 92.5 m; adding 9.1 m to bring this period up to a full year, these authors found the glacier movement to amount to 101.6 m/year. The comparison of the two reported full-year values suggests that Morsarjökull maintains a fairly uniform rate of movement throughout the year.

The same two authors have found that in its zone of highest dynamics the Skaftafellsjökull moved at a rate of 43 cm/day, equalling 156 m/year, and that for Svinafellsjökull these figures are 47 cm/day and 172 m/year.

All the results of the above cited measurements of glacier movement referred to zones at a considerable distance from the glacier snouts, hence zones of high dynamics. By comparison it is interesting to know what measurements

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near glacier snouts have revealed. This sort of study was made by Price (1969) on Breidamerkurjokull in the time from August 24, 1965 to August 14, 1966. During practically a full year this part of the glacier moved barely 24 m, equal to 6 cm/day.

GLACIER THINNING AND REGRESSION

On glaciers with a negative balance of ice masses to which class the Iceland glaciers also belong, the ultimate result of ablation is that the glaciers grow thinner and retreat. The author calculated the thinning, in other words, the surface lowering of Skeidararjokull in its zone of ablation for a 41 year period, with hypsometric maps dating from 1904 and 1945 as basis. He determined the effect of these processes from studying three profiles: the longitudinal profile No. 1 and the two transverse profiles Nos. 3 and 4 (Fig. 14); the results obtained are presented graphically in Figs. 20 and 21.

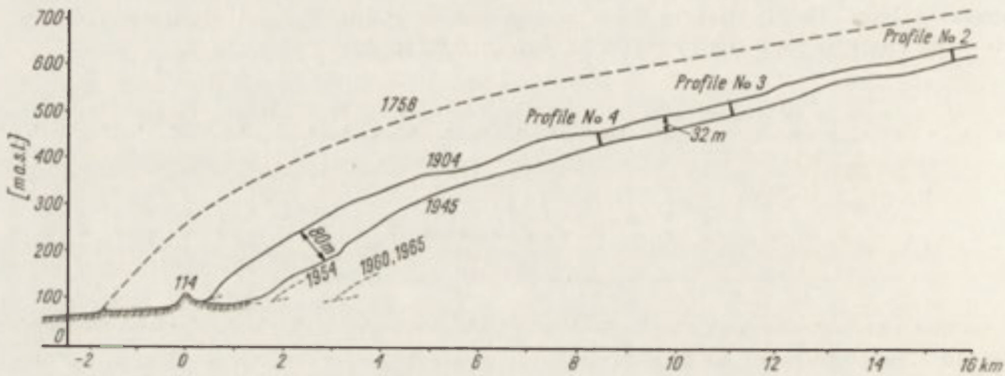

Fig. 20. Supposed Skeidararjokull surface in its maximal extent in historical time (latter part of the 18th century), and the surface in 1904 and 1945, and extent of snout in different years (along Profile No. 1)

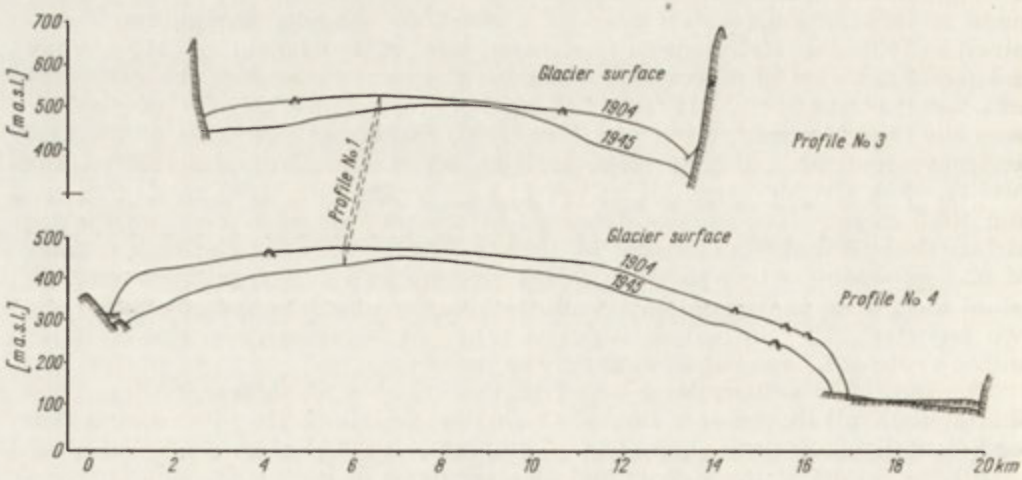

Fig. 21. Skeidararjokull surface in 1904 and 1945 along transverse profiles (Nos. 3 and 4)

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From the way the lines run (which for 1904 and 1945 indicate the altitude of the glacier surface along the longitudinal profile) it appears that the drop in thickness of the glacier surface and therefore its thinning decreases with increasing distance from the snout. From 1904 to 1945, the ice layer which melted was 80 m thick in the snout part; in some 8 km from the position of the snout, where the longitudinal profile crosses profile No. 4, there melted 42 m of ice and at the crossing with profile No. 3 — 28 m of ice. On the whole, this pattern of glacier thinning tallies with that of the rate of ablation, as discussed above, regarding the longitudinal profile. In the snout part, the drop of the glacier surface averaged 2 m/year in the years 1904–1945. This is by no means much, considering, for instance, the value for ablation during the summer season (let us remember that from June 21 to August 22, 1968, the ablation removed an ice layer 450.5 cm thick at Point T_{I}). The thinned-down glacier surface took on its new shape under the influence of periodical advances which led to a thickening of the glacier and to its surface being raised.

Figure 21 illustrates the changes the glacier surface underwent along the transverse profiles Nos. 3 and 4; here the lowering of the ice surface proceeded much more rapidly in the marginal zones than in the middle zone of the glacier. In profile No. 4 we note, for instance, that the glacier surface fell 100 m in the western, and 60 m in the eastern marginal zone, whereas in the middle zone this decrease was only 20 m. Clearly visible here is the identical behaviour of the ablation rate observed in the Summer of 1968 as it is shown in Fig. 16. In comparison, the authors King and Ives (1955) calculated that for the 1904–1953 period the decline of the Morsarjökull surface must have been 90 m. And these same authors established from traces of the glacier margin left on rock walls (the so-called "trim-lines") that during the same period the eastern part of Skeidarárjökull decreased 55 m in height in the Nordurdalur region.

A further effect of the prevalence of ablation over accumulation is the retreat of the glacier snout. Most of the information regarding the changing position of the snouts of Icelandic glaciers are found in papers published by Ahlmann, Eythorsson and Thorarinsson (1937, 1949, 1943). From these the author quotes only some of the data which are available and which seem to be sources, the most important. The above authors maintain that in historical time Skeidarárjökull, like most of the Iceland glaciers, had its farthest extent in the latter part of the 18th century. From this time are dated the moraines which up to now have survived in the Skeidarárjökull forefield, and which on maps have been marked as Sandgigur. After Thorarinsson (1943), the glacier margin was situated in 1756 some 1700 m south of its 1904 position. Both Ahlmann (1937) and Thorarinsson (1943) also report on the position of other glacier snouts as they were situated in later years, up to 1938.

In 1932 Eythorsson (1949) started his methodical annual measurements of how far the snouts of all most important glaciers on Iceland were extending. With regard to Skeidarárjökull, Eythorsson gives data separately for the western and the eastern part because of the unequal rate of retreat of particular parts of this glacier. According to measurements made by this author, the western part of Skeidarárjökull in 1929 made a particularly abrupt advance and in 1932 its snout margin was 390 m ahead of its 1904 position; for the eastern part this advance covered only 260 m. In 1932–1933 the glacier was stagnant, its recession amounting to a bare 3 m. Next, from 1933 to 1938 the glacier retreated at the average rate of 43 m/year in its western and 21 m/year in its eastern part. The 1939–1940 period again brought a recession, 3 m in the

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western and 2.5 m in the eastern part. This was followed by 5 years of a rapid recession, the mean annual values being 217 m in the western and a mere 28 m in the eastern part of the glacier snout. Then, in 1946–1947, came a further but less rapid recession.

In Figure 14 the author indicated the ranges of the snout margin of Skeidarárjokull, copied from the 1904 map of the Danish General Staff, from a map compiled in 1945, and from aerial photographs taken in 1960 and 1965; additionally shown is the line of the snout margin reported for 1954 by Todtmann (1960). The author shows all these marginal lines in Fig. 20 as points marked in the line of profile No. 1. Along this profile the Skeidarárjokull glacier retreated 900 m between 1904 and 1945, a further 400 m from 1945 to 1954, and an additional distance of 1300 m in the time from 1954 to 1965. As a result the period from 1904 to 1965 shows in profile No. 1 a recession of 2600 m, and a mean annual recession of 42.6 m. It will be noted from Fig. 17, that in a certain section of the eastern part of the glacier snout, in 1965 an advance of the glacier margin is shown (compared with the 1960 position), while in the remaining sections a slight recession took place.

The author determined by planimeter readings the size of the particular areas which for the periods mentioned were uncovered by successive deglaciations; these areas are as follows: between 1904 and 1945 the glacier retreated from 15.5 km², and between 1945 and 1960 from a further 20.4 km², while from 1960 to 1965 an area of 0.6 km² was again lost due to renewed glaciation, because in this period part of the snout margin regressed exposing 1.7 km² of forefield, and part of it advanced covering 2.3 km². In total, between 1904 and 1965 the Skeidarárjökull retreated from an area of 35.3 km².

In conclusion, the author wishes to draw attention to an undeniably interesting problem: the asymmetry in recession of the Skeidarárjökull front. While from 1904 to 1965 the recession of the western part covered 2600 m, the eastern section only retreated 500 m. The author ascribes this surprising fact to differences in climatic conditions governing the snout zone and these conditions, in turn, to the particular pattern of local air circulation. He dealt with these matters in greater detail in his report on meteorological investigations, and therefore they will be treated herein very briefly. In daytime, when solar radiation predominantly covers the entire glacier forefield, the winds eminating from the sea, pass in their direction toward the glacier over the sun-heated outwash plains and moraines. These winds come into collision with a glacier wind descending from the glacier snout. But, in the western part of the glacier the warm wind from the ocean is more powerful: it strikes the glacier and here the face of the snout is under its influence. Conditions are different in the eastern part of the glacier. Here the glacier wind is joined by a much stronger catabathic wind descending from high ice-covered parts of the Oraefa massif; together they reach far southward into the glacier forefield and constitute an effective barrier obstructing the penetration of warm winds from the sea and from the heated outwash and moraine plain. This makes the climate of the proglacial zone more severe in its eastern part than it is in the western zone, and this is why in the east the ablation processes are much less intensive. Ultimately this is also the reason for a slower retreat of this eastern part of the glacier snout.

The range covered by winds from the sea is also indicated by a characteristic pattern of cumulus clouds; during daytime, with growing wind intensity, this cloud string gradually moves inland. About noontime these chains of cumulus clouds approach the snout margin in the western part. But farther east,

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i.e., the nearer the high Oraefa massif we move, the greater remains the distance of these clouds from the glacier margin.

There are also some floral traits which signify differences in the climate of the proglacial zone. Observations made by the author, though cursory and rather amateurish, indicate that in the western part of the proglacial zone the vegetation is more abundant and demanding better climatical conditions than in the eastern part: in the western part there are more flowering plants, whereas in the eastern one there grow rather more mosses.

Fig. 22. Structural soils in the Sulutindar near the middle part of ablation zone of Skeidararjökull (*ca.* 650 m a.s.l.). Photo by G. Wójcik

It may be that the slower rate of recession of the eastern part of the glacier is due to a more rapid movement of the eastern part of the glacier toward its snout. Thus we see that the concept put forward by the author is only hypothetical in character. And the problem of asymmetry might be solved only by further detailed studies of ablation and glacier movement, separately for the western and the eastern part, combined with simultaneous continuous observations of climatic conditions in their forefields.

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REFERENCES

Ahlmann, H. W.: son, 1937, Vatnajokull in relation to other present day Iceland glaciers, *Geogr. Ann.*, 19.

Ahlmann, H. W.: son, 1937, Previous investigations of Vatnajokull, marginal oscilla-

http://rcin.org.pl

tions of its outlet-glaciers, and general description of its morphology, *Geogr. Ann.*, 19.

Ahlmann, H. W.: son, 1939, The regime of Hoffellsjokull, *Geogr. Ann.*, 21.

Ahlmann, H. W.: son, Thorarinsson, S., 1938, The ablation, *Geogr. Ann.*, 20.

Eythorsson, J., 1949. Variations of glaciers in Iceland, 1930–1947, *J. Glaciol.*, 1, 5.

Eythorsson, J., 1953, French-Icelandic Expedition to Vatnajokull, 1951, *Polar Rec.*, 6, 45.

Fisher, J. E., 1950, Ice pyramids on glaciers, *J. Glaciol.*, 1, 7.

Glaciers bands. Conference on terminology, 1953, *J. Glaciol.*, 2, 13.

Ives, J. D., King, C. A. M., 1954, Glaciological observations on Morsarjokull, SW Vatnajokull. Part I — The ogive banding, *J. Glaciol.*, 2, 16.

Ives, J. D., King, C. A. M., 1955, Glaciological observations on Morsarjokull, SW Vatnajokull. Part II — Regime of the glacier, present and past, *J. Glaciol.*, 2, 17.

Kalesnik, S. W., 1963, *Ocherki glatsyologii* (An outline of glaciology), Moskva.

Keller, R., 1965, *Vody i vodnyy balans sushi* (Waters and water balance of dry territories), Moskva.

King, C. A. M., Ives, J. D., 1955, Glaciological observations on some of the outlet glaciers of south-west Vatnajokull, Iceland, 1954. Part I — Glacier regime, *J. Glaciol.*, 2, 18.

Kosiba, A., 1938, Kilka zagadnień z morfotektoniki i glacjologii Islandii (Some remarks on the morphotectonic and glaciology of Iceland), *Czas. Geogr.*, 16, 4.

Okko, V., 1955, Glacial drift in Iceland, its origin and morphology, *Bull. Comm. Geol. Finl.*, 170.

Price, R. J., 1969, *Moraines, sandar, kames, and eskers near Breidamerkurjokull, Iceland*, Trans. Inst. Brit. Geogr., 46.

Thorarinsson, S., 1937, The main geological and topographical features of Iceland, *Geogr. Ann.*, 19.

Thorarinsson, S., 1939, Hoffellsjokull, its movements and drainage, *Geogr. Ann.*, 21.

Thorarinsson, S., 1939, The ice-dammed lakes of Iceland with particular reference to their values as indicators of glacier oscillations, *Geogr. Ann.*, 21.

Thorarinsson, S., 1943, Oscillations of the Iceland glaciers in the last 250 years, *Geogr. Ann.*, 25.

Thorarinsson, S., 1953, The Grimsvotn Expedition June–July 1953. *Jokull.* 3.

Thorarinsson, S., 1953a, Some new aspects of the Grimsvotn problem, *J. Glaciol.*, 2, 14.

Thorarinsson, S., 1964, Sudden advances of Vatnajokull outlet glaciers 1930–1964, *Jokull*, 14.

Todtmann, E. M., 1960, Gletscherforschungen auf Island (Vatnajokull), *Univ. Hamburg, Abh. Auslandsk.*, Reihe C, 65, 19.

Wójcik, G., 1970, Ablational processes on the Skeidararjokull (Iceland), *Bull. Acad. Pol. Sci., Ser. Sci. geol. geogr.*, 18, 4.

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HYDROGRAPHIC FEATURFS OF THE PROGLACIAL AREA OF SKEIDARÁRJÖKULL

ZYGMUNT CHURSKI

PREFACE

The hydrographic investigations were concentrated upon the proglacial area of Skeidarárjokull which is representative of a glacier lobe that at present is in retreat. This glacier descends to the coastal plain covered by sheets of glacial and glaciofluvial deposits. The base of this glacier lies at some 95 m a.s.l. In the forefield of the Skeidarárjokull a dense network of streams and lakes has developed, fed by waters derived from glacier ablation, from precipitation, from ground water sources and, likewise, from melting dead ice blocks buried in the glacial accumulation deposits.

The purpose of the hydrographic examinations undertaken by the author was to compile a hydrographic map of the forefield of the Skeidarárjokull, to collect material for a classification of the streams of this area, to understand the dynamics of these streams, and to understand the part which running water has had in forming this young glacial relief. Careful attention was also given to studies of postglacial lakes, especially of their origin, their thermal regime, and the effect which the lakes have upon water level oscillations in the rivers.

Observations were also maintained of the oscillations of the highest groundwater table, and its temperature. In addition for drawing up a water balance, measurements were made of the volume of flow in the rivers draining the Skeidarárjökull. The available aerial photos and some cartographical material made it possible to illustrate the evolution of the fluvial network as it took place in the period from 1946 to 1968.

The rivers and lakes occurring in the proglacial areas of Vatnajökull have repeatedly been an object of interest to hydrologists. A considerable amount of information, especially about flow volume, was collected by Rist (1956) who, among other data, reports on flow conditions in the Sula River. This river has also been studied by the present author. The water runoff in the forefield of Hoffellsjökull was an object of investigation by a Swedish expedition in 1951–1952. At that time, hydrological observations were made with Hjulström (1952, 1955) in charge, who himself concentrated his attention upon the dynamics of the ground water in the Hoffellssandur area, while the investigation of the rivers and lakes of this area was in the hands of Arnborg (1955) and Jonsson (1955). For the most part, these Swedish hydrologists paid attention to conditions of water runoff, to the quantity of ablation water carried off, and to the origin of the frontal lakes.

With regard to the rivers draining the forefield of the Skeidarárjokull

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glacier the majority of data collected dealt with the Skeidará River. These data consist mainly of records of the huge water volume carried off during the period which in Iceland is known by the term *jökullhlaup* (Thorarinsson 1953, 1954, 1955, 1967). On the other hand, no observations seem to have been recorded referring to the great number of lakes scattered over the Skeidarárjökull forefield, or to the Sandgigjukvisl River, one of the most conspicuous streams of this forefield.

Apart from the publications listed above, a good deal of information on the characteristics of the fluvial system has also been obtained from cartographical material, especially from aerial photographs; all of this source material has been fully utilised in the present paper. However, within the limited scope of our paper it became only possible to present the most important data derived from analytical studies of the source material and, in particular, of the aerial photographs taken successively by the U.S.A. Photogrammetric Service in 1946, 1960 and 1965.

Taking into due account all the available source material in the problem of the area under discussion, the present paper reports for the most part the results obtained from field examinations made in the period from June 20 to August 25, 1968. While our hydrographic studies were admittedly carried out in a relatively short period of time, they did enable us to investigate in a comprehensive way the hydrographic conditions ruling the Skeidarárjokull forefield. Incidentally, it might be claimed that what we observed in Iceland might prove useful in the palaeohydrographical research now under way in Northern Poland.

GENERAL HYDROGRAPHIC CHARACTERISTIC OF THE SKEIDARÁRJÖKULL FOREFIELD

The rivers draining the ice cap of the Vatnajökull escape radially in all directions. However, these rivers by no means resemble each other. Their divergence stems partly from differences in the geological structure of the area, a river crosses, and partly from their different distance from the ocean. The majority of the rivers which drain tongues of the Vatnajökull force their way to the ocean across basalt elevations, which form many waterfalls. Other rivers start their course in the coastal lowland adjoining glaciers and they usually flow over outwash sheets, forming in their course a hundred branches and flooding wide areas, especially in their estuaries. To this latter group belong the rivers draining the Skeidarárjökull.

Today the meltwater from Skeidarárjökull is drained by three big rivers which flow into the ocean. Two of them (the Sula and Skeidará Rivers) are what is called lateral rivers, while the middle river, Sandgigjukvisl, carries off the waters from the principal, i.e., the central part of the glacier front (Fig. 1). Several minor streams also occur in the three drainage basins, but they disappear in the outwash sands near the front of the glacier. The base of the glacier snout lies at about 95 m a.s.l.; its distance from the ocean is 20 to 30 km. This indicates that the main river gradient is some 3 to 4‰.

The Sula River starts its course from Lake Graenalon; until recently this lake had been drained by a neighbouring river, the Nupsa. Along the margin of the Sulutinder massif the Sula runs mostly underneath the ice; in this way it drains the various lateral lakes lying next to its course. This has had the effect that the water level has dropped considerably in all lateral lakes and also in Lake Graenalon, as seen from surviving cliffs and high lake terraces.

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Fig. 1. Hydrographic sketch of Skeidararjokull

1 — surface of the Skeidararjokull (number shows the height a.s.l. in metres), 2 — substratum of glacier (number shows the height a.s.l. in metres),* 3 — river net, 4 — lakes, 5 — basalt massifs, 6 — subglaciel river, 7 — river gauging station, 8 — limnograph, 9 — meteorological station, 10 — limit of Skeidararjokull, 11 — groundwater measuring station, 12 — moraines, 13 — vegetation, 14 — marshes, 15 — farms

* According to seismic measurements made by the French-Icelandic Expedition in 1951.

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In the case of Lake Graenalon this subsidence of the water table is about 100 m.

From under the ice the Sula River starts its surface course in the form of an immense mass of upwelling water, with a flow volume of some 50 m^3/s. After thus being forcefully ejected from the ground, the waters flow in several mutually interlaced branches; the branches of this river are from some 15 to 50 m wide and 1 to 2 m deep. The rate of flow is 2 to 3 m/s; the water carries suspended matter in a quantity of 10 g/l and is capable of tumbling rock fragments of the size of a human head over the river floor. Some of the Sula branches join the nearby River Nupsa. Near the ocean the branches of the Sula spread out, flooding the wide outwash plain and leaving behind most of the suspended matter they had carried.

For the most part the Sula River is fed by meltwater streams, but it also carries atmospheric precipitation, or waters from spring-melt arriving from the area situated near Sulutinder; here the Suludalsa River is the most substantial Sula tributary draining this area. These different sources of water flow are important when the water balance is considered, or the changes observed in the water level of the Sula.

This complex system feeding the Sula River tends mainly to smooth away differences in the water level. On sunny and warm days the river is mostly fed by glacier waters and shows the diurnal fluctuations characteristic of glacier rivers, while on rainy and cool days when ablation grows less, the inflow of precipitation waters is more effective. This is why the water level in the Sula River is subject to wide variations, depending on the quantity of glacial water it receives, on the quantity and frequency of precipitation, and on the eventual runoff from lateral lakes. After a measurement made on August 18, 1968, the flow was about 48 m^3/s, and this figure may be considered characteristic of mean conditions for the summer of 1968.

The water flow in the Sula River differs sharply from the flow observed in the nearby Nupsa River which carries clear water, has a rock-strewn floor, and receives its water only from precipitation. Hence the oscillations in the Nupsa water level are merely affected by the quantity of rainfall, and differ from the oscillations recorded for the Sula River.

The eastern flank of the Skeidararjökull is drained by the Skeidará River. The springs of this river probably lie under the ice cap of the Vatnajökull, east of the Grimsvötn crater. The Skeidará takes advantage of a longitudinal depression underneath the Skeidararjökull, and at the same time drains the subglacial lakes and the lateral lakes situated along the eastern rim of this glacier.

The River Skeidará originates from two powerful artesian type ice springs from which two river branches start: a western branch which forms the true Skeidará channel, while the eastern branch is called the Kaldakvisl. Close to the glacier margin the channels of these two rivers are from 30 to 50 m wide and 1 to 3 m deep; the rate of flow is some 3 m/s. Rivers flowing as rapidly as this carry not only material from the waning glacier; they also pick up enormous quantities of material from their banks which are being eroded and transport it downstream.

Some 2 km from its origin the rate of flow of the Skeidará River subsides markedly; the river splits up into a number of narrower branches which only in the estuary area join again, forming wide flooded areas. From the east, several minor tributaries join the river, arriving from the Skaftafellsjökull

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and Svinafellsjokull glaciers; so do a number of mountain creeks escaping from the Skaftafels massif.

Apart from these streams, three further minor glacier streams join the Skeidara; their springs lie south of the main artesian type ice-springs.

In view of the fact that the Skeidara River is for the most part fed by subglacial waters and by water accumulated in lateral lakes, the volume of flow in this river depends mainly on the conditions of this kind of runoff, not on the intensity of ablation. And therefore the diurnal oscillations of the water level do not show the same pattern of recurrence here as are observed for other glacial rivers. Based on approximate calculations, the discharge of all branches of the Skeidara River, including the Kaldakvisl, is estimated to be about 164 m^3/s.

A characteristic feature of the lateral rivers, particularly in evidence in the Skeidara drainage basin, are huge floods brought about by the eruption of the subglacial Grimsvotn volcano. Among other sources, this phenomenon was known, too, and has been reported by Thorarinsson and Hjulström who call it *jokullhlaup*.

During heavy type *jokullhlaup* floods the volume of the Skeidara flow may well rise to 50,000 m^3/s (Thorarinsson 1954). With this huge volume of water to carry the river grows to a width of 8 km and to 30 m depth. During these floods the river destroys the forefield area and spreads enormous masses of ice all over it. This feature must be kept in mind, when reflecting upon the origin of the outwash plains occurring in the outer parts and of the enormous number of small kettles which might have originated from melting-out of ice blocks scattered and partly buried during *jokullhlaup* periods.

In the marginal rivers which carry the overflow from the numerous subglacial frontal and lateral lakes, similar floods may also occur due to an abrupt draining of these lakes. Floods of this kind do not equal *jokullhlaup* floods in volume; but, to cite an example, when Lake Graenalon lost much of its water, the flow in the Sula River must have been very great, considering the fact that the water level in that big lake dropped some 100 m.

The middle part of Skeidararjokull is drained by the Sandgigjukvisl River. The upper reach of this river consists of several hundred small rivulets which issue on the glacier surface, some only a few, others as many as a dozen kilometers or so from the glacier snout. For the most part these watercourses do not reach the glacier margin because they disappear in glacier tunnels and appear outside the glacier in the form of inglacial streams or artesian wells (Fig. 2). Due to the motion of the glacier, both the channels of the glacial streams and the inglacial tunnels undergo continuous dislocation. Changes of this sort also affect the ice-springs and the springs with free outflow.

The water of streams flowing over the glacier surface is very clear. During the day the temperature of this water is near 0°C, and during the night the streams change into ice unless they flow rapidly. Near the glacier flanks the streams undergo pollution by material from the ablation moraine. An example of this kind of a glacier stream is shown in Fig. 3.

Of somewhat different character are glacier streams flowing down the scarps of the glacier. They flow along a steep gradient; their width usually varies from 15 to some 50 cm, their depth from 20 to 30 cm. In their run over the scarp of the glacier snout these streams form numerous steps which the water overcomes by small waterfalls (Fig. 4).

The streams flowing over the glacier surface and over its flanks form the upper part of the Sandgigjukvisl drainage basin. At the time our observations

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Fig. 2. River flowing over glacier surface. Photo by Z. Churski

Fig. 3. Glacial brooks passing over Skeidararjokull scarp. Photo by Z. Churski

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were made, the number of such streams draining into this basin was about 120. These streams are important due to their incising the flanks of the glacier and in this way speeding up its retreat.

Often the streams flowing from the glacier base cannot run off freely; therefore they accumulate their water between the side of the glacier and the ice-moraine ridges (ice-cored moraines), forming minor lakes from which the water escapes by overflowing the moraine barrier, or by underground flow.

Fig. 4. River outflow from ice tunnel under glacier. Photo by Z. Churski

In contrast to conditions at the lateral Sula and Skeidarȧ Rivers, the melt-water streams leaving the glacier within the Sandgigjukvisl drainage basin form, ahead of the glacier snout, a dense network of several hundred rivulets which, at some distance from the glacier snout, unite into one large water-course. But, before joining the main stream, many of these rivulets pass through some of the great number of lakes situated in the glacier forefield.

Again, differing from the Skeidarȧ and Sula Rivers, the Sandgigjukvisl River is mainly fed by ablation waters. However, in view of the low altitude of the glacier snout, this entire region lies in the zone of fairly abundant precipitation (some 2000 mm annually), and this fact greatly affects the way the rivers draining the Skeidarȧrjökull are supplied with water. Hence the Sandgigjukvisl River is characterized by a high degree of flow variance and by an irregular course of its water levels.

From this pattern it might be concluded that Sandgigjukvisl is ruled by a regime that might be called glacial-pluvial. Rivers of this kind of regime are a rarity now, because few modern glaciers lie in zones of ample rainfall. Still, one might assume that this is the character of the rivers which once were fed from retreating Pleistocene ice-sheets in the northern parts of Poland where, apart from ablation streams, waters from precipitation were playing an important part in feeding the rivers. This consideration was one

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of the reasons why the Polish Expedition undertook detailed examinations in the drainage basin of the Sandgigjukvisl River.

In addition to the three river systems mentioned which drain the Skeidarárjokull glacier, there are a number of further minor streams in the watershed region between Sandgigjukvisl and the two lateral rivers. But they are short, and they are in operation only in daytime, during the period in which ablation is of high intensity. For the most part the water carried off by these streams does not join any of the main rivers, instead, usually this water vanishes into the outwash sands at a distance of a few hundred meters from the glacier snout. However, cartographic material reveals that at the time the glacier snout was farther south, these streams were larger and they joined the main rivers, as can be seen from their well preserved old channels. An example of a "dying" river is the Blautakvisl which issues east of the Sula river. Further, in the watershed area between Sandgigjukvisl and Skeidará a number of undrained frontal lakes can be observed.

The annual course of water levels in the rivers mentioned bears the features of a glacial regime for which the highest levels occur in June, July and August, and the lowest in the winter months. But is must be remembered that this course is disturbed by rainfall which, sometimes coinciding with high meltwater flow, may cause short-lived floods. Most often this type of floods happen in autumn. All these general data regarding oscillations of the water levels, obtained from available information, were confirmed by the examinations made by Arnborg (1955) and Rist (1956) who made this sort of examinations on other glacial rivers, the Ytry-Rangá and the Austurfljot. But while during the winter large rivers show merely markedly lowered water levels, smaller streams and brooks vanish altogether and often reappear only in spring, sometimes at a different place. Low water levels in the rivers during wintertime also lead to the same phenomenon in the lakes, and near river channels they may also cause the lowering of the groundwater table.

DETAILED HYDROGRAPHIC CHARACTERIZATION OF THE FOREFIELD OF THE SKEIDARÁRJOKULL (STREAM CLASSIFICATION)

The author bases his characterization of the hydrographic network, combined with a classification of the rivers and lakes situated in the forefield of the Skeidarárjökull, on his field studies, principally undertaken in the Sandgigjukvisl drainage basin. The evolution of the fluvial system in this basin depends, first of all, on the way ablation proceeds and how the land surface is profiled ahead of the glacier snout. In the Sandgigjukvisl basin the *jokullhlaup* floods do not cause changes as large as in the drainage basins of the two lateral rivers. Evidence of this is that the low-lying landforms of glacial accumulation have not been washed away and that here well preserved lake hollows have survived. On the other hand, rain showers are important, especially those falling in autumn. This shows that the conditions under which the fluvial system of the Sandgigjukvisl basin has been developing greatly resemble those ruling the period of recession of the Scandinavian inland ice.

In spite of the nearness of the glacier which is here the main source of water the hydrographic system of the forefield of the Skeidarárjökull is highly diversified. Several types of watercourses and several types of lakes can be seen here, each in a different stage of evolution and each playing a different part in shaping the glacial relief. In the present paper the author intends to briefly discuss the different genetic types of watercourses and lakes, and

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to present the results of his observations with regard to thermal conditions and dynamics of the surface waters. Further, his intention is to report the most important results obtained from investigating the groundwater of the discussed region, and to supply certain data on the water balance of the rivers draining the Skeidarárjökull. The attached hydrographic map of a fragment of the Skeidarárjökull forefield has been compiled in conformity with official directives regarding the hydrographic map of Poland (Fig. 5; see the annex).

STREAMS

As a result of his detailed hydrographic mapping of the forefield of the Skeidarárjökull glacier the author distinguishes four types of streams, each type having a separate regime. Most abundantly seen in this region are glacial streams, that is those which obtain their water directly from glacier ablation.

The glacial rivers show vastly different features, depending on the way they issue from the glacier. Upon leaving the glacier they meet no obstruction, the glacier streams are usually of smaller size, say 1 to 3 m wide, and they run very close to each other; they appear rather in the shape of braided arms joining or even crossing each other (Fig. 6). When ablation is more

Fig. 6. Example of runoff of glacial waters in close vicinity to glacier. Photo by Z. Churski

intensive, these streams spread out covering wide surfaces and eroding the material of the area over which they pass. On the other hand, where runoff is in some way obstructed, the waters accumulate in front of the glacier snout and, after breaking through the ice-moraine ridge, they continue their run forming larger rivers. Finally, when it comes to streams which originate from artesian wells, these form alluvial cones in the glacier forefield, with rapidly

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changing systems of channel forms (Fig. 7). About 1 m from the glacier margin these small glacial streams join into single larger arteries and, after breaching the zone of high end moraines, they spread out over the outwash surface until they reach the sea. However, before combining to form a river, many glacial streams first pass a number of ice-dammed lakes where they deposit the debris they have been carrying (Fig. 5).

The density of the fluvial network fed by glacial waters depends on the conditions under which melting takes place, and on the nature of the relief in the glacier forefield. When ablation increases, the changes occurring in the density of the fluvial system can be observed from hour to hour. Most often these changes are produced by the rapid infilling of the fluvial channels with

Fig. 7. Glacial river originating from upwelling artesian spring. Photo by Z. Churski

debris. But, apart from these processes, the evolution of the pattern and the density of a fluvial network is also affected by the melting of dead ice blocks and by oscillating movements of the glacier front.

The rate of flow of glacial rivers depends on the topographical conditions of the proglacial area. In this area the rivers which are fed by meltwater and issue directly from the glacier, run at a rate of 2–3 m/s. Only those streams which are spread out over wider surfaces and locally form outwash fans, run at a much lower velocity not exceeding 1 m/s. However, it must be kept in mind, that this flow velocity varies greatly and primarily depends on the depth of the flow channel. Flowing at rates as high as that, the glacial rivers obviously carry large quantities of material. Those with a flow velocity of 1–2 m/s transport 10–15 g/l of debris and slide or roll enormous loads of material along their bottom. Rivers flowing at 2–3 m/s are easily capable of displacing football-sized large boulders.

The debris accumulated in glacial rivers is for the most part derived from the washing down of the ablation moraine. But it must be admitted that much

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of this material may also be derived from lateral erosion. Because the banks of the river channels consist of loose material not compacted by vegetation, some glacial rivers are capable of widening their channels at the rate of about 1 m/day. Even those rivers which pass through lakes and leave most of their suspended matter there stranded, are apt to undercut their banks after they leave the lakes, and to carry the loosened material downstream towards their mouths.

In view of the fact that the rivers passing a glacier forefield lack a graded profile, banks of sandy deposits develop here and there in river sections where the gradient grows flatter and the flow welocity decreases; but sooner or later these banks are displaced downstream (Fig. 8). In other sections of the river

Fig. 8. Formation of sand-gravel banks in glacial river. Photo by Z. Churski

where the gradient is steeper, processes of bottom erosion increase and this often causes changes in the position of these channels (Fig. 9).

A second type of streams observed in the Skeidarárjökull forefield are streams fed by groundwater inflow; this inflow arrives from the decaying glacier or by infiltration from marginal lakes. The streams fed by groundwater are usually small. Their source is pools occurring in smaller or greater numbers, from which the water joins to form small brooks 1–2 m wide and 0.5–1.0 m deep. The water of these brooks is clear, well filtered and devoid of suspended mineral matter.

The high transparency of these waters, sharply contrasting with glacial waters, causes during the summer a rich vegetation to develop, mostly of algae and mosses. The algae occur here mainly in the shape of threads forming what is known as cotton balls; or they can be found in the shape of gelatinous globules. Four species of algae occur here (after R. Bohr [1]); most common are the

[1] These algae were identified by Dr. R. Bohr of the Institute of Biology at the Nicholas Copernicus University of Torun.

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bright-green accumulations of thalluses of *chlorophycea Tetraspora lubrica* (Roth.) C. A. Agardh. This species is characteristic of cold-water streams; ecologically, biologically and, even, morphologically it resembles *Hydryrus foetidus* Kirch. which grows in mountain brooks. A second species are thalluses of *chlorophycea Ulotrix zonata* (Weber et Mohr) Kuetz, which form olive--green accumulations strongly attached to their base. This very variable species is cosmopolitan in its occurrence, but it probably also develops ecological varieties some of which may be linked with cold waters.

The accumulations formed by the thread-like thalluses of an alga of the *Xauthophyceae* family, probably of species *Tribonema* are the largest. On the other hand, the loose gelatinous globules attached to stones in the stream bottoms of groundwater streams are thalluses of the *Cyanophycea* algae, of species *Nostoc*. Most often a compact vegetation cover grows on channel banks; it even occurs that some streams are fully overgrown by this sort of vegetation.

Rivers fed by groundwater outflow occur in the Skeidararjökull forefield in two zones. The first lies mostly in the western part of the Sandgigjukvisl drainage basin, some 1 km from the glacier snout. Rivers of this type, situated as near as that to the glacier, are fed by meltwater which saturates the loose material of glacial accumulation and appears on the surface at natural depressions in the land relief. The streams thus fed by groundwater flow escape in this region into neighbouring glacial rivers. In shaping the land relief their role is negligible. But they certainly represent an excellent environment for the growth of vegetation — a remarkable feature in view of the close vicinity of the glacier margin.

The second zone with streams fed by groundwater lies in the outwash sheet, sometimes only a few, and more often about a dozen kilometers away from the glacier margin. Here the source is groundwater seeping into the ground from frontal glacial lakes, or glacial water arriving as groundwater from under the moraine area. But precipitation also plays a considerable part in feeding these streams. The streams passing over this outwash area are more important in their relief-forming capacity; they are capable of incising channels 1 to 5 m wide and about 1 m deep (Fig. 10).

The third type of rivers occurring only during the summer season in the forefield of the Skeidararjökull are small streams originating from melting dead ice blocks. These streams occur both in the end moraine zone and on the outwash sheets. They issue from different altitudes and therefore they may differ in both gradient and flow velocity. The number of this type of streams grows with rising air temperature and intensity of insolation. On the whole, these are merely brooks carrying barely a few or, at best, a dozen liters of water per second. This is also the reason why not all of them reach the glacial rivers flowing in their vicinity. This refers mainly to streams issuing from the zone of the high end moraines; they may run some or even several hundred meters, and then they vanish in the moraine sands.

In a hydrological sense the last-discussed streams fed from melting dead ice blocks are of little importance. Compared with true meltwater streams the quantity of water they carry is negligible. However, due to the fact that these brooks flow only during daytime, especially on fair-weather and sunny days, they contribute to raising water levels during afternoon hours. Still, they do play an important part in shaping the relief of this area. At their springs they form distinct sinks or depressions (Fig. 11). Where the flow is copious, earth cracks and mass movements occur. Brooks flowing over loose material on steep

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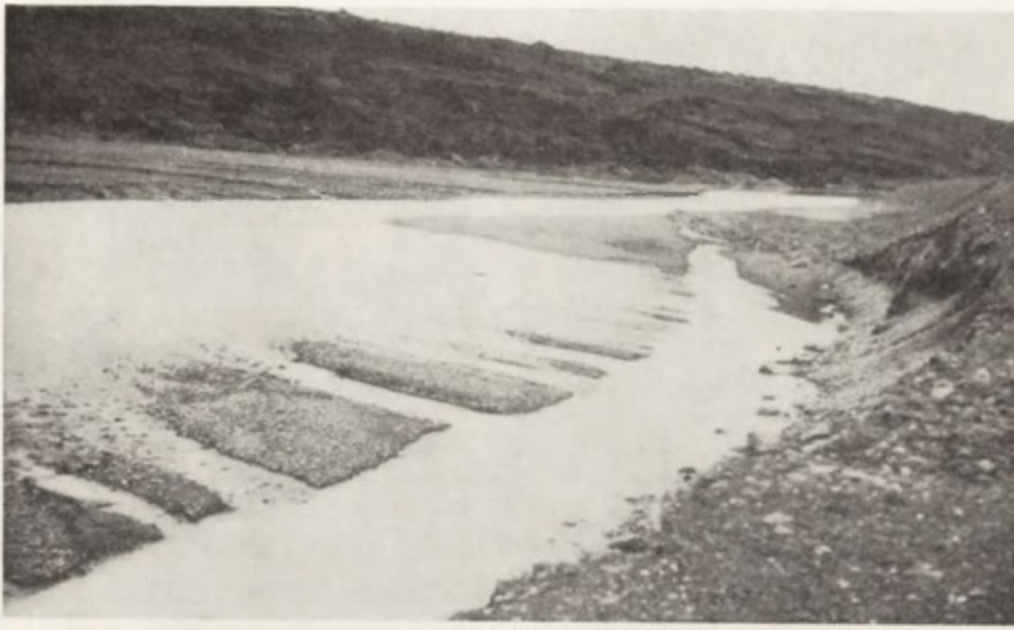

Fig. 9. Glacial river gliding over sand-gravel banks. Photo by Z. Churski

Fig. 10. River fed by groundwater in outwash sheet. Photo by Z. Churski

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scarps of moraines or valleys erode fairly deep gullies (Fig. 12). Those brooks which fail to join rivers and vanish in sand sheets, usually from some kind of classical fans of various sizes (Fig. 13).

The fourth and last group of streams observed in the Skeidararjokull forefield is not directly connected with the melting of the glacier. However, in

Fig. 11. See page cave at spring of a brook, fed by meltwater from buried dead-ice block. Photo by Z. Churski

view of the part they play in relief evolution it seems opportune to mention at least their most important features.

Thus, to the fourth group should be assigned those streams which are fed by water derived from springmelt. They flow only during spring while snow is melting very intensively, and while the ground surface underlying the snow is still frozen. During this time the meltwater cannot seep into the ground and has to flow over its surface.

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Fig. 12. Channel of stream fed from melting dead-ice blocks. Photo by Z. Churski

Fig. 13. Alluvial fan formed by stream fed from dead-ice meltwater. Photo by Z. Churski

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Because the surface of the area crossed by such streams is not fixed by vegetation, the water easily erodes wide flat-bottom valleys in the frozen ground, and where the water reaches a more level area it accumulates the material it had carried into alluvial fans (Fig. 14). In the summer these channels are dry; even after prolonged rain periods they carry no water. The valleys of these streams remind one of the dry lateral valleys which developed in Northern Poland during the glacial period.

All of the types of rivers discussed also take part in carrying off the waters of atmospheric precipitation, mostly rain, although it must be noted that the precipitation water is by no means the principal factor ruling the regime of these rivers. Except for the last-mentioned group of rivers which run only in the spring snowmelt periods, the remaining types, i.e. those fed by ablation

Fig. 14. Valley in outwash sheet eroded by meltwater flow. Photo by Z. Churski

streams, by groundwater, and by water from melting buried dead ice blocks, are connected with conditions of icemelt and are most active in the warmest months of the year, in June and July. They are also affected by the diurnal oscillations characteristic of a glacial regime, caused by differences in ice melting during the day and the night.

To understand these oscillations in detail observations of the water level were started in the Sandgigjukvisl River. A limnograph (water level recorder) was mounted at the place where the river breaks through the moraine zone (Fig. 15). In this section the river has a narrowed channel and reacts much better to changes in ablation conditions than it would in other sections. In order to understand the interdependence between the agencies affecting the ablation and the water level, the air temperature and the degree of insolation were also recorded at the same time, as well as each case of precipitation and its duration. All these data which in a combined form are presented in Fig. 16, give a detailed analysis of the regime of the water level during different stages of ablation; they also reveal the effect the examined agencies have upon

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the diurnal oscillations of the water level. From a detailed study of the graphs for the different agencies shown in Fig. 16 a number of conclusions can be drawn as to the interdependence between the investigated agencies. The very first correlation of the individual graphs indicates that glacial rivers react promptly to every change in conditions of ice- and snowmelt; hence the water level may be looked upon as a sort of faithful indicator of the course of ablation.

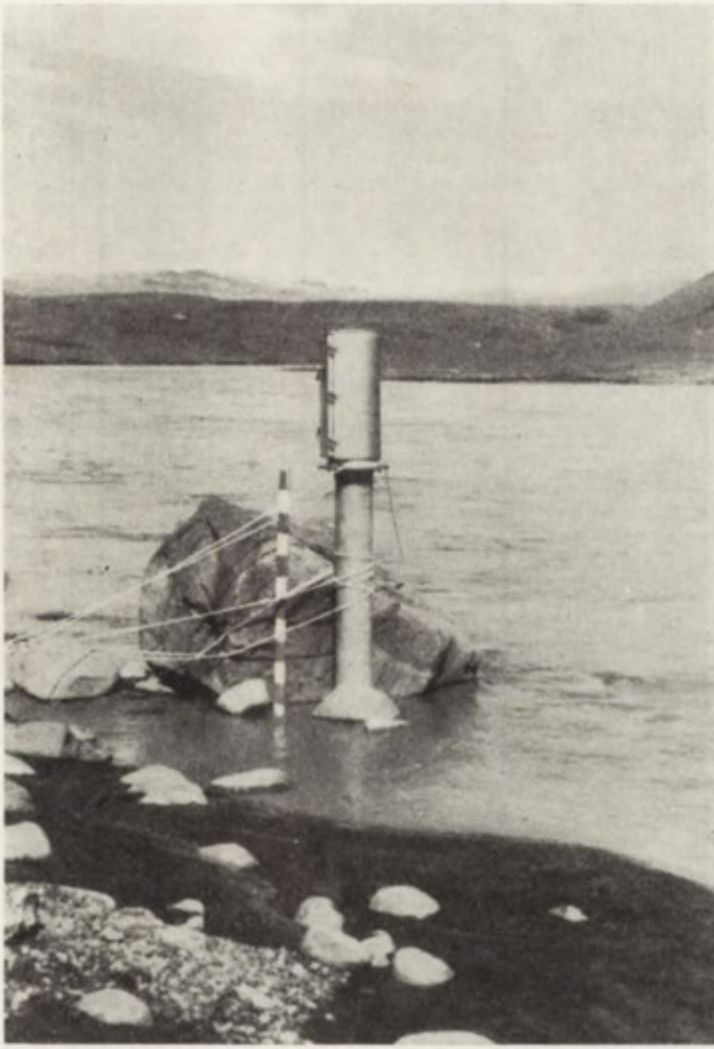

Fig. 15. Limnograph station on the Sandgigjukvisl River. Photo by Z. Churski

The most typical diurnal pattern of the water level in glacial rivers can be observed in fair days, with sunshine. On days like this the oscillation curve of the water level runs in conformity with the curve of air temperature fluctuations (July 5–12, 1968). This sort of fair weather brings an increase in the water level from about 9 a.m. on; highest is the level in the afternoon between 6 and 7 p.m. From 8 p.m., to 8 a.m. the limnograph recorded a subsidence of the water table. The period in which during the warmest months the water level

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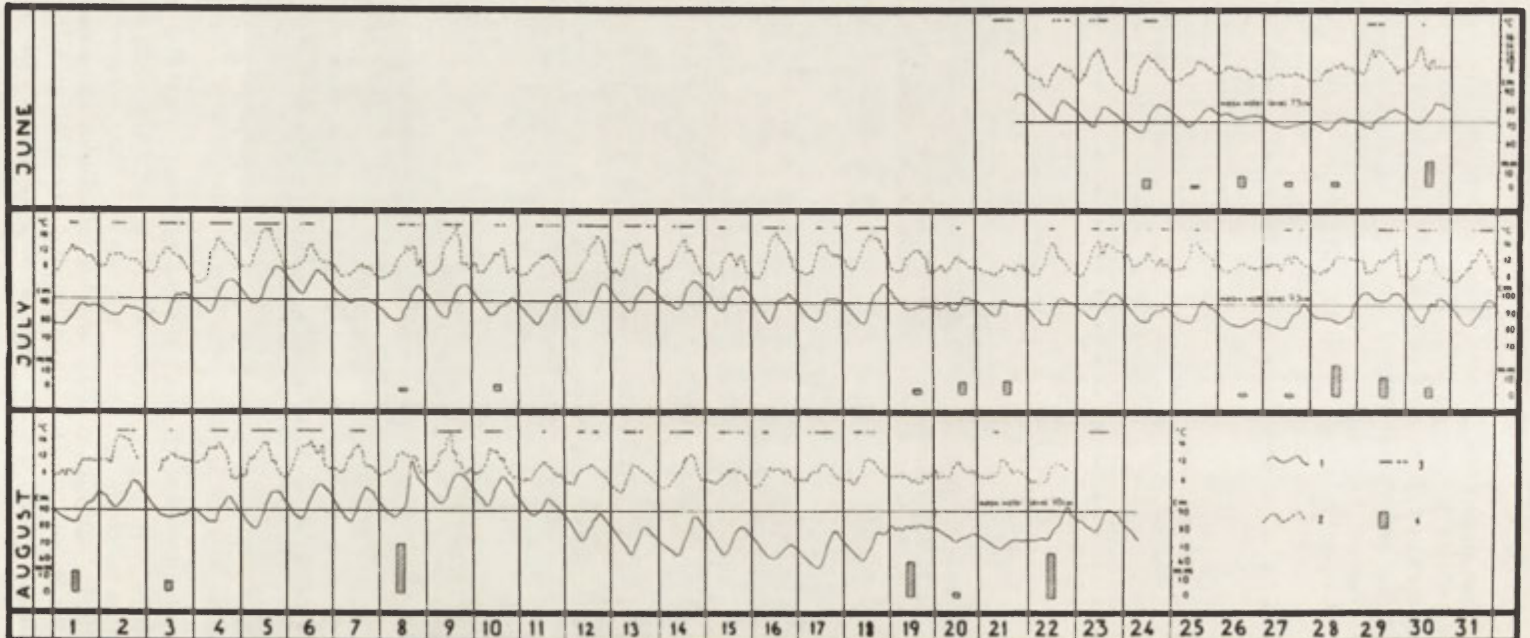

Fig. 16. Graph indicating Sandgigjukvisl water level in relation to air temperature, insolation and amount of precipitation
1 — water level in cm, 2 — air temperature in °C, 3 — insolation, 4 — amount of precipitation in mm/day

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was rising, was equal to the period of the drop in the water level; this lasted about 11 to 12 hours; but a rise in air temperature appears in the records to have set in between 4 and 5 a.m., and the highest temperature occurred in the time between noon and 3 p.m.

The diurnal oscillations of the water level were greatest on warm, sunny days when the air temperature reached 17°C; in cases like this the water level rose some 30 cm during a 12-hour period. It should be noted that a 30 cm rise is relatively little considering the amount of escaping water. On days equally warm and without rain, but with transient cloud, the diurnal oscillations of the water level were never as marked as on sunny days; at times it even happened (July 6–7, 1968) that the water level in the river dropped considerably. It is also worth mentioning that even short breaks in insolation caused by slight clouding have been recorded by the limnograph as periods of lessened flow (for instance on July 15 and 18, 1968).

Under conditions of total cloud and of minor transient precipitation of no more than 10 mm daily, the water level in the river was varying by barely a few centimeters (June 25 to July 5). When the period of light rains continued for several days, the decrease in ablation would cause a gradual lowering of the water level in both the glacial rivers and in the lakes they pass. And only continuous rain showers of more than 20 mm/day caused the water level to rise and to disturb the normal diurnal regime of oscillations. Under atmospheric conditions of the latter type the oscillations of the water level grew very irregular, varying within limits of 20 to 30 cm.

When precipitation becomes more plentiful the highest water levels occur in the afternoon hours. At this time the ablation waters are supplemented by rainwater; the result is an abrupt rise in the water level, reaching as much as 50 cm with regard to what the level was in the night before (Aug. 8, 1968). It should be stressed that the rise of the water is sudden but that, in contrast, the drop of the water level proceeds very slowly.

High floods have also been observed on sunny days following torrential rains. This must be ascribed to the intensified ablation which always increases after the material of the ablation moraine has been washed off the glacier surface, and after the surplus of groundwater has escaped the level which was also raised as a result of the heavy precipitation. After a new cover of an ablation moraine had been developed on the glacier front by the effect of a few sunny days, the water levels observed were no longer as high as they had been shortly after the rain showers.

The oscillations of the water level in the Sandgigjukvisl River are relatively slight probably because of the numerous lakes situated in the drainage basin of this river, the two largest of which have a surface area of about 3 km^2. During hours of intensified ablation these two lakes are able to retain during one day as much as 0.9 million m^3; this huge retention makes the water level in both lakes rise only some 20 to 30 cm. The effect of this storage of ablation water is, in the first place, a decreased amplitude of diurnal oscillations; further, these waters stored during the day in the lakes find their way into the river during the night. In order to illustrate the interdependence between the course of the water levels in the lakes I and II and the level in the Sandgigjukvisl River, the author prepared a diagram showing his observations made on August 5 and 6, 1968 (Fig. 17).

The influence of lakes upon the runoff system of the glacial waters is highly favourable. Not only do lakes fail to disturb the daily rhythm of oscillations of the water level; they rather mitigate this rhythm, and in this way

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they contribute to maintain some sort of equilibrium in surface runoff. Were the water, instead of being retained in the lakes, to reach the river directly, the daily water level oscillations in the Sandgigjukvisl River might reach figures as high as 200 cm.

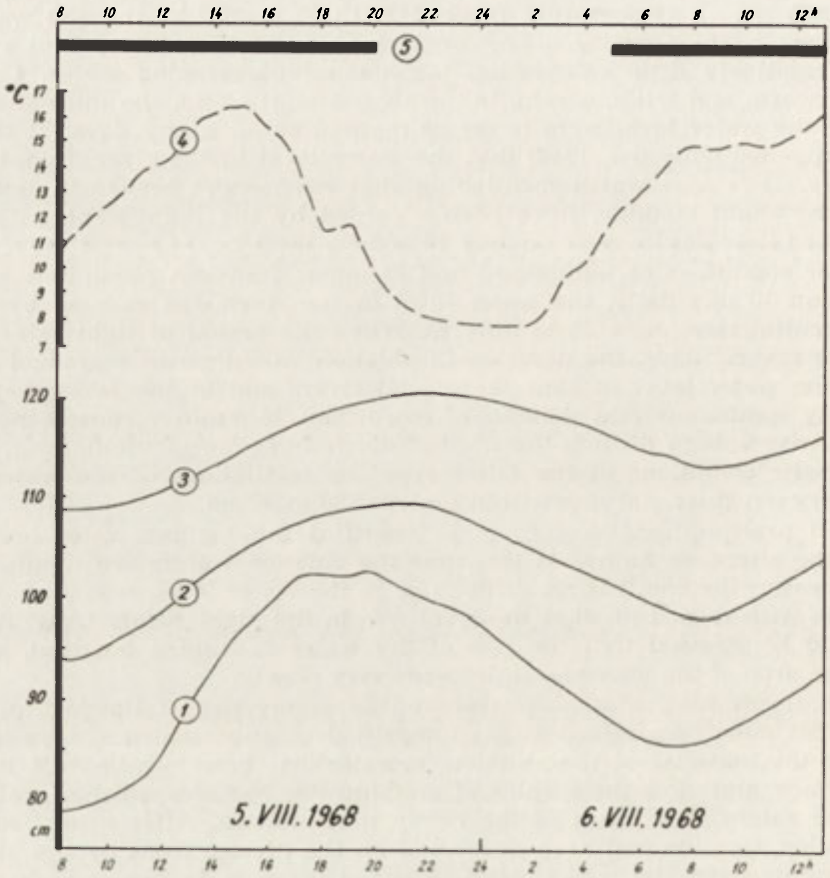

Fig. 17. Course of water-level fluctuations in the lakes and in the Sandgigjukvisl River as compared with air temperature and insolation changes
1 — course of water-level fluctuations in the Sandgigjukvisl River, 2 — course of water-level fluctuations in the first lake, 3 — course of water-level fluctuations in the second lake, 4 — air temperature in °C, 5 — insolation

Also of marked importance in the investigated area concerning runoff of the glacial waters is the groundwater which is held in the glacial and glaciofluvial material. Because these deposits consist of loosely spread glacial drift, the fluvial water can, while the high water level lasts, easily penetrate that part of the deposits which lies close to the river channel; afterwards, when the water level in the river drops, the water thus retained in the ground flows back into the river, forming numerous erosional gullies in the banks of the river channel (Fig. 18).

The rhythmic water level oscillations — though they are slight in the case of the Sandgigjukvisl River — bear considerably upon the structure of the

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river banks. With rising levels the water carries much more material and is more destructively active; hence processes of washing down and undercutting the river banks are of frequent occurrence.

The varying course of the water level described above for the Sandgigjukvisl River presents, to some extent, an average picture of what takes place in all the minor branch streams of which this river system consists. Considering, however, any single small rivulet of which several hundreds cover the Sandgigjukvisl drainage basin, one might assign any one of them to one of the different types of streams discussed above. On the whole it may be said that the diurnal rhythm of level oscillations described for the Sandgigjukvisl River applies more or less distinctly to all the distinguished types of rivers.

A remarkable problem met in the area under discussion is the thermal conditions of running water. So far this problem appears not to have been investigated in this region, although its solution seems to be important, if only

Fig. 18. Groundwater inflow into the Sandgigjukvisl Channel. Photo by Z. Churski

for establishing the ecological conditions under which vegetation develops in these streams. The author had to limit his thermal examinations to tests made only in selected periods of time; in spite of this, the data collected in the examinations of some of these streams throw light upon the general thermal conditions as they occur in running water.

Numerous measurements of the water temperature made at different hours of the day revealed that the water of springs of glacial waters and of melting dead ice has a temperature near 0°C. This determination assisted the author in establishing the source of numerous minor streams encountered in the end moraine zone. The greater the distance of some watercourse from its spring the higher usually is its temperature; it even happens that only a few bundred meters away from the point of its outflow, in the afternoon hours, water will show a temperature of 12–14°C.

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However, the rate at which the water temperature rises varies greatly; it depends upon the quantity of running water, on the topography of the channel, on flow velocity, and on the amount of suspensed material carried. Small streams widely spread over a large area warm up in a very short time and in the noon hours they easily reach 10 to 15°C, while great streams flowing rapidly grow warmer at a much slower rate; rarely their temperature rises as high as 10°C.

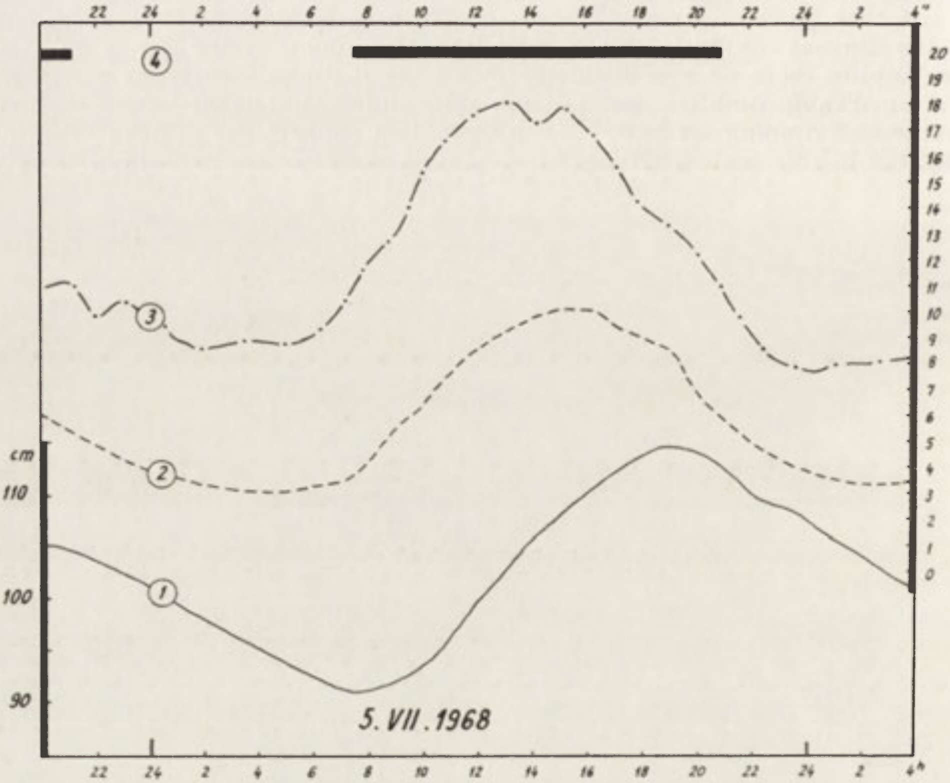

Fig. 19. Diurnal course of water temperature in River Sandgigjukvisl
1 — course of water-level fluctuations in the Sandgigjukvisl River, 2 — water temperature in °C, 3 — air temperature in °C, 5 — insolation

Irrespective of the thermal differentiation caused by flow volume and conditions of runoff, in all streams the water temperature also undergoes diurnal oscillations. For presenting evidence of this phenomenon the author prepared a diagram picturing the course of the water temperature during July 5, 1968 in the Sandgigjukvisl River near the limnograph station, in correlation with the course of air temperature, water level, and insolation (Fig. 19). This diagram discloses convincingly how much the water temperature depends on the temperature of the air; it also shows that the highest water temperatures follow highest air temperatures with a time lag of about 2 hours.

Somewhat different are the temperature conditions for streams fed from groundwater sources. Here the temperature values rather depend on the depth of the groundwater level from which the streams are fed. Groundwater reaching up to near the ground surface draws warmth from the overlying

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sandy deposits which on sunny days may absorb much heat, even up to 30°C. This shows that streams fed in this way may reach high temperatures even near their springs. On the other hand, streams fed from deeper-seated groundwater reveal at the places from which they issue rather low water temperatures, of the order of up to 4°C.

Also different are the thermal conditions in streams which arrive from lakes. The water temperature is to a high degree uniform, without the considerable changes observed in the water of streams derived from other sources. Hence the water temperature of streams flowing from lakes depends closely on the temperature of the lake water.

The above comment on the thermal properties of water determined in the streams of the Skeidarárjökull forefield calls attention to the wide thermal diversity of running water, brought about by differences in origin and in runoff conditions; it also throws light on the course of diurnal oscillations in water temperature. The data reported also indicate that in daytime some of the streams, especially the smaller ones, are apt to attain water temperatures higher than 10°C. These are small streams, especially those issuing from groundwater sources, which create conditions most favourable for the development of vegetation.

One of the foremost problems in glacial hydrology is to determine how much of the mass of escaping water is derived from ablation. For the area under discussion this was of particular importance, since at the same time measurements of ablation were made on the Skeidarárjökull (Wójcik 1970), and thus it became possible to compare the amount of water flow in the rivers with the decrease of the ice mass of the glacier. Even if these calculations are bound to be to some degree erroneous, their value with regard to understanding the regime of glacial rivers and understanding the ablation balance is undoubtedly past dispute.

For determining the quantity of meltwater escaping from the glacier, the author made hydrometric measurements in all rivers draining the Skeidarárjökull. Lacking more suitable instruments, flow rate measurements were made

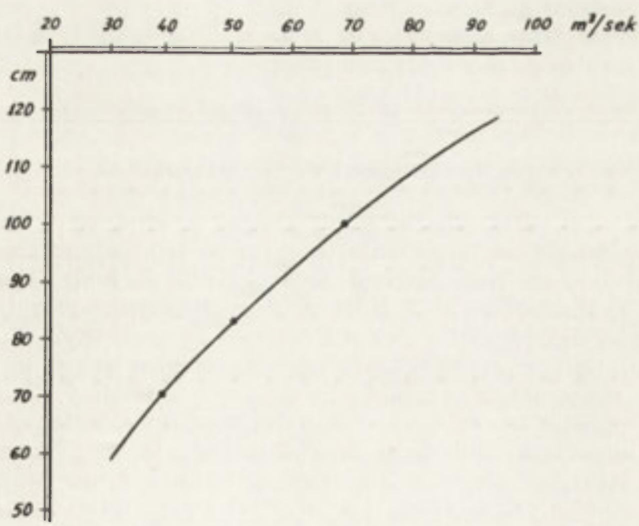

Fig. 20. Curve indicating discharge in the Sandgigjukvisl River

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in those river sections where it was possible to enter a river with rotary flow meters. On account of the very uneven floor of the river channels, the measurements in the channel cross-sections were made at 2 m intervals, and each time the measurements were repeated 3 to 5 times. This shows that the work was done as accurately as possible.

In the River Sandgigjukvisl measurements were made three times, at different water levels in the river (Fig. 20). Since within the drainage area of this river ablation measurements were at the same time made on the glacier snout (Wójcik), the equation of the water balance for this part of the drainage basin should be fairly accurate. The figures of the water balance for the whole Skeidarárjokull are certain to be less exact since in the Rivers Sula and Skeidará flow measurements were only made once, and because only approximate data were on hand as to the area covered by the water balance. In Table 1 are presented the results obtained from our hydrometric investigations; also added are certain factors referring to the Sandgigjukvisl water balance and to that of the Skeidarárjokull as a whole.

TABLE 1. Some data concerning the water balance of the Skeidararjokull in the period June 21 to August 23, 1968

Specification	Data
The area of Skeidararjokull	1726 km^2
The area of accumulating part of Skeidarárjokull	1165 km^2
The area of ablating part of Skeidarárjokull	561 km^2
Mean flow of the River Sandgigjukvisl June 21 to August 23, 1968 (mean water level — 88.7 cm)	56 m^3/s
Mean flow of the Sula River	48 m^3/s
Mean flow of the Skeidara River	164 m^3/s
Total flows at the mouth of the Sandgigjukvisl River. June 21 to August 23, 1968	309,657,600 m^3
Total flows at the mouth of the Sula River	265,420,800 m^3
Total flows at the mouth of the Skeidará River	906,854,400 m^3
Total surface runoff June 21 to August 23, 1968	1,481,932,800 m^3
Surface runoff on the ablating part of Skeidarárjokull	2641 mm
Total precipitation, June 21 to August 23, 1968	207 mm

STAGNANT WATERS (LAKES, ICE-DAMMED WATER BASINS)

The forefield of the Skeidarárjokull glacier is studded with fifty lakes of various sizes. It would be impossible to describe here all of the lakes observed; they are shown on the attached hydrographical map (Fig. 5). The author's detailed examinations enabled him to distinguish several types of lakes which differ genetically and show differences in their hydrological features.

The most numerous group embracing the largest lakes in this region is what is called marginal lakes, and among these one may distinguish lateral and frontal lakes. The lateral lakes lie along the lateral sides of the glacier, in between the glacier and the rocks confining the glacier. These lakes are fed by meltwater streams entering the lakes from numerous transverse glacier crevasses. Most often these lakes are drained by subglacial streams running along the lateral glacier margin. Due to the movement of the glacier and to changes in the way they are drained, these lakes undergo frequent transfor-

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mations. The water of the lateral lakes is relatively pure; its temperature is between 0 and 2°C. A picture of a lake of this type is shown on Fig. 21.

The most interesting, both as to hydrography and geomorphology, are those frontal lakes which lie next to the glacier snout. The author also paid special attention to them since traces of this type of lakes can also be seen in Northern Poland.

The forefield of the Skeidararjokull contains a dozen or so frontal lakes; the largest of them has an area of 1.6 km^2 and a depth of 14 m, while the smallest, of a few square m in extent, is only 1–2 m deep. All frontal lakes lie at sites where the escape of meltwater streams is or was, impeded by some obstruction. The origin of the basin containing a frontal lake is a complex problem. In his investigations of the marginal frontal lakes of the Hof-

Fig. 21. Lateral marginal lake near Sulutinder scarp. Photo by Z. Churski

felsjökull Jonsson (1955) arrived at the conclusion, that all larger frontal lakes have developed where there now is, or formerly was, a point of outflow of an englacial or subglacial river. In his opinion it was at such places that caves or deep hollows were formed in which water accumulated after the glacier had retreated.

More detailed examinations of these basins led to the assumption, that they might also have been the result of a number of other agencies such as: exaration, erosion by water flowing along the glacier margin, the breaking-off of larger ice blocks from the glacier snout, melting of ice buried below the glacier margin, or some blocking by ice-moraine walls of the water escaping from the glacier.

The origin of the frontal lake basins had to some degree its effect upon their shape and upon the morphology of the lake bottom. And this is why in this respect no uniform features can be observed. Most often the lake shape is oval. The lowest part of the bottom lies nearer the ice margin. But there

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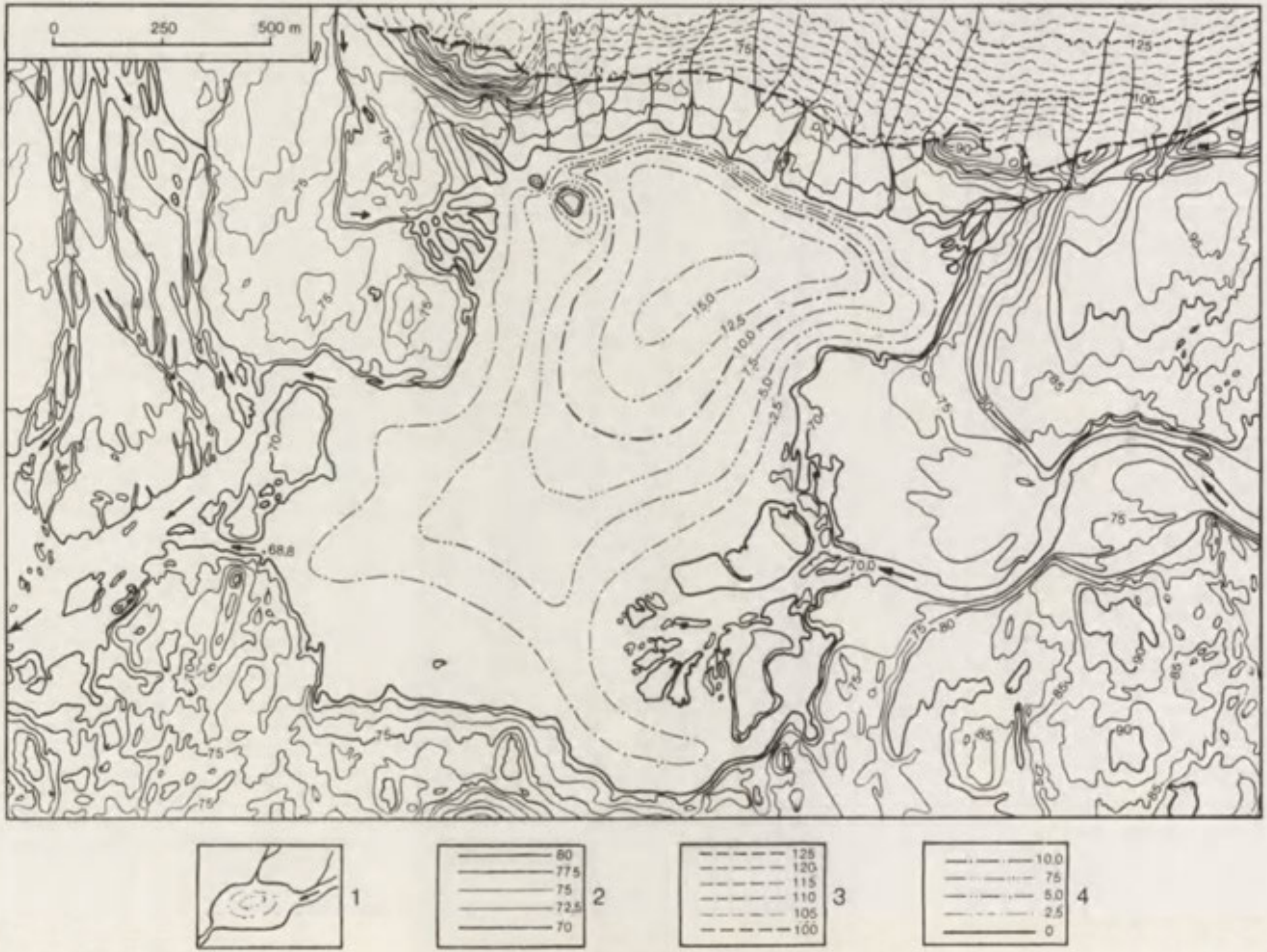

Fig. 22. Isobath map of Lake I
1 — river net, 2 — contour lines, 3 — surface glacier (number shows the height a.s.l. in metres), 4 — isobaths

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are many lakes where near the glacier margin ice-moraine ridges are submerged, the crests of which can at times be seen protruding from the water surface. The diagrammatic picture of the bottom of a lake of this type is shown in Fig. 22.

Frontal lakes receive their water directly from glacier streams and this explains the high degree of turbidity of the water in these lakes which is practically opaque. The surplus of the water accumulating in these lakes escapes by streams which run away from the lakes or which are eroded in the glacier margin. But there are also a number of frontal lakes with only underground drainage. Because these lakes are fed by glacier waters, their water levels follow the rhythm of diurnal oscillations. But due to the annual changes in the quantity of water feeding them, the frontal lakes are also subject to seasonal changes in their water levels.

The second type of lakes encountered in the forefield of the Skeidarárjökull are ground moraine lakes. They lie at a certain distance from the glacier margin, occupying either natural depressions in the moraines, or the basins of former frontal lakes. As a rule they are shallow. But the principal feature by which these lakes differ from those discussed before is that they obtain their water from a different source. No glacial waters reach them, and they are mainly fed from the groundwater. For the most part these lakes contain clear water, or only slightly turbid water due to wave action. They do not undergo any rhythmic oscillations, and no larger rock fragments occur in their bottom deposits. For all this, these lakes are subject to seasonal changes, and they show the effect of heavy rain periods. In their majority these lakes lack a surface outflow.

To the third group of lakes are assigned the smallest but, at the same time, the most numerous in this region: the kettle hole lakes. These lakes originate from melting dead ice blocks. As a rule these lakes are oval and their deepest part lies in the middle of the basin (Fig. 23). In the forefield of the Skeida-

Fig. 23. Lakes developed from meltwater kettles. Photo by Z. Churski

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rarjokull this type of lakes can be observed in various stages of development. During periods of intensive dead ice melting the surface of lakes of this type is apt to grow larger from day to day.

In the region under discussion, kettle hole lakes can be encountered everywhere: in moraine areas, in the outwash plains, in valley terraces and in the floors of present-day rivers. Often such lakes occur in groups of several or of dozens, near each other.

For the most part the kettle hole lakes are fed by groundwater; but depending on where they lie, river water may also find its way into them. When receiving only groundwater, these lakes are perfectly transparent; in their bottoms lies the identical material which forms the ground surface, and nowhere have been observed any of the fine-grained bottom deposits so characteristic of small lakes of different origin. No diurnal oscillations of the water level occur, but noticeable changes follow those in the groundwater table, as well as changes caused by the effect of heavy rainfall. Similar to the lakes discussed earlier, are also felt seasonal changes in the water level. In view of the transparency of the water, during summertime organic life thrives in this type of lakes.

The above characteristic shows that for subdividing stagnant waters no uniform criterion has been applied. Principally taken into consideration has been the diversity of the regime, and those features have been emphasized by which the distinguished types of stagnant waters clearly differ. However, this division has left no room for the stagnant waters which, being very small, rather resemble waterlogged regions and probably are kettle hole lakes in the making. These land forms are individually so tiny that their occurrence cannot possibly be indicated on hydrographic maps (Fig. 24). For the most part such waterlogged regions are fed by groundwater or precipitation. The water held in the many small hollows is very clear and is an environment favourable to vegetation.

It commonly occurs that cliffs, bank-near shoals, shore ridges and lacustrine terraces start to form in all the lakes mentioned, directly after they come into existence. These forms develop in a relatively short time due to wave action, to surface currents caused by winds, or by fluctuations of the water level. In particular lakes these forms may be seen in different stages of development, and this gives the opportunity of observing successive stages in their growth. On the whole, bank-near shoals in the larger lakes show a clearly marked abrasive platform-terrace and a sandy shelf, and the banks of such lakes vividly resemble the banks of glacial lakes seen in Poland (Fig. 25).

An intersting topic with regard to limnology and ecology are the thermal conditions of stagnant waters. In this respect stagnant waters show diversities even greater than those encountered in running waters. In particular lakes the water temperature depends mainly on the size of the lake, on its source of water, and on atmospheric conditions, and it is these three agencies which largely cause temperature differences in the individual basins. In summertime the diurnal oscillations of the air temperature have the effect that the water temperature in the lakes suffers diurnal changes similar to those in rivers.

From very many measurements which it would be impossible to enumerate in detail, one can arrive at a number of conclusions with regard to the thermal conditions in the investigated area. The first conclusion is that the lake water attains high temperatures only on warm and sunny days. Water heating is quickest in small and shallow lakes, lacking surface outflow and showing a high transparency. In August, around noon, the water of this kind

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Fig. 24. Initial stage of formation of the lake basin from meltwater kettles. Photo by Z. Churski

Fig. 25. View of the bank of Lake I. Photo by Z. Churski

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of lakes reaches temperatures from 14 to 18°C; but, under adverse weather conditions these lakes rapidly cool. However, if such small lakes are in contact with dead ice or if they are fed by water from melting ice blocks, their temperature is fairly constant and in the summer months it is 4–5°C no matter what time of day. In larger lakes without surface runoff, water heating proceeds at a slower rate and on sunny days their water shows in the hours from 3 to 6 p.m. temperatures up to 12°C.

Frontal lakes fed by glacial water usually show highly diversified temperatures. Part of the lake directly adjoining the glacier has a temperature of 2–4°C, but at the opposite shores the temperature may be as high as 7–9°C. Even small frontal lakes containing many floating ice fragments show on sunny days temperatures of only 3–5°C. On cloudy, rainy days the temperature of lake water is more uniform and depends on the air temperature: on the whole, on such days the lake water is several degrees cooler than the air, oscillating between 3 and 9°C.

To illustrate the stratigraphy of lake water, the author made a number of thermal cross-sections across the deepest and largest frontal lake; two of these cross-sections are shown in Fig. 26. The data obtained in this way reveal a very interesting thermal condition, typical of near-glacier lakes. Two mas-

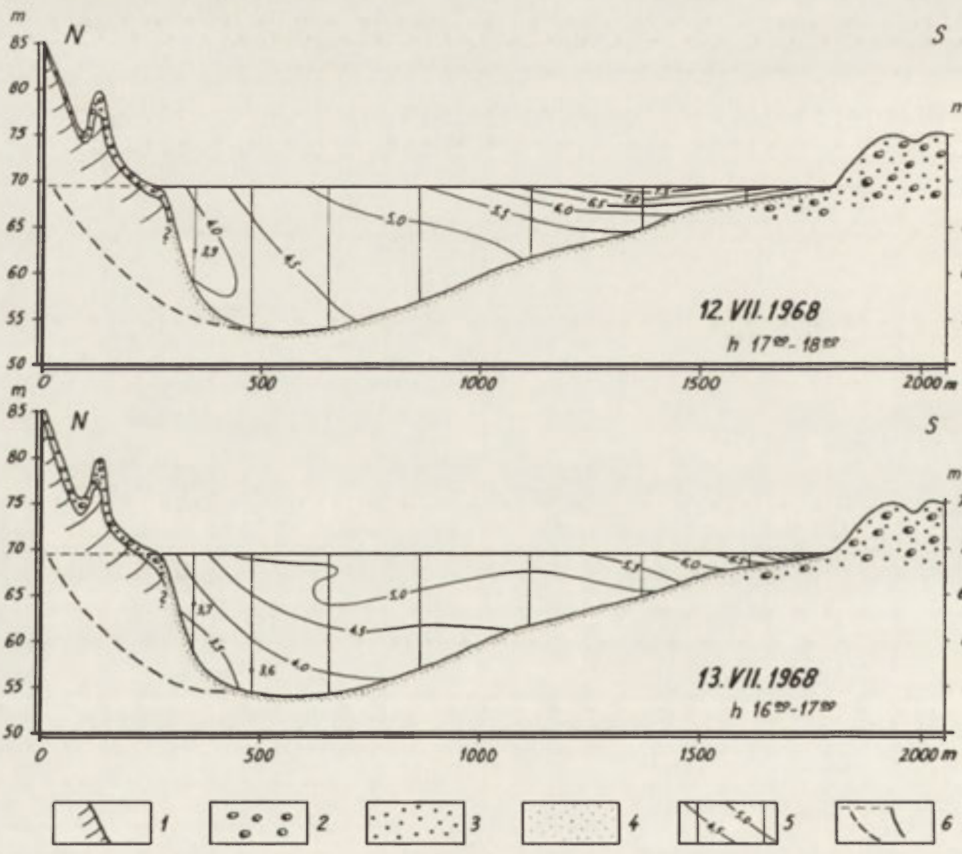

Fig. 26. Thermal cross-sections across Lake I

1 — glacier, 2 — gravel, 3 — sand, 4 — mud, 5 — water temperature, 6 — probable limit of the lake in 1960

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ses of water can be distinguished, markedly differing in temperature. In the northern, deeper part of the lake which adjoins the ice and receives numerous brooks escaping from the glacier, the water body has a temperature of 4–5°C. In the southern, more shallow part of the lake, the water temperature varies between 7 and 9°C.

These temperature differences between the northern and the southern part of the lake are so distinct that the same features appear even in the river branches leaving these two parts of the lake. At the time the author made his measurements, the river leaving the northern part had a temperature of about 4°C, while it was 7°C in the southern branch. Frontal lakes of smaller size with ice blocks floating on their surface show a more uniform temperature throughout the water body. But let us always keep in mind that all the data given here refer to the summer season in which our observations were made.

The diurnal temperature oscillations of lake water equal those recorded for water flowing in streams. Here our frequently repeated observations dis-

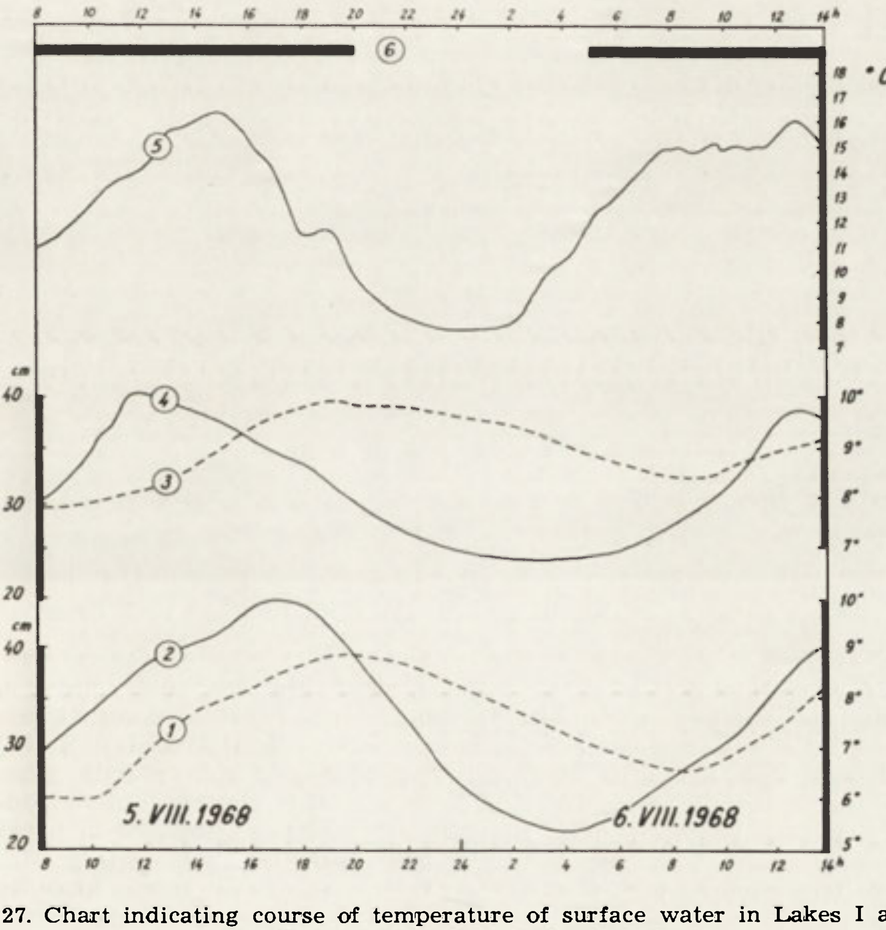

Fig. 27. Chart indicating course of temperature of surface water in Lakes I and II
1 — course of water-level fluctuations in Lake I,
2 — course of water temperature in Lake I,
3 — course of water-level fluctuations in Lake II,
4 — course of water-temperature in Lake II,
5 — course of air temperature, 6 — insolation

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closed that temperature amplitudes are higher in small and shallow lakes, and that at night the water temperature may drop to 4°C. On the other hand, larger lakes retain warmer temperatures for longer periods, the more so since usually fewer glacial waters pass into these lakes.

To gain a clear picture of these conditions the author made on August 5 and 6, 1968 for 24 hours a series of continuous control measurements in the lakes marked I and II. In Fig. 27 the results of these tests have been shown. Studying the curves of water temperature one notes that the highest values occur in the time from 2 to 4 p.m., that is before the highest water level starts to subside. The lowest temperatures are recorded for 2 to 4 a.m. However, it must be added that the above temperature changes refer to fair and sunny days, as shown by the curves for air temperature and insolation in Fig. 27. When the day is cloudy and rainy, the diurnal changes in the water temperature are barely 1 to 2°C.

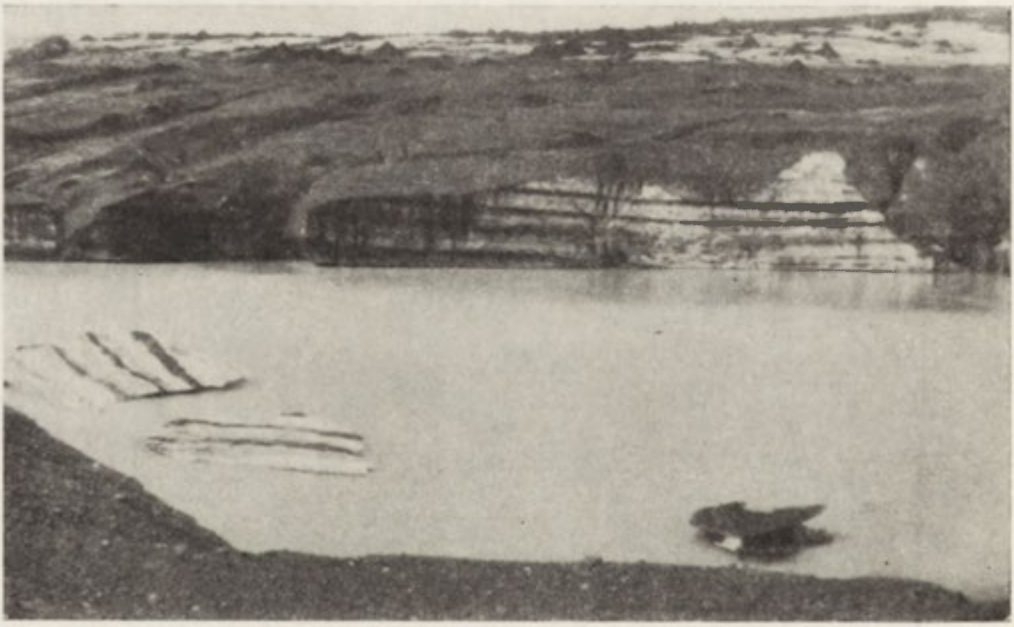

Fig. 28. Ice blocks floating on surface of frontal lake. Photo by Z. Churski

A further limnological problem of interest in the area under discussion is the study of the bottom deposits developing in the frontal lakes. Greatest is the amount of this type of deposits accumulated by glacial streams in those lakes which carry as much as 10–15 g/l suspended material, mainly silts and clays. But in some of these lakes coarser material is also encountered, such as sand, and even boulders which must have either been brought in by glacial rivers or melted out from ice blocks floating on the lake surface (Fig. 28).

The lake deposits lack floral or faunal remnants, but minor quantities of tree pollen and other plant pollen have been observed. In a single water sample taken from the depth of 10 m in Lake I, were found the following pollen: 14 of birch (*Betula*), 9 of pine (*Pinus*), one each of alder (*Alnus*), goosefoot plants (*Chenopodiaceae*) and of *Cyperaceae*, and two of unidentified trees.

One may assume that these pollens have been brought in from far away.

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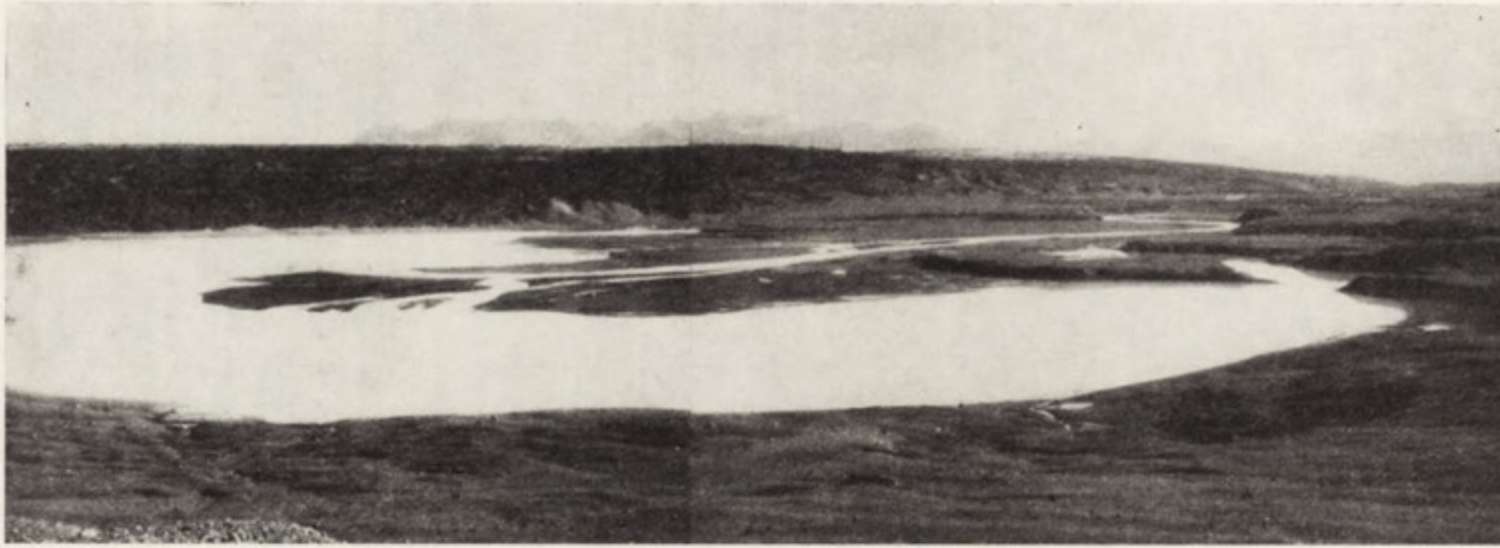

Fig. 29. Delta under development in Lake I. Photo by Z. Churski

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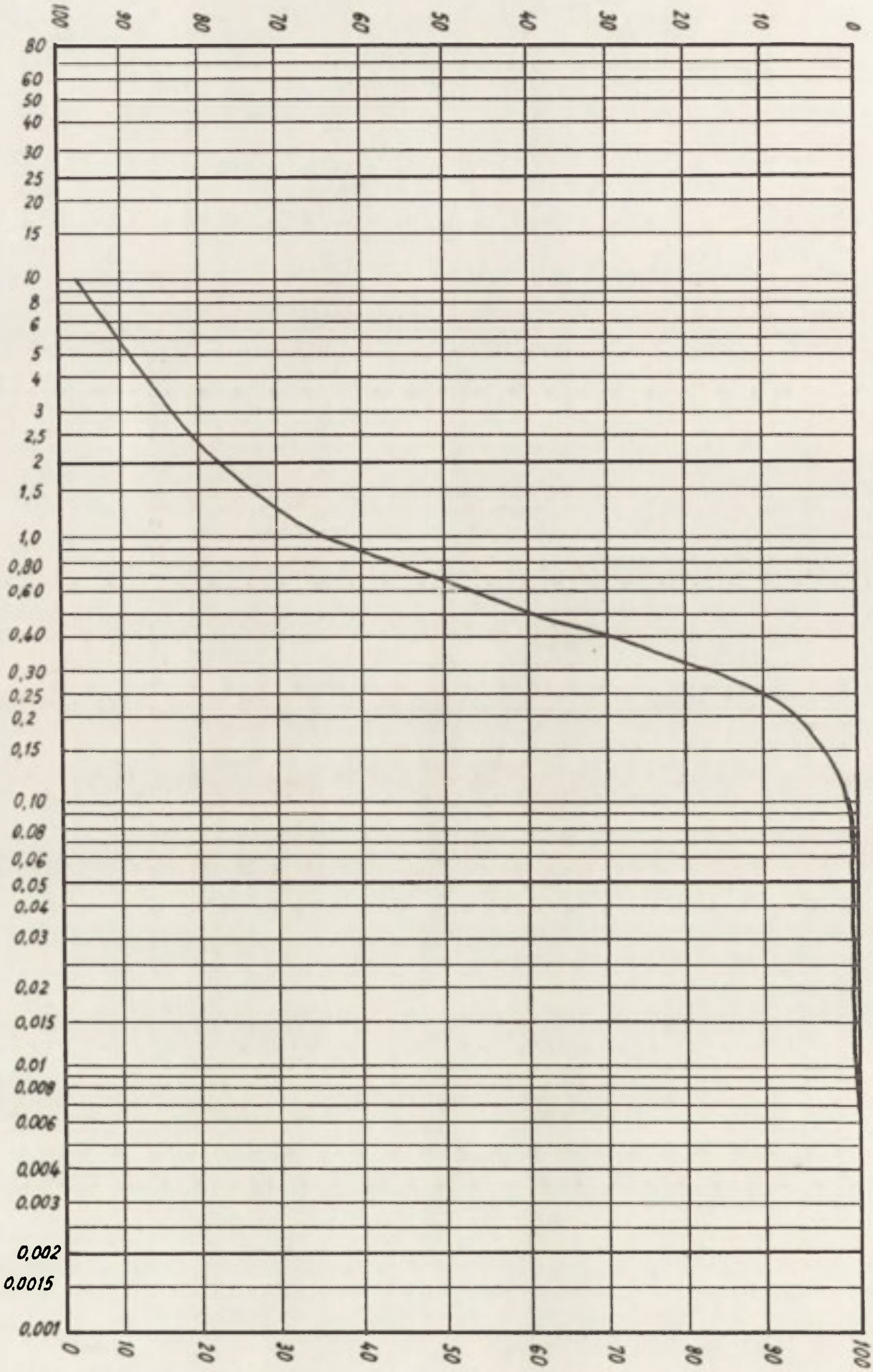

Fig. 30. Grain size curve of deltaic deposits

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In view of the fact that Lake I is a very young lake, only some ten years old, the number of wind-borne pollens found must be considered as high.

In the frontal lakes deposits accumulate at a rapid rate, but not always do these deposits remain unaffected after the lakes vanish. Frequently the lake bottoms are filled-in by rivers forming enormous deltas which completely overspread the deposits of former lakes (Fig. 29). The delta desposits mantling the old lakes consist for the most part of well rounded basalt sands (Fig. 30). Some of these deltas grow very quickly; for instance, the delta shown in Fig. 29 was formed within a period of only three years.

Sometimes a complete filling-in of lakes occurs by deposits of glacial waters, and in cases like this geological studies are the only means of determining where a once ice-dammed lake has been. This same situation is also encountered in reconstructing former Pleistocene lakes in Northern Poland, where the frequently observed ice-dammed sediments are often overlain by fluvial deposits.

In the Skeidararjokull forefield lakes often disappear within a short time — in much the same way as must have taken place when the Scandinavian inland ice was retreating. Former lakes are very numerous in the region under discussion, as shown by the great number of lacustrine deposits and dry lake hollows with well preserved outlines of former lake shores. The reasons why lakes have vanished differ. As mentioned before, a general tendency towards lake infilling can be seen here. Many of the lakes disappear because the glacier margin retreats and decreases in altitude. Lakes dependent on groundwater disappear when the groundwater table subsides. Many other lakes are drained by encroaching river channels. However, alongside this large-scale disappearance of lakes, frequently one observes the formation of new lakes. Thus, here we have an area characterized by continuous changes in hydrographic features.

GROUNDWATER

In the forefield of the Skeidararjokull the glacial material consists of loose deposits, mainly of sands, gravels and boulders. This composition is favourable to the storage of large quantities of water as well as to easy dislocation of stored groundwater.

Because the accumulation of glacial deposits lacks intercalated continuous strata of impermeable material, this area contains practically only one water--bearing horizon. The only kind of deposits which might disjoin this horizon, are dead ice lobes buried in the ground, or layers of lacustrine deposits extending within the sand beds. However, this type of impermeable sediments occur merely here and there, and they are unable to divide the water-bearing horizon into several partitions; only at isolated places may this sort of impermeable beds form a layer capable of retaining overlying lenses of water.

The groundwater of the Skeidararjokull forefield is fed from a variety of sources. In its nearest vicinity to the glacier margin it is directly fed by meltwater streams. In the end moraine zone the water-bearing horizon is additionally supplied with considerable quantities of water from precipitation and from melting dead ice blocks. The groundwater held in outwash sheets, situated a number of kilometers from the glacier front, obtains its supply for the most part from the rivers passing these sheets and from precipitation. But a certain amount of water may also reach this groundwater from the marginal zone of the glacier, as seems to be indicated by the general southward trend in the gradient of the groundwater table.

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The height of the groundwater table depends mainly on the altitude of the margin of the glacier snout and on the water level in the rivers. The groundwater table is highest near the glacier margin. With increasing distance from the glacier snout the groundwater table subsides. It is only in water lenses held back by dead ice lobes or by clayey lacustrine deposits that water may occur slightly higher than the basal level of the groundwater.

Next to the glacier, the groundwater lies usually at a shallow depth, within 0 to 2 m below the ground surface. It is at a greater depth only when it is underneath isolated mounds or massive moraine walls. However, it is exactly in this zone that there are apt to occur isolated groundwater lenticles upheld by impermeable lacustrine deposits or dead ice lobes. This explains the occurrence of the numerous though tiny lakelets situated at about 100 m a.s.l., thus at about 15 m above the mean water level in the rivers.

In the wide outwash sheets extending south from the end moraine zone, the groundwater table lies nearer the ground surface with increasing distance from the end moraines. Next to the moraines the groundwater table is probably 10 to 15 m below ground surface, but some 8 km farther south this depth is only 1 to 2 m; farther south, the groundwater reaches or even covers the ground. On account of changing water levels in the rivers it often happens that temporarily the groundwater table shows an east-west or a west-east inclination.

Fig. 31. Vegetation growth stimulated by high groundwater table. Photo by Z. Churski

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The discussed variability of the groundwater table has a pronounced influence upon the development of vegetation on the outwash plains. Where the groundwater lies deeper than 1.5 m below ground surface only that kind of vegetation can grow on the surface which is fed by rainwater; for the most part this is small, widely scattered plant growth, because fair weather in this region causes the surface deposits to be easily displaced and to be subject to aeolian processes. Where the groundwater lies higher, especially in hollows and depressions of the outwash sheet, larger plants like reeds and rushes appear, with their roots extending down into the groundwater (Fig. 31). These relatively high plants occurring mostly in clusters, retain the sand particles

Fig. 32. Outwash sheet waterlogged due to high groundwater table. Photo by Z. Churski

carried by the wind over the ground surface, and in this way they often start the development of larger aeolian landforms. On lower ground where the groundwater table lies at less than 20 cm below the surface, no vegetation can develop, and isolated depressions contain only bare and waterlogged patches of ground (Fig. 32).

In the area we are dealing with which is so very diversified in its hydrology, the oscillations of the groundwater table vary considerably. They depend upon changes in the water level in the rivers and upon precipitation. Diurnal oscillations occur only near the glacier and in the vicinity of the larger rivers; they show high amplitudes. Farther away from the rivers, no diurnal oscillations are observed in the groundwater table; but seasonal changes do occur, in conformity to water quantity and climatic conditions.

In order to determine the oscillations of the groundwater table of the area between the Sula and Sandgigjukvisl Rivers, the author installed nine piezometers from which the level of the water table and the groundwater temperature were systematically recorded. The location of these nine stations, as

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well as the measurements of the water level were controlled by geodetic instruments, with T. Konysz M.Sc. acting as instrument operator.

From the compiled results of these measurements of the groundwater table (Fig. 33) it appears that in the outwash sheet between Sula and Sandgigjukvisl the oscillations of the groundwater table closely depend on precipitation, and that the most sensitive in this respect is groundwater at 0.5 to 1.0 m depth. It further came to light from the diagrams mentioned (Fig. 33) that on the whole the groundwater oscillations are unimportant, being of the order of about 10 cm. Similar results in this were obtained by Hjulström (1955), who observed groundwater oscillations in the area of Hoffellssandur; but this author ascribed these changes rather to changes in the water level of nearby rivers.

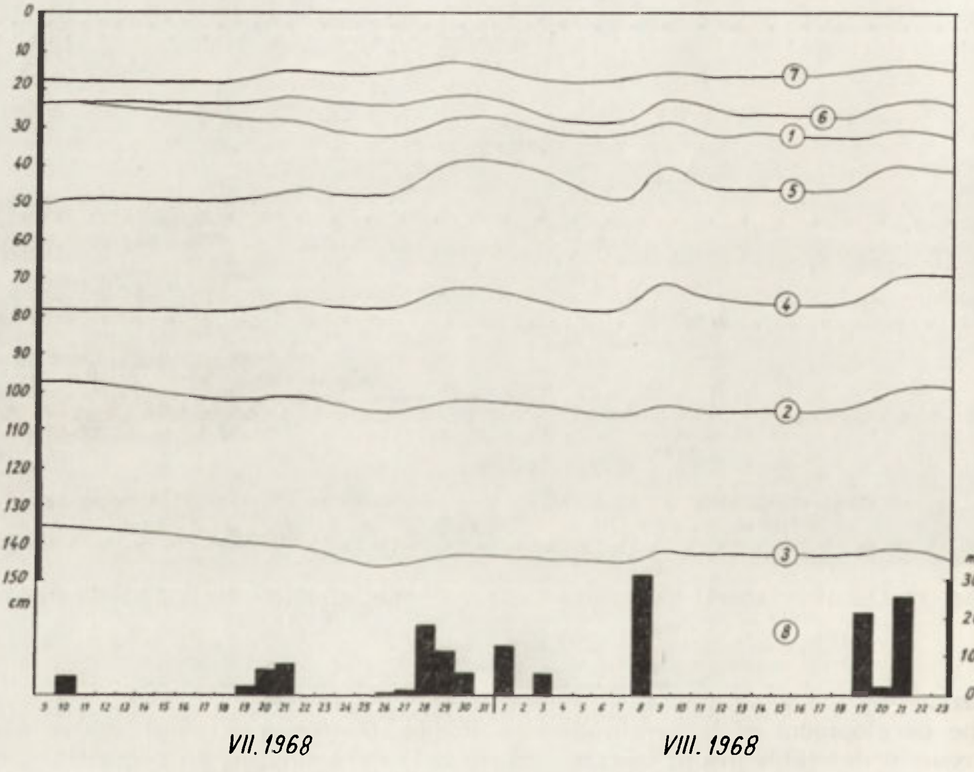

Fig. 33. Oscillations of groundwater table in outwash sheet
1–7 — Groundwater oscillations in groundwater measuring station in cm,
8 — precipitation in mm

By accurate measurements of the water level in test pits the author ascertained that in the area under investigation the groundwater table is inclined toward the Sandgigjukvisl River. This means that in the whole part of the outwash between Sula and Sandgigjukvisl the groundwater is fed by Sula waters which travel underground in an eastward direction. This has also been confirmed by numerous groundwater springs observed at the base of the Sandgigjukvisl channel.

With regard to the vegetation which develops near places where groundwater observations were made, temperature measurements of the groundwater

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were also carried out. They revealed an interdependence between the temperature and the depth at which the groundwater lies; they also registered the water temperature changes occasioned by atmospheric conditions.

The observations illustrated in Fig. 34 show that the highest temperatures are reached by groundwater whose level extends near the ground surface, and that for this water the amplitudes of temperature oscillations are highest. The lower the groundwater table, the cooler will be the water and the narrower will be the amplitudes of temperature oscillations. An exception proved to be well No. 2 in which at 1 m depth the water was cooler than it was at 1.5 m depth. This abnormal fact may be explained by the probability that well No. 2 was sunk into a former river channel in which the groundwater flow has been more intensive (see the annex).

In our tests the highest water temperatures of +18°C were determined in shallow wells with a water table 0.50 m below ground surface. But temperatures as high as this occur only on fair and sunny day about midday. On such days the outwash sand will be warmed up to 28–29°C. But at night the water of such shallow wells drops to 10–12°C.

During cloudy and rainy periods, the water temperature in these shallow wells was also 10–12°C. In deeper wells the temperatures are lower but, at the same time, more uniform. And this explains why vegetation develops most lavishly in that part of the outwash plain where the groundwater is relatively warm, without greater diurnal oscillations of its water table.

THE EVOLUTION OF THE FLUVIAL SYSTEM ON THE PROGLACIAL AREA OF THE SKEIDARÁRJÖKULL

In the same way as this is taking place near the margins of modern and former glaciers, the fluvial system observed in the forefield of the Skeidararjokull undergoes continuous and rapid changes. When reflecting upon the relief evolution of a given region it is of great importance to understand these changes, because in the initial period of the formation of a glacial landscape running water plays an essential part. The causes of changes in the fluvial system of the area under investigation are very complicated. On the one hand these changes are determined by oscillations of the glacier snout and the amount of meltwater runoff and, on the other, they are dependent on processes of erosion and accumulation caused by the rivers and by dead ice melting.

The most successful method to throw light upon this problem is the interpretation of aerial photographs taken at different periods of glacier standstill. Admittedly, field examinations would also permit the reconstruction of the location of former lakes and rivers, but determining the evolution of a river system in successive years merely on the basis of field studies would be extremely difficult.

To illustrate the evolution of the fluvial system the author made use, apart from his field examinations, of certain cartographic material, especially the aerial photographs taken in 1946, 1960 and 1965, and the photogrammetric survey made by T. Konysz M.Sc. in 1968. The hydrographic maps compiled on the basis of all this source material are shown in Figs. 35, 36, 37 and 5. For all this, it must be admitted that from the data gained from the aerial photographs not all stages of the changing fluvial system can be reconstructed; but, at any rate, this material makes it possible to clearly understand the changes which have taken place, including their rate and scope.

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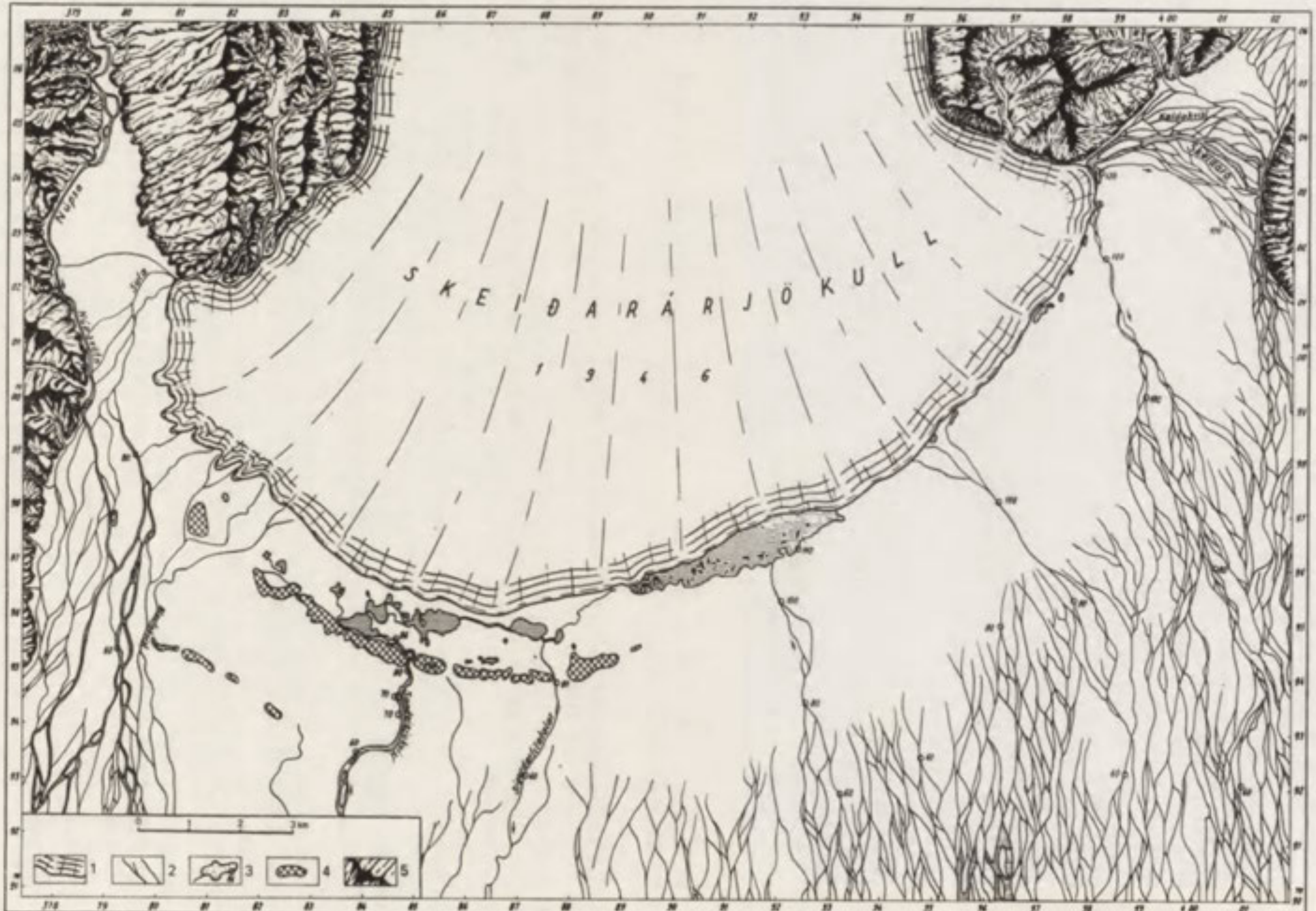

Fig. 35. Hydrographic map from 1946
1 — glacier, 2 — rivers, 3 — lakes, 4 — end moraines, 5 — basalt massifs

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Fig. 36. Hydrographic map from 1960

1 — glacier, 2 — rivers in the area of acting sandy tracks, 3 — lakes, 4 — traces of larger river humid beds, 5 — traces of larger river dry beds, 6 — end moraines, 7 — basalt massifs

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Fig. 37. Hydrographic map from 1965
1 — glacier, 2 — rivers in the area of acting sandy tracks, 3 — lakes, 4 — traces of larger river humid beds, 5 — traces of larger river dry beds, 6 — end moraines, 7 — basalt massifs

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Based on evidence from field studies and on examinations of the aerial photographs it can be asserted that the principal factors affecting the evolution of the fluvial system have been changes in the situation of the glacier snout and, particularly, changes in the altitude of the glacier base from which the meltwater escapes from the glacier.

In scrutinizing the successive stages of the glacier retreat one notes that every change in the position of the glacier was accompanied by a lowering of the base of the glacier snout. This in turn had a variety of consequences: river channel incision, subsidence of groundwater table, lake draining, retreat of lakes following the receding glacier, changes in runoff system along glacier margin. And because each successive glacier standstill brought a different altitude of the glacier base, the direction of water escape along the glacier margin suffered continuous changes.

All the changes occasioned by the glacier retreat are usually greatest in closest vicinity to the glacier margin. In the area farther away from the glacier front the changes of the fluvial system are less marked; for the most part their causes are changes in the groundwater table and dead ice melting.

Pictures of the fluvial system as it was in 1946, 1965 and 1968 are shown in the attached maps. In 1946 the base of the glacier snout had an altitude much higher than today. In spite of this, the runoff of the glacial waters met obstructions and for this reason extensive ice-dammed lakes were formed between the glacier margin and the end moraine zone. During summer the overflow from these lakes went off by several streams which carved their way across the end moraine zone, taking advantage of some of the then existing glacier gates.

Lake II was much more extensive than it is now and its altitude was some 20 m higher. In the lateral lakes the water level was also higher, as indicated by traces of their former extent in the form of well preserved cliffs and sandbanks near the lake shores. On the other hand, both the lateral Sula and Skeidará Rivers had more or less the same shape as in 1968.

In 1960 the retreating glacier halted at a distance of about 1.5 km from its 1946 position. The base of the glacier snout was lower than it had been in 1946; as a result those lakes which had not been filled in, were either drained or they lost their water due to the subsidence of the groundwater table. Following the glacier retreat, Lake II gained an eastward outlet for its waters and its water table dropped some 15 m or so. All the successive evolutionary stages of this lake are recorded in its numerous terraces (Fig. 38). But in front of the glacier snout a new frontal lake came into existence.

Five years later, in 1965, the fluvial system had not undergone changes as great as in the 1946–1960 period (Figs. 36 and 37). However, a close study of the 1965 air photo shows a number of fundamental changes. First of all, in the eastern and central part of the area under discussion the glacier, far from retreating, had moved forward some 50 m and had occupied some of the northern part of Lake I. The smaller ice-dammed lake which in 1960 had been situated between Lakes I and II has been filled in. Also changed in their courses appeared a number of minor brooks escaping directly from under the glacier snout. Finally, some further changes have taken place as a result of dead ice melting.

By 1968 the glacier had again retreated some 50 metres, and further changes had occurred in its forefield. The fluvial system as it was in 1968 is shown in the hydrographic map (Fig. 5), and its principal elements have been discussed earlier.

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All the above facts indicate that in near vicinity to the glacier the fluvial system suffers considerable transformations. But it must be stressed that these changes affect only the situation and the dimensions of particular elements of this system. And irrespective of where in the discussed area the glacier snout happens to have come to rest, the genetic types of rivers and lakes distinguished before will maintain their character.

Fig. 38. Lake terraces in scarps of Lake II. Photo by Z. Churski

CONCLUSIONS

Our hydrographic studies of the Skeidarárjökull forefield, whose results have partially been reported in the present paper, lead us to a number of conclusions; the most important among them will now be recalled.

From our work in Iceland we consider one of the most important parts to be the compilation of the hydrographic map of a part of the Skeidarárjökull forefield; in preparing this map we applied the directives issued in Poland for compiling Poland's hydrographic map.

The description of the types of rivers and lakes distinguished in Iceland, and the definition of the part they had in shaping the relief of this glacial region may prove to be of great importance for all further palaeohydrographic and geomorphological research in Northern Poland.

From a hydrological point of view the understanding gained of the regime and dynamics of the rivers draining the Skeidarárjökull glacier, especially the Sandgigjukvisl River, was extremely valuable. One can well imagine that the rivers escaping in Northern Poland from the waning Pleistocene glaciers must have behaved in a similar way because, as the Sandgigjukvisl, they were fed by glacial waters and by precipitation, and because the lakes in the glacier forefield and the groundwater accumulated in the glacial and glaciofluvial

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deposits must also have contributed to equalizing the diurnal oscillations of the runoff.

Our groundwater studies have disclosed that on outwash sheets which are always very porous and mostly lack vegetation, precipitation is readily absorbed into the ground and causes changes in the level of the groundwater table.

From our observations of thermal conditions we arrived at the conclusion that in summertime small and shallow water basins situated near the glacier might reach water temperatures up to 20°C, and that in both rivers and lakes the water temperature is subject to diurnal oscillations similar to those of the air temperature.

The fluvial system of the Skeidararjokull forefield undergoes rapid transformations. These mainly affect the position and the dimensions of particular elements of the fluvial system. And the genetic types of the rivers and lakes discussed may be expected to continue occurring in this region, until the time arrives when the glacier vanishes completely and the regime of the fluvial system changes radically.

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REFERENCES

Ahlmann, H. W.: son, 1940, The relative influence of precipitation and temperature of glacier regime, *Geogr. Ann.*, 22.

Ahlmann, H. W: son, Thorarinsson, S., 1943, *Vatnajokull. Scientific results of the Swedish-Icelandic Investigations 1936/37/38*, Chapter I–XI. Reprinted from Geografiska Annaler 1937–1940, 1943, Stockholm.

Arnborg, L., 1955, Hydrology of the glacial river Austurfljot, *Geogr. Ann.*, 37, 3–4.

Arnborg, L., 1955a, Ice-marginal lakes at Hoffellsjokull, *Geogr. Ann.*, 37, 3–4.

Arnborg, L., 1957, Erosion forms and processes on the bottom of the river Angermanalven, *Geogr. Ann.*, 39.

Ashwell, I. Y., 1966, The pleistocene history of some rivers in Western Iceland, *Geogr. Ann.*, 48 A.

Bogacki, M., 1970, Eolian processes on the forefield of the Skeidararjokull (Iceland), *Bull. Acad. Pol. Sci., Ser. Sci. geol.-geogr.*, 18, 4.

Churski, Z., 1970, Results of observation of water-level fluctuations in the Sandgigjukvisl River (Iceland) in Summer 1968, *Bull. Acad. Pol. Sci., Ser. Sci. geol.-geogr.*, 18, 4.

Davidov, L. K., Pronin, A. T., 1967, Vodnyy ryezhim ryek vytekayushchikh iz lednikov Fedchenko i Zerevshanskogo i ikh verkhnikh pritokov (Water regime of the rivers flowing out of the Fedchenko and Zerevshan glaciers and of their upper tributaries), in: *Krupneyshye ledniki Srednyey Aziy*, Leningrad.

Eythorsson, J., 1953, French-Icelandic expedition to Vatnajokull, 1951, *Polar Rec.*, 45, 6.

Galon, R., 1970, Mechanism and stages of retreating of the Skeidararjokull (Vatnajokull, Iceland), *Bull. Acad. Pol. Sci., Ser. Sci. geol.-geogr.*, 18, 4.

Galon, R., Szupryczyński, J., 1970, Wyprawa Polskiego Towarzystwa Geograficznego na Islandię (Sum.: The Iceland expedition of the Polish Geographical Society), *Czas. Geogr.*, 41, 1.

Hjulstrom, F., 1952, The geomorphology of the alluvial outwash plains (Sandur) of Iceland and the mechanics of braided rivers, in: *Proc. XVIIth Congress IGU*, Washington.

http://rcin.org.pl

Hjulström, F., 1955, The groundwater, *Geogr. Ann.*, 37, 3–4.
Howarth, P. J. I., Price, R. J., 1969, The proglacial lakes of Breidamerkurjökull and Fjallsjökull, Iceland, *Geogr. J.*, 135, 4.
Jónsson, J., 1955, On the formation of frontal glacial lakes, *Geogr. Ann.*, 37, 3–4.
Jónsson, J., 1957, Notes on change of sea-level on Iceland, *Geogr. Ann.*, 39.
Keller, R., 1962, *Gewässer und Wasserhaushalt des Festlandes*, Leipzig.
Konkina, N. G., Pronin, A. T., 1960, Vodnyy ryezhim ryeki Seldary (Water regime of the River Seldara), in: *Gidrologicheskiye issledovaniya Pamirskoy Ekspeditsyi L.G.U. po programie Miezhd. Geof. Goda 1957–58*, Izd. Leningr. Univ.
Krigström, A., 1962, Geomorphological studies of sandur plains and their braided rivers in Iceland, *Geogr. Ann.*, 44.
Østrem, G., Brigge, C. W., Rannie, W. F., 1967, Glacio-hydrology, discharge and sediments transport in the Decade Glacier area. Baffin Iceland N.W.T., *Geogr. Ann.*, 49 A.
Parde, M., 1960, Les facteurs des regimes fluviaux, *Norois*, 27.
Rist, S., 1955, Skeidarárhlaup 1954, *Serprentun úr Jökli*, 5.
Rist, S., 1956, Islenzk vötn, Reykjavik.
Rist, S., 1968, Jöklabreytingar 1965/66, 1966/67, og 1967/68, *Serprentun ur Jökli*, 18.
Rist, S., 1961, Rannsóknir á Vatnajökli 1960, *Jökull*, 11.
Serebriannyy, L. R., 1969, *Islandiya* (Iceland), Moskva.
Stenborg, T., 1970, Studies of the hydrological characteristics of glaciers, *Acta Univ. Upsal.*
Sundborg, A., 1956, The river Klarälven. A study of fluvial processes, *Geogr. Ann.*, 38.
Szupryczyński, J., Kozarski, S., 1970, Relief of the marginal zone of Sidujökull (Iceland), *Bull. Acad. Pol. Sci., Ser. Sci. geol.-geogr.*, 18, 4.
Thorarinsson, S., 1939, The ice-dammed lakes of Iceland with particular reference to their values as indicators of glacier oscillations, *Geogr. Ann.*, 21.
Thorarinsson, S., 1953, Some new aspects of the Grimsvotn problem, *J. Glaciol.*, 2.
Thorarinsson, S., 1954, Athuganir a Skeidarárhlaupi of Grimsvotn 1954, *Jökull*, 4.
Thorarinsson, S., Rist, S., 1955, Skaftarhlaup in September 1955, *Jökull*, 5.
Thorarinsson, S., 1967, Forvitnilegir jökullgardar vid jadar Skeidarárjökulls, *Jökull*, 17.
Tryggvason, E., 1960, Earthquakes, jökullhlaups and subglacial eruptions, *Jökull*, 10.
Wójcik, G., 1970, Ablational processes on the Skeidarárjökull (Iceland), *Bull. Acad. Pol. Sci., Ser. Sci. geol.-geogr.*, 18, 4.

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GLACIAL FORMS AND DEPOSITS IN THE SIDUJÖKULL DEGLACIATION AREA

Stefan Kozarski and Jan Szupryczynski

INTRODUCTION

Sidujökull is one of the three outlet glaciers of the south-west part of Vatnajokull (Fig. 1). They occur between Kerlingar and Hágöngur. The other two are Skaftárjökull and Tungnaárjökull (Fig. 2). The names Skaftárjökull (Skaptárjökull) and Sidujökull were in the past used interchangeably to denote the large south-west lobe of Vatnajökull, or they were used to refer to various neighbouring lobes, with Wadell (1920) and Nusser (1935) using the name Sidujökull to denote the small lobe between Graenalón and Jökullklettur. The lobe was still unmarked on Thoroddsen's map (1906). It should be deduced that different names for the outlet glaciers were connected with changes in the position of their margins, and what follows, in the distinctness of the glaciers themselves. In the present article we are going to use the names for the outlet glaciers as they were used by Thorarinsson (1964).

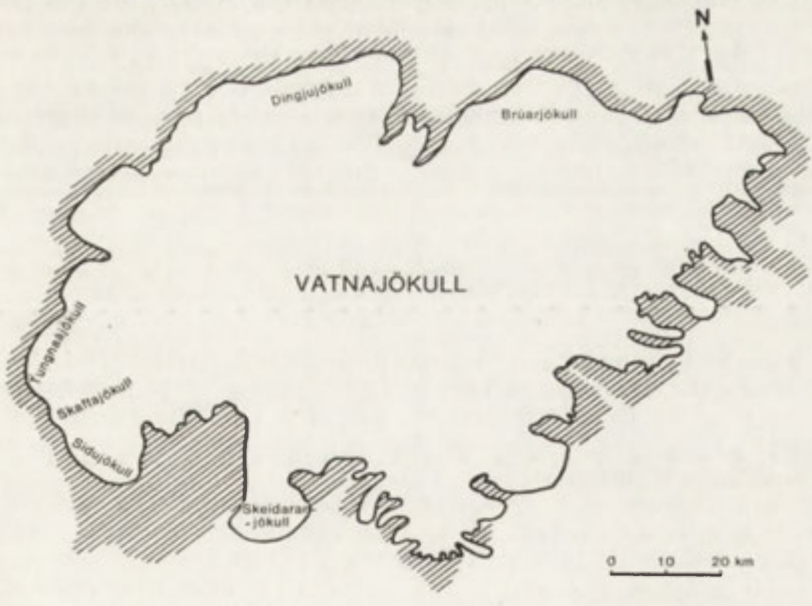

Fig. 1. Location of the investigated area

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He means by Sidujokull the clear-outlined lobe, also very well seen on the U.S. Air Force air photographs taken in 1960, which extends between Hagongur and Trollhamar. Both in the east, near Hagongur, as well as where the famous Laki volcanic line occurs (Trollhamar) the south-west part of Vatnajokull has deep embayments which set off Sidujokull from the neighbouring Skaftarjokull and from the small lobe between Hagongur and Graenalon. Besides its actual outline on air photographs, Sidujokull can be marked off as an ice stream with independent motion, which was testified by Thorarinsson (1964), who reports two sudden advances. The first occurred in 1945, the other in 1963–1964. Both advances had similar maximum range about 0.5 to 0.6 km.

Sidujokull, and the glacial deposits and forms in its recessional area has not been, so far, an object of special studies. In the description of Iceland ice caps Thoroddsen (1906) does not write anything on Sidujokull in the recent spatial meaning of this name. His report (Thoroddsen 1905, p. 188–189) is to the effect that he reached the foreland of Tungnaarjokull and Skaftarjokull, to the Hverfisfljot springs. During his journey he also visited Graenalon and Skeidararjokull, that is those parts of Vatnajokull that are east of Sidujokull. The expeditions that went to Vatnajokull to carry out general, glaciological or volcanological observations willingly used the route to Vatnajokull in the Djupa River valley and via Sidujokull. This was done by the Wadell (1920), Austrian (Nusser 1935, Thorarinsson 1964) and Danish (Nielsen 1937) expeditions. Reports from those expeditions, however, give only marginal, mostly glaciological, information about Sidujokull.

Lack of any information on the relief and glacial deposits in the foreland of Sidujokull is reflected in the 1:250,000 geological map complied by Kjartansson in 1962. The whole area beyond the zones of braided rivers is treated as a domain of lava occurrence. Two belts of end moraines only were distinguished by Thorarinsson (1964) on the basis of reconnaissance flights. The outward belt of end moraines marks, according to the author, the position of Sidujokull in historical times, while the inward one marks the position of the glacier after the 1945 transgression (Thorarinsson 1964, p. 84).

An analysis of air photographs that was made a year before our expedition allowed us to find a rich inventory of glacial forms and deposits in the foreland of Sidujokull. They became, together with the processes in the frontal zone, the object of our investigations in the 1968 summer season (June 20 to August 21).

GENERAL REMARKS ON THE AREA INVESTIGATED

The length of the glacier is about 38 km. Its front ends at about 700 m a.s.l., and the firn field extends up to 1500 m a.s.l. The ice front forms a lobe over 30 km long (Fig. 2). Our investigations included only the south-east part of the foreland 13 km long, because it was impossible to cross the Djupa river in the east and the tributary of Hverfisfljot in the west.

To the west Sidujokull borders on Skaftarjokull, and to the east on Skeidararjokull. It is difficult to mark off exact borders between them. South of the ice front occurs a basalt plateau 1000 m a.s.l. (Kalfafellsendi). The plateau is built of older grey basalts formed in late Tertiary and Pleistocene. Close to the ice front, in its eastern part, are younger basalts from Pleistocene. In the western part of the foreland there is a sheet of a Pleistocene palagonite for-

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mation and basalt lava from the 1783 eruption (Kjartansson 1962). In the central part of the foreland volcanic calderas occur that were probably active in historical times.

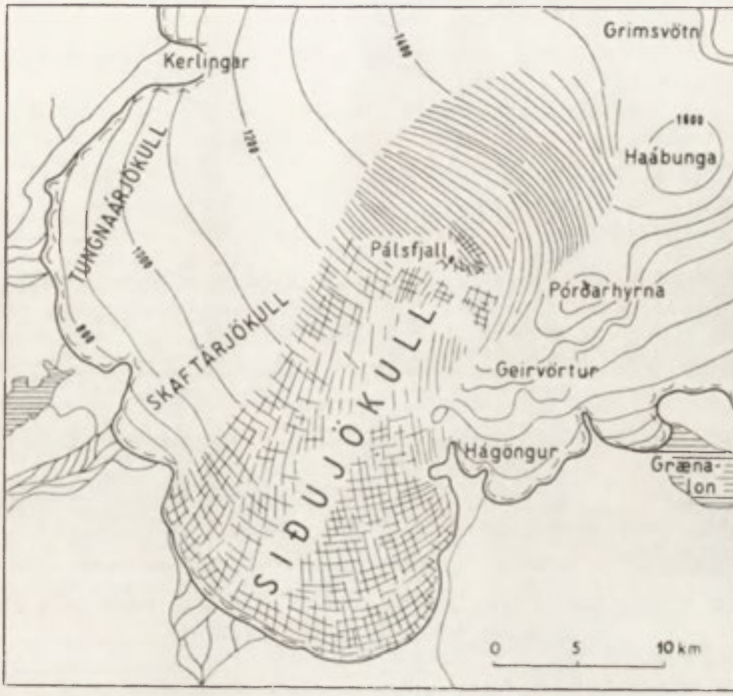

Fig. 2. Sketch map of Sidujökull 1963–1964. Based on aerial photographs — S. Thorarinsson 1964

GEOMORPHIC AND SEDIMENTARY EFFECTS OF GLACIATION AND DEGLACIATION PROCESSES FROM THE DECLINE OF THE 19TH CENTURY TILL 1968

ADVANCE OF THE GLACIER AT THE DECLINE OF THE 19TH CENTURY AND THE OUTERMOST END MORAINE BELT

The outermost belt of end moraines extends across the Djupá valley to enter the Brunná River valley at the foot of the basalt plateau. The moraines of this zone in the Djupá valley are about 2.5 km from the present ice front. The maximum range of this zone in the Brunná valley is marked off further south at a distance of about 5 km from the ice front. Unfortunately, we do not have any air photographs of this area and therefore we could not delimit its range cartographically in the drawn map (Fig. 3; see the annex).

End moraines of the 1st zone have the form or ridges with relative height up to 10 m. Distinctly elongated forms of the ridges follow more or less the shape line of the basalt plateau edge. The plateau barred any southward transgression of the glacier, while the Brunná valley made a southward transgression possible by a narrow ice lobe.

The morainal ridges are mostly made up of scoracious basalt lava. Apart from lava debris in the moraines and nearby, we also come across slabs of

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Fig. 4. Plates of grey basalt lava put up vertically as the result of a glacier transgression. Photo by J. Szupryczyński

Fig. 5. Plates of grey basalt lava put up vertically as the result of a glacier transgression from the year of about 1890. Photo by J. Szupryczyński

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Fig. 6. Push moraines from 1890 on Sidujokull forefield. Photo by J. Szupryczynski

Fig. 7. Fine eolian material on the surface of push end moraines from the year of 1890. Photo by J. Szupryczyński

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this lava pushed up from former nearly horizontal position. Some slabs are even upright (Figs. 4 and 5). In one of ridges in the Djupa valley we found a imbricate-like arrangement of slabs inclined proximally, that is northward (Fig. 6). Pushing up and putting up of the slabs originated, in our opinion, under the impact of bulldozing of the transgressing glacier. Apart from lava debris and slabs on the top of the moraines, there are also small amounts of outwash material, waste and fine eolian material (Fig. 7).

In the vicinity of the moraines we have a bare bedrock in the form of covers of scoracious, horizontal or nearly horizontal lava. The bedrock was bared as a result of glacier erosion and erosive activity of fluvioglacial waters. Activity of those waters forms in effect also small outwash cones that occur in the foreland of the end moraines. They are short cones built exclusively of scoracious basalt lava cobbles, gravels and sands. Their bases join with the foot of the end moraines or the proglacial valleys outlets which break through the end moraines. The outwash zone interdigitates with the plateau-des-

Fig. 8. Overlapping inflow of the cones descending from the plateau slopes (left) with outwash cones from the year of about 1890. Photo by J. Szupryczyński

cinding cones along the plateau slopes. This fact is very well underlined by the petrographic differences between the cones and outwashes. The first are built mainly of grey basalt debris, the second exclusively of dark scoracious basalt lava (Fig. 8). In the cones area the basalt bedrock crops up in some places over surfaces of the cones, which is marked in enriching the cones with very angular material.

A large number of closed hollows occur in the eastern part among the moraines of the first belt. In the distal part of the moraines are short proglacial walleys. In the central part of this zone an old marginal valley was preserved

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which occurs directly at the rear of the extreme belt of moraines or in the foreland of this belt. The valley was an outflow way of meltwater from this zone into the Brunná valley. In the western part the end moraines are directly adjacent to the marginal valley which has here the form of a canyon.

At the rear of the extreme belt are numerous short end moraine belts and outwash cones bound up with them. The belts sometimes take lobate outlines. All the observations made over the 1st zone lead us to the following conclusions:

(1) The end moraine complex of this zone was formed as a result of advances of the Sidu glacier onto its foreland which was covered with horizontally or nearly horizontally lying scoracious basalt lava.

(2) During the transgression lava was eroded by the glacier, hence numerous smoothed roche mountonnée surfaces, and thrusted, which found its expression in putting the lava slabs to upright position in the maximum range of moraines in this zone.

(3) During the maximum expansion, meltwater in this zone flowed in the central part in the marginal valleys eastward to the Djúpá and westward to the Brunná valleys. At the places where the ice front was in the open valleys of the Djúpá and the Brunná, meltwater flowed away centrifugally.

(4) A striking feature of the end moraines and outwashes in this zone is their petrographic homogeneity. They are built exclusively of scoracious basalt lava. In a sense, perhaps, the homogeneity could be explained by the fact that the moraines formed during the transgression in consequence of the exaration-bared bedrock accumulation. But it is hard to assume that only this glacier ice which contained chunks and crumbs of scoracious lava advanced. The question, then, must still remain open.

(5) The complex of short end moraine belts in the rear of the extreme position line of this zone testify the oscillating nature of the glacier recession.

(6) The outermost belt of end moraines has never been identified until now. On the basis of detailed field cartographic work, the belt could be mapped for the first time. In our opinion, the end moraines of this zone mark off the extreme position of Sidujökull in historical times, whereas Thorarinsson (1964) identifies, on the basis of air photographs, the glacier extreme position in historical times with the 2nd belt of end morain according to our nomenclature.

According to the information gained from Björn Stefánsson, one of the oldest farmers in Kålfafell, the outermost end moraines could be dated back to about 1820. The most outermost end moraines on the foreland of the neighbouring Skaftárjökull are dated, according to Thoroddsen's observations (1906), from 1889–1890, Thoroddsen reached the foreland of Skaftárjökull in 1889, and the ice front at that time was directly adjacent to the end moraines. In 1889–1890, most Iceland ice caps reached their maximum extent (Thorarinsson 1969). These years saw in Iceland the greatest transgression of ice caps in historical times. In some cases transgression was very sudden, for example in case of Bruárjökull and Eyjabakkajökull. Brúarjökull advanced as many as 10 km from 1889 till the spring of 1890, pushing in many places the old moraines formed during its 1810 transgression (Thorarinsson 1938). Eyjabakkajökull advanced several kilometres and formed end moraines pushed to 25 m in height, together with numerous fold structures. The Sidujökull transgression was presumably, also sudden. This could explain, we suppose, pulling out from the bedrock and putting upright the scoracious lava slabs and homogeneity of moraines of this belt.

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END MORAINES AND ICE-CORED MORAINES FORMED DURING GLACIER RECESSION IN THE 20TH CENTURY

The next two belts of end moraines are connected with the glacier recession, and run at a distance of about 1.2 km parallel to the ice front (in 1968). The distance between the 2nd and the 3rd belt ranges between 50 and 350 m. The moraines within the second belt are distinctly shaped but are small in size. Their mean relative height is only 2.5–3.5 m. The highest hillocks reach 5.5 m (Figs. 9, 10 and 11). The lowest relative height occurs in the eastern part, where it reaches only 1–1.5 m at places. Slopes of these moraines are gentle and do not generally exceed 20°. The moraines are rounded and gentle-sloped probably due to the melting out of the ice buried in the debris. There are numerous traces of subsidence on their slopes.

The moraines are mostly built of grey basalt debris (Fig. 9). Only in the middle of this zone could we find fragments of scoracious lava which occurs in the immediate foreland of the moraine that also builds the bedrock. In some sections, scoracious lava constitutes quite heavy admixture in the morainal debris. There is a considerable amount of stratified fluvioglacial deposits in the moraine-forming material.

The whole belt of moraines lacks any larger outcrops. The forms are probably accumulation moraines, which is evidenced by their small height and internal structure. Parallel to the 2nd belt runs the very distinctly shaped 3rd belt of end and ice-cored moraines. Ridges of moraines in this belt reach a considerably greater size, and relative height of individual forms reaches 4–5 m in the most eastern part. The highest are of 8 to 10 m. Only in short sections small forms with relative height 1–2 m are met with. The moraines within the third belt show far more fresh relief than those in the belt mentioned above. Their slopes exceed 20° and often 40°. Their crests are very sharp; at some places they are not more than 20 cm in width. Their summits are shaped in the form of sharp pyramids typical for the relief of ice-cored moraines. The moraines here show geological structure similar to those of the second belt (Fig. 10). No larger exposures were found among the moraines from the Djúpá river (Fig. 3) to as far as the volcanic calderas. But sharp, pyramidal summits and numerous tension cracks on their slopes, slumps and small hollows speak for the presence of glacial ice in the interior of some forms.

In the rear of the 3rd belt we often find zones of low hillocks with geological structure identical to that of the main belt moraines (Fig. 12). Among the hillocks often occur hollows which do not have any outlet and which are often filled with seasonal lakes (Fig. 13). In comparison to the rear of the 2nd belt, the presence of a great number of closed depressions and ponds is a characteristic feature for the proximal side of this belt.

The belt finds quite a different morphological expression west of the volcanic calderas. Here, the belt is formed into a massif consisting of a complex of ice-cored moraines. The width of the whole massif reaches 300 m. The relative height of individual moraines reaches 30 m in relation to the outwash plains in the foreland, and 50–80 m in relation to the ice-dammed lake extending in the rear.

Within the massif the summits of individual moraines are pyramid-like and slope values exceed 30°. Numerous, large outcrops allowed us to establish that the interior of any moraine contains dead glacial ice with very well preserved foliation and shear planes. Glacial ice is covered with a thin mantle of debris of a 1 m thickness. As a result of the melting out of buried glacial ice, various slumps, cracks and landslides of different size develop on the

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Fig. 9. Accumulation moraines within the limits of the second marginal zone of Sidujokull, July 1968. Photo by J. Szupryczynski

Fig. 10. Accumulation moraines of the second marginal zone of Sidujokull, July 1968. Photo by J. Szupryczynski

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Fig. 11. Accumulation moraines of the third Sidujokull marginal zone, July 1968. Photo by J. Szupryczynski

Fig. 12. Ice-cored moraine relief within the limits of the third zone W of the volcanic calderas, August 1968. Photo by J. Szupryczynski

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Fig. 13. A small lake at the background of the third moraine zone. Photo by J. Szu-pryczynski

Fig. 14. A small lake at the background of the third moraine zone. Photo by J. Szu-pryczynski

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slopes. Also numerous quite fresh kettle holes are significant, filled sometimes with water (Fig. 14). Ice-cored moraines in this place allow for the conclusion that it was here that the glacier frontal zone, at least 300 m wide, stagnated here during recession.

The above-mentioned belts of end and ice-cored moraines developed c. 1920 and/or 1940 according to the information given by Björn Stefánsson. It is from these years that he remembers the margin position. It is worth of note that Stefánsson's estimate of the age of moraines coincides with Thorarinsson's views on the age of the 2nd belt of end moraines. They both assume that the belt formed in the 1940s. Thorarinsson's photograph does not include the areas where we noted the first belt of the pushed moraines which date back from *ca.* 1890 (Fig. 15).

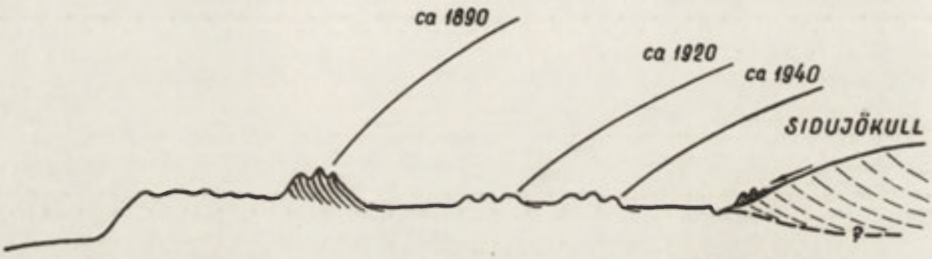

Fig. 15. Scheme of the Sidujokull recession

The neighbouring Skaftárjökull has, according to the investigations by Hauk Thómasson and Elsu G. Vilmundórtáttur in 1965 and 1966, another two distinct moraine zones apart from the 1889–1890 moraines. But those authors description includes neither the topography nor geological structure of these forms on account of their preoccupation with the discovery of hydrographic changes in Langisjór Lake. The younger moraine belts in Skaftárjökull foreland developed, in our opinion, at the same time as the recessional moraine belts in Sidujökull foreland did. Thus the deglaciation process was similar in these neighbouring areas.

GLACIER DRAINAGE SYSTEMS DURING RECESSION

The drainage system during the Sidujökull recession period between 1890 and about 1940 can be reconstructed on the basis of outwash forms and deposits and extremely numerous proglacial valleys that occur within all three moraine belts described above.

The foreland of the 1st belt of end moraines retained short outwash cones built of basalt lava debris. Similar to the moraines of this belt, also the cones reveal homogeneous geological structure: they contain only fine, crumbled, scoracious lava debris. Their geological structure and dark colour distinguish them from the sand forms formed in the later phases of deglaciation. During the maximal glacier advance about 1890, melt water flowed away from the moraines lying along the basalt plateau within the marginal valleys which have survived in the foreland of the moraines or in their rear until now. It flowed westward to the Brunná and eastward to the Djúpá valley. In these wide valleys water flowed centrifugally southward. Similar drainage system developed probably also during the formation of the moraines pushed in the

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rear of the outermost belt. Fragments of old outwash forms are still preserved in this belt. We may suppose that profuse outflow of melt water in later phases of deglaciation destroyed part of the pushed moraines and outwash patches in the rear of the outermost belt (Fig. 3). We did not find any major exposures in the outwash deposits bound up with the 1st belt. Most probably the thickness of stratified drift is small.

The foreland of the 2nd belt retained large outwash cones built by melt water that flowed for about those 20 years when the glacier was stagnant here. The cones were later on built up due to accumulation of sands and gravels transported here by melt water when the glacier was stagnant at the 3rd belt. Particularly well developed were the outwash forms on the Djúpá river and west of the volcanic calderas (Fig. 3). In the vicinity of the calderas is emplaced a whole system of interdigitating outwash cones which were formed by melt water when the glacier was stagnant at the line of the 2nd and 3rd belt. Extremely numerous kettle holes appear in the proximal part of the cones. In effect, there is a pitted surface at the foot of the ice-cored moraine massif.

Down the river that flows out of the ice-dammed lake in the rear of the ice-cored moraines occur 4 distinct outwash levels separated from one another by sharp erosion edges over 2–3 m in altitude. Thickness of outwash deposits here is over 10 m.

There are numerous proglacial valleys within the 2nd and 3rd belt. The map of Sidujökull marginal zone marks the occurrence of morainal forms by means of short sections of continuous line. All the breaks in end moraines of the 2nd and 3rd belt are the places where proglacial valleys occur. We distinguished two types of them:

(1) *Hanging proglacial valleys*. These are short and dry valleys which terminate in the north with a distinct step and generally with a short sand cone in the south. They are from several to tens of metres long. They developed when the glacier front was immediately adjacent to the end moraines. Melt water which flowed down the front dissected the end-moraine ridges and formed short valleys within them. The front surface was covered, just as it is today, with ablation moraine. Melt water came down off the glacier surface washing away the ablation moraine cover. The sands and gravels from the destroyed ablation moraine were accumulated at the mouth of proglacial valleys and formed short outwash cones. The excavation made within the proximal part of a short cone revealed the following deposit sequence:

(a) stone and gravel debris 10–30 cm in thickness,
(b) sand and gravel series 1.0 m in thickness,
(c) series of diverse-grained sands with gravel inserts 1.0 m in thickness,
(d) an insert of ash-grey loam 3–4 cm in thickness,
(e) series of diverse-graines grey-brown sands 30 cm in thickness.

Individual layers directly point out to the character of ablation process on the glacier front. Coarse sands and gravels were transported away from ablation moraine and accumulated in the cone during intensive ablation. The layer of silt developed during weak ablation, when the water discharge was very small. The thickness of deposits in those short outwash cones is generally small. It reaches 2–5 m in the proximal part. Due to the ablation, the glacier front was lowered and its margin retreated from the end moraines. Proglacial valleys lost their contact with the front and became dry.

The hanging valleys with a step at the glacier side and widening at their mouth or terminating with a short outwash cone were described from many

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end moraine zones of present-day and Pleistocene glaciation (i.a., Gripp 1929, Klimaszewski 1960, Troll 1957, Todtmann 1960).

(2) *Proglacial gaps.* Two subtypes were distinguished here:

(a) those related to the younger phase of the glacier recession, now dry (Fig. 16) which carried away melt water through the 2nd belt when the glacier was stagnant at the 3rd belt of end moraines,

(b) active valleys filled at present with melt water streams flowing from the glacier front through the 2nd and 3rd belt of end moraines.

Both dry and active valleys are from several to tens of metres wide and have steep slopes. On some of the slopes intensive denudational processes are active. Bottoms of dry valleys are built of stone (Figs. 13, 14), which speaks for an intensive flow of melt water.

The melt water outflow system at the time of glacier stagnation at the 2nd and 3rd belt of end moraines was similar to the recent hydrographic network. Melt water flowed westward and eastward, to the Brunná and the Djúpá respectively. Where the calderas are, water flowed in a deep rocky canyon

Fig. 16. A proglacial gap within end moraines of the third zone. Photo by J. Szupryczyński

between the calderas to the Brunná, while part of it flowed westward to the tributary of Hverfisfljot. Shallow stretches of marginal valleys were preserved along the moraines immediately in their foreland or rear. Some of them are dry and dead now, while others joined the present melt water streams network.

FLUTED MORAINE

Fluted moraine in Sidujökull foreland occurs between the third belt of end moraines and ice-cored moraines formed at the actual glacier front. The fluted

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moraine on the air photographs taken by U.S. Air Force in July and August 1960 reveals a far greater expansion than the one we established during the 1968 investigations. The air photographs (Fig. 17) record the existence of fluted moraine in the rear of the 3rd belt of end moraines from the calderas in the west and to the Djupa in the east, whereas in 1968 the well preserved fluted moraine occupied only the central part of the investigated area (Fig. 3). It was destroyed elsewhere by melt water and by rain water. Sidujokull foreland is in an area which receives 3000 mm of precipitation annually. That is why the influence of water coming from precipitation is so considerable in remodelling the foreland topography.

The fluted moraine has the form of parallel ridges with furrows between them (Fig. 18). They all run perpendicular to the glacier front, and their morphological axis is oriented from N 190° to N 200°. The ridges range from several cm to 0.3 m in altitude. Locally old ridges on the moraine surface appears merely in the form of a stripe of stones. The width of a ridge or a stripe ranges from 0.5–4.5 m. Individual ridges and furrows are from tens to 150 m long. The longest ridge (stripe of stones) reaches 340 m. The ridges

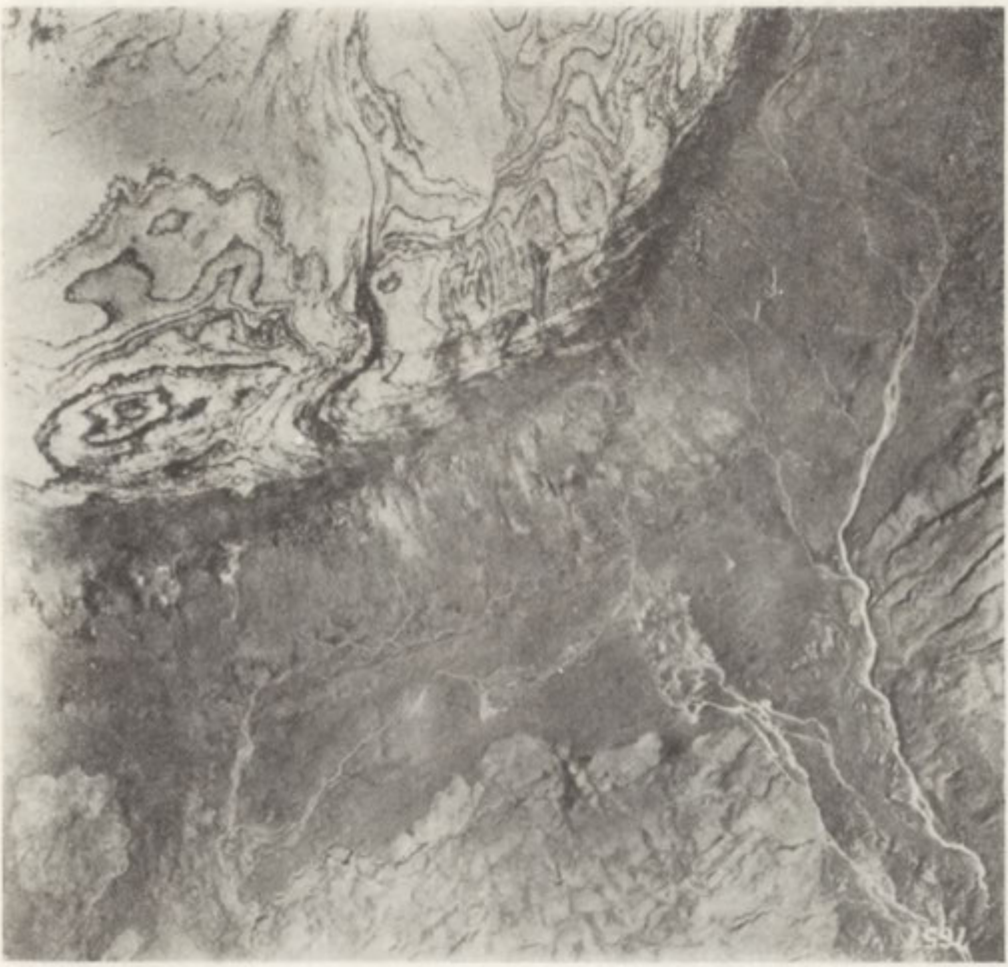

Fig. 17. Fluted moraine at the Sidujokull (air photograph taken in 1960)

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and furrows begin at the distal slopes of ice-cored moraines and run southward. The closer to the glacier front the more distinct the ridges are. Geometrical dimensions of ridges and furrows within the fluted moraine approximate the ones described so far in the literature on the subject (Baranowski 1970, Dyson 1952, Hoppe and Schytt 1953, Schytt 1959, Szupryczyński 1963, Todtmann 1960). The largest ridges within a fluted moraine were reported from Iceland by Todtmann (1960) from Brúarjökull foreland in the north-east part of Vatnajökull, where they reached 2.0 m in altitude and were several hundred metres long.

In all those cases a fluted moraine occurred immediately at the glacier front. The ridges in the foreland pass into distinct ridges under the glacier. Similarly situated were the fluted moraines in the foreland of Isfallglaciären in Swedish Lappland (Schytt 1959, 1963) and in front of Werenskiöld glacier

Fig. 18. Fluted moraine in foreland of Sidujökull, August 1968. Photo by J. Szupryczyński

in Spitsbergen (Szupryczyński 1963, Baranowski 1970). On the Isfall and Werenskiöld glaciers pits were dug into the ice front to expose the ridges underneath. Schytt (1959, 1963) and Baranowski (1970) found that the ridges under the ice are generally higher and have steeper slopes, and the material they are built of is frozen. It was also found that the fluted moraine in front of the glacier has a muddy consistence. Observations proved that the mud had not developed from saturation of formerly dry fluted moraine by ablational or rain water, but that the moraine already had such water content under the glacier (Schytt 1959, Baranowski 1970).

The fluted moraine in Sidujökull foreland is separated from the glacier front by ice-cored moraines and has already been drained, transformed and

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emplaced. It has no contact with the front anywhere, and that is why we could not study its formation *in statu nascendi*. But we took care of examining its geological structure.

The ridges and furrows differ in their surficial geological structure. The surface of ridges is covered with stones, while fine debris prevails in the furrows. A little further away from the glacier front, the grain size differences are also underlined by patches of moss which grows in furrows but is absent on the ridges. Differences in mechanical composition are of some value for it was found in the foreland of some glaciers that both ridges and furrows were built of homogeneous material (Baranowski 1970, p. 69). As to Sidujökull foreland grain-size differences between the ridges and furrows at the surface are very distinct.

In order to study the geological structure of the fluted moraine, an exca-

Fig. 19. An excavation within a fluted moraine. Photo by J. Szupryczyński

vation was made 3.0 m long, 1.1 m wide and 1.6 m deep (Figs. 19 and 20). It was made within a ridge, and was parallel to the morphological axis of the latter. Geological structure of the fluted moraine exposed in the excavation is presented in Fig. 21. There is a stone layer at the top. These are mainly pebbles 0.2 m in diameter. Sporadically, boulders to 0.5 m in diameter appear. Below lies a strongly compact, moraine deposit with a tabular structure. Underneath follows a thin layer of very strongly pressed slate-like material (layer 3 in Fig. 21), then a moraine deposit with numerous cobbles and pebbles (layer 4); deeper gravels and sands occur (layers 5 and 6). These are distinctly stratified and certainly represent older glaciofluvial series. The bottom of the exposure containes basalt lava waste. The fluted moraine thickness in the

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Fig. 20. Inner structure of the fluted moraine, August 1968. Photo by J. Szupryczynski

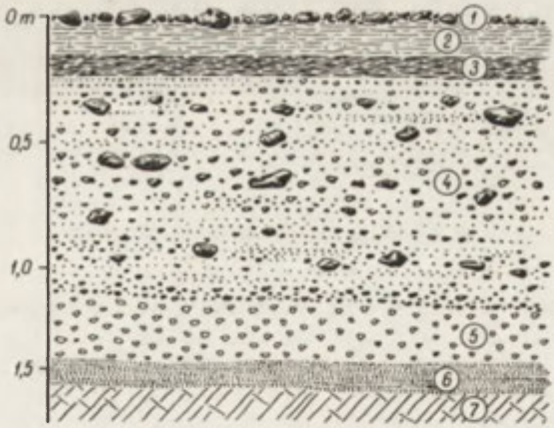

Fig. 21. Exposure in the fluted moraine

1 — A layer of pebbles and boulders up to 0.2 m in diameter. Behind the exposure the diameter of boulders comes up to 0.5 m, 2 — Grey moraine deposits, 3 — a thin strongly loamy layer, 4 — moraine material as in No. 2 with many pebbles and boulders up to 0.2 m in diameter, 5 — gravelly-sandy deposits-glaciofluvial, 6 — clearly stratified deposits of a yellow-brownish colour, 7 — rock-waste of basalt lava of a yellow-brownish colour

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excavation is small: about 1 m. Maximal thickness 1.3 m was noted in the western part on the Brunná River.

The geological structure presented in Fig. 21 is typical for the whole area where the fluted moraine occurs. Mechanical composition of individual layers of the moraine is given in Table 1. The first three samples were taken from the excavation. Sample 1 was taken from layer 2, sample 2 from layer 4 (Fig. 21). Both samples show great similarity as regards their mechanical composition, with the reservation that the morainal deposit in layer 4 has a heavy boulder content absent in layer 2. The most peculiar in its mechanical composition is the thin layer 3 (sample no. 3) with its overwhelming majority of fine, 0.002–0.5 mm particles.

TABLE 1. Mechanical composition of the deposits from the fluted moraine

∅ in mm	Percentage					
	Sample no. 1	Sample no. 2	Sample no. 3	Sample no. 4	Sample no. 5	Sample no. 6
>10	6.566	9.400		3.671	7.440	
10–9	0.753	1.160		0.354	0.615	
9–8	1.270	1.163		0.800	0.912	1.300
8–7	1.540	1.463		1.294	1.150	0.314
7–6	1.046	1.602		0.923	1.160	0.535
6–5	1.650	1.248	0.920	1.050	1.153	0.642
5–4	3.044	2.152	0.573	2.049	1.920	1.025
4–3	2.750	2.090	0.600	2.872	2.008	1.195
3–2	5.020	4.096	1.862	3.802	3.784	2.700
2–1.5	0.241	0.154	0.155	0.080	0.056	0.182
1.5–1.25	4.880	3.352	2.640	2.705	2.952	4.056
1.25–1.02	0.240	0.320	0.250	0.200	0.150	0.451
1.02–0.75	4.500	5.300	2.400	5.300	5.600	4.900
0.75–0.6	2.600	3.200	2.500	3.100	3.500	3.200
0.6–0.5	2.400	2.700	3.200	2.500	3.400	2.400
0.5–0.3	7.100	9.300	11.800	9.900	8.900	9.200
0.3–0.2	4.800	7.900	7.300	7.700	6.300	10.100
0.2–0.12	9.700	7.600	10.000	9.100	8.300	9.600
0.12–0.05	7.800	9.300	10.200	10.500	10.000	10.500
0.05–0.02	11.400	9.300	13.000	11.300	14.600	10.500
0.02–0.006	12.100	9.300	12.100	12.000	7.700	12.300
0.006–0.002	3.600	2.900	8.400	4.800	4.600	4.400
<0.002	5.000	5.000	12.100	4.000	3.800	10.500
	100.000%	100.000%	100.000%	100.000%	100.000%	100.000%

Samples no. 4, 5 and 6 were taken in the exposure of ground moraine on the Brunná River (Fig. 22). They were taken from the same moraine layers as those in the excavation. Samples no. 4 and 5 were taken from a depth of 0.2 mm and 1.0 m from the surface. Their mechanical composition is almost the same as that of samples no. 1 and 2; whereas sample no. 6 was taken from a layer analogical to the one in the excavation (layer 3), and its mechanical composition approximates that of sample no. 3. Samples no. 3 and 6 contain predominantly fine-graines, less than 0.1 mm material. In the samples analyzed, the material less than 0.2 mm takes 40–60% of the overall sample volume.

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With respect to their mineralogical-petrographic composition, the samples taken from various layers of the ground moraine are very similar. The morainal material is decisively rich in basalt debris whose content in some samples reaches 97.8% of the overall volume. Apart from that, obsidian debris was found, whose content reaches 32% of the overall volume in some samples. Only small content of quartz, olivin and magnetite was noted (Table 2).

The ground moraine debris contains only components of volcanic rocks which occur in the bedrock and environment of the glacier. No publications available to us give a detailed geological structure of fluted moraine (Baranowski 1970, Dyson 1952, Hoppe and Schytt 1953, Raitt 1962, Schytt 1959, 1962,

Fig. 22. Fluted moraine on the Brunna River, August 1968. Photo by J. Szupryczynski

1963, Todtmann 1957, 1960). They are usually limited to an analysis of the topography of fluted moraine surface and speculations as to its origin are based on this. The character of topography is no doubt crucial when we consider the origin of fluted moraine, but it is surely far more important to identify the character of sedimentation of this type of ground moraine as well as its mechanical composition. Less important here is its mineralogical-petrographic composition which will depend in many areas on the inventory of the rocks that make up the bedrock and environment of a glacier.

The character of sedimentation in this case is indirectly attested by pebbles orientation in individual layers. To prove this, measurements were made of orientation of longer axes of pebbles lying on ridge surfaces or stripes of stones. Measurements were also made of the orientation and dips of pebbles in deeper layers of the excavated ground moraine. The results are shown in

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TABLE 2. Mineralogic-petrographical composition of the ground moraine in the Sidujökull recessional area

Sample no. 1

⌀ 2.0 mm = 20.4%

bits of basalt	= 95.9%
bits of volcanic glass	= 3.2%
others	= 0.9%
	100.0%

⌀ 2.0–0.5 mm = 19.1%

bits of basalt	= 66.6%
bits of volcanic glass	= 32.6%
quartz grains	= 0.3%
others	= 0.5%
	100.0%

⌀ 0.5–0.1 mm = 25.1%

bits of basalt	= 85.2%
bits of volcanic glass	= 13.1%
quartz grains	= 0.8%
olivine	= 0.2%
magnetite	= 0.1%
others	= 0.6%
	100.0%

⌀ 0.1 mm = 42.6%

Sample no. 2

⌀ 2.0 mm = 19.8%

bits of basalt	= 87.9%
bits of volcanic glass	= 11.4%
others	= 0.7%
	100.0%

⌀ 2.0–0.5 mm = 15.8%

bits of basalt	= 65.4%
bits of volcanic glass	= 33.2%
quartz grains	= 0.5%
others	= 0.9%
	100.0%

⌀ 0.5–0.1 mm = 29.0%

bits of basalt	= 82.6%
bits of volcanic glass	= 15.8%
quartz grains	= 0.8%
olivine	= 0.2%
magnetite	= 0.1%
others	= 0.5%
	100.0%

⌀ 0.1 mm = 35.4%

Sample no. 3

⌀ 2.0 mm = 2.6%

bits of basalt	= 97.6%
bits of volcanic glass	= —
others	= 2.4%
	100.0%

⌀ 0.5 mm = 11.9%

bits of basalt	= 87.3%
bits of volcanic glass	= 12.0%
quartz grains	= 0.3%
others	= 0.4%
	100.0%

⌀ 0.5–0.1 mm = 16.5%

bits of basalt	= 92.1%
bits of volcanic glass	= 6.0%
quartz grains	= 1.0%
olivine	= 0.3%
magnetite	= 0.1%
others	= 0.5%
	100.0%

⌀ 0.1 mm = 69.0%

Sample no. 4

⌀ 2.0 mm = 15.2%

bits of basalt	= 97.8%
bits of volcanic glass	= 1.4%
others	= 0.8%
	100.0%

⌀ 0.5–2.0 mm = 14.3%

bits of basalt	= 77.5%
bits of volcanic glass	= 20.4%
quartz grains	= 0.8%
others	= 1.3%
	100.0%

⌀ 0.5–0.1 mm = 27.0%

bits of basalt	= 89.1%
bits of volcanic glass	= 9.1%
quartz grains	= 1.1%
olivine	= 0.2%
magnetite	= 0.1%
others	= 0.4%
	100.0%

⌀ 0.1 mm = 43.5%

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Sample no. 5

⌀ 2.0 mm = 17.5%		
bits of basalt	=	93.6%
bits of volcanic glass	=	5.7%
others	=	0.7%
		100.0%
⌀ 2.0–0.5 mm = 17.1%		
bits of basalt	=	48.6%
bits of volcanic glass	=	48.6%
quartz grains	=	1.2%
others	=	1.6%
		100.0%
⌀ 0.5–0.1 mm = 25.8%		
bits of basalt	=	78.4%
bits of volcanic glass	=	17.6%
quartz grains	=	1.9%
olivine	=	0.2%
magnetite	=	0.1%
others	=	1.8%
		100.0%
⌀ 0.1 mm = 39.6%		

Sample no. 6

⌀ 2.0 mm = 6.1%		
bits of basalt	=	89.8%
bits of volcanic glass	=	8.3%
others	=	1.9%
		100.0%
⌀ 0.5–2.0 mm = 15.1%		
bits of basalt	=	73.1%
bits of volcanic glass	=	25.7%
quartz grains	=	0.5%
others	=	0.7%
		100.0%
⌀ 0.5–0.1 mm = 37.0%		
bits of basalt	=	93.3%
bits of volcanic glass	=	5.5%
quartz grains	=	0.6%
olivine	=	0.3%
magnetite	=	0.1%
others	=	0.2%
		100.0%
⌀ 0.1 mm = 41.8%		

the diagrams. The first two (Figs. 23 and 24) reveal the orientation of pebbles on a ridge surface whose morphological axis runs N 190°. Orientation measurements were carried out on a 1 m² testing area in the central part of the ridge whose width was up to 1.5 m. Most pebbles, for as much as 56% (Fig. 23), lie parallel to the morphological axis of the ridge or in a sector close to this orientation — N 350° to N 30°. Only some pebbles, about 10%, have their longer axis perpendicular to the morphological axis, which may be conditioned by their shape or by later processes that led to their reorientation. Orientation

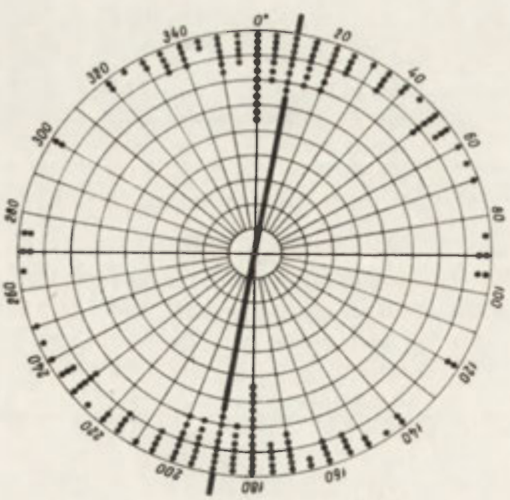

Fig. 23. Diagram of the arrangement of pebbles in the middle part of the small ridge of a fluted moraine. The morphological axis of the ridge is N 190°

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measurements were also carried out in the most eastern and western parts of the ridge on its contact with a furrow. The figures are presented in one of the selected diagrams (Fig. 24). The pebbles there, about 1–1.5 m from the morphological axis of the ridge, have most probably been reorientated during the stabilization of its form. It was probably in this part of the ridge that most pronounced orientation changes took place in consequence of gravitational

Fig. 24. Diagram of the arrangement of pebbles in a fluted moraine 1–1.5 m from the morphological axis of the ridge

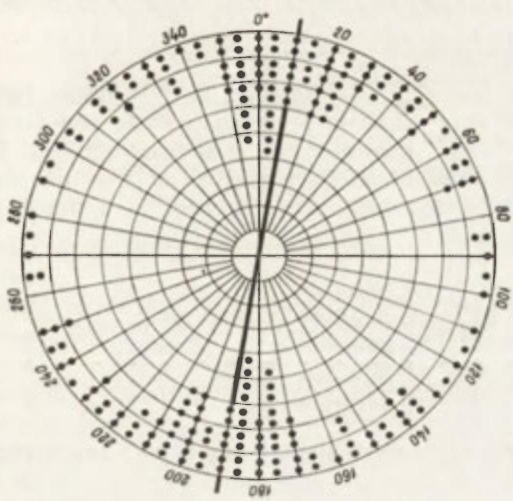

Fig. 25. Diagram of the arrangement of pebbles in the middle part of a small ridge of fluted moraine

sliding of pebbles toward the neighbouring furrows. Orientation measurements on another ridge surface on a 1 m² testing area (Fig. 25) also reveal that longer axes of pebbles are orientated parallel to the morphological axis of this ridge. The situation repeats itself in all the cases we examined. Still more pronounced orientation of longer axes of pebbles was found in the deposits building deeper layers of the fluted moraine. The orientations and dips in the excava-

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tion were measured at depths of 0–0.18 and 0.22–0.7 m beneath the surface (within layers 2 and 4 in Fig. 21).

So far Hoppe and Schytt (1953) and Szupryczyński (1963) have stated their views on the orientation of pebbles within fluted moraines. The first two authors stated that they could not observe any preferred orientation at fluted moraine ridges of Iceland glaciers. But they did not present their measurements in tables or diagrams. Szupryczyński (1963, p. 74–76) found that boulders and pebbles were orientated parallel to the morphological axis of individual ridges in the fluted moraine of Werenskiöld glacier foreland. But even in this case no figures were given due to the few measurements made.

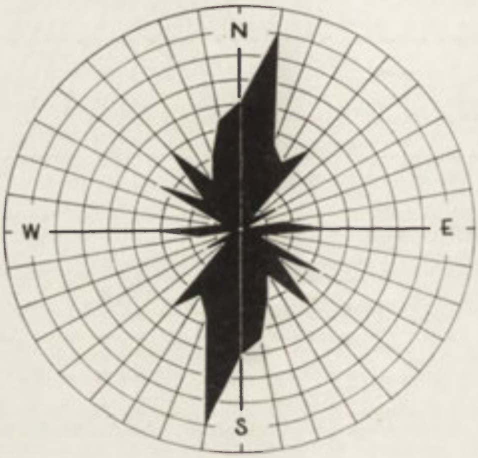

Fig. 26. Arrangement of longer axes of pebbles in a fluted moraine at a depth of 0.18–0.22 m

The orientation of pebbles within the ground moraine in Sidujökull foreland proves that the shifts of the morainal deposit corresponded to the direction of the glacier movement. It was found in both cases that (Figs. 26 and 27) longer axes of pebbles are orientated close to the orientation of the morphological axis of the ridge but their dips are different. Pebbles at a depth of 0.18–0.22 m (Fig. 8) have the following dips:

2–20°	—	65%	of pebbles
21–30°	—	16%	of pebbles
31–40°	—	12%	of pebbles
41–50°	—	5%	of pebbles
above 51°	—	2%	of pebbles

Thus most pebbles express a relatively small dip-within 2–30°. Inclination of above 30° is expressed by only 10% — in that 2% are above 50°.

The measurements of pebble dips in the morainal deposit at a depth of 0.22–0.68 (Fig. 27) revealed the following dip values:

2–20°	—	63%	of pebbles
21–30°	—	12%	of pebbles
31–40°	—	7%	of pebbles
41–50°	—	6%	of pebbles
above 51°	—	12%	of pebbles

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The prevailing value is 2–30°, for as much as 75% of pebbles express dips in this range.

Hoppe and Schytt (1953) and Flint (1957) suppose that the boulders in the glacier bedrock influenced the formation of tunnels in the glacier basal part. Next, morainal debris was pressed into the tunnels under the pressure of ice mass and formed narrow ridges there. The authors formulated this view on the basis of their observations in Brúarjökull foreland, where they found a very large number of boulders within the ridges. With reference to Isfallglaciären (Swedish Lappland), Hoppe and Schytt (1953) assume the pressing of saturated bottom moraine debris into hollows in the lee of boulders. Release

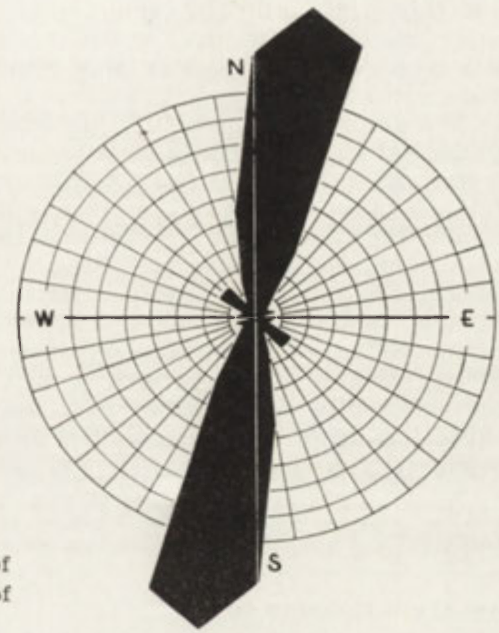

Fig. 27. Arrangement of longer axes of pebbles in a fluted moraine at a depth of 0.22–0.68 m

of pressure could later cause the moraine to be trapped by basal ice so that the former was carried along with the ice while more recent debris was again pressed in the lee of boulders. Todtmann (1960) states on the basis of her observations on Iceland ice-cap forelands that there are in many cases large areas of fluted moraine where no large boulders can be found. She assumes thus that the origin of this type of moraine is different. She asserts that ridges (*Grundspaltenwälle*) form in the basal crevasses of a glacier. A crevasse in the bottom of a glacier develops, according to her, as a result of tension in the ice mass passing from the phase of transgression (or recession) into the one of stagnation. Wet and plastic fluted moraine is wedged by pressure into the formed crevasses.

The ridges and furrows in the fluted moraine were also regarded as forms which developed in result of glacier or melt water erosion (de Quarian, Schnitter 1920, Ray 1935, Strøm 1963). This type of moraine was also regarded as a specific form which developed as a result of the joint action between exarating and accumultating processes of a glacier (Dyson 1952). Recently, Baranowski (1970) has attempted to explain the origin of fluted moraine modified by ice motion in terms of frost heaving in the glacier bedrock. He thinks that

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fluted moraine develops in specific subglacial periglacial zones in consequence of changes in the thermal regime in the glacier bedrock.

None of these hypotheses explains in a satisfactory way the topography of a fluted moraine, that is, the regularity in the distribution of its ridges and furrows. The hypotheses have been based, in our opinion, on too narrow observational material. Baranowski's (1970) hypothesis would have to be confirmed by observations in thermal conditions in the bottom of many glaciers. It should be admitted, however, that his hypotheses attempts to give a more complex explanation of the origin of fluted moraine topography.

We assume on the basis of our own observations that the role of large boulders that are to influence the formation of crevasses in the glacier bottom and ridges in the former cannot be regarded as the only decisive factor in the origin of fluted moraine. For the area of fluted moraine in Sidujökull foreland has a very small number of large boulders. And in many places large boulders are absent.

We gathered from Sidujökull foreland observations on the topography, geological structure and orientation of pebbles within the fluted moraine. The material gathered does not confirm any of the hypotheses about the origin of fluted morain nor does it permit any decisive explanation of the problem.

All the observations concerning this form in various areas have proved that fluted moraine is a specific geomorphic facies of ground moraine which develops in subglacial conditions in the bottom of a glacier during its motion. Pebble orientation measurements in the morainal material and on the surface of the fluted moraine ridges in Sidujökull foreland allow us to assume that there exists a distinct concordance between the pebbles orientation and glacier movement direction. It allow us at the same time to discard the conception of squeezing up of ground moraine, because that would preclude any strong orientation of pebbles.

DEPOSITS AND FORMS IN THE ACTUAL ZONE OF INTENSIVE ABLATION

ABLATION MORAINE COVER

Sources of Material

We can observe as a common phenomenon the occurrence of a cover of unsorted deposits in the lowest part of the Sidujökull slope. Their mechanical composition is very differentiated and ranges from silts to large boulders with thickness from a few centimetres to several metres. Thickness grows, in general, down the glacier front, and is especially considerable at the places close to the ice-cored moraines. But this applies only to the cases in which no superglacial stream runs in the trough between the front and an ice-cored moraine forming out of the former. Lack of a deposit cover on the surface of the lowest part of Sidujökull's front is an exception. First of all, the cover does not occur at the places where the front has the form of an ice cliff. This situation can be observed at the south-eastern margin of Sidujökull which is undercut by the Djupa river. The vertical wall of the glacier front prevents formation of any deposit cover, while at the same time it is conducive to the formation of scree dumps at the glacier-toe which are composed of the debris that melts out from the ice cliff (Fig. 28).

The formation of the deposit cover in the lowest part of Sidujökull's front which constitutes the ablation moraine cover is a complex process. The deci-

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sive factors involved here are both various sources of the debris as well as elementary processes responsible for the concentration, redeposition and accumulation of deposits on the glacial surface.

The material of the ablation moraine cover comes from three sources which can be easily identified in the case of Sidujokull. The first two are in the glacier itself; the third is in the extraglacial zone. The material which emerges from the interior of the glacier in the zone of intensive ablation is of double nature: it is either volcanic ashes or debris. The ashes left after volcanic eruptions cover large areas, among them also glacier surfaces. This sort of phenomenon is commonly observed in Iceland. Information about this is given, for instance, by Wadell (1920) or Thorarinsson (1949). The volcanic ashes deposited on glacier surfaces are next covered with snow and, as the accumulation of snow advances, they penetrate into the lower parts of the firn and glacial ice. Repeated volcanic ashfalls give, in effect, numerous layers which are revealed in glacial ablation zones in the form of black lines that resemble contour lines in their distribution. They can be well seen on aerial photographs of the Sidujökull (Fig. 29). Unfortunately, we could not establish their thickness. A petrographic-mineralogical analysis does not leave a shade of doubt that the material coming from the darker bands in the zone of intensive ablation of Sidujokull is volcanic ashes (Kozarski, Szupryczyński 1971). It consists chiefly of volcanic glass (obsidian) with a small admixture of basalt and palagonite debris and clay balls which are secondary in this material, for they formed the admixture as a result of transport of the volcanic ashes on the glacier surface. Of vestigial occurrence are: quartz, basalt debris with

Fig. 28. Moraine damps at the ice cliff on the Djúpá River. (SE Sidujokull margin). Photo by S. Kozarski

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Fig. 29. Volcanic ash bands in the ablation zone of Sidujokull (air photograph taken 1960)

volcanic glass, ilmenite, olivine, magnetite, apatite, iron oxides and iron hydroxides as well as zeolite.

Prevalence of obsidian (81.6%) in relation to the other components is an eloquent proof of the origin of the material; it also decides about the dark colour of the material. A mechanical analysis proves that volcanic ashes belong to sand fractions with the prevalence of fine-grain fraction. The mechanical composition like this speaks for a great susceptability of volcanic ashes to transport by superglacial waters.

The debris which is transported by the glacier and that comes up to the surface of its front under the influence of ablation reveals quite a different character and mode of occurrence. Usually it is unsorted and is composed of both silts and large boulders over 1 m in diameter, with a whole range of intermediate fractions. There are, however, places — attached to the shear planes only — which is indicated by the phenomenon of overthrusting, in which the debris reveals outstanding homogeneity in mechanical and petrographic-

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-mineralogical composition. Analyses point out to the fact that 57% of the debris is of a 0.002 mm fraction and that only 0.7% is of a fraction 0.5 mm in diameter, with the coarsest material reaching a 0.75 to 1.02 fraction and constituting only very small admixture (0.1%).

X-ray analyses of the clayey substance, which constitutes 85.5% in the petrographic-mineralogical composition of the whole material, proved that it belongs to the same group as montmorillonite does. Taking into consideration the composition of the part left after desludging of the clay, as well as mode of occurrence and formation of clayey particles and their reaction to water (high plasticity) we are allowed to infer that the analysed substance is bentonite loam developed from old volcanic ashes under the influence of chemical weathering. The type of weathering determines the conditions in which it took place. The conditions could not be an englacial environment. The process of weathering took place in subaerial conditions, on the bedrock, before glaciation of the analysed area that is now covered by Sidujokull. It was only later that products of chemical weathering were dragged into the moving glacier along the shear planes and became in this way one of the sources of material for the formation of the ablation moraine covers at the glacier front.

Debris occurs and is transported inside Sidujokull in two ways: disseminated and concentrated. The disseminated occurrence and transport comprise a wide range of fractions: from absolutely extreme, that is from loamy substances of 0.002 mm in diameter to large boulders 1.6 m in diameter.

The disseminated debris occurs mostly in the basal part of the glacier. In the close-to-surface glacial ice layers, as has been proved by examination of the cleavages and cracks in the surface of the glacier, the debris is scarce. Results of our observations are compatible in this respect with the observations made on the glaciers in the north-western part of Spitsbergen (Boulton 1967). It is worth noting that fine-grain fractions are dominant in the debris. This debris covers the surface of the ice with a film during ablation which takes place on the vertical walls of the glacier margin. The film is not able to blur the structure of the glacier base.

The concentrated occurrence of the debris consists in that the debris forms compact bands which are well seen in the ice cliff on the Djúpá River (Figs. 30 and 31). In the basal part the debris bands are nearly parallel to the base and foliation of the glacial ice. Over the basal part they are more and more upturned to the surface of the glacier. The thickness of the bands observed in the ice cliff on the Djúpá river reaches from several cm to about 1 m. The top surface of the bands is remarkably uneven. At the places where crevasses occur, the band-forming debris penetrates into them in the shape of wedges (Fig. 31). Rising of the wedges from a solid band makes us suppose that their formation is connected with squeezing up of the debris into the crevasse from underneath. Despite complete freezing of the debris we should accept this way of formation of wedges. The process of wedge formation must have taken place and was conditioned by heavy pressure of the glacial ice which, when conveyed on to the frozen-debris bands, caused the debris to acquire the consistence of a plastic body. The phenomenon which has been characterized above has nothing to do with *Lehmmauern* from Spitsbergen as described by Gripp (1929) whose way of formation as reported by that author has recently been criticized by Boulton (1967, 1970). The debris bands consist of a completely unsorted and structureless clay-and-sand mass with a heavy content of pebbles and boulders.

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No place has been found in which a debris band would reveal stratification and variety in the mechanical composition such as ascertained by Boulton (1967) in Sørbreen. The whole of material in the debris bands is identical with a typical basal till.

Bad conditions for an examination of the cliff and its base did not allow for a closer study of the nature of the debris bands and their relation to the base at that. Hence, there are difficulties in explaining formation of these bands. But one, directly observable, phenomenon is worthy our attention; name-

Fig. 30. Debris band in the basal part of the ice cliff. Photo by S. Kozarski

ly, that the bands reveal lack of conformity with the ice foliation, cross it already a little above the glacier-toe, and that this lack of conformity is the greater the closer we get to the surface of the glacier (Fig. 28). Therefore, the situation in our case is different from the one reported by Hooke (1970) from near Thule (Greenland), where the debris bands are a component of the ice foliation. Having this phenomenon in view, despite the criticism (Weertman 1961) of Bishop's views (1957) on the formation of the so-called shear moraines, it should be judged that the presence of thick debris bands in the Sidujökull is connected with the process of shearing. It seems that it is difficult at present to propose, in reference to Sidujökull, another adequate theory of the incorporation of the debris bands. The freezing model formulated by Weertman (1961) cannot be applied to the Iceland ice-caps, because they do not fall within the category of cold ice sheets to which the model refers. It is also difficult to accept Boulton's (1970) conception of the incorporation of subglacial material into englacial position through thrusting in view of a lack of any dis-

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tinct ice deformation in the cliff on the Djupa river. Another way of debris incorporation given by that author and derived from Lliboutry's conception (1964–1965, p. 689): that is, through basal freezing, would require here, in the case of Sidujökull, investigations over the physical state of the glacier base, or it would at least require investigating the presence of external traces of functioning of this mechanism in the form of existence of definite ice and debris band stratification together with reversing of its primary sequence which refers to the bedrock.

Fig. 31. Debris band with a wedge-like structure in the basal part of the ice cliff. Photo by S. Kozarski

The thick debris bands crop out as low debris dykes on the glacier surface in the zone of intense ablation (Fig. 32). They are the major source of the debris for the formation of the ablation moraine cover on the glacier surface. Similar to Spitsbergen glaciers (Boulton 1967) the dyke-forming material is frozen, but due to a very strong ablation, the interstitial ice vanishes rapidly from the dykes; therefore, the dykes are low: they reach a height of 0.5 m at the utmost and are not so sharp in outline as those described by Boulton (1967).

The third debris source for the formation of ablation moraine covers lies in the extraglacial zone. This is the least productive source and has to do only with the finest material from the category of dust fractions. The source we speak about is eolian transport. During our research work we had the oppor-

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Fig. 32. Debris-dyke on the Sidujökull surface. Photo by S. Kozarski

Fig. 33. A dust storm in the marginal zone of Sidujökull. Photo by S. Kozarski

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tunity to observe in the foreland of the glacier several dust storms which transported material on to the surface of the glacier (Fig. 33). The storms would come after several rainless days, and then outwash deposits underwent rapid transportation. Particularly susceptible to deflation are silts which cover the bottoms of abandoned channels of braided rivers. These silts are the material which was transported by rivers as a suspended load and which later underwent decantation. Since the material is blown out from large areas and disseminated on the glacier surface, we cannot establish in what proportion it is an admixture of the whole mass of mineral deposits from which ablation moraine covers are formed.

Results of our observations supplement the remarks which have already been made before by Lewis (1940), Swithinbank (1950) and Bout (1956) as well as Okko (1956) on the eolian transportation of fine debris on to the surface of Iceland ice caps. These results, however, do not constitute a basis for drawing the conclusion that eolian deposits may be concentrated on a glacier surface in a degree which would render possible the formation of covers that condition the development of ablation cones (Kozarski, Szupryczyński 1971).

The First Phase of Debris Concentration on the Glacier Surface and Ephemeral Relief

The zone of intensive ablation on Sidujökull is clearly bipartite. In the lowest part of the front one part of it is a continuous ablation moraine cover whose width reaches 50–70 m on the average and 120 m at the maximum. The other part of the zone extends up the glacier surface. The observations made in the period between 24 June and 6 July 1968 enabled us to establish its width as 300 to 350 m, and those made at the beginning of August established a 500–600 m width. The zone is a belt of maximum ablation and in the case of gentle sloping of the glacier surface (about 12°) is conducive to the formation of ablation cones (Kozarski, Szupryczyński 1971). Its role in the formation of the continuous ablation moraine cover is very crucial, for it is in this part of the zone of intensive ablation that mineral deposits are temporarily concentrated and then deconcentrated, and transported on to the lowest part of the glacier front. We can, therefore, say that the zone plays the role of a transit surface.

The mineral deposit cover in the part of the zone of intensive ablation already discussed is discontinuous and varies highly in its thickness: from a few millimetres to 1.3 m. The highest thickness can be noted on ablation cones tops, that is, at the places where the transitional concentration of deposits was initially most intense. Discontinuity of the cover and its varied thickness lead to the formation of ephemeral relief on the glacier surface. The relief appears in the form of a great quantity of ablation cones which should be regarded as a distinctive feature of maximum ablation belts on Iceland ice caps.

Ablation cones occur only where the front of the receding Sidujökull is gently sloped. They do not occur where the steep front occurs: at the ice cliff of the southern part of Sidujökull undercut by the Djúpá river. Above the ice cliff the glacier surface is dissected with a system of deep crevasses which show the way of the surface drainage. They lead to a complete elimination of superglacial drainage and so they do not allow for a concentration of volcanic material and debris redeposed by superglacial waters that give rise to the ablation cones.

A preliminary and perfunctory observation of the Sidujökull front leads to the wrong conclusion that the ablation cones are distributed in a chaotic way. Yet a closer analysis of the mode of their occurrence allows for establish-

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ing regularities in their distribution and sources of these regularities. The ablation cones follow clearly-cut lines which generally intersect at the right angle. The most frequent direction of orientation is that of being perpendicular to the glacier front. This direction is clearly related to the system of narrow radial crevasses and narrow superglacial stream channels. The second direction is that of being parallel to the glacier front and is conditioned by the arrangement of debris band outcroppings and shear planes. Where the crevasses and channels are most densely distributed, we can note the greatest frequency of occurrence of the ablation cones per square unit. In order to establish the extent of this phenomenon in figures, exact measurements were carried out at the tested areas. They were arranged into belts 100 metres long which were perpendicular to the glacier margin and divided into 10 by 10 metre squares. Two to fifty-four ablation cones were noted in individual squares. The average figures were 8 to 28 cones per m^2. The forms under discussion take, more often than not, the shape of a triangular-base cone if they are advanced in development. In the initial stages of development they have the shape of ridges extended in accordance with the lines of material concentration. So, these are ridges either perpendicular or parallel to the glacier margin. Among many clearly shaped, triangular-base cones we could observe three slopes: two lateral, eastern and western and the third always facing south, but investigations established that the presence of the southward slope is caused by the most favourable conditions for the action of insolation as well as a most intensive down wash of the material covering these parts of the cones by heavy rains. At the time when rainfalls were especially abundant, the southern cone slopes entirely lost their cover.

The height of ablation cones as measured in various parts of the Sidujokull front oscillates between 0.2 and 4.5 m. Cones over 5 m high were noted sporadically. The highest cones were connected with the occurrence of dead glacial mills in which there were the best conditions for the sedimentation of thick series of deposits at the stage of decreasing water outflow.

The examined cone slopes show various inclination. Most frequently it oscillates within the 30–50° range. In the top parts of the cones, where the covering material is of greatest thickness, the slope reaches 70°. This is caused by the lack of any ice core in the top. Any steep, almost vertical slope, is formed exclussively in the wet covering material. Although we often met with a symmetrical arrangement of the angles of the eastern and western slope, it should be stressed that this is not a general rule. In many cases the angles of these slopes were unequal, and differences up to 10° spoke for the existence of a clear-cut asymmetry. We did not find any regularities in the exposition of the gentle and steep slope. The distribution of the gentle and steep slope here was quite accidental. We should then conclude that the distribution is directly connected with the amount of the creeping material on one or the other slope. This phenomenon is brought about by temporary water content in the covering material and its friction.

All the ablation cones examined have an ice-core formed of glacial ice and covered with a coat of deposits (Fig. 34). Any ice core still contains foliation and crevasses which gave rise to debris deposition. The debris coat does not mantle the ice core regularly. Greatest thickeness was noted at tops of ablation cones if a cone was formed in the place of a former glacial mill: the thickness was 0.6 to 1.3 m, maximally it was over three metres. The slope cover is one to eight centimetres thick, and twenty centimetres in the extreme.

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In the first period of our research, between June 24 and July 6, the deposit cover was completely frozen at its contact with the ice core. It speaks for the lack of a seasonal development (summer) cycle of the ablation cones, e.g., from their formation till their disappearance. Our observations in this respect confirm the opinions of Wadell (1920) and Bout (1956) which were based on different criteria, as well as Swithinbank's (1950) view. This part of the cones which does not disappear at the end of summer will last under the cover of winter snow irrespective of the stage of development they are in.

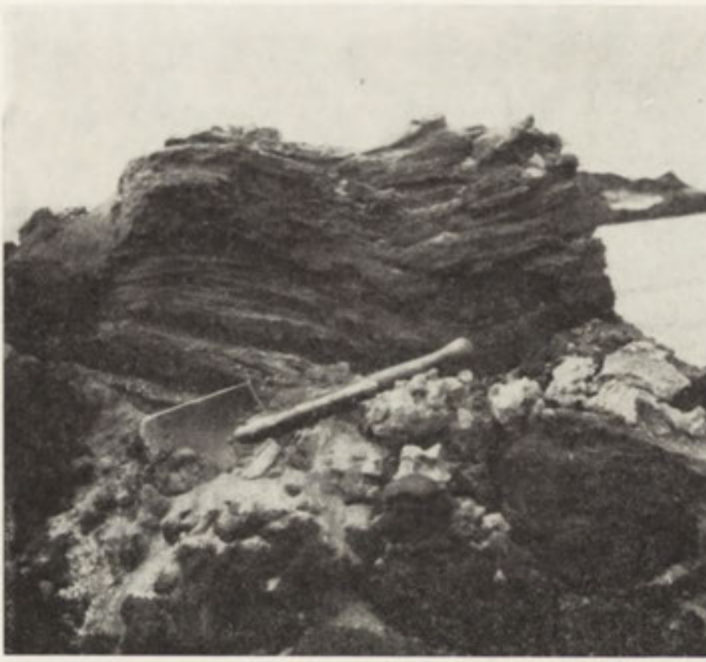

Fig. 34. Internal structure of deposits covering the summit of an ablation cone. Photo by S. Kozarski

It follows from the facts which have been presented already that the ablation cones at the Sidujökull front were formed only where the concentration of deposits had taken place previously. Insofar as we discovered two sources for the concentration they imply two ways of the formation of ablational cones. It should be explained here that differences in the formation of the cones are pertinent only to the initial stage and their distribution, but not to the way of development.

The first way of the formation of the cones is closely bound with the existence of superglacial channels and radial crevasses that are either perpendicular or slant to the glacier front. The characteristic feature here are ablation cones built on the channels and crevasses plane with a cover composed exclusively of materials whose structure is characteristic for running-water deposits (Fig. 34). We can distinguish four phases in the development of ablation cones formed in this way:

1st phase—deposition of the debris transported by superglacial streams either in a channel or in a crevasse; depending on the amount of debris and the depth of channel the deposited material may be over one metre thick;

2nd phase — formation of the central sandbar in the channel leads to branching off of the stream or to a change in its flow direction; thermal action of

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the stream water and, above all, ablation of the glacial ice which is not covered with any mineral debris cover causes a gradual lowering of the glacier surface round the cover; and in this way the cover comes up to the surface; its insulating influence on the glacial ice underneath is the cause for the formation of the ice socle which in its level outline exactly corresponds to the shape of the sandbar;

3rd phase — appearance of the slopes causes the gravitational movement of the debris covering the ice socle; the joint action of ablation on the glacier surface round the cone and the gravitational creeping are the main stage in

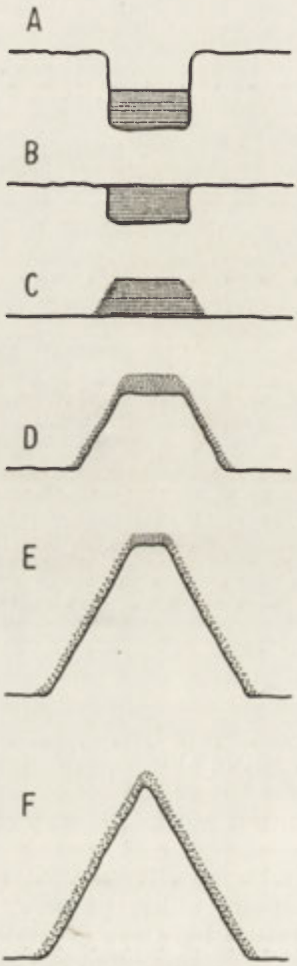

Fig. 35. Scheme showing the main mode of ablation cones development on Sidujökull

the development of a cone; this phase lasts as long as the amount of debris coverning the cone sufficiently insulates the ice core against the external thermal influences;

4th phase — when the amount of debris on the cone summit is too small to insulate the slopes sufficiently then comes a gradual disappearance of the cone; on Sidujökull the gravitational transport of the cone-covering debris down slope is supported in a significant way by abundant rainfalls.

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The development of ablation cones described above applies to most of the forms examined by us on Sidujökull (Fig. 36). Only sporadically could we meet with occurrence of other cones lower, however, which preserved a crevasse, instead of a flat surface, in the ice-core summit. These are undoubtedly forms which developed from the water-deposited debris accumulated in very narrow crevasses where the conditions were not right for the formation of central sandbar. This way of initiating cones was described by Lewis (1940) and Swithinbank (1950). More rare are also those ablation cones that develop under the influence of debris deposition on the banks of superglacial stream channels.

Fig. 36. An ablation cone with a crevasse in the summit. Photo by S. Kozarski

In the lower parts of the glacier front which is not strewn with an ablation moraine cover we have to do with flooding and rapid deposition of the transported material when the waters reach the high level and transport an abundance of solids (Fig. 37). This phenomenon is characteristic for the beginning of summer, when the superglacial stream channels are not deep enough and are to a large extent filled up with snow. The debris deposited on the banks in the shape of miniature *levee* gives rise to irregular ablation cones which disturb the system of distribution of the cones founded on the crevasses plane. This mode of debris deposition and cone formation was regarded by Bout (1956) as basic for the initial stage of the formation of a superglacial stream network for Hofsjokull.

The second way of ablation cones formation on Sidujokull is connected with the outcroppings of thin debris bands and shear planes. The debris that gets out of them wraps their closest neighbourhood insulating in this way the glacier surface against the thermal influences of insolation and air. This phe-

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nomenon has recently been described by Boulton (1967) in reference to the Sørbreen glacier (Vestspitsbergen). The process of cones formation started in this way is continued later on analogically to the development of forms with a superglacial stream deposit cover. Besides their origin and character of their covers, the cones formed in this way differ from other cones in that they are distributed parallel to the glacier front; furthermore, they are closely arranged one to another. This has already been noticed by Swithinbank (1950), who examined the ablation cones on Skaftárjökull. The distance between one cone and another is often so small that the cones join into a line of ridges. Similar to Lewis (1940) and Swithinbank (1950) we also observed that cones of this type are often formed at the lines of shear planes together with the phenomenon of overthrusting.

Fig. 37. *Levees* and growing ablation cones on the sides of supraglacial stream channels. Photo by S. Kozarski

At the places where the debris bands outcrop, there is no trace of the formation of ablation cones of bigger size. The reason for this lies in the heavy content of interstitial ice in the debris bands whose rate of melting is conducive to considerable saturation of the debris with water and causes rapid deconcentration of the former. In spite of this phenomenon, the retarding action of the debris on ablation close to the debris bands outcroppings is marked near them by the convexity of the glacier surface.

Formation of ablation cones, independent of the source of the covering material, should be regarded as a trace of temporary, but important and first phase of deposit concentration on the glacier surface. The first phase of concentration is introductory for the second that takes place in the lowest part of the front, where the ablation moraine covers are formed, which lead to durable geomorphic effects directly at the glacier margin.

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The Phase of Debris Deconcentration and its Transport on the Glacier Slope

The advancing process of ablation leads to a gradual disappearance of ablation cones and interstitial ice in the outcroppings of the thick debris bands. As a consequence, the debris concentrated in the form of covers on the ablation cones or in the thick bands undergoes scattering over the glacier surface. There are three major elementary processes responsible for this deconcentration which work on the surface with varying rate and productivity in transport.

The dominating role among these processes is played by superglacial waters which activity can be compared with rill- and sheet wash (Fig. 38). They carry away all the debris from the covers that mantle the disappearing abla-

Fig. 38. Thick debris concentration on the glacier surface. Photo by S. Kozarski

tion cones as well as silts and fine-grained sands from the debris bands. Limited transport capacity of the rill- and sheet wash leads to a celection of debris. Only silts and fine-grained sands undergo transportation, while coarse material is left (Fig. 38). Great mobility of the waters causes that, especially beneath the outcroppings of thick debris bands, the finest material is spread over quite regularly and it covers the glacier surface with a film. The presence of this film, which is of a very dark colour on Sidujökull, considerably accelerates ablation. This phenomenon, physically conditioned has been known for quite long, e.g., in alpine glaciers (Heim 1885, p. 228) and more precisely identified through the application of quantitative methods by Östrem (1959) during his experiment en Isfallglaciären (Swedish Lappland) as well as by Ward (1952) during his investigations at the Central Baffin Island ice cap. During our investigations on Sidujökull we identified the phenomenon through the qualita-

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tive method only by viewing the amount of working waters and greater progress in lowering of the glacier surface in the belt of the occurrence of the continuous fine-debris film.

In contrast to melt waters running down the glacier surface, the other processes conditioned by gravitation do not operate selective and transport the debris regardless of its size. The processes discussed here include falling down of debris clumps from the low dykes when the interstitial ice thaws, morainic flows as well as sliding and rolling down of cobbles and boulders.

The Second Phase of Debris Concentration on the Glacier Surface and Formation of the Ablation Moraine Cover Proper

The bulk of the debris transported by the elementary processes which have been mentioned above accumulates in the lowest part of the gently sloped glacier surface. The first to undergo accumulation is that debris which was delivered by gravitational movement. Superglacial melt waters accumulate the debris only partially. Its bulk, above all silts as a suspended load and very

Fig. 39. Ice-cored ridges and a thick ablation moraine cover in the trough. Photo by S. Kozarski

fine-grained sands, is carried over to streams and undergoes accumulation in the extraglacial zone. Constant delivery of superglacial debris to the lowest part of the glacier margin, at the time when the front remains unchanged or when there is a slow recession, causses concentration of debris and the formation of a continuous cover of unsorted material which is the cover proper of the ablation moraine. Its width reaches 50 to 70 m on the average when measured from the glacier-toe upwards. The thickness of the cover, which is from several cm to several metres, generally increases down the glacier slope. Greatest thickness was noted at the places where the cover reaches

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to the depressions between the active glacier and developing ice-cored ridges unless melt water runs down them (Fig. 39); while the lowest thickness was found near the outlets of subglacial tunnels (Fig. 40). In places like these a greater, as a rule, slope of the glacier surface and connected with it a more intensified movement of superglacial debris direct into a stream is not conducive to debris accumulation. For the same reasons the cover is less thick, too, in the neighbourhood of bigger superglacial streams.

The ablation moraine cover in the lowest part of the Sidujökull margin in spite of the fact that, in general, it reveals the same features locally show a considerably variability. This phenomenon is true, above all, in the mechanical and to a smaller degree in the petrographic composition which changes sporadically. There are various reasons for this variability in the mechanical composition that finds its expression in a general change in the proportion between the coarse debris content in the form of gravel, cobbles and boulders

Fig. 40. A thin ablation moraine cover at the mouth of a subglacial tunnel. Photo by S. Kozarski

and the fine debris represented by sands and silts. The reasons for this lie either in the unequal distribution of the coarse debris in the debris bands, which, after its release on to the glacier surface decides about its considerable participation in the formation of the ablation moraine cover and then we can speak about primary differences or differences brought about by the action of superglacial melt waters. These waters carry away the finer debris and cause a local enrichment of the ablation moraine cover with the coarse debris. This process can also occur already during the formation of the cover, which we have mentioned already, or in the advanced stages of its growth. The first

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instance is characterized by the presence of a considerable amount of coarse debris in the profile of the whole cover, while the second is marked with a heavy content of cobbles and boulders on the surface (Fig. 41). The pointed here differences in the proportion between the coarse and fine debris in ablation moraine covers are of prime importance for the other gravitation-conditioned movements which the cover undergoes. It is closely connected with water content. The differences also decide about the degree of cohesion in the cover-forming deposits. When cohesion is small in the case of a cover with a very coarse debris content, the possibilities for a longer preservation of the ice on which the cover rests are limited and vice versa.

Fig. 41. Ablation moraine cover with a high content of cobbles on its surface. Photo by S. Kozarski

Depending on the rate of the glacier recession, deposition of the ablation moraine cover leads to various geological and geomorphic effects on the foreland of the glacier. This is a direct consequence of the degree of debris concentration and, what follows, of the degree of debris thickness in the marginal part of the glacier surface. When the process of recession is rapid and relatively steady, the ablation moraine does not bring about any geomorphic effects in the glacier foreland. The moraine is deposited on the surface of the basal till in the shape of unsorted and loose debris 20 to 50 centimetres thick. This causes the formation of characteristic and known (Flint 1957) bipartition of moraine deposits in geological profile (Fig. 42). When the recession is slow then good conditions are created for the debris concentration in the lowest part of the glacier. Thick ablation moraine covers are formed at that time and play the role of a layer insulating the glacier against external thermal conditions. Their presence leads to permanent geomorphic features directly at the glacier margin in the shape of end-moraine hillocks, but before they are shaped into their final form, they undergo the ice-cored moraines phase.

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Fig. 42. Basal till with a thin ablation moraine in the top part. Photo by S. Kozarski

ICE-CORED MORAINES

Mode of Occurrence and Geomorphic Features

The actual Sidujokull margin is accompanied by a belt of sharp-out-lined hillocks and ridges with relative height of the order of several to 16 m (Szupryczyński, Kozarski 1970). They form an almost unbroken chain at the south-eastern margin of Sidujokull; to the ice cliff near Djúpá river and to a deep embayment in Sidujokull front at the caldera line in the southern part of the glacier (Fig. 3). Stretches of the continuous chain in front of the southern margin are broken in two long sections. There are either small hillocks which could not be mapped in the scale of the geomorphic map or they do not occur at all. Nor could we notice any hillock chains at the glacier margin which takes the shape of an embayment at the caldera line. According to our observations, the lack of hillocks and ridges directly in front of the glacier margin has various reasons. As basic here should be considered the slope of the glacier front and melt water action. At the places where the glacier front is very steep, as has already been pointed out in reference to the ice cliff on the Djúpá river, good conditions do not exist for debris concentration and formation of an ablation moraine cover on the glacier surface. Debris falls down from the steep front and gathers at its foot in the form of low dumps. Situation like this was noted in the glacier front embayment at the caldera line. The activity of melt waters decides in two ways about the formation of long breaks in the belt of hillocks and ridges at the glacier margin. The first way is connected with superglacial waters. If there is a dense network of melt water streams on the glacier surface, their intensive transporting work precludes any concentration of superglacial debris. Debris down washing accelerates

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Fig. 43. Ice-cored moraines at the SE Sidujokull margin. Photo by S. Kozarski

Fig. 44. A niche in the central part of an ice-cored moraine. Photo by J. Szupryczynski

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ablation and precludes any possibility for the formation of a thick ablation moraine cover. The second way is connected with the work of meltwaters directly at glacier margin. Wherever large areas covered by outwash deposits occur, no hillocks or ridges were noted there. That means that melt waters carried away debris from the glacier margin and prevented its deposition. The formation of numerous, narrow and short gaps in the hillock and ridge ranges is also related to meltwater work. But these gaps will be discussed later, when we will be considering the genesis of the forms under discussion.

The hillocks and ridges are of more or less the same shape all along the examined stretch of Sidujokull margin. The hillocks are usually marked with cone-shaped outlines with a clear-cut summit (Fig. 43). When there is a cavity in the central section of a hillock (Fig. 44) a double summit occurs. The summit is a product of processes which transformed a hillock of a primarily regular cone shape. The ridges are also sharp in outline and significant for the relief of the marginal zone. They are marked with the parallel-to-the-front orientation that is easy to notice because each ridge has a clear-cut crest-line. The crest-line is blurred only when the central part of a ridge similar to a hillock, is deformed by a cavity. Depending on the stage of development of the cavity, the crest-line either divides into two in the initial phase or disappears completely if the cavity reaches as far as the foot of the ridge.

Hillock and ridge slopes reveal, generally speaking, the same inclination close to 40° or a little less. This value corresponds to the angle of repose of gravel or sand-and-gravel deposits which mostly occur on the surface of the forms under discussion. When an overwhelming predominance of sands over gravels can be noted, the slope correspondingly descreases to 30 or 28°. When one examines individual sections of slopes, they seldom show uniform inclination as it might seem to follow from the figures given above. When hillock and ridge transformations brought about by secondary phenomena are considerable, the slope within which various cavities occur is changed into a not uniform surface.

Internal Structure

The hillocks and ridges which have been characterized above have the same type of internal structure in cuttings. There is to be observed in the profile a bipartition. In the top part a mineral deposit cover occur, while underneath we can observe the ice core only (Fig. 45). This type of internal structure allows us to count the hillocks and ridges occurring directly at Sidujokull margin amoung the category of phenomena termed as ice-cored moraines (Goldthwait 1951, Östrem 1962, 1964, Hook 1970). They are quite commonly noted and described at the recent margins of Iceland glaciers (i.a. Ahlmann 1938, Okko 1956, Spethmann 1912, Todtmann 1960, Woldstedt 1954), but in spite of this they have not been given enough notice as far as their formation mechanism goes reconstructed on the basis of studies of the processes in the marginal zone of glaciers and the internal structure of the forms themselves.

The mineral deposit cover resting on the ice core is of a thickness 0.2 to 1.0 m. A thickness over 1 m is something rare. The most frequently noted thickness was 0.2 to 0.5 m. It means that the ice core, when we take into account present height of the ice-cored moraines, is the dominant element in them. The main bulk of the deposits covering the ice core is completely unsorted. The deposits are made up of both very fine material represented by silts as well as sand-and-gravel fractions, cobbles and sometimes boulders. The character of the whole cover is identical with the ablation moraine resting

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Fig. 45. Internal structure of an ice-cored moraine. Photo by S. Kozarski
A — ice core, B — ablation moraine cover

Fig. 46. Stratified fluvioglacial deposits on an ice core. Photo by S. Kozarski

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in the lowest part of the active glacier margin. Only sporadically, in a few places, could we note a stratified sand-and-gravel deposit cover on ice cores (Fig. 46) or a cover with traces of stratification. These are short transported outwash deposits that have been accumulated earlier on the glacier margin and built short superglacial outwash cones.

The cover-forming material on the ice-cored moraines is very mobile. A heavy content of gravels, sands and lean silts with lack of any clayey particles at the same time is not conducive to water retention in the cover. This feature, resulting from the mechanical composition, coupled with a very dark colour of the covers causes the surface part of the covers to dry up easily on rainless and windy days. Hence, unbound, loose, cover debris easily undergoes gravitational processes and slides down ice-cored moraines. We could observe this phenomenon each time we examined the form mentioned above. Gravitational movement of debris causes its constant redeposition which leads to sorting of rock grains according to their size. Redeposition is intensified if an ice core starts to melt more intensively. Then debris slides under gravitation not only down the external slopes, but also down the newly-formed internal slopes. The process of sliding down is further reinforced by sliding of whole cover parts.

The main element in cuttings through hillocks and ridges, the ice core, decides about the shape of the ice-cored moraines existing at the recent glacier margin. The highest ice core thickness found was about 15 m when measured in the culminating part of the hillock which was 16 m high altogether. It was not feasible to carry out any crystallographic investigations of the ice that builds ice cores which would allow for a deeper insight into its nature, as it has been proved by Ostrem's (1963) and Hook's (1970) investigations. In spite of this real deficiency it seems to us that the structure of the core-forming ice

Fig. 47. Upturned foliation of ice core in the ice-cored moraine at SE Sidujokull margin. Photo by S. Kozarski

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Fig. 48. Upturned foliation of an ice-cored moraine at S Sidujokull margin. Photo by J. Szupryczynski

Fig. 49. Upturned thick debris band in the ice-cored moraine. Photo by S. Kozarski

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as observed in cuttings is clear enough to allow formulation of inferences about its origin. The inferences are also futher supported by observation of processes in the marginal zone that condition the formation mechanism of ice-cored moraines. In all the locations investigated the core ice has an unusually distinct foliation (Figs. 47 and 48). Foliation dips at a 20 to 50° exclusively towards the glacier margin. A comparison between the core ice foliation and that from the margin of the active glacier leads to the conclusion that they are identical. We always found the up-glacier dipping foliation in radial tensional crevisses and that on the walls of perpendicular to the glacier-margin superglacial channels as well as foliation from the ice cliff near Djupá river to have dip values similar to those in the ice cores close to the glacier surface. The dips show constant tendency to decrease up the glacier. The upturned foliation in the marginal zone is undoubtedly a proof for the upward-flow of ice even in case of lack of a flow-hindering factor (Hook 1970). We can judge then that the upturned foliation in ice cores of the ice-cored moraines which constantly dips up the glacier at the same time proves that these cores are built of glacial ice.

This conclusion is further confirmed by another phenomenon found in the ice cores. This is the presence of debris bands also upturned just as it is on the glacier surface in the zone of intensive ablation or in the ice cliff near Djupá river (Fig. 49). They dip up-glacier. The ice-core debris bands, as it should be deduced from other publications (Boulton 1967, Goldthwait 1951, Hook 1970) are a good argument in favour of the view that these cores are made up of glacial ice. There is no reason to suppose that the Sidujökull margin ice cores have been formed in any other way, for example, through recrystallization of the distal debris-covered snow banks, which fact was recorded in Swedish and Norwegian glacier marginal zones (Ostrem 1962, 1963, 1964) or through the incorporation of superimposed ice formed from wind-drifted snow wedges during the advance of an ice cap as is the the case in Greenland (Hook 1970). For in the mild maritime climate of Iceland winter snow patches resting in the Sidujökull marginal zone disappear very rapidly. There were only thin snow patches on the glacier slope at the beginning of the main part of ablation season. They disappeared before any noticeable release of debris from the glacier could start.

The debris bands in the ice cores add new material to the overlying morainic deposit cover. This causes growing thickness of the cover down the band and also strengthens the insulating influence of the cover on the ice core at the same time. Since the material in debris bands is quite loose, it undergoes an immediate incorporation into the very mobile cover. This does not create any conditions for the formation of secondary cones on an ice-cored moraine slopes which are known from the recently deglaciated areas of Spitsbergen (Boulton 1967) and Greenland (Hook 1970). The debris, mainly fine-grained, which is dispersed in the ice-core interior does not in a great extent enrich the thickness of the overlying cover on account of its small volume.

Origin and Development

Complex studies of the superglacial debris concentration process on the glacier margin, of the hillocks and ridges occurring near the margin together with their internal structure allows us to formulate a conception as to the origin of the ice-cored moraines, a conception which is valid, with one exception, for the whole investigated Sidujökull margin stretch. A particular case

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of an irregular ice-cored moraine formation does not fall within the scope of the general conception and will be discussed separately.

It has already been pointed out that the core-building ice in the hillocks and riges is of glacial origin. It means that the origin of the ice-cored moraines should be looked for in the zone upwards from the ice-toe. This zone is the line of contact between the ablation moraine cover and the dirty ice surface. It is here that the differentiation in ablation rate is most pronounced. Ablation is accelerated on the dirty surface and retarded under the ablation moraine cover (Boulton 1967, Goldhwait 1951, Hook 1970, Östrem 1959, Ward 1952). In consequence, an initial trough is formed on the glacier surface in the zone under discussion, parallel to the glacier margin (Fig. 50B). The trough gives rise to separation of the stagnant ice-body from the active glacier margin

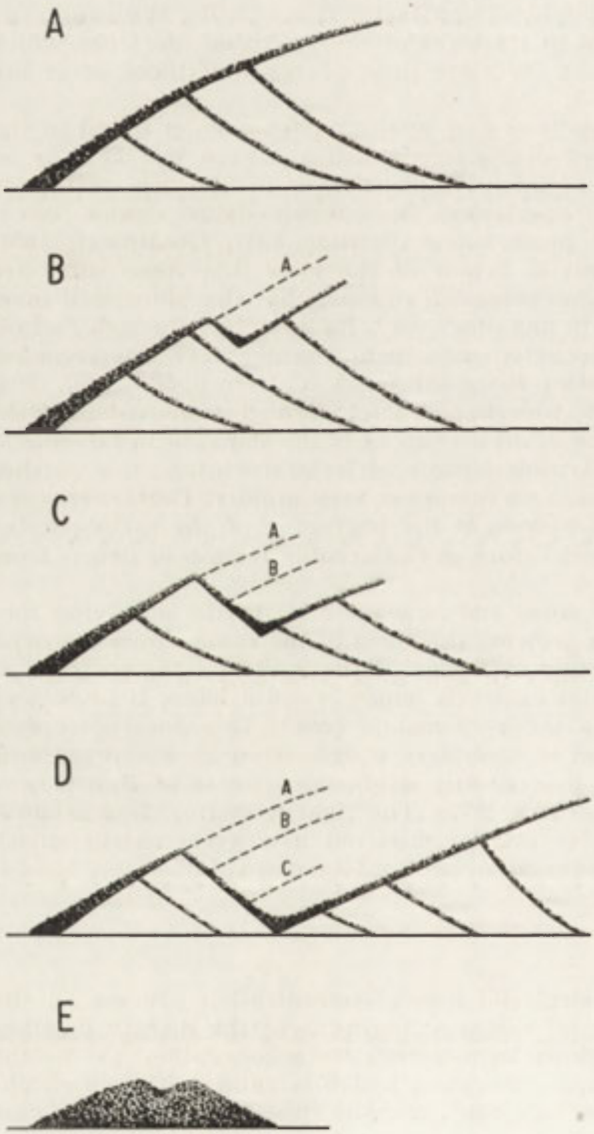

Fig. 50. Scheme showing the origin and development of ice-cored moraines at Sidujokull margin

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which, in the more advanced stages, will constitute an ice core. Formation of the initial trough brings about formation of the back-slope at the glacier side. Original differences in the ablation moraine cover: greater thickness down the glacier and smaller up it, is the reason for the back-slope back-wearing. Its crest-line moves together with it. The back-slope back-wearing is faster than the change rate in the external slope as long as respective thickness of the overlying covers is more or less the same.

Formation of the initial trough is the most important moment in an ice--cored moraine formation. For deepening of the trough leads to complete separation of an ice-cored moraine from the glacier front. The process of deepening is a complex one. Various decisive factors at work here are given from other glaciated areas where ice-cored moraines are now being formed (Goldthwait 1951, Klimaszewski 1960, Kolomits 1960, Szupryczyński 1963, Ward 1952). The most complete list of them is given by Goldthwait. Amongst them, as the most crucial for the object of our investigations we would like to consider on the basis of observations from various points of the Sidujökull margin, with certain restrictions though, the contrast in ablation rate between an ice-cored moraine and the black-ice slope, stream work and superglacial water ponding in closed hollows. The restrictions in reference to the first factor result from the fact that, as it has already been mentioned above, we should also take into account the black-ice slope back-wearing. Therefore, deepening and widening of the trough is determined both by intensive ablation in the black-ice slope as well as back-wearing of a forming ice core under the impact of ablation. Deepening and widening of the trough is impossible without the superglacial water work. This water, when flowing down the glacier slope in scattered fashion, forms into parallel to the ice margin streams. A stream formation phase is preceded with ponding because of the existence of hollows in the trough bottom. Dammed water accumulates heat and transmits it to the bedrock melting in this way the ice there. The bulk of work is done by the streams. Less active here is corrosion mentioned by Goldthwait (1951) because it is limited by a low debris friction coefficient on the ice surface in a flowing-water environment. Most significant here is debris transport away from hollows, erosion of the back- and glacier slope which conditions crawling of debris on them as well as thermal influences on the ice connecting the glacier--toe with an ice core. Melt water streams flow in the trough parallel to the glacier margin for a long distance before they can reach a gap in the ice-cored moraines. Therefore, the area on which melt water work and the geomorphic effects brought about by it are the greatest in the ice-cored moraine formation process. In the final stage of the trough development, when the ice sole between an ice core and the glacier margin disappears entirely, we often have to do with reponding (Fig. 51). But this does not any longer have any influence on the ice-cored moraines.

When there is a complete loss of contact between the ice-cored moraines and the glacier margin, their further development depends on the reduction process of the ice cores and gravitational movements of the ablational moraine cover released by this process. As the reduction of the ice cores advances constant redeposition of debris through sliding and falling occurs. Observations over these phenomena lead to the conclusion that measurements of longer axes of pebbles cannot yield any results in investigations over the influence of glacier movements on the forms under discussion and the terminal moraines associated with them. The mode of low terminal moraine ridge formation without any traces of pushing that occur in the ice-cored moraine

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Fig. 51. Ponding up of meltwaters between the ice front and ice-cored ridges. Photo by S. Kozarski

Fig. 52. Hollow with a low rim of ablation moraine. Photo by S. Kozarski

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phase is rather common on Iceland; that is why the results of the investigations over orientation of pebbles obtained from the Breidamerkurjokull terminal moraines by Price (1969) are so striking. It is not unlikely that fabric in those terminal moraines is a secondary phenomenon.

The final effect of ice cores disappearance in the hillock and ridge belts is deposition of debris on the bedrock (Fig. 40E). Debris also takes the form of a low, 3 to 5 m, sometimes even less, hillocks and ridges with a sharp, mostly double crest-line and numerous closed depressions. Disappearance of ice-cores does not last very long, but involves at least twenty-odd years. We base this conclusion on the fact that we still found in 1968 fresh traces of melting out of ice cores in the 1940–1945 end moraines of the 3rd belt east of the calderas. The process of melting lasts of course much longer if ice-cored moraines are very high, that is of ice cores are of greater thickness. This phenomenon occurs in the ice-cored moraines west of the calderas, on the ice-dammed lake, in the moraines that partly belong to the 3rd belt too, and still contain considerable and easily identifiable amount of dead ice.

In the case of ice-cored moraines on which the cover of ablation moraine was not thick enough to form in effect a morainal hillock, a concave form develops which is circled with a low rim of ablation debris (Fig. 52). The forms give an illusion of morphological features of disappeared pingos.

Occasionally, ice-cored moraines develop in a different way at Sidujökull margin than that suggested by the mechanism presented above. Separation of an ice block from the glacier front was brought about by the Djupá river due to the formation of a tunnel in the front. With time passing, the tunnel ceiling melted. We could watch the final phase of this process in July and August 1968. It was in this way that a near-the-front block of dead ice with distinct foliation and debris bands was formed. From the latter, debris abundantly emerged which formed the cover on the surface and low dumps at the foot of the ice block. It should be assumed, then, that further reduction of an ice-core formed in this way will not be conducive to formation of a distinctly shaped hillock or ridge, but to a complex of completely chaotically arranged end moraine hillocks. Probably, this mode of development of ice-cored moraines may explain the occurrence of single or small groups of irregular ice-cored moraines unrelated to the main end moraine belts in the recession area.

ACTUAL DRAINAGE SYSTEM OF THE GLACIER MARGIN

We found during our investigations that Sidujökull margin is drained in two ways: one superglacial, the other subglacial. The few englacial tunnels in the margin were out of function. Therefore, we should reckon also with this mode of margin drainage before 1968.

Superglacial drainage may be called dispersed. It involves a very rich system of small meltwater streams. In spite of their small size, the streams play an important role in the development of the marginal relief. Apart from forming the trough, and significantly influencing in this way the formation of the ice-cored moraines, meltwater is also responsible for the development of most short breaks in their belts, and later for the development of end moraines. The breaks, if advanced in their development, give rise to greater part of proglacial valleys.

The breaks are formed from the beginning of the development of ice-cored moraines. If there is a change in the flow direction from perpendicular to parallel at a part of a stream, a short hanging proglacial valley is formed that

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has a small outwash cone at its mouth in the foreland of the end moraines. This is an evidence for earlier centrifugal outflow of water. Bigger superglacial streams cut ice-cored moraines down to the bedrock and form proglacial valleys which often have a system of outwash cones or terraces on the foreland at their mouths.

Subglacial drainage is concentrated. Meltwater flows out of the tunnels in two ways: either gravitationally or under hydrostatic pressure (Fig. 53). Both give rise to wide rivers already directly at the ice margin. Particularly wide, oven over 60 m, are the rivers fed by the tunnel complex with outflow caused by hydrostatic pressure. From underneath the ice-toe, pillars of water squirt up to 1 m in height which have great erosive power. Hence, repeated observations at the place of particularly strong outflows, about 0.5 km east of the embayment in the southern margin, allowed us to ascertain strong lateral erosion and alterations of the stream bed and in the outflow direction at the

Fig. 53. Meltwater issue under hydrostatic pressure (artesian type ice spring) at the Sidujökull margin. Photo by S. Kozarski

same time. Ascertaining outflow under hydrostatic pressure or morphological effects of their activity is an important contribution to the discussion centred around the question of drainage of the last ice-sheet in Polish-German Lowland (Kozarski 1967).

Subglacial drainage system is responsible for major geomorphic changes in the marginal zone. Rivers of this system cause development of large gaps in the ice-cored moraines, and if they flow parallel to the margin for longer distances, they bring about total destruction of these forms through lateral erosion. This fact, more than anything else, explains long stretches without ice-cored moraines at the margin which are covered with outwash deposits and where abandoned channels of the braided melt water streams occur.

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West of the calderas line, the marginal drainage system is enriched with a new hydrographic element, a large ice-dammed lake. It lies in the lowering between the ice-cored moraines of the 2nd and 3rd belt formed back in 1920 and 1940–1945. The lake has features of considerable stability. It already had in its present shapeline, which we identified in 1968, back in 1960, which is shown in the air photographs. Its outflow channels, however, are transformed. The outlet in the western part is still active. It is a river, a tributary of the Hverfisfljot. Cutting of this tributary into outwash deposits and deepening of the formed gap in end moraines, which is evidenced by a terrace complex, caused lowering of the lake water level. Considerable series of lake deposits observable on its shores and lake terraces are an evidence for that. A further consequence of water-level lowering was the abandonment of the outlet in the eastern part. The large river, a tributary of the Brunnà, still seen on the air photographs taken in 1960, was already absent in 1968; while relief still preserved its dry bed very well.

Observations over the activity of melt water streams or over the changes in the drainage system enabled us to explain better and understand the influence of fluvial processes dynamics on the development of the marginal zone relief in Sidujökull recession area, that were discussed in a former chapter.

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REFERENCES

Baranowski, S., 1970, The origin of fluted morain at the fronts of contemporary glaciers, *Geogr. Ann.*, 52A, 1.

Bishop, B. C., 1957, Shear moraines in the Thule area, northwest Greenland, *U.S. Snow, Ice and Permafrost Research Establishment, Research Report*, 17.

Boulton, G. S., 1967, The development of a complex supraglacial moraine at the margin of Sørbreen, Ny Friesland, Vestspitsbergen, *J. Glaciol.*, 6, 47.

Boulton, G. S., 1968, Flow tills and related deposits on some Vestspitsbergen glaciers, *J. Glaciol.*, 7, 51.

Boulton, G. S., 1970, On the origin and transport of englacial debris in Svalbard glaciers, *J. Glaciol.*, 9, 56.

Bout, P., 1956, Les cones de poussieres des glaciers Islandais, *Rev. Geomorph. Dyn.*, 7–8.

Dyson, J. L., 1952, Ice-ridged moraines and their relation to glaciers, *Amer. J. Sci.*, 250, 3.

Flint, R. F., 1957, Glacial and Pleistocene geology, New York.

Gripp, K., 1929, Glaziologische und geologische Ergebnisse der Hamburgischen Spitzbergen Expedition 1927, *Naturw. Verh. Hamburg, Abh. Naturw.*, 22, 2.

Goldthwait, R. P., 1951, Development of end moraines in east-central Baffin Island, *J. Glaciol.*, 59, 6.

Heim, A., 1885, Handbuch der Gletscherkunde, Stuttgart.

Hook, R. L., 1970, Morphology of the ice-sheet margin near Thule, Greenland, *J. Glaciol.*, 9, 57.

Hoppe, G., Schytt, V., 1953, Some observations on fluted moraine surfaces, *Geogr. Ann.*, 35, 2.

Kjartansson, G., 1962, *Geological Map of Iceland*, scale 1:250,000, sheet 6, South-Central Iceland.

http://rcin.org.pl

Klimaszewski, M., 1960, *Studia geomorfologiczne w zachodniej części Spitsbergenu między Kongs-Fjordem a Eidem-Bukta* (Geomorphological studies of the western part of Spitsbergen between Kongsfjord and Eidembukta), Zesz. Nauk. UJ, 32, Prace Geogr., Ser. Nowa, 1.

Kolomits, E. G., 1960, Present periglacial landscapes of Novaya Zemliya, in: *Periglacial phenomena on the territory of the USSR*, Moscou.

Kozarski, S., 1966/1967, Origin of the subglacial channels in the North Polish and North German Plain, *Bull. Soc. Amis Sci. Pozn.*, Ser. B, 20.

Kozarski, S., Szupryczyński, J., 1971, Ablation cones on Sidujökull, Iceland, *Norsk Geogr. Tidsskr.*, 25, 2.

Lewis, W. V., 1940, Dirt cones on the northern margins of Vatnajökull, *J. Geomorph.*, 3, 1.

Lliboutry, L., 1964–1965, *Traite de glaciologie*, vol. 1–2, Paris.

Nielsen, N., 1937, A volcanoe under an ice-cap, Vatnajökull, Iceland, 1934–1936, *Geogr. J.*, 90, 1.

Nusser, F., 1935, Bericht über die österreichische Island Vatnajökull Expedition, *Mitt. Geogr. Ges. Wien*, 78.

Okko, V., 1956, Glacial drift in Iceland, its origin and morphology, *Acta Geogr.*, 15, 1.

Östrem, G., 1959, Ice-melting under a thin layer of moraine and the existence of ice cores in moraine ridges, *Geogr. Ann.*, 41, 4.

Östrem, G., 1962, Ice-cored moraines in the Kebnekajse area, *Biul. Perygl.*, 11.

Östrem, G., 1963, Comparative crystallographic studies on ice from ice-cored moraines, snow-banks and glaciers, *Geogr. Ann.*, 45, 4.

Östrem, G., 1964, Ice-cored moraines in Scandinavia, *Geogr. Ann.*, 46, 3.

Price, R. J., 1969, *Moraines, sandar, kames, and eskers near Breidamerkurjökull, Iceland*, Trans. Inst. Brit. Geogr., 46.

Quervain, A., Schnitter, E., 1920, Das Zungenbecken des Bifertengletschers, *Dankschr. Schweiz. Naturf. Ges.*, 55.

Raitt, W. L., 1962, A patterned marginal plain in Norway, *J. Glaciol.*, 4, 33.

Ray, L. L., 1935, Some minor features of valley glaciers and valley glaciation, *J. Geol.*, 43.

Schytt, V., 1959, The glaciers of the Kebnekaise-Massif, *Geogr. Ann.*, 44, 3.

Schytt, V., 1962, Notes on glaciological activities in Kebnekaise, Sweden 1962, *Geogr. Ann.*, 44, 3–4.

Schytt, V., 1963, Fluted moraine surfaces, *J. Glaciol.*, 4, 36.

Strøm, K., 1963, A patterned marginal plain in Norway, *J. Glaciol.*, 4, 34.

Swithinbank, Ch., 1950, The origin of dirt cones on glaciers, *J. Glaciol.*, 1, 8.

Szupryczyński, J., 1963, *Rzeźba strefy marginalnej i typy deglacjacji lodowców południowego Spitsbergenu* (Sum.: Relief of marginal zone of glaciers, and types of deglaciation of southern Spitsbergen glaciers), Prace Geogr. IG PAN, 39, Warszawa.

Szupryczyński, J., 1968, *Deglaciation in Southern Vestspitsbergen. Polish Spitsbergen Expeditions 1957–1960.* Summary of scientific results.

Szupryczyński, J., Kozarski, S., 1970, Relief of the marginal zone of Sidujökull (Iceland), *Bull. Acad. Pol. Sci., Ser. geol.-geogr.*, 18, 4.

Thorarinsson, S., 1938, Über anomale Gletscherschwankungen mit besonderer Berücksichtigung des Vatnajökullgebiets, *Geol. Foren. Stockh. Förh.*, Mai–Oct.

Thorarinsson, S., 1949, Some tephrochronological contributions to the volcanology and glaciology of Iceland, *Geogr. Ann.*, 31, 1–4.

Thorarinsson, S., 1964, Sudden advance of Vatnajökull outlet glaciers 1930–1964, *Jökull*, 14.

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Thorarinsson, S., 1969, Glacier surges in Iceland, with special reference to the surges of Bruarjökull, *Canad. J. Earth Sci.*, 6, 4.

Thoroddsen, Th., 1905/1906, Island, Grundriss der Geographie und Geologie, *Petermanns Mitt., Ergänzungsh.*, 152, 153.

Todtmann, E. M., 1957, Am östlichen Rand des Bruarjökull, Nordrand des Vatnajökull (Island, August-September 1956), *Neues Jb. Geol. Paläont.*, 7/8, Stuttgart.

Todtmann, E. M., 1960, Gletscherforschungen auf Island (Vatnajökull), *Univ. Hamburg, Abh. Auslandsk.*, Reihe C, 65, 19.

Thómasson, H., Vilmundardóttur, E. G., 1967, *Nakkrar athunganir vid Langasjó Reykjavik.*

Troll, K., 1925, Die Rückzugestadien der Würmeiszeit im nördlichen Vorland der Alpen, *Mitt. Geogr. Ges. München.*, 18.

Wadell, H., 1920, Vatnajökull. Some studies and observations from the greatest glacial area in Iceland, *Geogr. Ann.*, 2.

Ward, W. H., 1952, The physics of deglaciation of Central Baffin Island, in: The glaciological studies of the Baffin Island Expedition, 1950, *J. Glaciol.*, 2, 11.

Weertman, J., 1961, Mechanism for the formation of inner moraines found near the edge of cold ice caps and ice sheets, *J. Glaciol.*, 3, 30.

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MAP ANNEX

PHOTOGRAMMETRIC MAP OF THE FOREFIELD OF THE SKEIDARARJÖKULL, SHEETS 1, 2, 3 BY T. KONYSZ

GEOMORPHOLOGICAL SKETCH TO SHOW THE EASTERN PART OF THE SKEIDARARJÖKULL FORELAND. FIG. 2 TO THE ARTICLE BY K. KLIMEK

SKEIDARARJÖKULL MARGINAL ZONE, FIG. 1 A AND 1 B TO THE ARTICLE BY S. JEWTUCHOWICZ

DETAILED GEOMORPHOLOGICAL MAP OF THE PROGLACIAL AREA OF SKEIDARARJÖKULL TO THE SECOND ARTICLE BY R. GALON

HYDROGRAPHIC MAP OF THE FRAGMENT OF SKEIDARARJÖKULL FOREFIELD. FIG. 5 TO THE ARTICLE BY Z. CHURSKI

GROUNDWATER TEMPERATURES IN 1968 DETERMINED AT SELECTED PERIODS. FIG. 34 TO THE ARTICLE BY Z. CHURSKI

MAP OF THE SIDUJÖKULL. FIG. 3 TO THE ARTICLE BY S. KOZARSKI AND J. SZUPRYCZYNSKI

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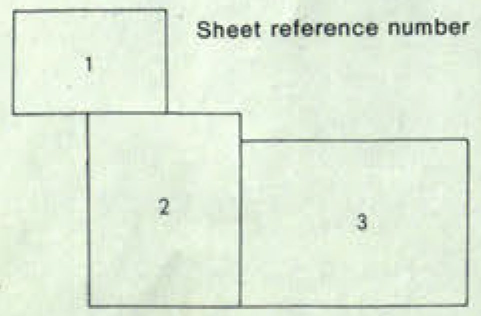
PROGLACIAL AREA OF SKEIDARÁRJÖKULL (VATNAJÖKULL)

Sheet 1



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Local system of geodetic axes x, y, z

Contour interval: 5 m (on glacier surface)

1.25 m (on glacier foreland)

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PROGLACIAL AREA OF SKEIDARÁRJÖKULL (VATNAJÖKULL)

Sheet 2

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Local system of geodetic axes x, y, z

Contour interval: 5 m (on glacier surface)
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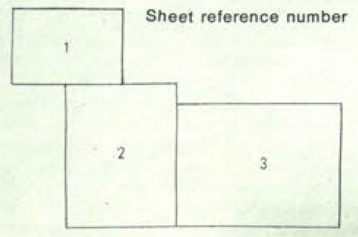
PROGLACIAL AREA OF SKEIDARÁRJÖKULL (VATNAJÖKULL)

Sheet 3



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26

MAP ANNEX

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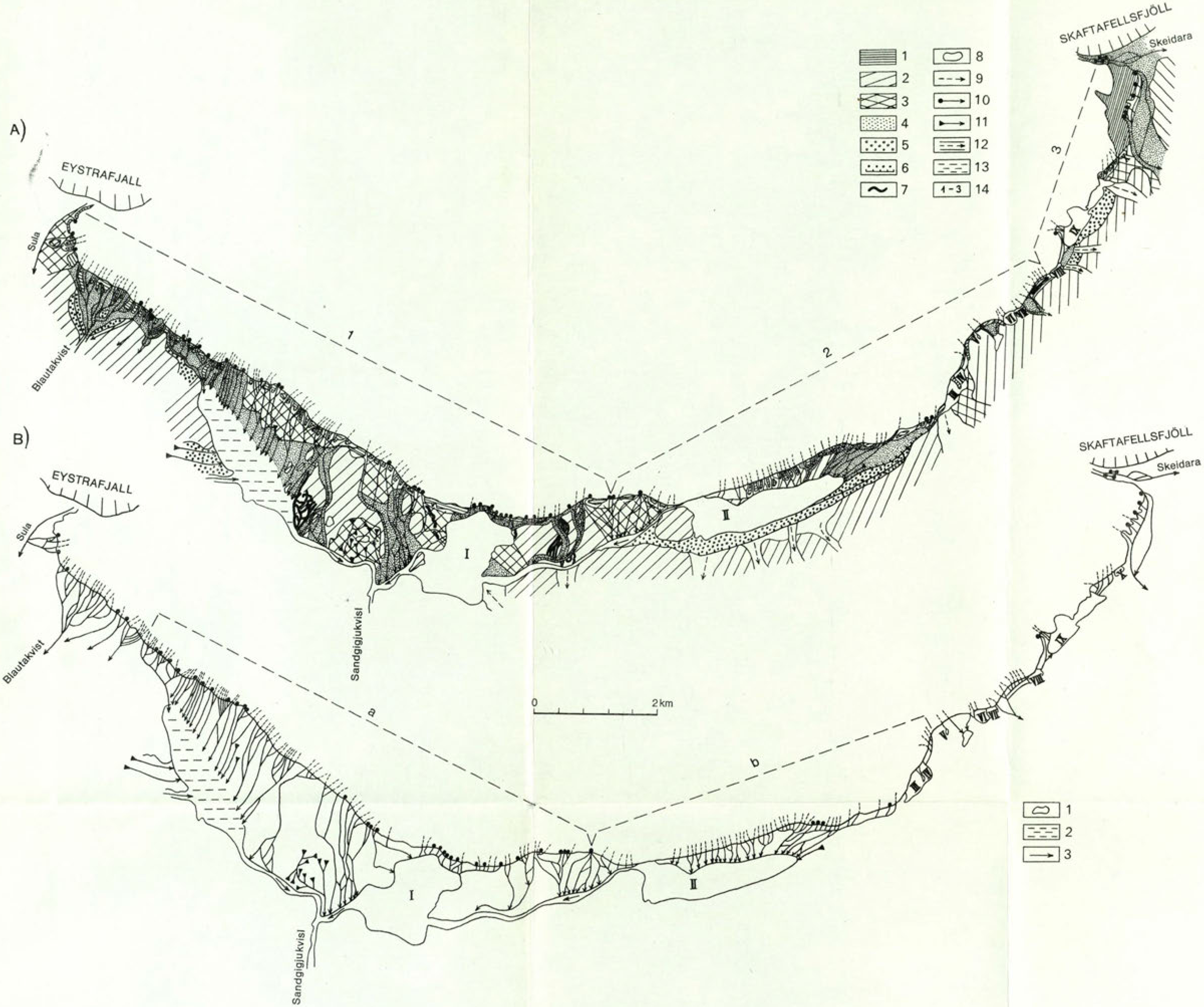

Fig. 1 A. Skeidarárjökull marginal zone

1 — ice-moraine ridge, 2 — older moraine, 3 — older and younger moraines, eroded, 4 — present-day sandur, 5 — older sandur, 6 — river terraces, 7 — eskers; 8 — lakes, 9 — glaciofluvial supraglacial streams, 10 — englacial streams, 11 — streams supplied with meltwater from buried dead-ice, 12 — abandoned older river valleys, 13 — ponded-lakes, 14 — numbers show segments of the marginal zone, 15 — numbers I-X indicate the marginal lakes

Fig. 1 B. Drainage of the Skeidarárjökull marginal zone

1 — lakes, 2 — ponded lake, 3 — rivers: (a) older segment of the meltwater channel, (b) younger segment of the meltwater channel, numbers I-X indicate the marginal lakes

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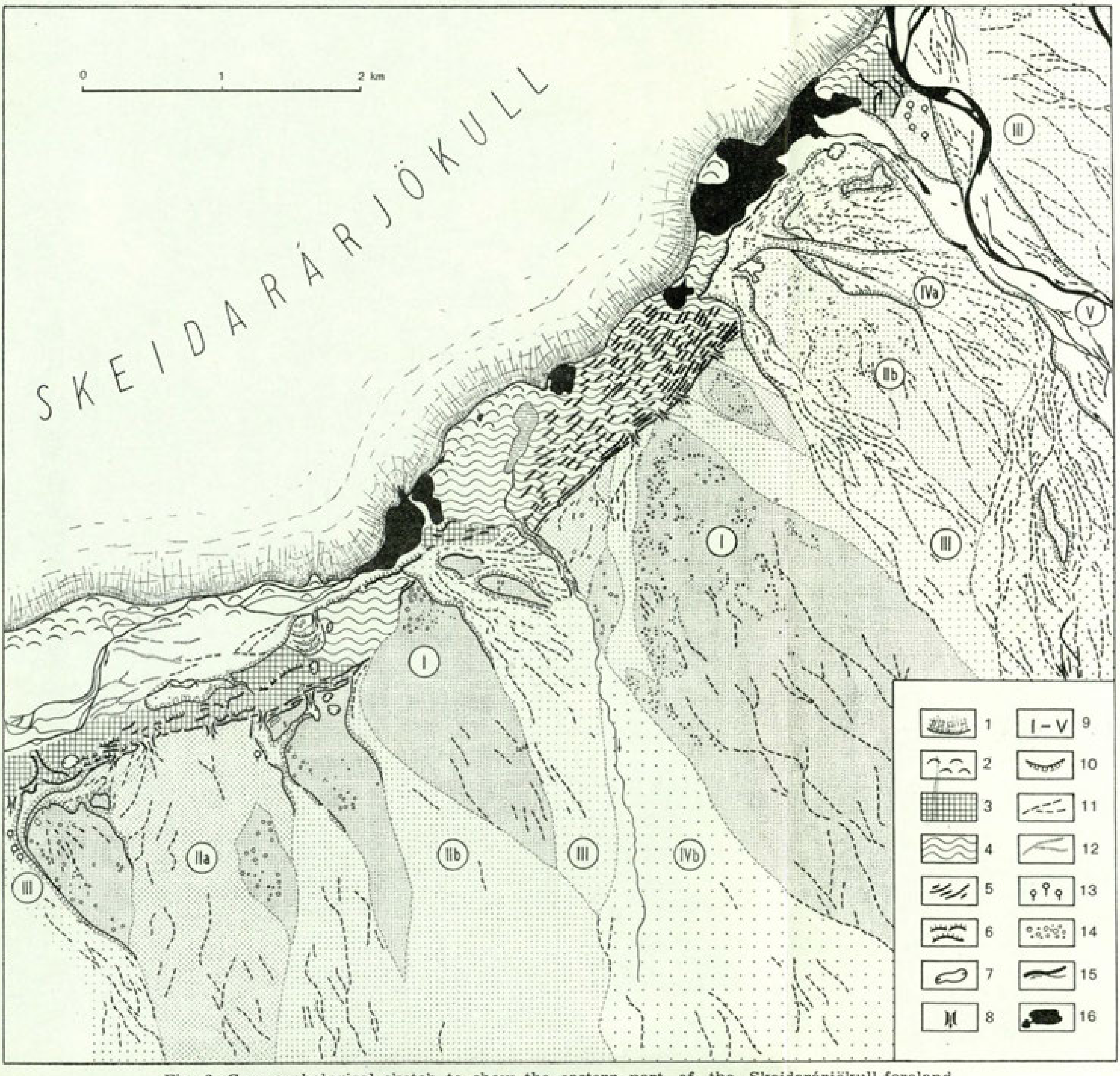

Fig. 2. Geomorphological sketch to show the eastern part of the Skeidarárjökull foreland
1 — ice margin in 1960, 2 — ice-moraine hillocks, 3 — hummocky end moraine, 4 — undulating end moraine, 5 — greater moraine ridges, 6 — ice-contact slopes, 7 — greater thaw basins in the end moraine zone, 8 — small outflows of meltwaters, 9 — glaciofluvial levels of different ages, 10 — meltwater cliffs, 11 — dry channels proglacial river with pavements preserved at the bottom, 12 — dry channels of proglacial river filled with sandy-silty materials, 13 — "gravel shadows" deposited downstream of ice-blocks in flowing water, 14 — pitted outwash plain, 15 — channels carrying water in 1960, 16 — ice-marginal lakes

Annex to the article by K. Klimek. Geographia Polonica 26, 1973

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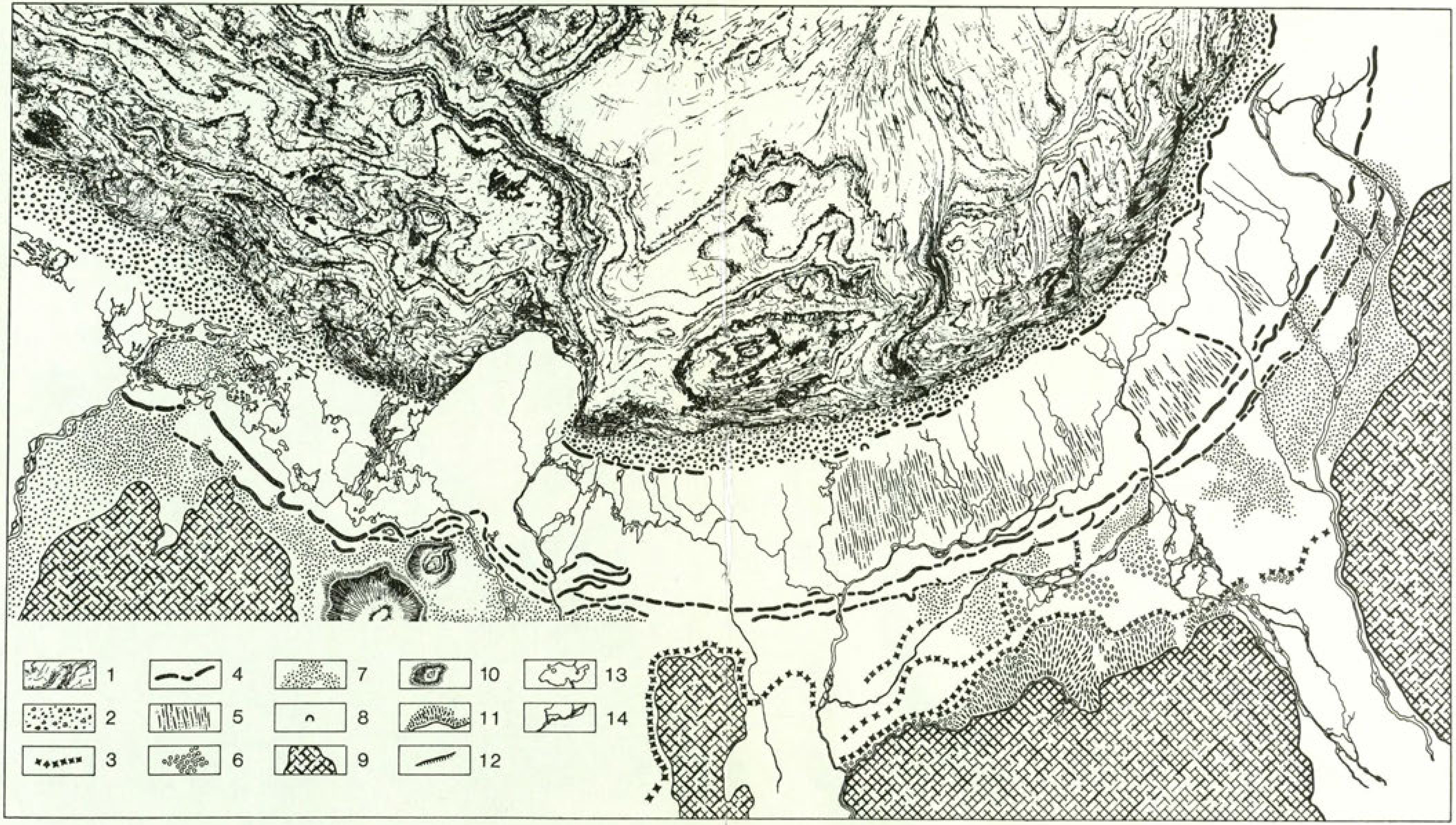

Fig. 3. Map of the Sidujökull marginal zone

1 — glacier, 2 — ablation moraine on the front of the glacier, 3 — first zone of the end moraines (push moraines) of 1890, 4 — end moraines and ice-cored moraines, 5 — fluted moraine, 6 — fluvioglacial deposits, connected with the first zone of moraines from 1890, 7 — younger fluvioglacial deposits, 8 — outflow of subglacial waters, 9 — basalt massifs, 10 — volcanic caldera, 11 — alluvial fans, 12 — glacier cliff, 13 — lake, 14 — drainage network

Annex to the article by S. Kozarski and J. Szupryczyński. Geographia Polonica 26, 1973

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Scale 1:26 800

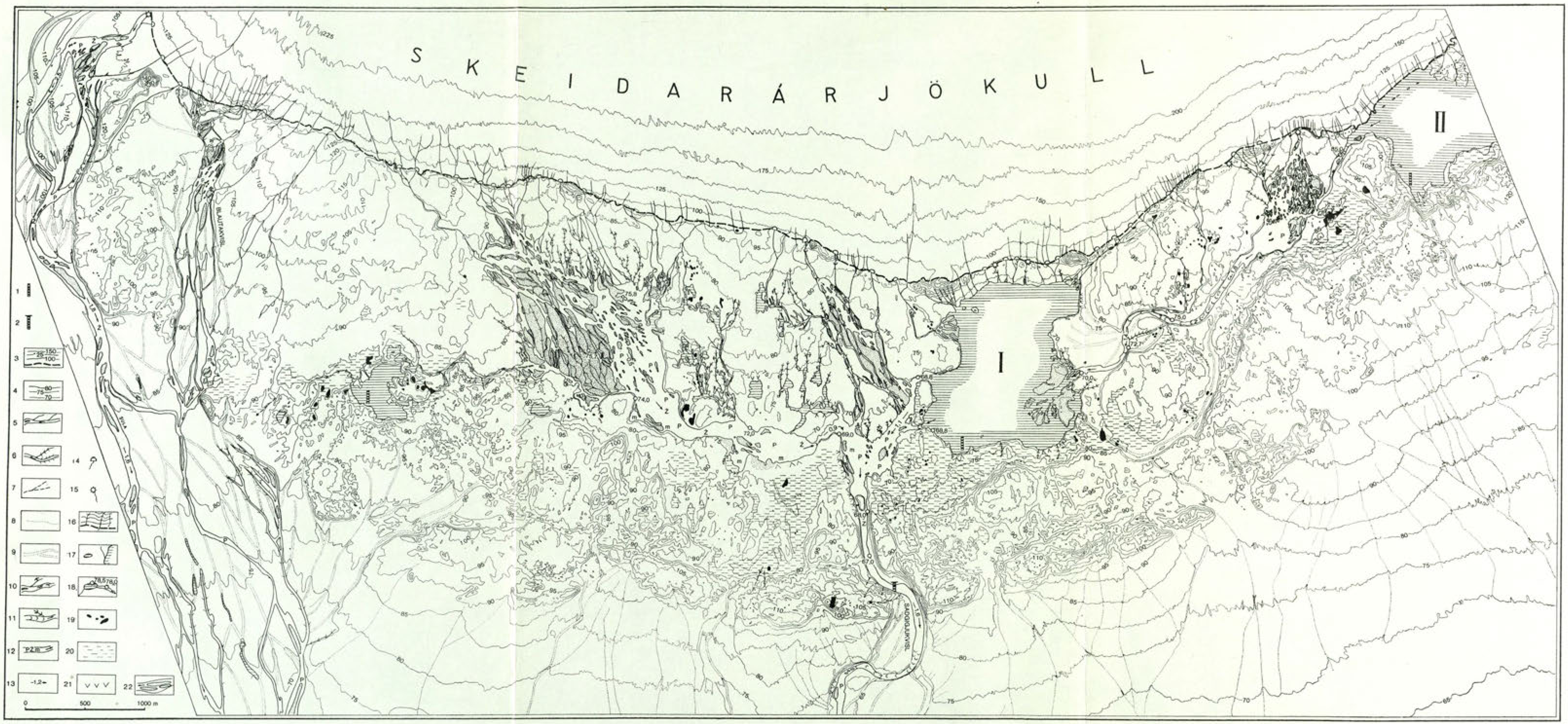

Fig. 5. Hydrographic map of fragment of Skeidarárjökull forefield

1 — river gauging station, 2 — limnograph, 3 — surface of glacier, 4 — contour lines, 5 — glacial streams, 6 — streams fed by groundwater outflow, 7 — periodic streams originating from melting dead ice blocks, 8 — periodic streams fed by water derived from springmelt, 9 — dry bottoms, 10 — stream with overgrown bed, 11 — rapids on the river-bed, 12 — character of the bottom: m — muddy, p — sandy, ż — gravelly, 13 — velocity of water flow in m/s, 14 — outflow of subglacial waters, 15 — pulsating outflow of meltwater streams, 16 — glacial streams on snout of Skeidarárjökull, 17 — lake and oxbow lake filled with water, 18 — height of mean water level a.s.l. in metres, 19 — small ponds filled in, 20 — terraces and bottoms of the disappeared lakes, 21 — leakage of groundwater with visible surface outflow, 22 — sand-gravel banks in glacial rivers and deltas

Annex to the article by Z. Churski, Geographia Polonica 26, 1973

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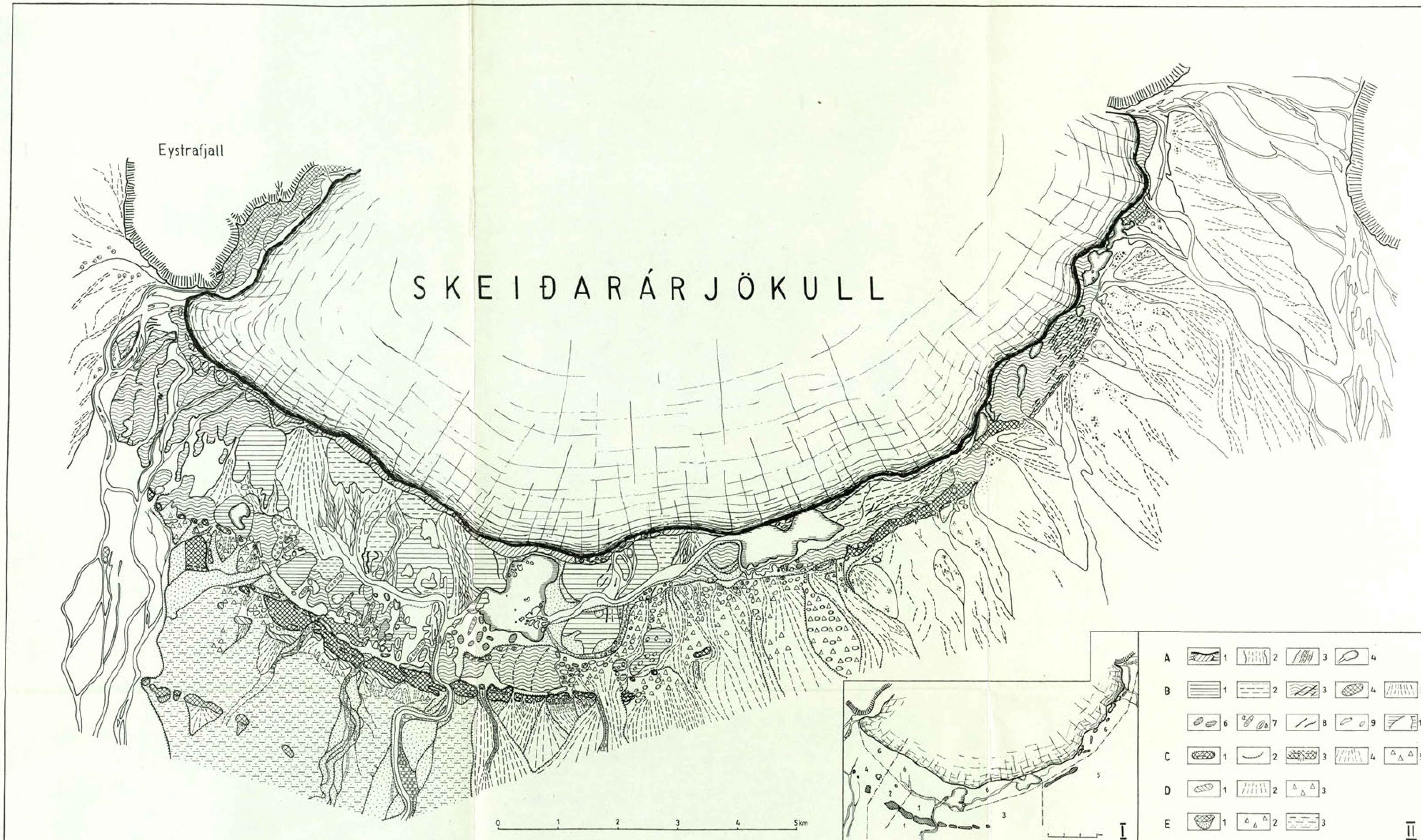

PWN Warszawa 1973

DETAILED GEOMORPHOLOGICAL MAP OF THE PROGLACIAL AREA OF SKEIDARÁRJÖKULL

ELABORATED BY R. GALON AFTER INVESTIGATIONS OF THE POLISH GLACIOLOGICAL EXPEDITION 1968 (SEE SKETCH MAP I)

I. Field work by: R. Galon (1), M. Bogacki (2, 3), K. Klimek (4, 5) and S. Jewtuchowicz (6)

II. Explanation of Signs:

(A) Youngest ice-marginal deposits and landforms:

1 — ice-cored moraine ridges and hummocks; 2 — outwashes; 3 — valley terraces and bottoms; 4 — proglacial lakes and meltwater streams

(B) Ice-marginal landforms from last fifty years:

1 — flat or undulant moraine plains; 2 — degradation plains; 3 — moraine hummocks, locally moraine ridges; 4 — moraine hills nd ridges; 5 — outwash plains; 6 — kames; 7 — crevasse fillinges (ridges): (a) gravels and boulders; (b) sand and silt; 8 — eskers; 9 — kettle holes; 10 — dammed lake bottoms (dry)

(C) Main end moraine zone from 1900—1910:

1 — end moraine hills and ridges; 2 — ice-contact slopes; 3 — moraine gaps; 4 — outwashes; 5 — accumulation of boulders (destroyed end moraines)

(D) End moraines and outwashes older than 1900:

1 — end moraine plateaus with small ridges; 2 — outwashes; 3 — accumulation of boulders (destroyed end moraines)

(E) Oldest glacial deposits and landforms (eighteen century):

1 — end moraines and outwashes; 2 — accumulation of boulders (destroyed end moraines); 3 — degradation and accumulation plain

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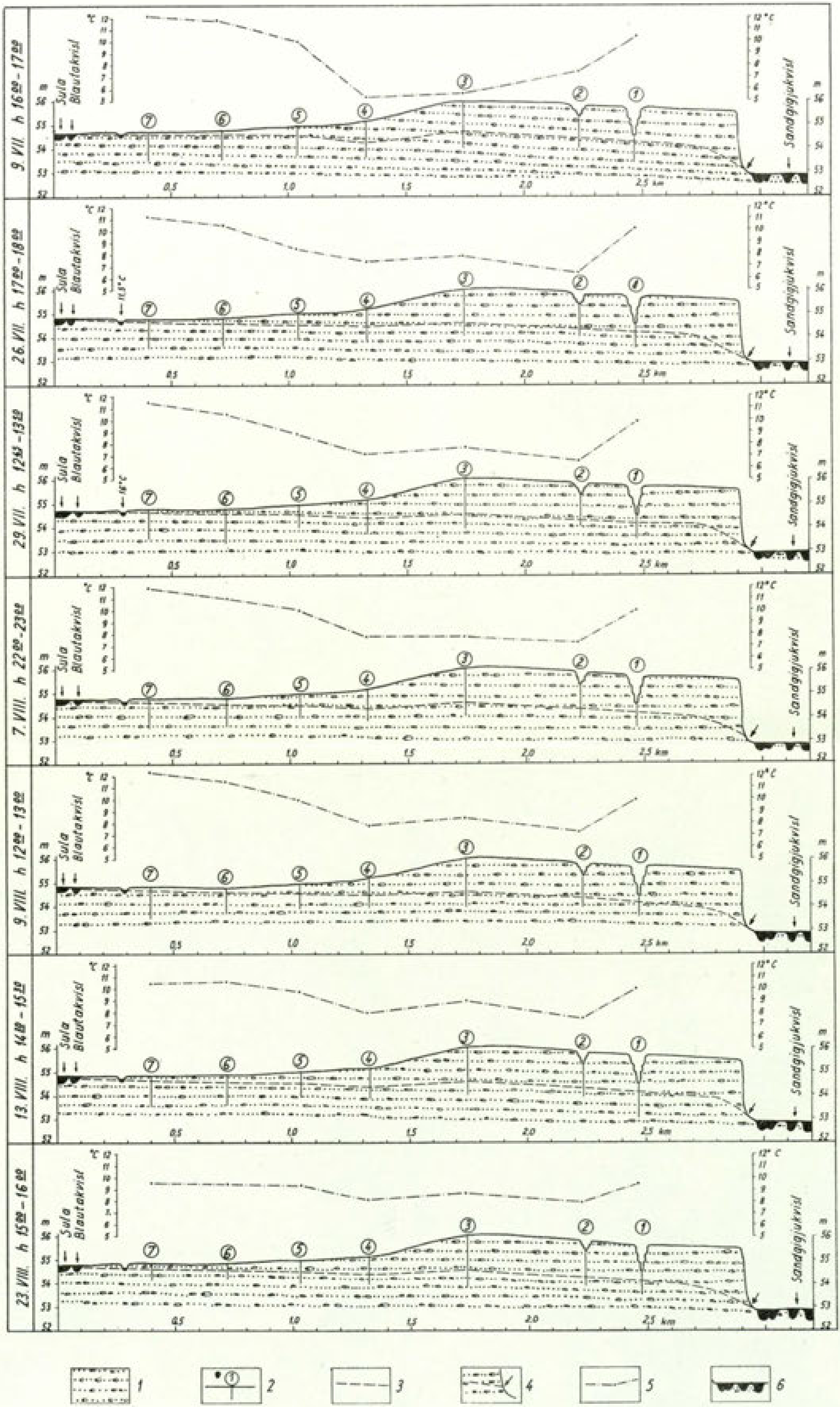

Fig. 34. Groundwater temperatures in 1968, determined at selected periods
1 — sand and gravel, 2 — groundwater measuring station, 3 — groundwater level, 4 — leakage of groundwater, 5 — groundwater temperature, 6 — rivers

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