Abstract

The author describes the physiography of the Prince Olav and the Prince Harald Coasts and discusses the geomorphic development of the coastal and the inland ice-free areas of the region based mainly on field surveys and air-photograph interpretation.

The Yamato Mountains are situated inland about 300 km away from the coast and consist of seven small massifs and some 20 nunataks with maximum elevation of 2490 m above sea level. They were once covered completely with the former ice sheet. The fall of the ice level revealed the glaciated topography including at least two cirques which had been formed by mountain glaciation prior to inundation of the former ice sheet. The trace of erosion by the wet-based ice sheet is found on a part of the bedrock. In the retreat of the ice from the highest to the present levels, three stages can be recognized; the Yamato, the Fukushima, and the Meteorite Glacial Stages for the maximum, intermediate, and near past to present standstills of the ice sheet. During the ice retreat outlet glaciers and detached ice masses fed partly by drifted snow modified the topography of the presently ice-free areas.

In the coastal region, the ice sheet retreated from the edge of the continental shelf prior to 30000 years before, leaving the glaciated ice-free areas on the present coast. Areal scouring of the former ice sheet produced the undulating surfaces with subsequent features controlled by geologic structures on the preglacially existed subdued topography. Selective linear erosion also cut the surface to produce precipitous cliffs and U-shaped valleys both in the ice-free areas and on the sea floor. Crooked glacial grooves and related minor topography suggest the erosion by the wet-based former ice sheet. Raised marine features after glaciation and radiocarbon dates give the information of crustal movement in time sequence.

1. Introduction

Some aspects of the Antarctic glacial geology and geomorphology: Much work has been done on the glacio-geological and geomorphological problems in Antarctica since the early stage of the Antarctic expedition history. In particular, survey of ice-free areas and ocean bottom surrounding Antarctica has progressed greatly since IGY of 1957–1958. However, smallness and sparse distribution of ice-free areas in the vast Antarctic Continent have prevented the elucidation of the Antarctic glacial history in a continental scale. Marine geological and geomorphological survey of the continental shelf and the continental slope also has been met with difficulty in many places, being hindered by sea ice.

The most convincing data on ice sheet fluctuation has been obtained from the McMurdo Sound region. PÉWÉ (1960) identified multiple glaciations for the first time by the geomorphological study in one of the "Dry Valleys" and Ross Island, and correlated tentatively these glaciations to the Quaternary glaciations in the Northern Hemisphere. After that, detailed studies of glacial geology including radiometric age measurements of volcanic rocks in the Dry Valley region provided the long history of fluctuation of the ice sheet and mountain glaciers (*e.g.*, DENTON *et al.*, 1970). Results are as follows: 1) The Dry Valleys are glacial trunk valleys of which main portions were formed by outlet glaciers from the East Antarctic plateau prior to 4 Ma B.P. Therefore, the ice sheet was bigger in the Tertiary Period than at present. 2) After that, four ice readvances into the valleys took place because of the rise of the plateau ice surface. But the extent of readvances became smaller successively from the older to the younger advance. A part of a valley floor remained ice-free in the last 4 Ma. 3) On the other hand, the Ross Ice Shelf expanded four times within the last 1.2 Ma when the grounding of the ice shelf took place, and invaded the Dry Valleys from the Ross Sea side. The latest expansion of the Ross Ice Shelf was contemporaneous with the Wisconsin Glacial in the Northern Hemisphere. 4) Mountain glaciers on high peaks, being maintained by local snow accumulation independently of the ice sheet, expanded three times during the last 1.2 Ma. But their fluctuation was out of phase with the ice sheet and the ice shelf fluctuations.

After these studies, finding of *in situ* Pliocene marine deposits in the Dry Valleys led to the new phase of Late-Cenozoic studies. Drilling of the sediments in the Dry Valleys (Dry Valley Drilling Project) provided the variable sedimentary sequence with intermittent erosional phases from the Mid- to Late-Miocene to the Recent (*e.g.*, WEBB and WRENN, 1977). The cored sequence shows first deposition in a deep fjord in the Miocene with subsequent erosion possibly caused by the crustal uplift and grounding of the Ross Ice Shelf, followed by shallow water deposition in the Mid-Pliocene further erosion, and then deposition in the Pleistocene from the grounded ice sheet and floating ice. From the topographic analysis of surface and subsurface profiles through the area west of the McMurdo Sound together with stratigraphic study of sediments and geological structure of the basement rocks it was inferred that the significant southward tilt of the Dry Valleys region and block movement took place in this area during the Post-Jurassic and the Pleistocene (WRENN and WEBB, 1977). Thus, the outstanding results of glacial geology have been obtained in this area on the basis of intensive research in a very advantageous geomorphic environment.

In other regions in Antarctica, however, glacio-geological and geomorphological studies have not been afforded by ample evidence, in particular, of chronology. In the Beardmore Glacier region, glacial drift was classified into two groups, one older thick lodgement till and several younger thin moraines (MERCER, 1972). The former till was attributed to a product by the wet-based ice sheet covering the Transantarctic Mountains. From the latter, three former stages of expansion of the Beardmore Glacier were identified, and correlated to the rise of the Ross Ice Shelf due to grounding. However, the correlation does not seem to be based on sufficient field evidence but rather on inference.

Circumstances for geomorphological investigations of regions far from the McMurdo Sound are much disadvantageous. In the Ellsworth Mountains, the origin and development of the present landforms speculated by geomorphological field work and air-photo interpretation are: 1) deformation and uplift of the mountains in the Jurassic, 2) fluvial erosion of the major valley systems, 3) the onset of mountain glaciation which has continued in the higher part of the mountains until the present, 4) continental glaciation due to inundation by the ice sheet on the lower part of the mountains, 5) partial deglaciation with a lowering of the level of the ice sheet by as much as 300 m, with only minor fluctuations up to the present (RUTFORD, 1972). Morainic deposits in the Queen Maud Land and the Mac. Robertson Land were surveyed in detail, and their lithology, degree of weathering and locations were interpreted in connection with fluctuations of the ice sheet (BARDIN, 1972, 1977). These studies provided important information for constructing the Antarctic glacial history. But the correlation of events with those in other regions of Antarctica is still problematical.

On the other hand, marine geological research in the ocean surrounding Antarctica brought much information on the onset and fluctuations of glaciation in Antarctica as well as on the paleo-oceanographic evolution of the ocean. A study of micropaleontology and sedimentology of the piston cores obtained by the m/s ELTANIN revealed the glaciation of the Antarctic Continent during the Eocene and the temperature trend in the Southern Ocean throughout the Tertiary (MARGOLIS and KENNETT, 1971). A further study on the DSDP cores elucidated the evolution of the Southern Ocean in connection with continental drift of the Gondwanaland, temperature fluctuation of sea water and change in ocean current (KENNETT, 1978). On the other hand, interpretation of ice-rafted debris in relation to the glaciation of the Antarctic Continent is sometimes ambiguous (WATKINS *et al.*, 1974, 1977). Therefore, the correlation between geomorphic development on land and sedimentary sequence on sea bottom awaits further studies.

Research on the geology and geomorphology of the continental shelf and slope is expected to provide much information on the glacial history of Antarctica. Topographic and geophysical investigation of the sea beneath the Ross Ice Shelf indicated that the sea bottom topography is characterized by a series of troughs and ridges, and geophysical survey suggests that the fundamental origin for sea bottom topography is caused not by glaciation but tectonic structure (ROBERTSON *et al.*, 1977).

Sparker and side-scan sonar investigation on the continental shelf of the eastern Weddell Sea revealed the submarine glacial deposits consisting of four major units and topography due to iceberg scours on rises (FOSSUM *et al.*, 1977). However, the correlation of these events with the coastal geomorphology of ice-free areas is left to future field work.

The above brief note on previous geomorphological and glacio-geological investigations in Antarctica indicates that much more information on regional geomorphology is necessary for the elucidation of the whole Antarctic geomorphic development and also of the regional characteristics in the present state of knowledge.

Background of the present research: The first reconnaissance flight of Lützow-Holm Bay was carried out by the Lars Christensen's Expedition (CHRISTENSEN, 1939), and a map on the scale of 1: 250000 was published (HANSEN, 1946) on the basis of oblique air-photos taken by the expedition. A part of Lützow-Holm Bay was also covered by air-photos taken in the U.S. High Jump Operation in 1947. However, no ground survey had been conducted before the Japanese Antarctic Research Expedition started in 1956.

In the First Japanese Antarctic Research Expedition (JARE-1) YOSHIKAWA and TOYA (1957) made a geomorphological survey on the Ongul Islands and the Langhovde ice-free area. They discussed glacial, periglacial and coastal geomorphology, which are concerned with 1) characteristics of glacial erosion of the ice sheet and extent of the former ice sheet, 2) crustal uplift after shrinkage of the ice sheet, and 3) periglacial phenomena such as exfoliation, formation of patterned ground, and wind erosion. This research provided the framework and the guidepost for the succeeding studies. During the first wintering at Syowa Station, TATSUMI and KIKUCHI (1959a, b) conducted geomorphological and glaciological surveys together whit the study of bedrock geology in the coastal area of Lützow-Holm Bay. An outline of glacial geology of the Langhovde, the Skarvsnes, and the Skallen ice-free areas was described. During JARE-2 and -3, some characteristics of sea ice were observed (MURAUCHI and YOSHIDA, 1959), and the thickness of the inland ice sheet was measured for the first time (NAGATA, 1961). In JARE-4, the thickness of the ice sheet was measured along the traverse route to the Yamato Mountains and geological and geomorphological surveys were conducted in the mountainous area (ISHIDA, 1962; KIZAKI, 1965). This was followed to some extent by the members of JARE-5. Glacial landforms of the Yamato Mountains and fluctuation of the ice sheet were discussed (YOSHIDA and FUJIWARA, 1963), and the morphology of the inland ice sheet was examined (FUJIWARA, 1964). In the coastal area, the topography and the sediments of raised beaches were analysed (MEGURO *et al.*, 1964), and glacial and periglacial landforms were discussed on the basis of field work in JARE-6 (KOAZE, 1964). In addition, the geological structure, movement of a glacier tongue, and some morphological characteristics of an ice-free island were examined by photogrammetric interpretation (NAKANO *et al.*, 1960; HATANO, 1961).

Characteristic features of submarine topography of Lützow-Holm Bay and off the Prince Olav Coast were also discussed, based on field surveys during JARE-3–JARE-5 and soundings from the expedition ship SOYA together with the data obtained from several kinds of the charts (YOSHIDA *et al.*, 1964). On the other hand, submarine geology and geomorphology were also investigated by analyses of sediments obtained from dredging and coring on board the expedition ship SOYA (SATO, 1964; KAGAMI, 1964; SATO *et al.*, 1965).

Since 1967, field surveys during summer seasons were carried out also in ice-free areas hitherto uninvestigated. Raised beaches, glacial landforms, periglacial phenomena, and saline and fresh water lakes have been studied extensively or intensively (YOSHIDA, 1970a, b; FUJIWARA, 1973; MORIWAKI, 1974, 1980; YOSHIDA *et al.*, 1975). Geomorphological observations on the occasion of geological survey of bedrock were also conducted during winterings in JARE-10, -11, -13 and -15 (ANDO, 1971; WATANABE and YOSHIMURA, 1972; ISHIKAWA, 1974). On the other hand, the detailed study of the submarine geomorphology has been proceeding since 1968 (FUJIWARA, 1971; OMOTO, 1975, 1976a, b; MORIWAKI, 1975, 1979). Depth-soundings have been conducted by an echo-sounder from the surface of sea ice during wintering (YOSHIDA, 1969). Sounding data by the expedition ship FUJI were also available, especially in JARE-9 when the ship could sail to the south near the Breidvågnipa ice-free area. After JARE-12, the ship could hardly sail into the bay, having been prevented by thick sea ice, so that sounding data by the ship have not been obtained since that time.

Besides the above-mentioned researches, many glaciological traverses have been carried out. These works are also closely related to the physiographic study of the region.

Objective of the present research: Geomorphological problems to be solved in the Prince Olav and the Prince Harald Coasts regions, as shown in the foregoing studies, are summarized as follows: 1) the role of the former ice sheet in producing the topography of deglaciated areas, 2) the mode and the age of fluctuation of the ice sheet, 3) the role of crustal movement and sea-level change in the geomorphic development, and their interrelationship, 4) current geomorphic agency such as the present glaciation and periglacial activity. Most of these problems are concerned with the Antarctic glacial history, and are difficult to be solved completely. To elucidate these problems, physiographic characteristics of the region are described and discussed in detail on the basis of the field work and air-photo interpretation.

2. Physical Setting

2.1. The outline of the land and sea around the Prince Olav and the Prince Harald Coasts

The Prince Olav and the Prince Harald Coasts form the easternmost part of the Queen Maud Land, from about 34° to 45° of east longitude. Japanese Antarctic station "Syowa" is located at the centre of these coasts in 69°00′S and 39°35′E. This region is characterized by the considerably large embayment of Lützow-Holm Bay bounded by the Riiser-Larsen Peninsula on the west and by the massive Enderby Land on the east (Fig. 1).

Physiographic properties of the Riiser-Larsen Peninsula are scarcely known to

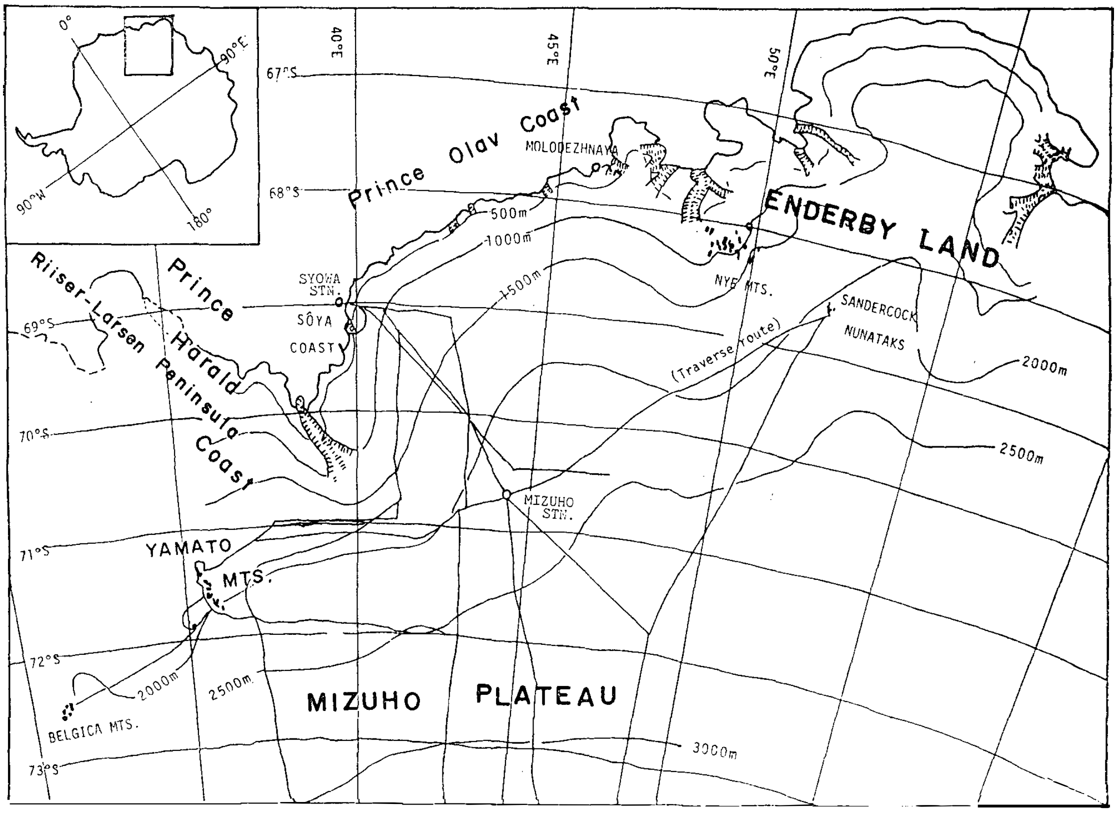

Fig. 1. Sketch map of the ice sheet from the Prince Harald Coast to the Enderby Land.

date. No ice-free rock exposure is detectable*, but a fairly clear crestline of ice can be traced on a LANDSAT image of the peninsula, indicating the steepness of slopes to the coast. A coastline west of the peninsula is more or less fringed by ice shelf. Some ice-shelf-like margins exist along the east coast of the peninsula. On the other hand, from the east coast of Lützow-Holm Bay to the Mac. Robertson Land the 1000 km long coastline is almost free from ice shelf. The Riiser-Larsen Peninsula occupies a transition zone from the ice-shelf-bound coast to the scattered-rock-outcrop-bound one.

The Shirase Glacier pours into the head of Lützow-Holm Bay and drains a considerable amount of ice (AGETA, 1971a; YOSHIDA, 1972; SHIMIZU *et al.*, 1978b). It exerts influence on the surface morphology of the ice sheet to some extent towards inland forming a convergent drainage basin behind Lützow-Holm Bay. The east coast of Lützow-Holm Bay, named the Sôya Coast, is characterized by the ice-free areas of relatively large size and several distinct outlet glaciers separating them.

To the east, the Prince Olav Coast turns its direction of extension, showing a somewhat straight coastline. Relatively small ice-free areas scatter along the coast, and small ice streams between them discharge ice from a rather divergent drainage basin (Fig. 2).

The ice-free areas are composed of gneissic and granitic bedrocks, their ages being Precambrian to Cambrian (TATSUMI *et al.*, 1964; MAEGOYA *et al.*, 1968). They were entirely buried under the ice sheet in its maximum stage, and subjected to glacial

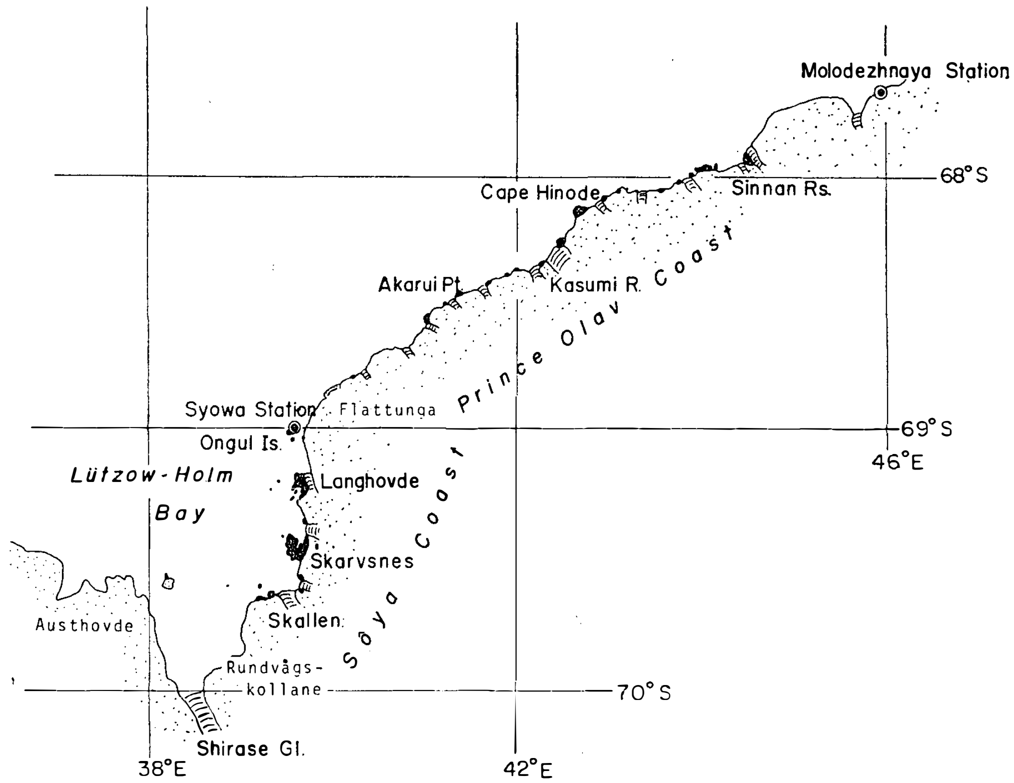

Fig. 2. Index map of the Sôya Coast and the Prince Olav Coast.

* A Russian chart "Karta Korennogo Reliefa Zemli Enderbi" (1977) indicates the presence of nunataks. But it may be erroneous.

erosion. After the retreat of the ice sheet they exposed their glaciated features, and since then, have been exposed to periglacial agency. Considerable parts of coastal areas possess raised beaches and/or marine terraces, indicating a regional uplift after shrinkage of the marginal part of the ice sheet.

Four inland ice-free areas, the Botnnuten, the Yamato Mountains, the Sandercock Nunataks, and the Belgica Mountains, were visited by the field party of JARE. All are supposed to have been covered with the ice sheet at its maximum stage. Among others, the Yamato Mountains, situated 200 km inland from the nearest coast, comprise many nunataks and provide tolerable information on the fluctuation of the inland ice sheet. The Glaciological Research Program was conducted in the Mizuho Plateau between the Yamato Mountains and the Sandercock Nunataks from 1969 to 1975. Physical properties of the present glaciation of the inland ice sheet were made clear to a greater extent (ISHIDA, 1978). Discovery of meteorites from the ice surface near the Yamato Mountains provided also some information on ice behaviour (NAGATA, 1978a).

The offing of the coastline is overspread by sea ice almost throughout the year. A flaw lead, popularly called "Otone Channel", appears north of the Prince Olav Coast at the end of December, its position being fairly coincident with the outer margin of the continental shelf. Southern limit of ice breakout (HEINE, 1963) during summer changes considerably from year to year in Lützow-Holm Bay. The breakout, however, rarely occurs south of 69°S, except a narrow polynya along the Sôya Coast between the Skarvsnes ice-free area and the Ongul Strait. Ice breakout along the Prince Olav Coast seems to take place more often than in Lützow-Holm Bay.

The condition of pack ice in summer is also highly variable from year to year according to weather. In general, the ice is heavy in the vicinity of Lützow-Holm Bay, which has a pretty large embayment. This heavy pack ice works not only for disturbing ship navigation but also for affecting marine agency near the coast.

Information on the nature of ice-covered sea is very limited because of technical difficulties. The current survey of submarine topography in Lützow-Holm Bay has revealed the existence of complicated drowned glacial landforms.

2.2. Climatic environment

Meteorological observation at Syowa Station has been continued over 20 years. Accumulating data enable us to outline the "mean state" of climatic condition in the vicinity of Syowa Station. Surface meteorological observation has been conducted also on the inland ice sheet during surveys of oversnow traverses and at the inland Mizuho Station (70°42'S, 44°20'E, 2230 m above sea level). Climatological data at Syowa and Mizuho Stations are shown in Table 1, though the period of record at the latter station is considerably short for describing the "mean state".

Characteristic features of the climate around Syowa Station, many of them were pointed out during the first wintering (MURAKOSHI, 1958, 1959), are as follows: 1) The annual mean temperature is the same as those in other coastal stations in nearly the same latitude, but the monthly mean temperature is a little high in autumn and low in spring compared with those in other stations, due probably to the broad pack ice zone during spring (MORITA, 1961), 2) even in the midst of winter, the highest tem-

Table 1. Climatological data of Syowa and Mizuho Stations.

Syowa Station

	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sep.	Oct.	Nov.	Dec.	Year
Mean air temperature (℃)	−0.9	−3.3	−6.2	−9.9	−14.1	−15.6	−18.5	−19.1	−18.7	−12.8	−6.6	−1.6	−10.6
Absolute maximum air temperature (℃)	9.5	5.5	3.6	0.4	−2.4	−0.7	−3.6	−3.9	−3.9	−1.6	3.5	8.1	9.5
Absolute minimum air temperature (℃)	−11.6	−17.0	−22.9	−29.1	−36.3	−36.3	−42.7	−39.6	−42.1	−29.5	−23.9	−12.2	−42.7
Mean humidity (%)	66	65	67	70	77	61	61	68	65	63	76	66	67
Mean wind velocity (m/s)	4.0	4.7	7.1	6.7	6.4	7.6	6.2	6.3	6.0	6.0	6.4	4.6	6.0
Absolute maximum wind velocity (m/s)	45.0	43.8	35.7	40.9	59.2	45.5	46.1	52.0	48.9	40.4	41.0	35.2	59.2
Mean cloud amount	6.3	6.9	7.7	7.1	6.5	6.8	6.3	6.8	5.8	6.8	6.8	5.7	6.6

Based on the results obtained during 1957–1975.

Mizuho Station

	Jan.	Feb.	Mar.	Apr.	May	June	July	Aug.	Sep.	Oct.	Nov.	Dec.	Year
Mean air temperature (℃)	−19.2	−24.2	−32.1	−35.6	−37.5	−40.4	−38.8	−40.3	−38.3	−33.6	−24.3	−18.5	−31.9
Absolute maximum air temperature (℃)	−4.8	−13.4	−18.3	−19.8	−19.0	−25.0	−18.7	−19.4	−19.4	−17.1	−7.1	−6.2	−4.8
Absolute minimum air temperature (℃)	−33.5	−37.8	−44.7	−49.3	−50.8	−57.7	−53.7	−54.7	−50.3	−49.2	−42.3	−31.2	−57.7
Mean wind velocity (m/s)	7.1	8.6	9.8	11.5	12.1	11.3	11.5	11.7	12.4	9.4	9.7	7.3	10.2
Absolute maximum wind velocity	14.5	16.5	20.5	21.0	24.5	18.0	25.4	22.2	24.5	24.0	18.0	16.0	25.4
Mean cloud amount	5.8	5.3	6.5	5.5	2.8	4.5	5.2	6.4	5.6	5.4	6.6	4.8	5.4

Based on the results obtained during 1977–1979.

perature exceeds $-10°C$, and high temperature during winter is closely related to high wind velocity, 3) wind direction is greatly controlled by the topography of the continent, and the prevailing wind blows from NE quadrant, 4) the number of stormy days is considerably large, and yet the mean wind velocity is relatively low compared with that in other coastal stations, due to the separation from the main land. High wind seems to be more frequent in the marginal part of the continent, as is indicated by wind facetted pebbles and dune sand in the Langhovde ice-free area. Sublimation of snow (and ice) is affected by humidity deficit and wind velocity (MORITA *et al.*, 1961), so that stable high wind may work effectively for the formation of a large bare ice (blue ice) zone of the ice sheet in the Sôya Coast.

High air temperatures exceeding 0°C are occasionally observed from November to February. Melting of snow, however, can take place in September at places sheltered from high winds in an ice-free area. Surface temperature of rock reaches frequently 30°C in midsummer. Therefore, the air temperature close to the ground and the surface ground temperature exceed 0°C at a considerably earlier stage in the annual march than does the normal surface air temperature. However, MORIWAKI (1976) showed that the frequency of freeze-thaw of the ground is not so high.

The climate of inland is more severe than that of periphery of the continent in many respects. In the marginal slope of the ice sheet the air temperature decrease is linearly proportional to the elevation increase (SCHWERDTFEGER, 1970). In the marginal slope katabatic wind is very remarkable and snow accumulation is generally high. In the vicinity of the Yamato Mountains, conspicuous melting of the ice surface does not take place except at specific places protected from high winds in the lee of ice-free areas. The ablation of ice, therefore, is caused mainly by wind action, as is indicated by the occurrence of the Yamato meteorites which are found sitting on the ice surface.

3. Geomorphology of the Yamato Mountains

The Yamato Mountains are situated at about 71.5°S, 35.7°E, 200 km south-southwest of Lützow-Holm Bay and were found in 1960. Since then geomorphological, geological and glaciological surveys in this and adjacent areas have been carried out. Most part of the ice-free area was covered by vertical aerial photographs taken in summers of 1969–1970, 1970–1971 and 1975–1976. Geodetic control surveying has been proceeding intermittently since 1973. Finding of the Yamato meteorites in 1969 (YOSHIDA *et al.*, 1971; NAGATA, 1975) and succeeding investigations (NAGATA, 1978b, 1979) are another aspect of the earth science of this region. Mechanism of meteorites accumulation on the ice sheet surface is closely related to the behaviour of the ice sheet around the Yamato Mountains.

The Yamato Mountains consist of many nunataks, rising 50 to 800 m out of the surrounding surface of the ice sheet. There is considerable difference in dimensions of seven larger nunataks and other smaller ones. Therefore, it should be appropriate to designate larger ones as massifs for description (YOSHIDA and FUJIWARA, 1963) (Fig. 3).

These massifs, with the heights of 2100 to 2400 m, extend over 60 km in the north-south direction, conformably with the trend of the major geological structure. The ice sheet flows from south-southwest to north and to west around and through the mountains. The large bare ice area near the mountains seems to indicate the stagnant nature of the ice flow. Measured upward motion of the ice sheet surface (NARUSE, 1978) may support this presumption.

No meteorological observations have been done during winter. But data obtained during the summer survey traverses have been accumulating since 1960 (YOSHIDA *et al.*, 1962; AGETA, 1971b; KOBAYASHI and NARUSE, 1975). Micro-meteorological condition near the mountains was observed and discussed in relation to formation and preservation of bare ice areas (KOBAYASHI, 1979). Meteorological data at Mizuho Station which is located inland at similar elevation above sea level and remoteness from the coast to those in the Yamato Mountains show that the mean annual temperature is −33°C and the absolute minimum temperature is close to −60°C. Even in the midst of summer, the daily mean temperature is around −10°C in the Yamato Mountains. Melting is rarely seen except in and around some parts of ice-cored moraines. This may be caused in part by cooling effect of strong winds. The occurrence of meteorites on the bare ice area of the ice sheet surface indicates that the

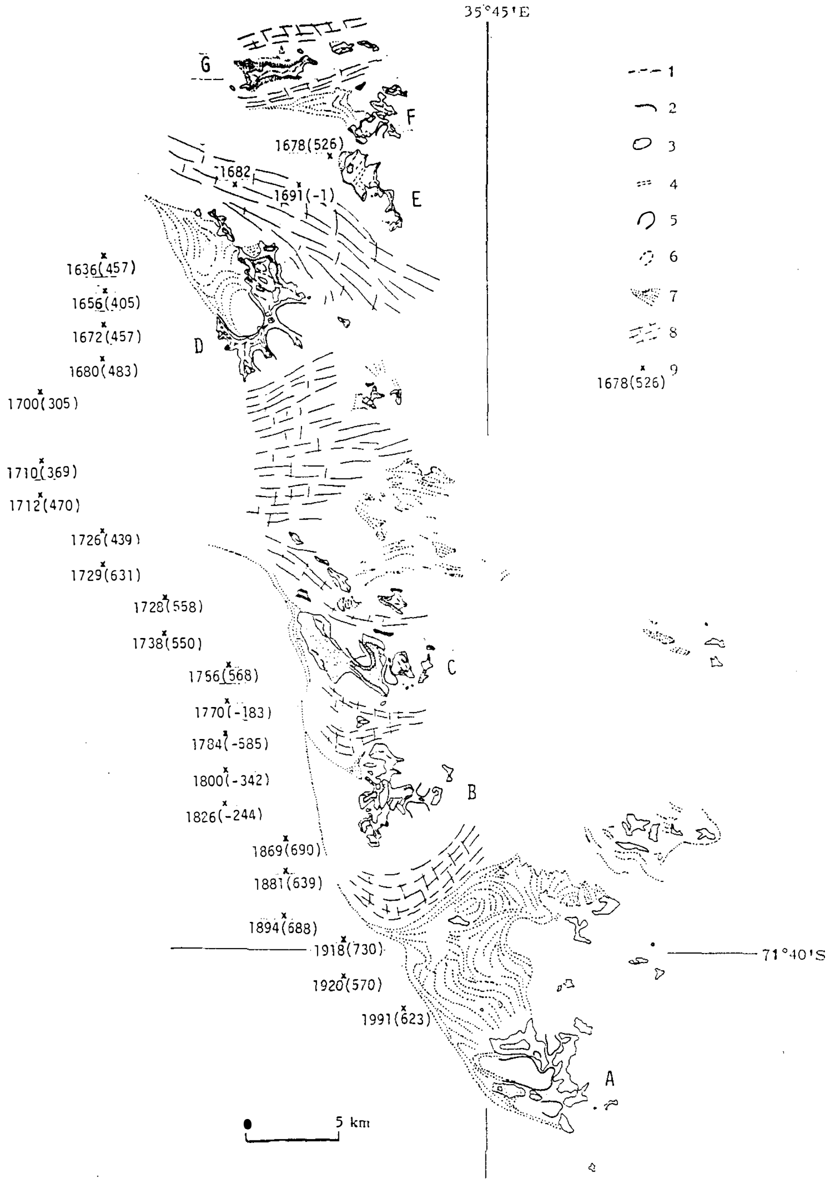

Fig. 3. Geomorphic map of the Yamato Mountains. (1; Main ridge usually associated with precipitous slope, 2; cliff, 3; flat surface and gentle slope with or without morainic deposit, 4; glacial trough, 5; cirque glacier, 6; cirque, 7; moraine on the ice sheet, 8; outlet glacier, 9; elevations of ice surface and subglacial rock surface in metre.)

winds prevent melting processes of ice such as the formation of cryoconite holes.

3.1. Bedrock geology of the Yamato Mountains and regional subglacial relief of Enderby Land and eastern Queen Maud Land

Bedrock geology: The bedrock geology of the Yamato Mountains was first investigated by KIZAKI in 1960 (KIZAKI, 1965). Geological surveys were also conducted in 1969, 1973, 1974, 1975–1976, and 1979–1980 (SHIRAISHI, 1977; YOSHIDA, 1979). The outline of bedrock geology is described according to these studies in relation to geomorphology.

The mountains are composed of various gneisses, granites and syenites together with aplite and pegmatite veins. Most of rocks are classified generally into two groups; a charnockitic group and a granitic group. The former was ascribed to the earlier metamorphism and the latter to the later one. Isotopic age determinations, though very few, showed that the later metamorphism took place around 5×10^8 years B. P. SHIRAISHI and KIZAKI (1979) designated this event as the "Queen Maud Orogeny", which is widely recognized in areas of the Eastern Antarctic Shield.

The trends of geological structures of large and small scales are generally conformable with the trend of extension of the mountains. The long axis of each massif is also concordant with the direction of foliation of gneisses. On the other hand, metamorphosed basic dikes and pegmatites are seen everywhere, cutting the foliation of country rocks. The thrust faulting from east to west is thought to have taken place concordantly with the trend of gneissic structure of country rocks at the later stage of the tectonic evolution. Smaller shear zones and joint systems cut obliquely or rectangularly the trend of foliation, giving the influence on the development of limbs of massifs and small nunataks.

The mountains have been believed to be geologically block mountains uplifted during the Tertiary Period (SHIRAISHI and KIZAKI, 1979). Each massif of the mountains has its peculiar geologic unit different from those in other massifs, suggesting the presence of a concealed fault underneath a glacier between two massifs. However, no evidence has been obtained as to the presence of faults related directly to the crustal movement. In this connection, and for the basic information of the region, subglacial topography of the area is briefly examined.

Subglacial topography: Data on ice thickness and subglacial relief of bedrock surface are still insufficient in this region for defining the reliable feature. KAPITSA (1964) made a survey traverse in the inland plateau behind the Enderby Land and the Princess Ragnhild Coast, between meridians 20°E and 110°E, and mentioned the presence of the subglacial mountains which continue to the Gamburtsev Mountains in the central part of East Antarctica (KAPITSA, 1964). This peculiar feature is shown as the Gamburtsev and the Vernadsky Mountains in the Atlas Antarktiki (1966) and reproduced in the Batimetricheskaya Karta Antarktiki (1974). KAPITSA has the view that the continuation of the Vernadsky Mountains extends to the Riiser-Larsen Peninsula, and the Yamato Mountains are situated on this "rooted mountains" of major geological structure in this region. Later studies, however, suggested that KAPITSA's view should be revised considerably. The existence of the Gamburtsev Mountains was confirmed by the radio-echo sounding but elevations of summits were reduced by about 800 m (ROBIN *et al.*, 1970). Data obtained by KOGAN (1969) and by ISHIDA (1970) are not coincident with KAPITSA's result. The present author (1976) has suggested that the Yamato Mountains extend not to the Vernadsky Mountains but rather to the Sør Rondane Mountains, on the basis of the meagre data concerning subglacial relief and disposition of the exposed mountain massifs. This seems to be supported to some extent by the revised Russian chart (Karta Korennogo Relief Antarktide, 1977). The idea that the Yamato Mountains constitute a part of the Vernadsky Mountains is still maintained in that chart. But the subglacial mountainous areas are detached from the Yamato Mountains and reduced in area, so that the use of the name Vernadsky Mountains may be inappropriate.

Radio-echo sounding at the ice sheet surface along the Japanese traverse routes provided passable information on the subglacial relief (NARUSE and YOKOYAMA, 1975; OMOTO, 1976a, b; SHIMIZU *et al.*, 1978a); some parts of the results and interpretation are problematical, as will be mentioned below.

OMOTO (1976a, b) summarized the subglacial topography based on measurements in 1969 and 1973 as follows:

1) Low-lying (near sea level) flat surface with somewhat deep troughs near the coast, at least as far as 120 km from the coast to inland. 2) Undulating surface up to 500 m above sea level to the south. 3) The existence of a deep valley continuing possibly to the Shirase Glacier which is the largest ice stream in this area (Shirase Subglacial Trough after OMOTO). 4) A mountainous area with the maximum height of about 1500 m above sea level around 72°S, 43°E (OMOTO considers this area as Vernadsky Mountains). 5) Broad low undulating surface (Mizuho Subglcaial Basin after OMOTO) between the Yamato and so-called Vernadsky Mountains. 6) Rather deep and rugged subglacial floor along the west side of the Yamato Mountains. SHIMIZU *et al.* (1978a) also dealt with the subglacial topography on the basis of almost the same data but including sounding in further northeast area, recognizing the difficulty in interpreting the record of radio-echo sounding.

Data by JARE are not always consistent with the Russian chart. MAE (1978) analysed multi-echoes of soundings, and showed that the deepest echo does not always indicate the reflection from the bedrock surface. Especially, most of the soundings were conducted at intervals of 2–5 km along the traverse routes, so that the interpretation of sounding records is more difficult than that of records obtained from continuous air-borne sounding. Thus, the depth of the so-called "Shirase Subglacial Trough" was calculated as shallower by 1500 m than before (MAE, 1978). It was shown that the corrected values of bedrock relief are fairly consistent with the distribution of free-air anomaly of gravity, though OMOTO (1976a, b) suggested the non-consistency of both trends at some places. Difficulty in interpretation of the JARE radio-echo soundings lies also on the insufficient areal coverage.

No reliable information is available on subglacial topography south to southwest of the Yamato Mountains. A broad bare ice area of the ice sheet south of the Yamato Mountains suggests the rise of subglacial rock floor. The Russian bedrock relief chart (1977) shows the rise of bedrock between the Yamato and the Belgica Mountains by inferred contour lines. But this chart seems to be not always accurate, for example, the existence of a very small and deep (*e.g.*, 1280 m below sea level) depression of inland is doubtful.

Evidently more data are necessary to draw a definite conclusion about subglacial topography and its relation to the major geological structure of the region in question. However, the following remarks can be made:

1) The Yamato Mountains are situated on the subglacial rise of rock floor extending from the Sør Rondane Mountains through the Belgica Mountains. The hypothesis that the Vernadsky Mountains together with the Yamato Mountains constitute a large mountain chain which indicates the major geological structure of East Antarctica should be reconsidered.

2) The large-scale topography of the Yamato Mountains has obviously been controlled by a geological structure including fault-lines. However, it is difficult to make it clear whether the crustal displacement due to faulting affected directly the shaping of the outline of the mountains. Similar features could result from selective erosion along fault-lines or other geologic structures susceptible to denudation. Difficulty in discerning the tectonic features from the erosional ones increases greatly when the terrain is widely covered with ice.

The Sør Rondane Mountains are said to be block-faulted mountains (VAN AUTENBOER, 1964). It has been considered that the upheaval due to faulting occurred in the Tertiary-Quaternary times, though the evidence is insufficient. Similarity in geology and in location on the East Antarctic Continent, and the topographic continuity between the Sør Rondane and the Yamato Mountains seem to support the view that the Yamato Mountains are the fault-block uplifted in the Late-Cenozoic time. Low elevations of the subglacial floor close to the western margin (OMOTO, 1976a, b) and northern margin (Russian map, see p. 13) also seem to support this view.

3.2. Some features of the ice sheet around the Yamato Mountains

The present author (1963) gave a preliminary description of the morphological properties of the ice sheet around the Yamato Mountains, referring to the general trend of surface slope, the distribution of bare ice areas, and flow characteristics. The Glaciological Research Program in the Mizuho Plateau during 1969–1975 provided considerable information on the ice sheet characteristics around the mountains (ISHIDA, 1978). According to this study, flow rates of the ice sheet east of the Yamato Mountains were between 0.2 and 21.3 m/year. It is a geomorphologically interesting fact that the rates decrease remarkably at a place about 30 km from the mountains and, at the same time, "positive" vertical displacement of the ice surface begins to occur towards the mountains (NARUSE, 1978). This displacement was determined by subtracting the thickness of net accumulation during four years from the difference in elevations between 1969 and 1973–1974. Therefore, positive value indicates uplift of the ice surface. This measurement has proved the existence of upward motion of the ice sheet on account of mountain obstruction. Exposure of internal moraines and concentration of meteorites on the ice surface are considered to have been caused by this type of ice flow and ablation (NAGATA, 1978a; YOSHIDA and MAE, 1978). The annual rate of upward motion of about 5 cm was almost the same as the annual amount of ablation in the bare ice area (by about 5 cm) near the mountains (NARUSE, 1975; SHIRAISHI *et al.*, 1976), which suggests that the mass balance of the ice sheet surface is in equilibrium at least in a short term. MAE and NARUSE (1978) also concluded that the thinning of the ice sheet is taking place in the Shirase drainage basin east of the Yamato Mountains, in spite of net positive balance of snow accumulation, and that the dynamic state of the ice sheet is unstable.

The distribution of bare ice areas on the ice sheet can be demarcated considerably clearly by the examination of the LANDSAT images (Fig. 4). The formation of a bare ice area is common around inland mountains, especially in the lee of them. A bare ice area indicates obviously the ablation area formed locally on the ice sheet, though the detailed mechanism of bare ice occurrence is not always clear. In the

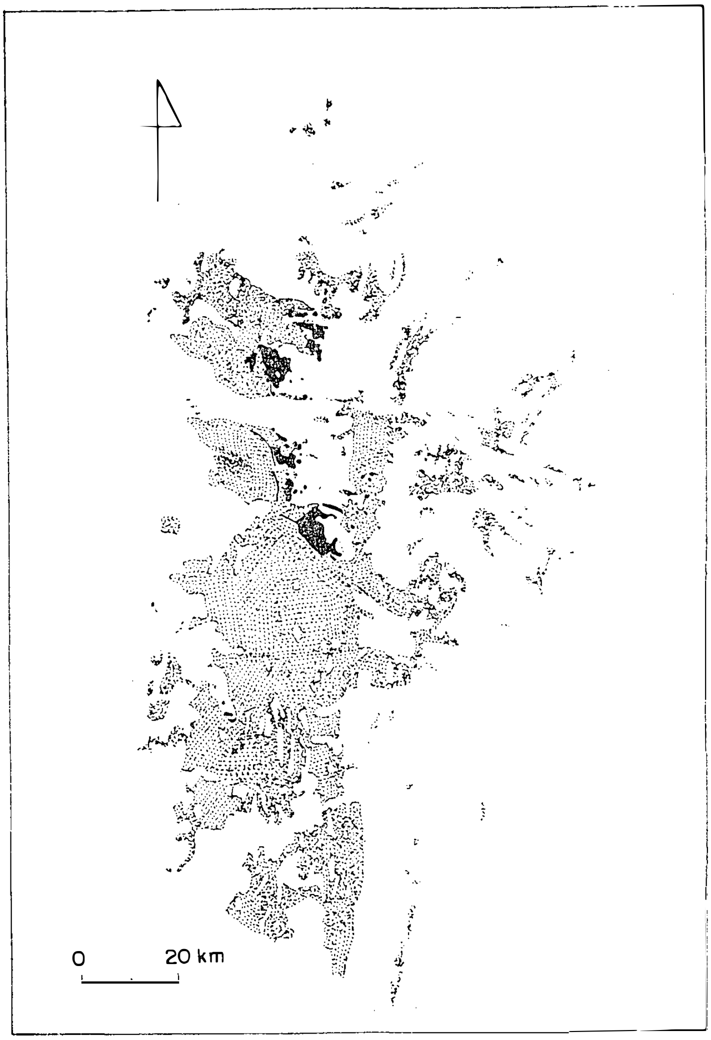

Fig. 4. Distribution of bare (blue) ice in the vicinity of the Yamato Mountains. (Drawn from the ERTS photograph taken on December 15, 1973. A dotted area and a blackened area indicate a bare ice area and an ice-free area including surface moraine, respectively.)

lee of mountains the snow accumulation is prevented by the topographic effect, and, as a result, ablation exceeds accumulation. As a rule, the direction of surface wind coincides roughly with the general trend of ice flow on the Antarctic Continent, because the dome-shaped topography determines the katabatic wind direction. Accordingly, the lee side area of the mountains corresponds to the downstream side of the mountain obstacle where the ice supply is reduced, and thus the formation of the bare ice area becomes active in that area.

Stagnation and upward flow of the ice in the upstream side of the mountains lead also to the formation of bare ice surface in some places. As stated above, it was found that the annual ablation of ice by about 5 cm was compensated with upward flow of ice near the Yamato Mountains. On the other hand, drifting snow accumulates in many topographically favourable places in the windward side of the mountain obstacles, though the snow layer is thin and discretely overlies glacier ice in places, suggesting the temporary nature of snow accumulation at these locations.

Micro-meteorological condition in bare ice areas differs from that in snow-covered areas. Turbulent flow of air and radiation effect on bare ice surface favour preservation of bare ice condition (KOBAYASHI, 1979).

In the region near the Yamato Mountains the ratio of bare ice areas to ice-free rocky areas is much larger than in other inland ice-free mountains. Bare ice areas occupy about 3250 km^2 of the region with the extent of about 125×70 km where the ice-free rocky areas of about 50 km^2 are included (YOSHIDA and MAE, 1978). This large group of bare ice areas probably extends to the south over 4000 km^2. The step-like topography developed on the ice sheet around the Yamato Mountains suggests the existence of the subglacial rise of bedrock, as is indicated by the crevassing on steeper slopes and several detached nunataks (Minami Yamato Nunataks and Kabuto Nunatak). Upward ice movement and little snow accumulation on steep slopes might have formed a very wide bare ice surface in this area, though direct evidence has not been obtained.

Based mainly on aerial photographic interpretation, the author distinguished minor features in the bare ice areas: 1) crevasse and linear structure, 2) furrow, 3) flow line streak, and 4) ice mound associated with scoop (YOSHIDA and MAE, 1978). Most of them seem to be formed in connection with the ice sheet flow. For example, the ice mounds associated with scoops which extend in a row over 20 km towards downstreams appear to have been originated in a heavily crevassed zone along the steep slope of the bare ice area in the upstreams, and then transported by ice movement (YOKOYAMA, 1976), altering their shapes under the influence of drifting snow.

The melting of bare ice surface rarely occurs in the Yamato Mountains region, probably because of low air temperature and strong wind which may cool the bare ice surface heated by insolation. Occurrence of a meteorite boulder, simply sitting on ice surface (YANAI, 1976), seems to support the above inference. Otherwise the boulder would sink in ice to some depth, forming a cryoconite hole. WATANABE and YOSHIMURA (1972) showed, by crystallographic study, that the surface bare ice in the vicinity of the Sandercock Nunataks in the Enderby Land is not superimposed ice but true glacier ice.

3.3. Glacial landforms

The glacial landforms of the Yamato Mountains have been described on the basis of field surveys in 1960 and 1961 (YOSHIDA and FUJIWARA, 1963). Since then many glaciologists and geologists have worked around the mountains. Aerial photographs and the LANDSAT images have provided the new information on morphological features of the mountains and the ice sheet. However, considerable parts of the YOSHIDA and FUJIWARA's study are still valid at present, and description and discussion presented here are based essentially on this study with minor modifications.

3.3.1. Outline of the topography

The mountains had been buried underneath the ice sheet at its "flood" stage, and subjected to glacial agency. The former ice sheet at least by 400 m on the eastern side and by 700 m on the western side of the mountains higher than the present level began to recede at unknown time. It is difficult to know the ice flow pattern at the maximum stage of the ice sheet, especially in the inland region. On the basis of glacial landforms of the Yamato Mountains, the glacial striation in the coastal ice-free areas, and the outer edge of the continental shelf the author tried to reconstruct the former ice level in the Prince Harald Coast region (YOSHIDA and MAE, 1978). The conjectured contour lines of the former ice sheet suggest that the former flow lines and the shapes of the drainage basins of the ice sheet were fairly different from those at present. The ice flow from southeast might have predominated over the Yamato Mountains region at that time.

In consequence of the emergence of the mountains from the ice sheet, the ice was transformed into outlet glaciers occupying depressions between massifs and nunataks. Local ice masses have been formed by separation from the ice sheet and by replenishment with drifted snow, as the lowering of ice sheet level proceeded. Outlet glaciers and detached ice masses have modified the configuration of the mountains, forming troughs, cirques, and local precipitous cliffs.

Morainic materials were also deposited in many places on and around the mountains. Some ridges are covered completely by morainic deposits. Several types of moraines can be seen on the ice surface, indicating the local flow patterns of ice. Morainic heaps have some characteristic features. Local periglacial environment is typical in sorted polygons and solifluction lobes poorly developed in places of ice-free areas. Patterned moraines are formed on degrading glacial ice. Many of these features suggest the differentiation according to time elapsed since deglaciation.

3.3.2. Landform of each massif

The Yamato Mountains are divided into three groups according to their configurations and locations for the convenience of description. They are the northern, the middle and the southern groups.

The northern group consists of D, E, F and G massifs. The common feature is the well-developed rugged topography, as is shown in many places by "aiguille" and "arete" topography. The G massif occupies the northernmost part and shows the most precipitous topography. The trend of the ridge which is fairly conformable with the trend of gneissic structure (N80°E) is somewhat different from that of a neighbouring massif, suggesting the existence of a structural boundary beneath the outlet glacier between G and F massifs.

Summit altitudes of about 2000 m of peaks in the G massif including that of a slightly isolated nunatak are very uniform, though flat surfaces are hardly seen. This level can be traced to the E, F and D massifs.

The F and E massifs are separated by a small outlet glacier of stagnant nature. Rather flat or gentle slopes bounded discordantly by precipitous cliffs occupy the upper parts of the massifs. The gentle slopes are partly covered with morainic and frost-riven materials. Erratic boulders are found also in the summit areas, indicating that the former ice sheet covered completely the massifs in the past. Moreover, ice-scoured surfaces are present in places at a level slightly lower than the summit level. This level, 2100 m$\pm$* above sea level, develops in many places and seems to be correlative with the summit level of the G massif.

The gentle slopes of the E and F massifs are clearly cut by small glacial troughs which are ice-free at present. Their trends are conformable with the direction of the main joint system. Ice-smoothed bedrock with glacial striae is seen on the bottom of the trough in the E massif. The western end of this trough was truncated by an outlet glacier between E and D massifs. These troughs were probably shaped by former small outlet glaciers during a receding stage of the ice sheet. The gentle slopes were also apparently excavated to some extent by cirque glaciation in some places. A typical cirque form with steep head wall and basin was not observed. But rather shallow and steep floors were cut into the gentle slopes.

Recent cirque glaciers and drift-snow ice, as detached or local ice masses, are exerting glacial action on the massifs to various degrees, according to their features such as location and amount of supply. For example, a cirque glacier facing to the southwest in the F massif exhibits rather typical surface features with accumulation and ablation areas separated by a firn line.

The D massif, named Mt. Fukushima, has the highest summit in the Yamato Mountains. The northwestern part of the D massif is considerably massive with somewhat gentle slopes bounded by steep cliffs. It is difficult to determine from direct evidence whether the summit was covered by the ice sheet in the past. However, a glacially smoothed flat surface exists at a shoulder close to the summit area at a level 50 to 100 m below the summit. The summit area also shows a dormant feature to be called nunakol (TAYLOR, 1922). From these facts, the D massif, hence the Yamato Mountains, seems to have been buried completely beneath the ice sheet some time in the past.

Rather flat parts of the gentle slopes seem to be divided into two levels. The higher and the lower levels are about 2300–2400 m and 2000–2100 m high, respectively. A fairly large (750$\times$1250 m) cirque develops between these two levels of gentle slopes, facing to the northwest. The cirque has no typical amphitheatre form. Rather shallow and steep bottom is surrounded on three sides by cirque walls which obviously had cut into the upper gentle slopes.

The southwestern part of the D massif is composed of the NE-SW trending and the NW-SE trending narrow ridges. The summit elevations of the "aiguille" peaks in the NE-SW trending ridge seem to be correlative to the 2300–2400 m level on the northeastern massif, and those in the NW-SE trending ridge to the 2000–2100 m level. In the latter, a small but clear flat surface exists on one peak summit. The NE-SW trend is conformable to the direction of gneissic structure, while the NW-SE trend is subsequent to the lineation direction of joint system and fractured zone.

These two ridges and the northeastern trunk massif constitute cirque walls of a northwest-facing peculiar cirque glacier (named Nizi-no-kubo). A small amount of ice and snow are supplied to the glacier by small hanging glaciers on the headwall and by drifting snow. But a bare ice nature over the considerably flat surface and a wide recessional surface moraine in the downstream of the glacier indicate predomi-

* This elevation was surveyed using the reference point which was determined by barometric altimeter; therefore, the absolute elevation is not so accurate at present.

nance of ablation. The wave-like pattern of moraine ridges bulging to the downstream might have been formed by ice flowage. However, the existence of a large amount of talus deposit covering the southwestern wall and continuing to the moraine field suggests that the ice movement in the cirque is very sluggish. It is difficult to consider that this cirque glacier was shaped by an outlet glacier flowing through the massif, judging from the presence of steep ridges with pinnacles as a headwall. The subglacial topography of the cirque is not known yet, but considerable parts of the lower cirque wall seem to be still concealed by the glacier ice in the cirque, judging from the topographic relationship between the cirque wall and the glacier ice. One probable explanation is that the prototype form of the cirque had been formed at the stage of mountain glaciation before the inundation of the ice sheet in this region. As the massif emerged from the ice sheet, the cirque has been modified to some extent by erosion of localized ice mass. Based on the results of radio-echo soundings of the ice sheet DREWRY (1975) suggested the development of valley glaciers antecedent to the ice flood in the subglacial Transantarctic Mountains and the Gamburtsev Mountains regions. This hypothesis may be applied more reliably to the Yamato Mountains region.

The middle group is composed of the B and C massifs. Their characteristic features are essentially the combination of gentle slopes and precipitous cliffs which abut on discordantly. In this group, however, gentle slopes are considerably predominant compared with those in the northern group.

The C massif together with a row of nearby nunataks in the north stretches in the direction of NW-SE, subsequently to the joint and fracture structures.

Gentle slopes are widely covered with morainic deposits including erratic boulders and frost-riven detritus. Patterned ground develops in places on these deposits. Two levels at 2000–2100 m and 1800–1850 m are distinguished in gentle slopes. The former is traced to the summits of nearby nunataks in the north. After the deglaciation of the ice sheet, the upper surface was modified in part by cirque erosion. The glacially polished cirque bottom with striation whose direction is quite different from the trends of the ice sheet flow clearly indicates that the former ice mass in the cirque scooped out shallow hollows on gentle slopes, though the cirque form is somewhat indistinct.

Two small cirque glaciers are situated on the upper surface. Both are being fed actively by drifted snow from the east. However, the lower part of one of them shows ablation surface and takes the shape of receding tongue, leaving thin terminal moraine on the ice surface continuing to an outlet glacier. It is difficult to know whether this glacier is shrinking at present or not. At least, however, considerable stagnation predominates in this particular area.

The lower surface constitutes a western broad ridge of the massif, and is separated by a narrow slope from the upper surface. It is fairly flat with the crest at the southwestern edge of the ridge and is covered almost completely with morainic deposit except a small peak projecting from the surrounding by 120 m. Conspicuous patterns such as furrow, pit and mound develop on the moraine heap. No particular bedrock topography is distinguished on this surface because of morainic cover.

A large cirque glacier, 750 m wide and at least 1250 m long at the foot of a cirque wall, cuts into the upper and a part of the lower surfaces. The surface of ice is flat and merges in an outlet glacier in the north. The feature is similar to the Nizi-no-kubo glacier in the D massif. However, a part of the ice surface encircled by a cirque wall is covered completely with morainic deposit, suggesting the stagnant or even degrading state of this ice. It is difficult to clarify the formation mechanism of this sort of cirque form at present. However, as in the case of Nizi-no-kubo, it is possible to consider that the cirque glaciation antecedent to development of the ice sheet formed the original cirque configuration.

Gentle slopes in the B massif are usually found at 2200–2300 m in height. The northern part of the main massif and two small nunataks in the east are widely covered with morainic deposits and frost-riven detritus with patterned ground. To the south the morainic cover is rather thin and weathered, glacially smoothed surface is exposed around the summit of 2380 m high. Rather gentle slopes of small areas are distributed in places on the western flank at a level of 2100 m high.

The western side of the massif is steep and a considerable part of it is occupied by hanging glaciers. A precipitous cirque wall also cuts into the western bedrock. Its situation is somewhat similar to that of the Nizi-no-kubo cirque glacier.

In addition, here, a small cirque glacier exhibits the receding nature, exposing polished bedrock in front of a glacier tongue. It is not known when the shrinkage took place. Local condition seems not to allow positive snow accumulation, resulting in recent reducing of this ice mass.

The southern group is the A massif and adjacent nunataks. The A massif is composed of many peaks separated by cirque glaciers or drift-snow ice, and shows complicated topography. Peaks have generally gentle slopes and are covered extensively with morainic deposits which were derived mostly from the inland area. A vast moraine field is also formed on the ice sheet north of the A massif. Sporadically exposed small bedrock in the moraine field suggests the shallow subglacial rock floor.

It seems difficult to distinguish the definite summit levels of gentle slopes at present. However, rather flat parts appear to concentrate at levels of 2400 and 2100 m in height. The steep cliffs develop mostly in very limited places where stream flow of the ice sheet is remarkable. Outlet glaciers only slightly affected the shaping of the A massif.

Cirque glaciers of various sizes and mass balance are seen as in the case of other massifs. A large glacier is thrusting its ablation area into the moraine field on the ice sheet, showing positive or zero mass balance, while a small tributary glacier leaves the clear terminal moraine in front of its snouts, indicating rather recent retreat due to negative balance.

3.3.3. Moraine and patterned ground

As was briefly described earlier, morainic debris covers some parts of the ice sheet and the ice-free massifs. Its characteristic features and topographic significance are mentioned briefly.

Moraines on ice sheet surface: Remarkable moraine fields occur on the ice sheet surface west of the F massif, northwest of the D, and north of the A massif. Morainic debris is composed of angular and rounded boulders, pebbles, and sand and silt.

Erratic boulders are very common in the debris. Complex flowline patterns are distinguished on moraine fields. These patterns reflect the ice flow of surrounding ice sheet, outlet, and cirque glaciers. Besides the flowline pattern, an area with many small hollows develops on moraine fields in some places. The hollows often contain frozen (almost all the year round) pools and drifted snow. This area may be designated as "pitted moraine". It is difficult to understand at present what factors caused the differentiation of "pitted" and "non-pitted" moraines. The thickness of morainic cover is usually very thin, and yet the cover is considerably effective for protection against ice wastage by insolation and wind ablation. The "pitted" pattern indicates that gradual wastage of underlying ice in this area proceeds more effectively than in the neighbouring flat areas. This sort of moraine fields is designated as "recessional" moraine. It is clear that the moraine-concentrated field with a flowline pattern is formed on the leeside of the massif where the ice movement is sluggish. The detailed mechanism of formation is not clear yet. Supply of materials, sluggish ice flow, and predominance of ablation are necessary conditions for formation. However, "recessional" nature of the ice sheet or outlet glaciers seems not always necessary. German term "Staumoränen" (KLEBELSBERG, 1948) may be more appropriate to describe them.

Morainic debris has been supplied basically from the basal till (ground moraine) brought by shearing of the ice sheet. Sporadically distributed moraines on the upstream side of the C massif may also be derived from the basal till transported to near the surface of ice. YOKOYAMA (1976) reported a flat surface with potholes on a nunatak belonging to the nearby Minami Yamato Nunataks. This seems to show that the wet-based ice sheet had produced this surface. MAE and NARUSE (1978) concluded that the ice sheet east of the Yamato Mountains is in the state of basal sliding, hence the bed of the ice sheet is probably wet. It is difficult to correlate these facts directly to the moraine field in the Yamato Mountains. However, it is possible that the wet-based condition was favourable at some time in the past for the formation of lodgement till which was supplied to the mountains in large quantities.

Moraines on ice-free areas: Moraines covering flat surfaces or gentle slopes of the massifs are essentially the same as those on the ice sheet. Morainic materials were derived mainly from the basal till of the ice sheet. Some frost-riven rock fragments were supplied from the nearby bedrock after deglaciation.

Moraine heap, designated usually as ground moraine, exhibits patterned ground in many places. The development of periglacial topography or phenomenon is common in ice-free areas of Antarctica, especially in coastal regions which are literally periglacial. Moreover, patterned moraines are often found even in mountains located inland at high elevations (*e.g.*, TRAIL, 1964).

Periglacial landforms are rather poorly developed in the Yamato Mountains, at least because of too low air and ground temperatures. The degree of development of patterned structure on moraines, however, differs with massifs on the whole.

In the A massif, most moraines are ice-cored, and thickness of debris is so thin that it can be described as "one boulder" thick (PRICE, 1973). The pattern on the morainic debris is small-scale hillocky unevenness of "pitted" features resembling those on the ice surface moraines. Therefore, patterned features in the A massif reflect directly the configuration of ice surface beneath the morainic debris.

Patterned ground develops relatively well on gentle slopes in the B massif. The morainic debris containing frost-riven blocks derived from nearly tor-like bedrock has the wavy surface pattern. This pattern consists of indistinct stone rings and stone stripes which shape solifluction lobes in some places. Horizontal sorting of rock materials is very poor, but slight vertical segregation is seen between upper coarser and lower finer materials. This patterned moraine has ice-core, but its surface pattern seems to be independent of the surface configuration of ice-core, though the thickness of the morainic debris is the order of several decimetres.

Patterned ground on the upper gentle slopes in the C massif is almost the same as but slightly less developed than that in the B massif. However, patterns on the morainic debris which covers the flat surface of the lower level are mostly flowline and pitted patterns as those in the A massif, suggesting that the pattern is controlled by ice-core features.

In the D, E and F massifs, morainic deposits over gentle slopes are rather thin and sparsely distributed compared with those in the B or C massif. In some places, thin debris covers bedrock without distinct ice-core. Indistinct stone circles were found on such debris free of ice-core.

The following inference may be drawn from the foregoing descriptions.

Difference in the amount of morainic debris derived mainly from the basal till between the northern group and the middle-southern groups reflects active ice flow in the north and sluggish ice flow in the middle-south in recent times and also in the past when massifs were buried in the ice sheet. This is also indicated by difference in landforms such as precipitous slopes and glacially polished surfaces. A long moraine bank which extends from the moraine field in the A massif to northwest of the C shows active ice sheet flow towards north on the west side of the mountains and sluggish flow of outlet glaciers in the A, B and C regions.

On the other hand, difference in the growth of patterned ground in morainic deposits on gentle slopes reflects difference in time elapsed since moraines have been exposed to subaerial condition and settled on each massif.

In the A massif, most of moraines on bedrock emerged from the ice at the latest stage as is indicated by patterned moraines expressing directly the wastage of underlying ice. In the B massif, moraines on the broad ridge have been subjected to periglacial process to some extent after shrinkage of ice from that area. Moraines on the upper gentle slopes in the C massif are similar to those in the B but less developed. But the morainic cover on the lower surface resembles those in the A massif, suggesting that the recession of ice was delayed some time in this area. In the northern group, the amount of morainic debris on bedrock is somewhat less than in other groups. Here, rather thin moraine together with frost-riven detritus covers bedrock in places without well-defined ice-core. Patterned features of morainic debris are less developed, though small stone circles are distributed sparsely. Among these facts, at least, greater wastage of ice-core seems to indicate that a period of time elapsed since deglaciation has been longer in the northern group than in other groups.

3.3.4. Some problems on glacial landforms and fluctuation of the ice sheet

Based on the description in the previous section, glacial landforms of the Yamato Mountains are discussed in relation to fluctuation of the ice sheet.

1) Early stage

The mountains had been buried under the ice sheet at its maximum stage, and subjected to ice sheet glaciation. Ice surface elevation above sea level and absolute age at that time are not known to date. Terrestrial ages of the Yamato meteorites found near the mountains were determined as the order of 10^4 years B. P. (NISHIIZUMI and ARNOLD, 1980). These dates cannot be related directly to the age of the former ice sheet, though systematic accumulation of such data may provide information on chronology of ice fluctuation in future. It is difficult to consider that the maximum elevation of the former ice sheet was much higher than that of the highest summit of the mountains, assuming that the uplift of the mountains since that time has been negligibly small. Because, even if the ice sheet expanded horizontally to the outer edge of the continental shelf at its maximum stage (it corresponds to about 4% expansion compared with the present state in the Lützow-Holm Bay region), increase of ice surface elevation inland would be considerably small, as was discussed by HOLLIN (1962) in other parts of the continent.

Nevertheless, a glacially abraded surface close to the highest summit in the D massif shows that the former ice sheet exerted areal scouring over the higher part of the massif. It is not well-known how effective is the erosion ability of an ice sheet. There was a controversy between the glacial erosion school of thought and the believers of the theory of glacial protection (COTTON, 1947). Recently, the geomorphological role of an ice sheet is considered to be both erosive and protective to the terrain according to the relief of an area, thickness and temperature of ice, etc. (*e.g.*, EMBLETON and KING, 1968). EVTEEV (referred to MARKOV *et al.*, 1970) estimated that the rate of lowering beneath the east Antarctic ice sheet was sufficient to have removed a layer of rock 1 km thick from the whole area. The estimated rate seems too large to apply to the Yamato Mountains region, because the remnants of preglacial topography are found in the region. It should be emphasized, however, that the areal scouring took place in some higher parts of the mountains beneath probably rather thin ice sheet. The great antiquity of the Antarctic ice sheet (*e.g.*, DENTON *et al.*, 1970; LEMASURIER, 1972) might have caused considerable glacial abrasion, even if the ice flow was quite sluggish. Wet-based glaciation may be another probable factor (SUGDEN and JOHN, 1976) for effective glacial scouring as is indicated by the glaciated surface with potholes in the Minami Yamato Nunataks, though evidence has not yet been obtained in the Yamato Mountains. The maximum stage indicated by the above glaciation is designated as "Yamato Glacial Stage".

The upper part of the mountains began to emerge from the ice surface at unknown time, as the former ice receded from this region. Receding of the Antarctic ice sheet would take place extensively on a regional scale, responding to climatic change and/or mechanical cause (*e.g.*, surging due to sea-level rise). Therefore, it might have affected the whole Yamato Mountains region. However, as is inferred from the difference in degree of development of patterned morainic debris, the northern group, D, E, F and G massifs, had been released from ice cover a little earlier than other groups, though the upper gentle slopes or flat surfaces with elevations of 2300–2400 m can be traced from the D massif to the B and the A massifs. It is possible to consider that the retardation of deglaciation was caused in part by sluggish ice flow around the middle and southern groups.

The upper gentle slopes including flat surfaces had been evidently subjected to areal scouring of the ice sheet. However, their undulating topography involving higher peaks is well-preserved. The ice sheet erosion moulded the gentle slopes to form ice-smoothed, partly rounded surface. It is difficult to obtain evidence indicating preglacial erosional process. However, it seems probable that the elevated gentle slopes had existed before inundation of the ice sheet. The fact that the peculiar cirque glacier in the D massif had been formed before the ice sheet buried the massif supports this interpretation. That is, the gentle slopes originated in the remnant of elevated undulating geomorphic surface formed by preglacial erosion.

2) Middle stage

In consequence of the emergence of the mountains from the ice sheet, distinct outlet glaciers were formed through depressions between massifs and nunataks, though some ice streams might have existed already before the emergence of bedrock. Glacial erosion by concentrated ice excavated troughs between massifs. Glacial corrasion of outlet glaciers together with frost shattering scooped out parts of the gentle slopes, producing precipitous slopes discordantly to gentle slopes.

During the lowering of the general ice level, small ice masses were formed by drifted snow or as relics of the former inland ice detached from the ice sheet. They occupied mountain slopes favourable to deposition of snow and ice, and gnawed the bedrock. SUGDEN and JOHN (1976) suggested that important cirque activity may not accompany the decay of an ice sheet in mid-latitude areas. Most of cirques in glacierized area have been thought to be formed before the glacierization of the area. In the Yamato Mountains region, however, some cirques affected obviously the gentle slopes glaciated by the ice sheet. The shallow and incomplete shapes of cirques in this area were caused by weak erosion due to severe climatic condition which does not allow intense frost shattering compared with that in lower latitudes. The present cirque erosion is evidenced by the formation of small terminal moraine in front of an existing cirque glacier and by the glacially polished bedrock adjacent to a glacier.

The small ice masses detached from the ice sheet surface are fed by drifted snow. Therefore, the regimen of such ice mass depends mainly on the drifting and deposition of snow. Various types of cirque glaciers are distributed in all massifs except the G in accordance with topographic situation which determines the accumulation-ablation relationship. Such relationship is related primarily to the surface level of the ice sheet and secondarily to the aspects of cirques in connection with wind direction, because drifted snow is supplied almost exclusively from the ice sheet surface. A cirque glacier in the east side of the D massif consists entirely of accumulation area. Cirque glaciers in the E, C and A massifs have upper accumulation and lower ablation areas with or without receding tongues. Cirque glaciers in the C and the D massifs which might originate in preglacial cirques are almost free from snow accumulation.

The relationship between the level of the ice sheet and the formation of small ice masses is not simple. It is safely said that the positive or balanced glacier regimen for small ice masses cannot exist for a long period of time after the ice surface lowered to some extent. The cirque glaciers in the mountains surrounded by the ice sheet change their balance in relation to the ice surface elevation.

It is difficult to know correctly whether a standstill existed during the retreat of ice from the maximum stage to the present. However, some vacant cirque cutting into gentle slopes seems to represent a standstill. In the D massif, the northwest-facing cirque slightly scooped out the edge of the upper surface and the lower surface. This topography suggests the following event: The ice sheet surface had receded to some extent, approaching to the lower surface after scouring of the upper surface. After that, the retreat had ceased for a certain period. During this interval, a cirque glacier was formed by drifted snow on a steep slope and a cirque was gouged. The ice sheet began to retreat again and the lower flat surface emerged from the ice. The cirque glacier survived for a certain period of time and scooped the lower surface slightly. As the ice retreat proceeded, extinction of snow supply resulted in disappearance of ice in the cirque.

A small glacial trough in the E massif which excavated the gentle slope might have been formed also at this stage. Ice flow into the glacial troughs in the E and the F massifs might have diminished during that time.

The significance of a standstill discussed above is not always clear in the history of fluctuation of the ice sheet. The level of 2000–2100 m high can be traced almost throughout the mountains, with peaks and benches having traces of sheet flow scouring. A pattern of flow direction changed considerably from that at the first stage, as is indicated by a small trough in the E massif.

These facts suggest that this standstill of the retreat was significant in the general fluctuation of the ice sheet in the Yamato Mountains region.

Therefore, "Fukushima Glacial Stage" is proposed on this standstill, as a hypothetical framework for further comparison of general fluctuations with those in adjacent mountainous regions. A "glacial stage" should represent usually the major advance of the ice. It is difficult to ascertain the existence of expansion of the Antarctic ice sheet during receding from the maximum to the present, especially in inland areas. However, the author believes that even discrimination of the standstill is meaningful in the present stage of knowledge.

3) Later stage

As a result of retreat of the ice sheet to the present level, outlet glaciers have become more confined to channels between ice-free areas, exposing the upper parts of precipitous rock walls of troughs. Radio-echo sounding on the outlet glacier between the E and the D massifs shows that the ice is about 1500 m thick (SHIMIZU *et al.*, 1978a; OMOTO, 1976a). Cirque glaciers might have responded to recent accumulation-ablation relationship, changing their configurations.

Extensive surface moraines on the ice sheet were formed mainly by ablation of basal ice upthrusted to the surface caused by stagnation of ice movement. Judging from the preservation of flow line and pitted patterns and from the altitudinal situation which is very close to the ice sheet surface, the morainic deposit on the lower surface of the C massif seems to have been formed in the same or a little earlier period than the period of ice surface moraines.

Glacier ice occupying the preglacially formed cirques in the D and the C massifs shrank to a level surface concordantly with the ice surface.

Measurements of upward movement and surface ablation of bare ice area indicated that the ice surface elevation is in equilibrium at least for a short time. The formation of moraine fields with patterned features probably required a certain period of time from the near past to the present. The extensive bare ice area carrying meteorite fragments on its surface (Meteorite Ice Field) south of the Yamato Mountains suggests a standstill or slight recession in ice sheet fluctuation.

Therefore, it seems appropriate to designate tentatively the formation of supraglacial moraines on the ice sheet together with the formation of the Meteorite Ice Field as "Meteorite Glacial Stage". The mode of occurrence of moraines is similar to that of Insel I Stage (BARDIN, 1972) at Mt. Insel in the Wohlthat Mountains, where patterned moraines with ice-core are distributed from 0 to 50 m above the present ice level.

In the Mt. Insel area (71°28′S, 11°30′E), Insel I, II, and III Stages (from late to early) have been distinguished by BARDIN (1972), based mainly on characteristics of moraines. These stages were correlated by him tentatively with glaciations in the McMurdo Sound region, but evidence seems insufficient. VAN AUTENBOER (1964) reported two glacial cycles in the Belgica Mountains 200 km southwest of the Yamato Mountains, and at least one fluctuation in the Sør Rondane Mountains 350 km west of the Yamato Mountains. The most difficult problems are finding or discrimination of readvance during retreat from the maximum stage to the present level and the chronology. These mountainous areas, including the Yamato Mountains, are important for elucidating the behaviour of the ice sheet in a regional scale because of the similar geographic situation on the continent where the mountains are located 100 to 200 km inland from the coast and hinder ice sheet flow from the south. The correlation of ice sheet fluctuation among these mountains is the principal problem to be studied in future.

4. Geomorphology of the Coastal Areas

The coastal areas of the Prince Olav and the Sôya Coasts consist of a marginal slope of the ice sheet, ice streams, scattered ice-free areas, and a peripheral sea ice zone. Problems to be discussed are concerned mainly with landforms of ice-free areas, but status of the present glaciation is also described concerning morphological characteristics.

4.1. Present glaciation

Knowledge of the present glaciation in the marginal zone of the ice sheet in the region is considerably less than that in inland, because many of active glaciers are inaccessible and glaciological work has concentrated on the traverse and station glaciology of inland ice sheet. The following descriptions and comments are presented on the basis of some ground surveys, visual observation from the air, and interpretation of vertical aerial photographs and LANDSAT images.

The Shirase Glacier: The Shirase Glacier is the largest ice stream in this region, cutting a channel into the ice sheet and pouring itself into Havsbotn of Lützow-Holm Bay (Fig. 5). The width is about 8 km near the snout and the height of a central portion exceeds 50 m at floating part. Flowline pattern on the surface can be traced upstream about 50 km from the mouth on LANDSAT images. Heavy crevasses and seracs and longitudinal chasm in the lower reaches indicate active flow with differential movement of the glacier.

The measurement of flow rate during winter provided a value of 5.9 m/day or 2.1 km/year in 1960 (NAGATA and YOSHIDA, 1962; FUJIWARA and YOSHIDA, 1972) near the snout where the glacier tongue seems to be afloat. The rate of 5.8 m/day was also obtained by measurement of ice flow for 4 days in August of 1975 at almost the same place (HAYASHI, 1977). NAKAWO *et al.* (1978) made a photogrammetric measurement of movement of icebergs near the snout on the vertical air photos taken in 1969 and 1975. This gave the flow rate of 2.5 km/year. The faster velocity seems to indicate the greater displacement of icebergs caused by expanding flow than that of the glacier snout. Recent photogrammetric measurement shows that the rates are 1.8 km/year on the right side of the glacier close to the ice-free area (Insteklep-pane) and 2.3 km/year on the left side of the snout between November 1975 and January 1977 (FUJII, unpublished). A reasonable flow rate for the glacier may be 2.2–2.3 km/year, which is the largest value among those reported so far for glaciers in

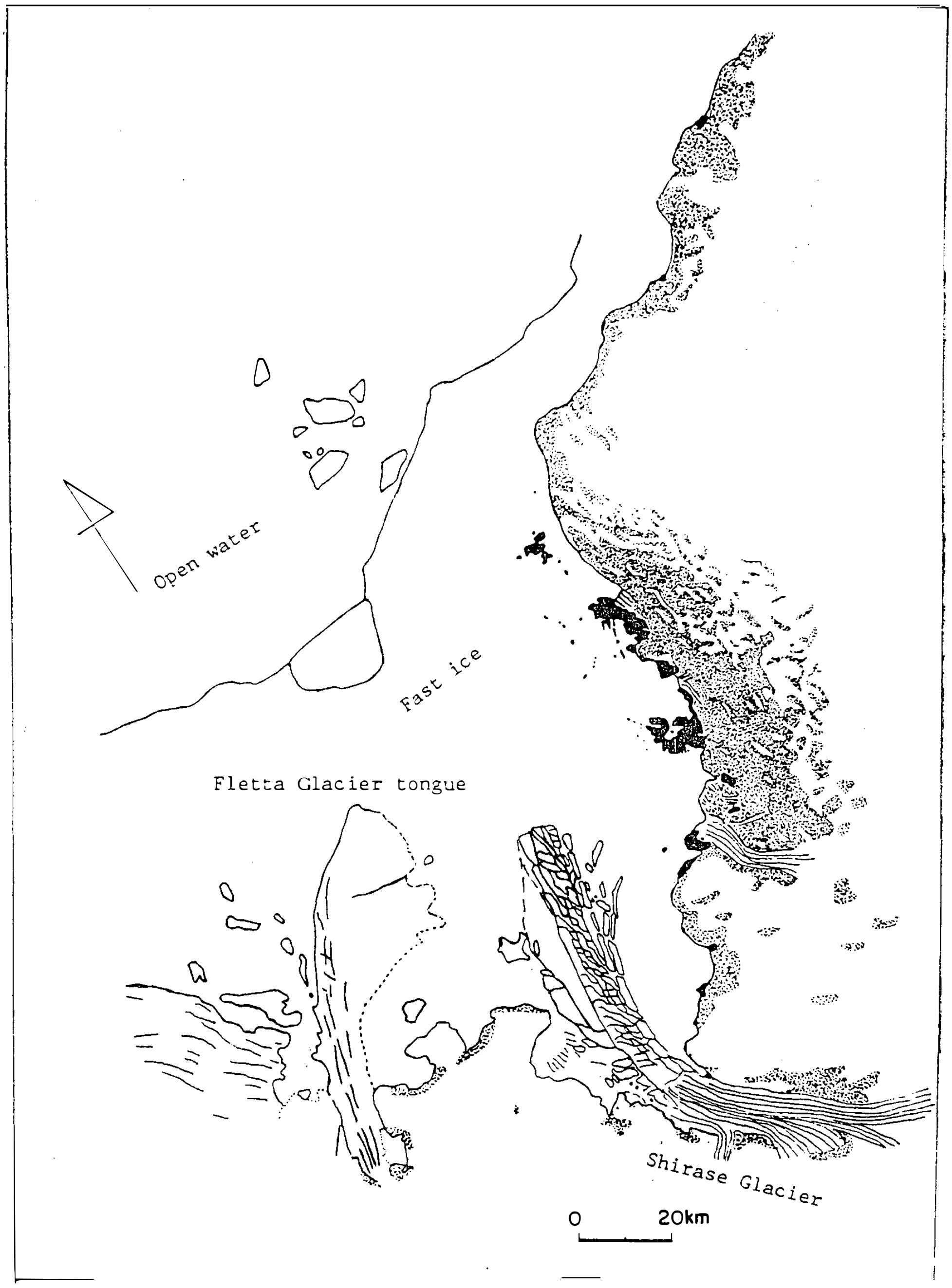

Fig. 5. Ice condition in the eastern part of Lützow-Holm Bay. (Drawn from the ERTS photograph taken on January 21, 1974. A dotted area represents a bare ice area.)

Antarctica.

This active thrusting of the glacier makes a conspicuous, floating glacier tongue which consists of a row of heavily-crevassed icebergs that are closely joined together. The position of a terminal of this tongue has changed considerably since 1960 when the author first observed the glacier tongue from the air. Around 1960 it was located fairly north. In 1967, the author found that it retreated to south by more than 15 km. After that, it advanced again toward north. The total amount of advance was esti-

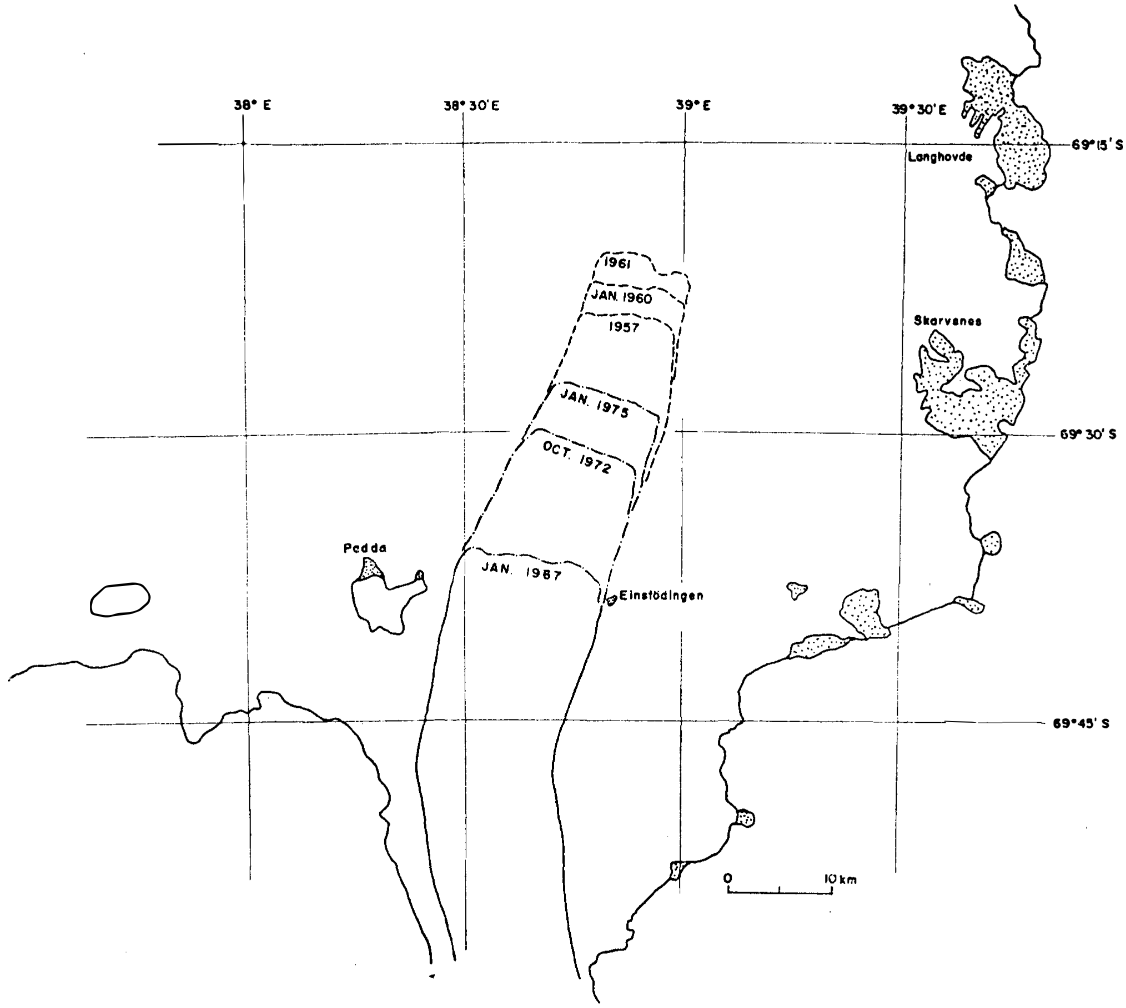

Fig. 6. Change of the terminus of the Shirase Glacier tongue.

mated at 15 km from 1967 to 1974, which corresponds to the measured flow rate of the Shirase Glacier (Fig. 6).

It is difficult to determine at the present stage of research what factors affected effectively the change of the glacier tongue. However, the breakdown of polar fast ice in Lützow-Holm Bay may be one of important factors in disappearance of a terminal part of the tongue, in view of the distribution of icebergs and its change.

The discharge of ice by the Shirase Glacier probably amounts to order of 4 to 5×10^9 t/year, assuming its thickness being 300 m (Ageta, 1971a; Yoshida, 1972). Nakawo *et al.* (1978) estimated the discharge at $7.4\pm1.9\times10^9$ t/year. The latter estimate is based on the annual rate of 2.5 km/year, 11.8 km glacier width, and 280 m ice thickness. The flow rate and glacier width seem to be slightly overestimated. In any case, the discharge by the Shirase Glacier contributes greatly to the draining of ice in the Shirase drainage basin in the Mizuho Plateau. Monitoring of the glacier tongue and more precise estimation of ice discharge have been continued by photogrammetric surveying.

Glaciers on the west coast of Lützow-Holm Bay: Geomorphological and glaciological information is very scarce on the west coast of Lützow-Holm Bay. This is due mainly to difficult access, rare occurrence of ice-free areas and difficulty in interpretation of featureless aerial photographs. It has been known since 1957 that the snow accumulation prevails on the west coast compared with the east coast (TATSUMI and KIKUCHI, 1959b), and that some parts of the coast are occupied by ice-shelf-like ice (aerial photographs taken in 1957; Report of the JARE-5 Wintering Party, 1962). The state of knowledge, especially on the basis of ground survey, has not been much improved since 1961. However, the following information was obtained by visual observation with helicopter flights and by interpretation of the LANDSAT images.

A floating glacier tongue was found to be extending its terminal some 50 km to north-northeast from an embayment named Fletta by the Norwegian Expedition (YOSHIDA, Y., 1975). A glacier which supplies continental ice to the tongue is tentatively named Fletta Glacier. The Fletta Glacier tongue expands its terminal part to form a flat and low tabular ice. The expanded part may be formed by the divergent movement of icebergs constituting the tongue and by plentiful snow accumulation on them. The author has pointed out that the peculiar shape of icebergs with dome-like side view at the Shirase Glacier tongue might be caused not by weathering or degradation of old iceberg but by deep snow accumulation on a heavily crevassed iceberg which indicates active movement (YOSHIDA *et al.*, 1962). This hypothesis can be applied to the Fletta Glacier tongue. Kaname Island, situated at the edge of the tongue and found in 1970, is likely to influence the maintenance and demarcation of the expanding part of the tongue. It exposes slightly its surface out of the glacier tongue, and yet reserves abundant vegetation composed of many kinds of lichen and moss (Report of the JARE-11 Wintering Party, 1971). This indicates the island has not been overridden by the glacier tongue for a considerable period of time, though the situation of the island seems to be critical for the covering by the glacier tongue. No information is available on the ice condition at the time of the Norwegian Expedition in 1937. However, the Norwegian place name "Fletta" means "plait". Therefore, it can be considered that a glacier tongue with complicated fracture pattern existed around the Fletta inlet at that time.

Towards northwest from the Fletta Glacier tongue the coast for some 120 km is fringed by the ice shelf of 60 km in maximum width. In some places vigorous calving of heavily crevassed icebergs is seen, suggesting that the regimen is considerably positive in the Riiser-Larsen Peninsula area. The small ice shelf is probably maintained by this active feeding, supported by the existence of small islands (Karamete Point) near the edge of the ice shelf.

Glaciers on the Sôya Coast: The Sôya Coast, which bounds the eastern side of Lützow-Holm Bay, is interspersed with ice-free areas of various dimensions. Therefore, most of ice flow take the shape of outlet glaciers between ice-free areas, and the sheet flow is rather sluggish (AGETA and NARUSE, 1971; FUJIWARA and YOSHIDA, 1972).

The following are the results of preliminary observation of the surface condition in the marginal zone of the ice sheet east of the Ongul Islands (YOSHIDA, 1972). Above the elevation of about 500 m the ice surface remains covered with snow even in the summer season, showing no remarkable melting to the naked eye. Meltwater formed on snow surface in the summer season is contained in snow as capillary water and sometimes makes a thin layer of ice crust at or near the surface. Therefore, sublimation and evaporation are the main agencies of ablation in this area. Bare ice begins to be exposed at about 500 m in height, and this level varies within a relatively narrow range of elevation throughout the year. Below 500 m a zone of snow accumulation alternates with a zone of bare ice except in the summer season, owing to the step-like, wavy relief of the ice sheet. In summer, many of snow fields below 350 m are subjected to remarkable melting. Meltwater channels are often formed below 150 m in height.

Examination of the LANDSAT images reveals that the bare ice zone develops extensively on the Sôya Coast especially from near the Skallen ice-free area to Mukai Rocks east of the Ongul Islands (Fig. 5). The cause of this feature is not yet well known. The following factors are thought to be responsible: the northwestern aspect of slope which receives much solar radiation, the existence of relatively large ice-free areas, the sluggish movement of the ice sheet, and the small amount of snow accumulation owing to the topography. The first factor seems most effective, because the same situation can be observed in other places, for example, in the southwest coast of the Riiser-Larsen Peninsula, on the LANDSAT images.

Apart from the cause, melting phenomenon in summer is notable in this area. On the basis of the preliminary observation of meltwater channels, significance of melting in the ablation of the ice sheet has been emphasized (YOSHIDA, 1972). Recent measurements at Cape Hinode and Cape Omega, where the development of bare ice zone is not so conspicuous as that on the Sôya Coast, seem to support this idea (MORIWAKI, 1976, 1980).

Stream flow in the Sôya Coast is influenced naturally by the ice-free areas and probably by the subglacial topography. It has been pointed out that the depressions occupied by outlet glaciers were formed possibly along the fractured zone such as fault-line (TATSUMI and KIKUCHI, 1959b). Recent studies on the submarine topography confirmed this view to a certain extent (FUJIWARA, 1971; MORIWAKI, 1975, 1979). Six outlet glaciers are actively producing glacier bergs on the Sôya Coast.

The Langhovde Glacier flows from SSE to NNW, confined between the Langhovde ice-free area and the sheet flow of inland ice, and has a small floating margin (YOSHIKAWA and TOYA, 1957). The flow rate was measured to be about 170 m/year near the terminal (FUJIWARA and YOSHIDA, 1972). A small unnamed ice stream 5 km north of the Langhovde Glacier protrudes its floating tongue toward NNW. Fluctuation of the position of its terminal is recognized by occasional observation (MORIWAKI, 1976) and aerial photography. Flow and calving of icebergs are inferred to be vigorous. These two glaciers flow parallel to each other and obliquely to the coastline, suggesting that a dominant subglacial structure susceptible to glacial erosion develops in that direction.

The Honnör Glacier has complicated surface features, indicating rather shallow and rugged subglacial bedrock. In 1937, its floating glacier tongue protruded to west some 8 km, and was named Honnörbrygga (Norwegian map, "Prince Harald Land", 1946). This was found to have disappeared before 1957 (YOSHIKAWA and TOYA, 1957). In 1960, fast ice in front of the Honnör Glacier did not melt away in summer, and was folded and sheared by push of a kind of small iceberg tongue (ARMSTRONG *et al*., 1966). At least after 1967, a polynya is formed frequently in summer. As a result, calved icebergs freely drift with the current. In such case, the formation of a long floating glacier tongue may be difficult. It is not well known what factors control the formation of a floating glacier tongue. In the case of the Shirase Glacier, the iceberg tongue has been advancing continuously at the same rate at least during the last 18 years. And yet the position of its terminal receded considerably in some year. This seems to show that the fast ice condition related to growth and breakdown is one of important factors to control the formation of a floating glacier or iceberg tongue, though the role of ice surge should not be excluded. To solve this problem the monitoring of fluctuation of ice margin has been continued intermittently since 1969. No sign of remarkable advance of the Hönnor Glacier has been found since that time. The flow rate of the Honnör Glacier is not measured yet, but it is inferred to be the order of 400 m/year from the deformation of sea ice in front of the glacier terminal (FUJIWARA and YOSHIDA, 1972).

The Telen and the Skallen Glaciers have floating glacier tongues with a length of 5 km. Their flow rates were estimated to be at least 400 m/year from the amount of dislocation of fast ice along shear cracks which were formed near the glacier tongue by the advances of glaciers (NAGATA and YOSHIDA, 1962). The flowline pattern on the Skallen Glacier can be traced upstream more than 20 km on the LANDSAT image (Fig. 5), indicating its active movement, while on the other outlet glaciers flowline patterns marked by crevasses, longitudinal ridges and depressions, and concave surface slopes are observed only within a short distance from the coast.

The form of the Rundvåg Glacier is somewhat obscure compared with those of other outlet glaciers. The glacier produces a small iceberg tongue composed of closely split icebergs. Formation of polynyas in summer in the coastal area south of the Skarvsnes ice-free area occurs much less frequently than in the northern coastal area. This difference in sea ice condition may contribute to maintain glacier or iceberg tongues of the Telen, Skallen and Rundvåg Glaciers. The flow rate of the Rundvåg Glacier was assumed to be 300 m/year for the calculation of ice discharge, though no data have been obtained yet (NAKAWO *et al*., 1978). This figure is somewhat larger than true one, because the effect on fast ice deformation is small.

The sheet flow is very sluggish on the Sôya Coast especially to the south of the Langhovde ice-free area, as is shown by flow rates of 3 to 10 m/year measured at several places (AGETA and NARUSE, 1971; FUJIWARA and YOSHIDA, 1972). In some places close to the ice-free bedrock almost stagnant sheet flow formed a moraine bank derived from upthrust basal ice. Around the Skallen ice-free area, the section of basal ice containing the till and overlain by clear ice with a sharp boundary plane can be seen at the edge of ice-free bedrock. On the other hand, the thin ice sheet without till covers bedrock directly at the other place. These situations may indicate that the upper, sliding clear ice can override the bedrock, leaving a lodgement till layer behind.

Glaciers on the Prince Olav Coast: On the Prince Olav Coast, the marginal condition of the ice sheet is considerably different from that on the Sôya Coast. The Prince Olav Coast is interspersed frequently with ice-free areas, similarly to the Sôya

Coast. However, dimensions of most of ice-free areas are small. And outlet glaciers develop well between these ice-free areas, forming rather small floating glacier or iceberg tongues. Thus, the proportion of length of coastline to stream flow, sheet flow, and bare rock coasts is fairly different from that of the Sôya Coast (Table. 2). It is remarkable that the coast of stream flow occupies 30 percent of total coastline, being twice as much as on the Sôya Coast. The flow rate of the Nishi-naga-iwa Glacier was estimated by interpretation of aerial photographs to be about 500 m/year at the floating glacier tongue showing divergent movement (NAKANO *et al.*, 1960). This glacier is inferred to be one of the fastest moving glaciers on the Prince Olav Coast from its configuration including calved icebergs and from the observation of sea ice deformation during the over sea-ice traverse from Syowa Station to Molodezhnaya Station in the winter season of 1967. No definite data on other glacier movements have been obtained on the Prince Olav Coast. However, rough estimate of flow rate can be made by reconnaissance of condition of glacier movement by visual observation from the air and by aerial photography. At least 5 outlet glaciers including the Nishi-naga-iwa Glacier out of 17 seem to have flow rates of the order of 500 m/year. Flow rates of 6 outlet glaciers out of 17 may be of the order of 400 m/year, and the rest may have those of the order of 300 m/year. Therefore, discharge of inland ice by these stream flows may be large in quantity against the drainage area. Assuming the flow rate as 400 m/year for stream flow coastline, ice thickness as 200 m and density of ice as 0.9 g/cm^3, the discharge is roughly estimated at 6×10^9 t/year for 4.0×10^4 km^2 drainage area (the area was measured using the ice divide shown in Fig. 1 in SHIMIZU *et al.*, 1978a). This is quite different from the estimations in other areas, 5 (or 7.4)$\times10^9$ t/year for 20×10^4 km^2 (SHIMIZU *et al.*, 1978a) at the Shirase drainage basin and $1.4\sim1.5\times10^9$ t/year for 2.2×10^4km^2 at the Sôya drainage basin (SHIMIZU *et al.*, 1978a). The large amount of discharge on the Prince Olav Coast may correspond in part to the abundant snow accumulation in this area (Report of JARE-15, 1975) (Table 2).

Table 2. Length of the coastline according to the mode of ice movement along the Sôya and the Prince Olav Coasts.

	Sheet flow	Glacier flow	Rock-bound	Total
Insteoddne–Mukai Rocks	49 km (36%)	21 km (15%)	68 km (49%)	138 km
Mukai Rocks–Sinnan Glacier	126 km (44%)	79 km (28%)	80 km (28%)	285 km

The sheet flow also seems to be more active on the Prince Olav Coast than on the Sôya Coast, as is indicated by frequent distribution of ice wall or ice front which shows the calving of icebergs from the ice sheet. In this connection, a portion of the ice sheet margin named "Flattunga" is worthy of attention. Flattunga is thought to yield a considerable amount of icebergs, though it does not form a glacier channel but only densely crevassed sheet flow.

4.2. Landforms of the ice-free areas

The ice-free areas along the Prince Olav and the Prince Harald Coasts are composed of crystalline bedrocks of several kinds of gneisses and granites including metabasites and a small amount of marble (sixteen sheets of the large-scale geological map have been published since 1974). The landforms are greatly controlled by the geological structures of these bedrocks. Superficial deposits such as glacial drift, fluvioglacial deposit and unconsolidated marine sediment are widely distributed on bedrocks. All the ice-free areas are inferred from their topographic features to have been covered with the former ice sheet and subjected to glacial agency.

4.2.1. Description of landforms of some ice-free areas

Sinnan Rocks: The Sinnan Rocks are situated at the eastern end of the Prince Olav Coast and bounded on the northeast by the Sinnan Glacier and on the southwest by the Nisi-Sinnan Glacier (a provisional place name). With the maximum height of about 200 m near its inner margin the undulating surface descends gradually to the sea. A shallow and broad glacial trough cuts the central area and its lower part is drowned to form a small embayment. The Sinnan Rocks are divided by this trough into the southwestern and the northeastern parts (Fig. 7).

The undulating relief of the southwestern part trends in the SE-NW direction concordant with the trend of gneissic foliation. Glacial striation also indicates the

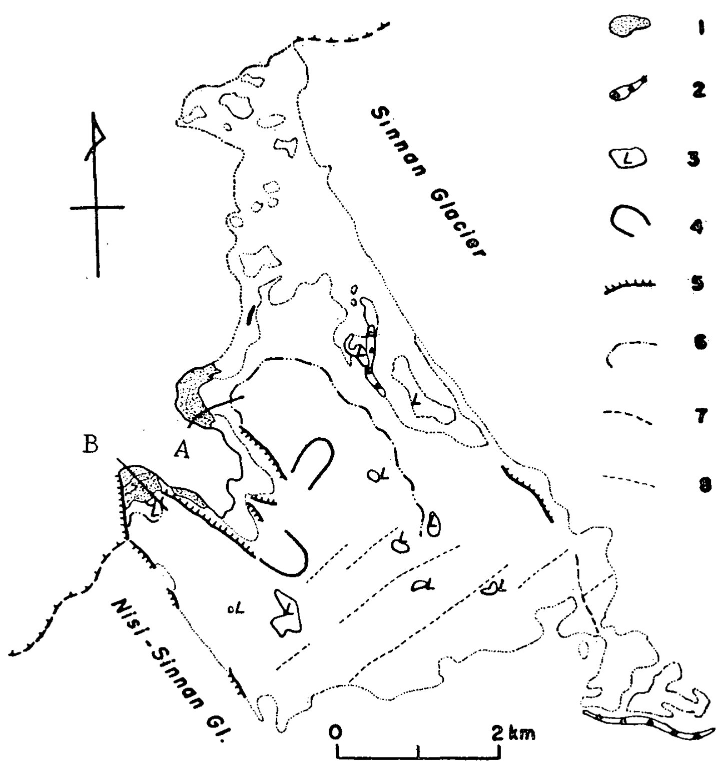

Fig. 7. Geomorphic map of the Sinnan Rocks ice-free area. (1; Raised beach deposit, 2; recessional moraine, 3; lake, 4; cirque, 5; precipitous cliff, 6; boundary of "pitted bedrock", 7; major linear depression, 8; minor linear depression along foliation of gneissic bedrock.)

same direction of the former sheet flow. On the other hand, the southern edge is being eroded by the Nisi-Sinnan Glacier to form a vertical cliff in the same direction. This cliff turns its direction to N5°W abruptly at the northeast terminal of the glacier, facing the sea. Therefore, after some thinning of the ice sheet, a NNW-trending stream flow had been formed, then it was transformed into the present-day Nisi-Sinnan Glacier as the retreat of the ice proceeded.

On the other hand, the northeastern part displays a complicated topography composed of small-scale rises and depressions which may be called "pitted surface of bedrock". To the south, the gneissic foliation perpendicular to the general trend of ice movement affected greatly the trend of relief. These features are inferred to have been shaped chiefly by glacial erosion, judging from the preservation of striated surface in some places, though periglacial agency modified the original relief to some extent after deglaciation. This fact suggests also that the sheet flow erosion was not intense enough to remove small-scale protuberance.

Raised beaches develop on the Kiridasi Point and the Maruhana Foreland, modifying glaciated surfaces (YOSHIDA, 1970b). Beach deposits consist of gravels of pebble to boulder in size mixed with sand. The form of these raised beaches can be designated as "elevated marine-boulder pavement" (NICHOLS, 1961). Some small depressions are considered to form "pitted beach" (NICHOLS, 1961). The remarkable features are shingle ridges resting horizontally on headlands. They occur at the height of 31–32 m and 20–21 m on the Maruhana Foreland and 27–29 m and 15–17 m on the Kiridasi Point. An indistinct one is also identified at a 10 m height on the latter (Fig. 8). It is difficult to consider that these ridges were formed as moraine ridges during the retreat of the ice sheet. They occur horizontally, surrounding partly a convex gentle slope in the form of an arch facing the sea, or fringing a shallow depression in the form of a slightly concave arch facing the sea. This mode of occurrence suggests that the sea ice pushed a beach deposit to form a shingle ridge. Therefore, this ice-pushed ridge (NICHOLS, 1968) represents an approximate elevation of a former sea level, though it is not an exact indicator of a sea level because of its deposition in the place slightly higher than sea level at that time. These features clearly indicate that raised beaches have not been affected by the readvance of the

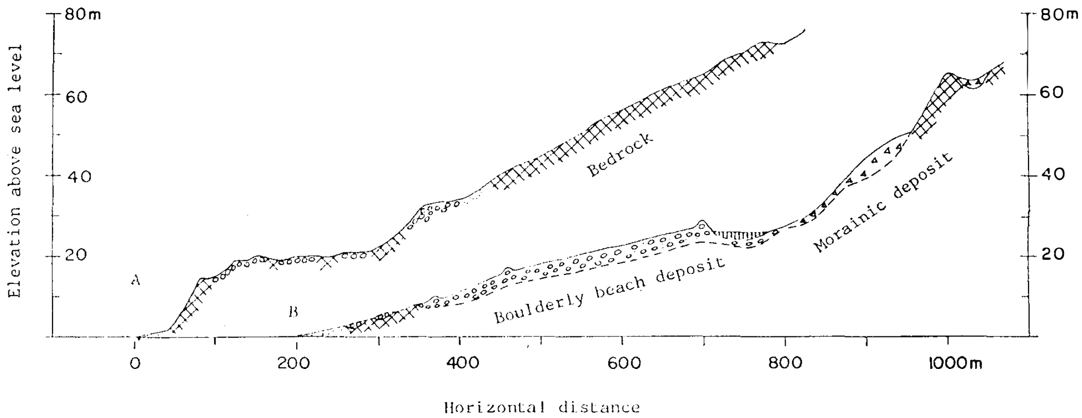

Fig. 8. Profiles of raised beach on the Sinnan Rocks. (Location is indicated in Fig. 7.)

ice sheet. The wave-washed bedrock with a small amount of rounded pebbles also exists below the 10 m height.

Cape Hinode: Cape Hinode is a small ice-free area with a low undulating surface of small rises and depressions. The roche mounotnée topography has survived to some extent on the undulating surface, though it has been subjected considerably to mechanical weathering, especially in its northern part. Disposition of rises and depressions is weakly controlled by gneissic foliation with a trend of NW-SE and lineation of pegmatitic vein and joint system with trends of N-S and NE-SW. The weakness of rock control is inferred to be due chiefly to the nearly vertical dip of banding of rather uniformly distributed granitic gneiss.

In Cape Hinode, patterned ground can be seen in some places on morainic detritus. Larger stone circles about 2 m in diameter associated with smaller circles less than 1 m in diameter develop well on the ground where abundant water is supplied from a pool and a superficial or subterranean melt-water stream. According to MORIWAKI (1976), segregation of gravels from finer materials is recognized in an active layer of 13 to 30 cm thick under the mud core of a stone circle. A frozen, non-sorted layer of gravels mixed with sand and silt exists beneath the active layer. Contraction cracks are also found in a fewer places where water supply is not sufficient. However, they display not polygonal networks but only crooked long lines associated with short branches. An exhumed wedge composed of gravels and coarse sand is 35 cm long and 20 cm wide at the upper part and a few centimetres wide at the lower part. It indicates the immature stage of development (MORIWAKI, 1976).

Raised beaches are characterized by the stepped topography composed of marine-boulder pavement partly associated with wave-washed surfaces and low sea cliffs. Step surfaces are recognized at 8–10 m, 15–20 m, 25–27 m, and 31–35 m above the approximate sea level. Obscure small steps also exist at elevations of 5–7 m and 23 m. Rather gentle sea cliffs with the relative heights of 5 and 3 m demarcate the upper limits of the 15–20 m surface and 25–27 m surface in some places. Wave-cut or wave-washed surfaces are also found in part at levels around 15 and 31 m high, continuous to the marine-boulder pavement. The existence of sea cliffs indicates that standstills of sea level occurred relatively to land at the elevations of 20 and 27 m in this area in the past.

The Kasumi Rock: The Kasumi Rock is a low-lying, small ice-free area situated between the Itime and the Kasumi Glaciers. The uneven surface with the maximum height of 45 m displays fairly "fresh" topographic features of polished and quarried surfaces with very small amounts of ground moraines and relatively recent beach deposits, except a narrow band of stranded moraines along the Kasumi Glacier. It was difficult to find evidence for conspicuous former marine agency during the survey for a short time. It may be due to 1) the area had no morainic detritus to be transformed into glacial marine deposit, or 2) the area was deglaciated quite recently so that the emergence of a recognizable amount of land has not taken place yet. The lack of lichen and moss in this area seems to indicate the latter case is likely. But evidence is not sufficient.

The Akarui Point: The Akarui Point is also a small ice-free area with the highest elevation about 150 m. Ridges and depressions subsequent to gneissic structure of the bedrock extend in the SE-NW direction, concordant with the direction of the former ice flow. Two linear depressions along structurally weak lines also intersect obliquely the general trend of topography. The bedrock is subjected to weathering considerably, but the polished surfaces with glacial striation are also found in some places. Weakly sorted patterned ground is distributed on morainic detritus in places where water supply is sufficient.

The coast is rocky for the most part, and evidence of former marine agency is scarce. However, the beach deposit is found near the mouth of a shallow glacial trough in the central part of the Akarui Point. The deposit consisting of sand and gravel and containing shell fragments forms a flat surface about 10 m high, abutting on the nearby bedrocks. Carbon dating of shell fragments supposed to be *Laternula elliptica* gave the age of 7730$\pm$110 years B.P. (Table 4f). A pocket beach, probably slightly raised, also exists in a small cove.

The Ongul Islands: The Ongul Islands are situated in the northeastern part of Lützow-Holm Bay 5 to 10 km away from the margin of the continent, and composed of many small islands. Among others, West and East Ongul Islands are larger ones. They show low-lying, undulating features below 50 m.

The landform of the Ongul Islands has been discussed by several researchers. YOSHIKAWA and TOYA (1957) made a geomorphological survey for the first time, and reached the following conclusion; 1) the Ongul Islands were once covered with the ice sheet and the depressions in the direction of E-W were formed by glacial erosion, though the geomorphic trend of a N-S direction subsequent to gneissic foliation is also remarkable, 2) periglacial phenomena, such as patterned ground, nivation, and wind action, are recognized, 3) the presence of raised beaches indicates that the area was uplifted 15 to 20 m after the shrinkage of the ice sheet, caused probably by the isostatic displacement due to unloading of ice. The succeeding surveys were carried out mostly along these lines.

The glacial erosion by ice flow from the east apparently scooped out ENE-WSW trending shallow depressions or troughs on bedrocks. Shallow depressions and rises trending in the N-S direction are also considered to have been sculptured by the ice sheet, and they present considerable rock control caused by dipping gneissic structure and different resistivity of various gneisses. However, MORIWAKI (1975) suggested that glacial scouring in the direction of N-S had also affected this area, because submarine depressions in the vicinity of the Ongul Islands have the shape of a glacial trough eroded by N-S trending ice flow. The deep Ongul Strait between the Ongul Islands and the continent is also considered to be a N-S trending, drowned glacial trough. Glacial striae remaining in a few places show two different trends of S35°–65°W and N15°–45°W, but it is difficult to estimate their chronological order. The problem of relationship between E-W trending and N-S trending glacial erosion in this area is to be solved in the future.

The mechanical weathering of bedrock in the Ongul Islands seems to be relatively intense compared with that in other ice-free areas. FUJIWARA (1973) noted that solifluction lobes composed of decomposed sand and breccias and morainic material develop to some extent in the Mizukumi Stream of East Ongul Island. Thickness of an active layer in this place was found to be 55 cm, but the segregation of coarser materials from finer ones was taking place in an upper active layer 20 cm thick at most. And the bedrock where the thickness of overlying detritus layers is less than 20 cm is thought to be subjected to congelifraction which resulted in weak cryoplanation. On the other hand, very few remnants of striated bedrock surface in East Ongul Island show that the flat surface has been worn away by 10 to 15 cm in thickness since the retreat of the ice sheet from the island (YOSHIDA, 1973). This fact indicates that the amount of denudation by periglacial processes including wind action is small even in comparatively well-weathered Ongul Islands. Nivation by drift-snow ice or snow-patch is also weak and affects the ground very slightly.

Patterned ground develops very poorly on the morainic materials. Especially, sorted patterned ground is hardly found in the Ongul Islands. This phenomenon is thought to be caused by low air temperature and thin thickness and coarser composition of unconsolidated deposit together with low content of water (KOAZE, 1964). MORIWAKI (1976) showed that the temperature of an active layer below 20 cm deep does not fluctuate crossing 0°C but remains above 0°C during summer (from November 27 to March 1), once it melts. This measurement well explains the non-effectiveness of freeze-thaw-cycle in the ground in this area.

Terrace-like surfaces and raised beaches are found generally below the level of 20 m high on the Ongul Islands. Raised beaches show stepped features in some places. The problems are 1) the highest altitude of the former sea level, 2) the significance of stepped features, and 3) the age of raised beaches.

MEGURO *et al.* (1964) analysed organic remains contained in the raised beach deposits on East Ongul Island, and discussed the radiocarbon datings of them and composition of Foraminifera in relation to raised marine features. The author pointed out from the radiocarbon ages of shell fragments and Foraminifera that the retreat of the ice sheet from the Ongul Islands took place at least over 23000 years B.P., but it might not be over 40000 years B.P., because the raised beaches had not been affected undoubtedly by the ice sheet as was pointed out by YOSHIKAWA and TOYA (1957). On the other hand, the fossil Foraminifera assemblages contained in the same deposits suggest that the deposits were deposited on the sea bottom at depths of about 100 m. This fact contradicts the fact that the raised marine features are hardly found beyond the height of 20 m above sea level on East Ongul Island. This problem still remains unsolved.

FUJIWARA (1973) made a detailed survey on the raised beach topography at Mizukumi Stream in East Ongul Island. He recognized marine terraces at five levels of 14.0–11.5 m (surface I), 10.0–9.5 m (II), 7.0–6.0 m (III), 4.5–3.0 m (IV), and below 2.0 m (V). He suggested that surface I had been formed during the Würm Interstadial sea level, surface III might correspond to a considerable standstill, and surface V had been formed during the post-Glacial high sea level.

On the other hand, OMOTO (1977) identified 22 levels of raised beaches and step landforms below 22.1 m on East Ongul Island. He also observed flat surfaces on Ongulkalven at levels of 18 to 20 m, 24 to 25 m, 27 to 29 m, and 33 to 35 m, and raised beaches and step landforms on Teöya at 12 levels below 19.4 m.

FUJIWARA's conclusion seems to contradict in part the fact that a beach deposit containing shell fragments with the radiocarbon age of 5850 years B.P. was found at a level of 16 m high on East Ongul Island (YOSHIDA, 1973). The significance of stepped features is not yet clear. Results obtained by OMOTO (1977) seem to indicate that at least relative lowering of sea level has been continuous as a whole below a level of 20 m above sea level, though levels distinguished by FUJIWARA seem to be more distinctive. On the other hand, stepped features above 20 m have no clear evidence of marine origin at least on East and West Ongul Islands. It is considerably difficult to distinguish marine sediments containing no organic remains or stratified bluish clay from morainic and decomposed materials, because most of marine sediments themselves originated in morainic materials. Erosional bedrock surfaces are also difficult to identify to be of marine origin. Calcareous marine algae adhered to bedrock is clear evidence of marine agency and found on East Ongul Island below 20 m.

In conclusion, the highest altitude of former sea level is considered to be about 20 m, and raised beaches below a level of at least 16 m and possibly 20 m have been formed during the "post-Glacial" age in the Northern Hemisphere as a result of regional uplift and sea-level change.

The Langhovde ice-free area: The Langhovde district is a large ice-free area next to the Skarvsnes district on the Prince Olav and the Prince Harald Coasts. It forms a hilly land of gneisses and granite, having an area of 52 km^2 and the maximum elevation of 500 m above sea level. The greater part of the eastern margin faces to the Langhovde Glacier flowing towards north. To the south, the ice sheet east of this district increases its surface elevation, resulting in formation of small ice tongues and an outlet glacier over the southern part of the Langhovde. The western side of the Langhovde opens directly to the sea with many indentations and islands. The Langhovde can be divided topographically by conspicuous glacial troughs into two parts, the northern and the southern (YOSHIKAWA and TOYA, 1957). The glacial troughs, one is larger and the other smaller having bottoms of less than 50 m high, had been the ways to drain the continental ice until a comparatively later stage of deglaciation. Terminal moraines near the eastern ends of the troughs suggest that the stagnation of short duration had occurred during shrinkage of the ice sheet. The moraines dam up meltwater from the Langhovde Glacier to form a glacial lake in which the fluvioglacial deposit forms a deltaic fan (Fig. 9).

The northern part of the Langhovde is composed of several mammilated peaks and undulating hilly land. Mt. Tyôtô, standing out some 200 to 300 m with steep cliffs from the surrounding hilly land, is thought to have been once covered by the ice sheet. A rather flat and narrow surface 200 to 250 m high in the form of a glacial bench is situated at the western side of the peaks. This level is accordant with the heights of some summits and rather flat shoulders in the northern and the southern parts of the Langhovde. The hilly land can also be divided into two levels, a rather flat area 30 to 50 m high and an undulating hilly area about 100 to 150 m high.

The relief is strongly controlled by the geological structures and rock types. Conspicuous cliffs surrounding Mt. Tyôtô run conformably with the synclinal structure of gneisses. Trends of long axes of wide depressions, headlands, and islands coincide with strikes of gneissic foliation. Discordant junction between rather flat surfaces and steep cliffs around Mt. Tyôtô occurs at the boundary between different

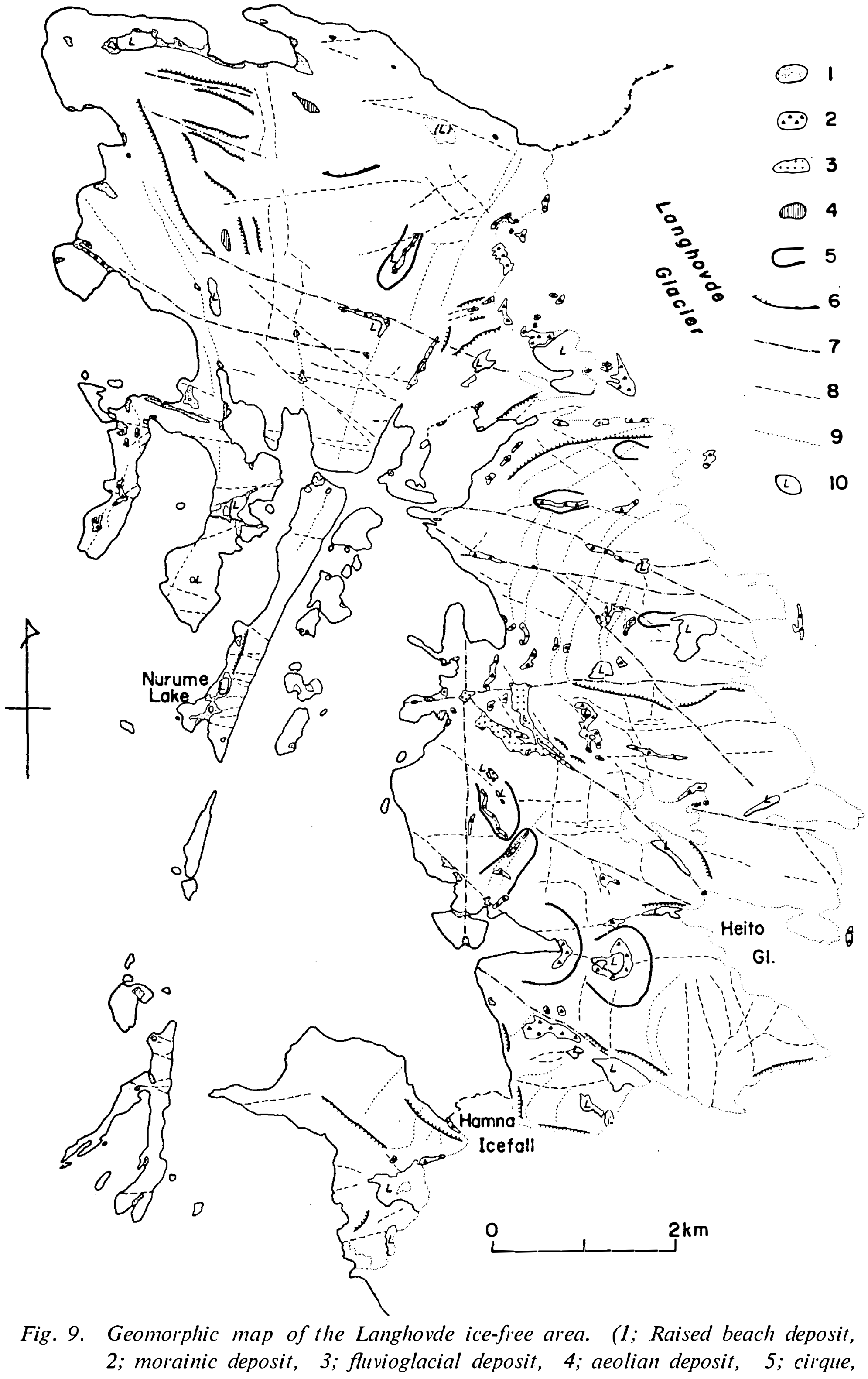

Fig. 9. Geomorphic map of the Langhovde ice-free area. (1; Raised beach deposit, 2; morainic deposit, 3; fluvioglacial deposit, 4; aeolian deposit, 5; cirque, 6; steep slope, mostly valley wall of a glacial trough, 7; major linear depression, 8; minor linear depression, 9; long depression along foliation of gneissic bedrock, 10; lake.)

kinds of gneisses. Massive pyroxene gneiss (ISHIKAWA, 1974) contributed to some extent to the formation of a very flat surface 30 to 50 m high in the northeastern part (MORIWAKI, 1976). On the other hand, rather narrow and linear depressions are formed along joints and fractures perpendicular or oblique to strikes. In some places, such relief has been somewhat exaggerated by periglacial erosion and deflation after cropping out from the ice sheet.

Deglaciation is considered to have proceeded somewhat earlier in the northern part than in the southern part judging from the degree of weathering. And as the northern part was more detached from the ice sheet, subaerial denudation, especially wind erosion, has worked more effectively in this area. Dikes of pegmatite protrude 30 to 40 cm from the surrounding gneisses in some places resisting to wind abrasion (ISHIKAWA, 1974), and several conical drifts of aeolian sand are distributed in the leesides of peaks. But the glacial drifts in many places and glacial striae in a fewer places clearly indicate that the amount of denudation after shrinkage of the ice sheet was fairly small.

During the retreat of the ice sheet a depression 2 km southeast of Mt. Tyôtô was occupied by a small ice mass independent from the ice sheet, and was slightly deformed into a cirque. The cirque has a typical terminal moraine. Similar but more distinct glacial cirques and adjoining small glacial troughs are found in the southern part of the Langhovde. This cirque glacier must have been fed by drifted snow from the ice sheet surface. Therefore, the regimen of such a glacier was controlled by its locality and aspect in relation to the ice sheet surface. YOSHIKAWA and TOYA (1957) suggested that these cirque glaciers had survived and eroded depressions until the stage when the ice sheet retreated to the level enough to cover presently exposed bedrock only below the level of 100 m above the present sea level. It is difficult to clarify the relation between the cirque formation and the ice sheet retreat. However, it is inferred most likely that there was a standstill of the ice sheet retreat during the formation of a cirque. Because snow supply to maintain an ice mass from the ice sheet must have been necessary for cirque erosion at least for certain period.

On the other hand, most of glacial drifts occur as thin ground moraines, indicating that any remarkable stagnation of the margin of the ice sheet did not take place on the presently ice-free area during the retreat of the ice sheet (KOAZE, 1964). However, the observation on recent shear moraines (MORIWAKI, 1976) suggests that the moraines are too thin to form a heap of debris of considerable size near the ice sheet margin which seems to be stagnant for a certain period at present. Somewhat noticeable recessional or ablation moraines can be found in very limited areas such as the Omega Point, Honnör-iwa (a tentative place name), and near the southeastern corner of the Skarvsnes ice-free area. These facts suggest the absence of the terminal moraine does not always mean that there has been no stagnation period during general ice sheet retreat, in the case of the Antarctic ice sheet whose shrinkage may proceed in the form of sublimation, evaporation, surface melting, and iceberg calving from ice cliff terminated in the sea. The ice sheet might have receded from the Langhovde ice-free area rather intermittently, as is indicated in part by cirques and some moraines such as those in the Naka-no-tani Valley.

Raised beaches are distributed along the coast mostly below the height of 20 m.

Many of them occur in the form of pocket beach occupying the depression. Elevated marine-boulder pavement is found in a few places. Most of deposits were derived from glacial drifts, and some include organic remains such as molluscan shells and worm tubes. Results of radiocarbon datings of these organic remains are shown in Tables 4c and 4d.

Several saline lakes or pools are distributed on the coast. An empty lake bottom covered with evaporites is also found. Elevations of lake surfaces are mostly below the present sea level. Some lakes had obviously been connected with sea, evidenced by marine sediments or raised beach topography in the basins. As a result of crustal upheaval, the connection between the lake and the sea was broken off by elevated rock threshold. If the amount of evaporation of lake water exceeds the amount of inflow, the lake level naturally lowers and the salinity of water increases. Therefore, lakes which are considered to have been once connected to the sea contain waters of chemical composition similar to but more concentrated than that of sea water. The elevations of rock thresholds of lakes showing the evidence of the previous connection with the sea are lower than 20 m. On the other hand, the Lake Akebi whose threshold to the sea is about 45 m high is a saline lake and its surface elevation is 4 m below sea level. But the concentration of chemical composition of its water is considerably lower than that of sea water. This fact is considered to indicate that the maximum height of former sea level had not risen over at most 45 m above sea level.

The southern part of the Langhovde forms massive hills with rather flat surfaces and precipitous cliffs, and the landform characteristics are much alike those in the northern part. However, the ice sheet east of the Langhovde increases its surface elevation towards south, and decreases the difference in elevations between ice and ice-free areas, as described before. As a result, the southern past is more influenced by the present ice sheet and more humid than the northern part. Preservation of glacial features such as striation, groove and quarried surface is also better than in the northern part.

Flat surfaces, or peaks with accordant height, though they are small in size, can be recognized at levels below 50 m, 100–150 m, 200–150 m, 350 m and 450 m. These surfaces develop stepwise to some extent, and the level of 200–250 m high is most remarkable.

Linear arrangement of the topographic elements is more conspicuous in the southern part than in the northern part. It is indicated mainly by NW-SE and E-W trending valleys, depressions and precipitous cliffs. Somewhat bending, N-S trending ridges and shallow depressions are also well developed along the gneissic foliation planes dipping to the east. Trends in the directions of NW-SE and E-W perpendicular to the gneissic foliation are apparently controlled more or less by the joints of gneissic bedrock.

The joints of bedrock, forming weak lines, were considerably effective for selective glacial erosion by the ice sheet, because their directions roughly coincide with the general trend of ice flow. Glacial valleys are generally narrow and deep. Relatively wide depressions, where lakes such as the Lake Yukidori are often distributed, are obviously formed at intersecting points of two or more lines of weakness. The

Kami-kama lake and the Simo-kama cove, large cirques located near the southern end of the district, are also good examples of this kind. Kami-kama is the intersecting point of three linear depressions, and the coastline of Simo-kama was determined by four linear structures. No distinct faultline has been found in the main Langhovde district. However, the Hamna Icefall, a small outlet occupying a trough between the southern end of the main district and Hamnenabben, is inferred to develop in a depression along a faultline, judged from the adjacent geological structure (ISHIKAWA, 1974).

Detached small ice masses and marginal tongues of the ice sheet supply melt-water to the valleys, and fluvioglacial deposits are more abundant in the southern part than in the northern part. Near the lower ends of the Yatude Valley and Yukidori Valley glacial valleys, deltaic fans have been formed on the coast, and a part of them was raised and dissected. One fluvioglacial deposit in a circular depression terminates in a cliff facing to the depression. This is inferred to have been formed by the disappearance of a small ice mass which had occupied the depression and blocked the fluvioglacial deposit. These fluvioglacial deposits represent melting phenomenon during a somewhat older stage than the present.

Raised beaches are also distributed along the shoreline, especially around the Simo-kama and the mouths of the Yukidori Valley and the Yatude Valley. However, the greater part of the coastline is rocky and bounded by steep slopes. Therefore, distribution of raised beaches is limited to fewer places in the southern part than in the northern part. OMOTO (1977) distinguished nine flat surfaces of raised beaches up to 27.4 m high in the Langhovde district. But the highest one is erosional in nature and almost devoid of deposits. It seems to the author that raised marine features with positive evidence develop only below 20 m.

The mechanical weathering by frost action is affecting surface features of the Langhovde district to some extent, as described before, while the development of patterned ground is very poor. Sorted patterned ground and contraction cracks are distributed on moraines near the Langhovde Glacier. The former is also found in a few places of the intertidal zone. Only incomplete contraction cracks are seen on morainic debris in the ice-free land area.

Well-developed sorted stone circles with the mean diameter of 1 m long are situated on the flat recessional moraine close to the ice-free bedrock facing to the Langhovde Glacier in the southern part. These circles appear to extend to the area now covered with snow and ice. In other words, their occurrence seems to indicate that they are exposed in part to subaerial condition by wasting of snow and ice. The formation of a sorted stone circle may not occur under the perennial ice (BERG and BLACK, 1966). Therefore, it is possible to infer that they were formed some time in the past when the ice retreated from the present margin or at least snow accumulation was less than that at present. Further study is needed to examine the validity of this interpretation. Raised beaches presently half buried under drift-snow ice can be seen on West Ongul Island. This phenomenon may be interpreted along the same line that there was a period of less snow accumulation in the near past.

The Skarvsnes ice-free area: The Skarvsnes district is the largest ice-free land in this region with an area of 63 km^2. It is bounded on the east by the ice sheet and on the west by the sea. The Skarvsnes is a hilly land with the maximum elevation of 400 m, but much of the area is lower than 300 m. Conformity of summit heights is not much striking, but levels of 200–250 m and of 100–150 m high are recognizable.

Glacial erosion by the ice sheet has smoothed the topography as exemplified by quarried surfaces and mammilated peaks, precipitous cliffs, and shallow linear depressions and ridges, as in the case of the Langhovde district. However, linear topographic features in this district are not so conspicuous as in the Langhovde, reflecting the complicated geological structure. Polyclinal foldings trending N-S superposed

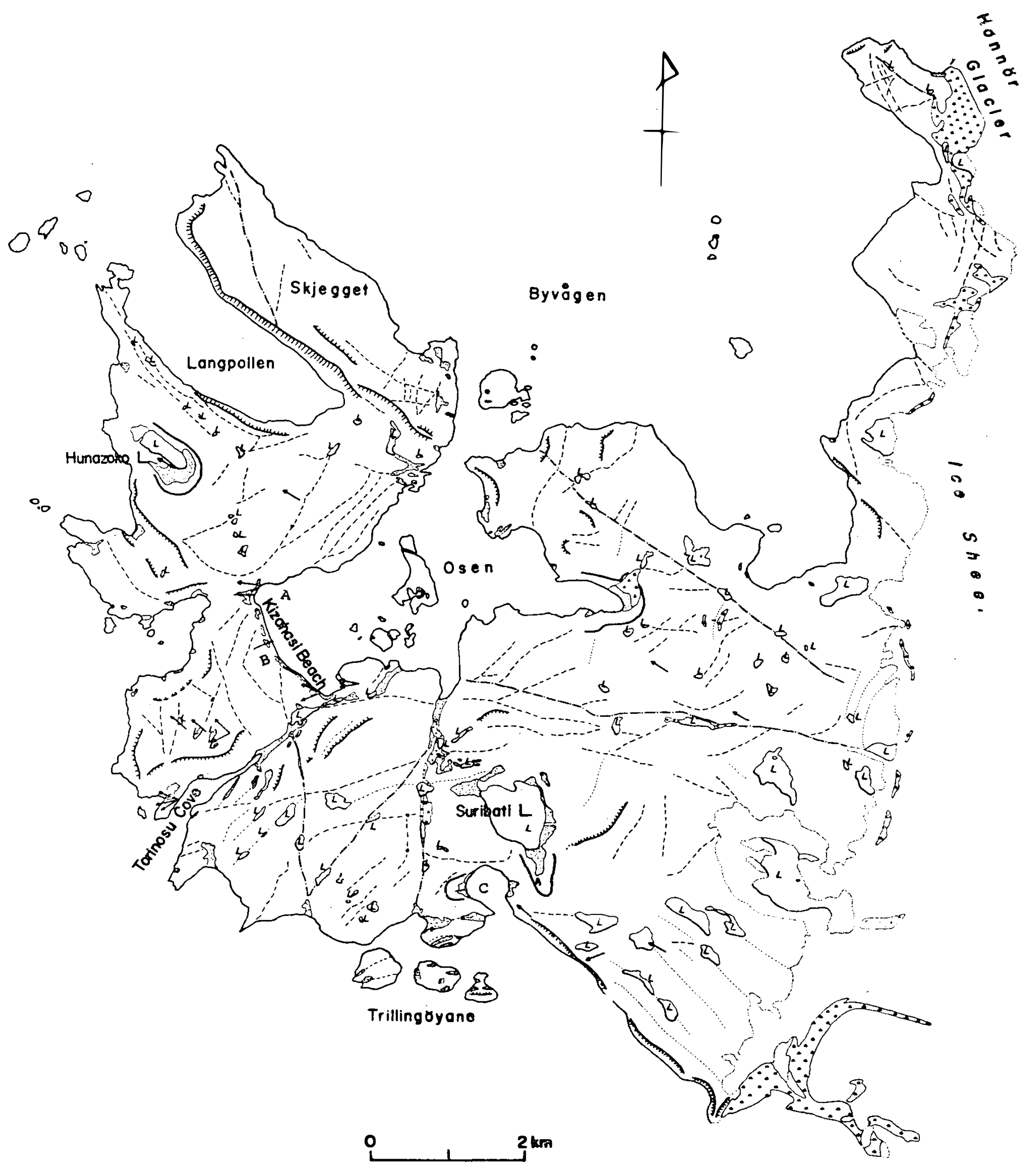

Fig. 10. Geomorphic map of the Skarvsnes ice-free area. (For legend, see Fig. 9. Direction of glacial striation is added to this map.)

by later foldings trending E-W and N-S occur throughout the district. Therefore, strikes and dips of banding and foliation of gneisses considerably vary all over the district. Low ridges and shallow depressions develop in parallel along strikes trending NW-SE in the southeastern part near Mt. Tenpyô, conformably with the general direction of the former ice flow which is inferred from glacial striations and grooves.

On the other hand, similar ridges and depressions trending N-S or NNE-SSW perpendicular to the direction of the former ice flow are found in the eastern part of the district near the Maruyama Peak and in the southwestern part near the Torinosu Cove, along the gneissic foliation. In the latter case, slopes represent the large-scale stoss-and-lee topography. Obscurity of linear topographical arrangement around the Maruyama Peak is probably caused by massive porphyroblastic gneiss showing a mushroom synform with very steep dipping (Fig. 10).

Fractures trending NW-SE and NE-SW represented by joint systems are also developed throughout the district. A nearly vertical cliff with relative height of 500 m including submarine 100 m cliff was carved by glacial erosion on the southwest side of the Skjegget peak. This cliff is considered to have been formed along the joint system trending NW-SE. Other linear depressions or cliffs are generally of a small scale, cutting slightly into bedrock.

Glaciated surfaces with striations and grooves in this district are preserved better than in the Langhovde, though their thin rock crusts with striations are upheaved slightly by exfoliation in some places. Striations are trending generally NW-SE to WNW-ESE. Intersection of striae with directions of NW-SE and WSW-ENE is also observed in several places. It is difficult to determine whether two directions of the former ice flow indicated by this intersection belong to different ice advance stages or to the same one. However, the latter case is more likely, because no significant topographic features indicating two different ice flows can be seen other than glacial striation.

Morainic deposits in the district are mostly thin ground moraines occupying depressions and erratic boulders. Exceptions are the cases of a cirque-like depression in the easternmost part of Osen bay and the east side of the northernmost part of the district. In the former case, a conspicuous morainic heap covers the greater part of the head of a horseshoe depression. During shrinkage of the ice sheet, stagnation of ice is inferred to have occurred at least in a local scale, occupying a glacial cirque below a level of 100 m high, and to form a morainic heap. In the latter case, fairly remarkable stranded moraines are formed between the ice-free area and a northerly flowing glacier, indicating slight ice retreat at present.

Periglacial processes are operating weakly on bedrock and unconsolidated materials. Wind action indicated by wind-facetted pebbles predominates in the eastern part of the district. "Honeycomb" weathering is also found in some places. Exfoliation and formation of a block field are taking place on quarried surfaces, but striations have survived to a considerable extent. Incomplete non-sorted polygons composed of contraction cracks can be seen on beach sand and gravel in a fewer places. Sorted stone circles of obscure shape are found only on the tidal flat of the Kizahasi Beach. This fact indicates that the wave action near the shore is not effective enough to move sand and gravel at present.

Raised beaches develop in many places along the present shoreline. Well-defined raised beaches and marine sediments are also found around the Lakes Suribati and Hunazoko which are situated in basins below sea level. The sorting of raised beach deposits is usually very poor. But the sand and gravel bed found near the Lake Funazoko has well-defined lamination indicating the effect of moving water. This bed contains fossil marine molluscan shells and is overlain by a matt of diatomaceous earth 5 to 10 cm thick. Mirabilite which is considered to be of marine origin (DORT and DORT, 1971) has also crystallized on the sediment. The water of these lakes is highly saline, and most of salt ions are inferred to be derived from sea water (TORII and YAMAGATA, 1973). These lakes had been obviously arms of the sea until some time in the past, connected by narrow channels. As a result of crustal uplift in this area the lakes were separated from the sea by the emergence of thresholds which are about 15 m high at the Lake Suribati and about 3 m high at the Lake Hunazoko at present. Radiocarbon dates of organic remains are a little younger in the Lake Hunazoko than in the Lake Suribati (Table 4f). The difference in radiocarbon dates seems to reflect the difference in the elevation of thresholds, that is, the difference in age of separation from the sea.

A considerably well-marked stepped topography develops on raised beaches at the Kizahasi Beach. Ten steps can be distinguished below the level of 18–19 m high. There is a small steep slope along the present strandline, cutting the lowest part of the raised beaches. The slope is inferred to have been formed by the slumping of beach deposit caused by the permeated sea water. Freezing and thawing of permeated water probably accelerate the slumping. NOGAMI (1977) concluded that the steps of raised beaches were formed not by wave action but by freeze-thaw of the permeated ground water or sea water. This conclusion is probable especially in the embayment where the wave action is hard to work. On the other hand, a rather flat raised beach deposit showing the incomplete shape of beach ridges is found at the mouth of the Torinosu Cove at an elevation of 10 m, suggesting the weak wave action. There is a possibility that a small cliff a few metres high along the Torinosu Cove and a cliff 20 m high behind the raised beaches at the Kizahasi Beach are sea cliffs carved into bedrock. However, no positive evidence has been found (Fig. 11).

OMOTO (1977) distinguished 19 levels of raised beaches or steps in the Skarvsnes district up to the elevation of 39 m. However, deposits containing *in situ* marine

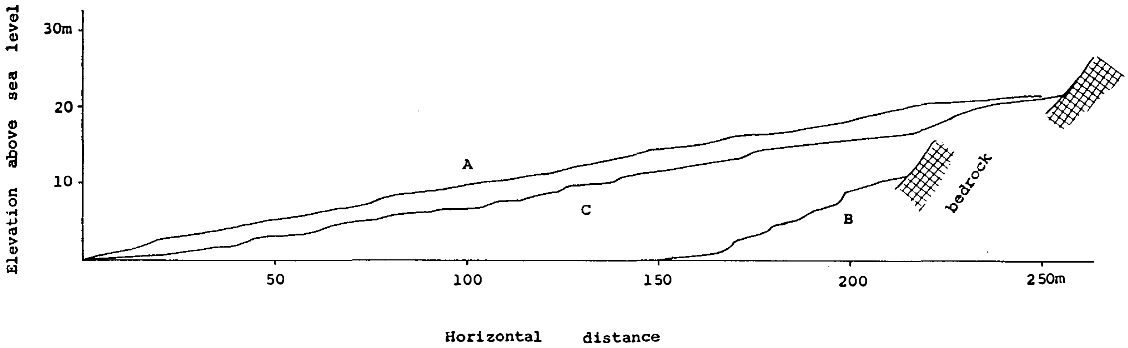

Fig. 11. Profiles of raised beach on the Skarvsnes ice-free area (Location is indicated in Fig. 10).

organic remains have been found almost exclusively at localities below the height of 20 m. The gap between elevations below 19 m and those above 28 m is illustrated in OMOTO's schematic cross section of raised beaches and step landforms (1977), though he does not touch upon the significance of this gap. Some of the step landforms or beach-like deposits above 20 m may represent former relative high stands of the sea level. However, deposits or steps above 35 m are considered not to relate to the former sea level, at least during "post-Glacial", judging from the incompleteness of topography and from the chemical composition of a saline lake (Hama Lake: a tentative place name by H. MURAYAMA) situated near the coast. The lake occupies a small basin whose threshold to the sea is 35 m high, and has no outlet stream. The lake water is saline, but the ionic composition and its concentration do not suggest the sea water origin, unlike the waters of the Lakes Hunazoko and Suribati (MURAYAMA, 1977).

Results of radiocarbon datings of fossil molluscan shells and worm tubes contained in raised beach deposits or marine sediments are tabulated in Tables 4b, 4e and 4f. One sample gives an age about 30000 years B.P. Others are younger than 9000 years B.P. Relationship between the elevations of localities above sea level and the dates is not very clear, because the habitats of the mollucs are not always close to sea level but of a wide depth range and the radiocarbon date itself has some uncertainty. However, the elevation of *in situ* fossil shells indicates the minimum height of the former sea level when the shells were alive. Taking the above into consideration, the relation of the dates to the elevations of localities seems to be significant for understanding the crustal uplift which amounts to at least 16 m since 6000 years B.P.

The Skallen ice-free area: The Skallen ice-free area consists of three districts, *viz*., Skallen, Skallevikhalsen and Hjartöy. The Skallen district displays a gently undulating erosional surface with an area of about 10 km^2 and the maximum elevation 186 m. The periphery about 23 m long is surrounded by sea except its southern edge 2.7 km long where the ice sheet terminates in a stagnant state (Fig. 12).

The glaciated surface marked with well-preserved striations, grooves, and small-scale quarried surfaces with relative height of 1 to 2 metres shows undulation trending generally in the direction of ENE-WSW to E-W, carved along the gneissic banding or foliation. The direction of dipping of gneissic banding changes from north to south and *vice versa*, according to major foldings which develop well in this district. Southerly dipping rocks form a stoss-and-lee topography with gentler stoss slopes and steeper lee slopes, but northerly dipping rocks form asymmetric ridges with rather steeper slopes on the stoss side. This fact indicates a strong control of the gneissic structure on glacial scouring.

Fractures are also developed obliquely to the direction of gneissic foliation. However, only shallow and straight depressions were formed along them. It is also noted that the major thrust fault concordant with the gneissic banding affected topographic features very slightly.

The direction of the former ice flow inferred from the directions of glacial striae and grooves is generally SE-NW, which is oblique to the general trend of major relief. Two intersecting striations are seen in places, suggesting the change of direction of ice flow probably from SE-NW to SSE-NNW during the retreat of the ice

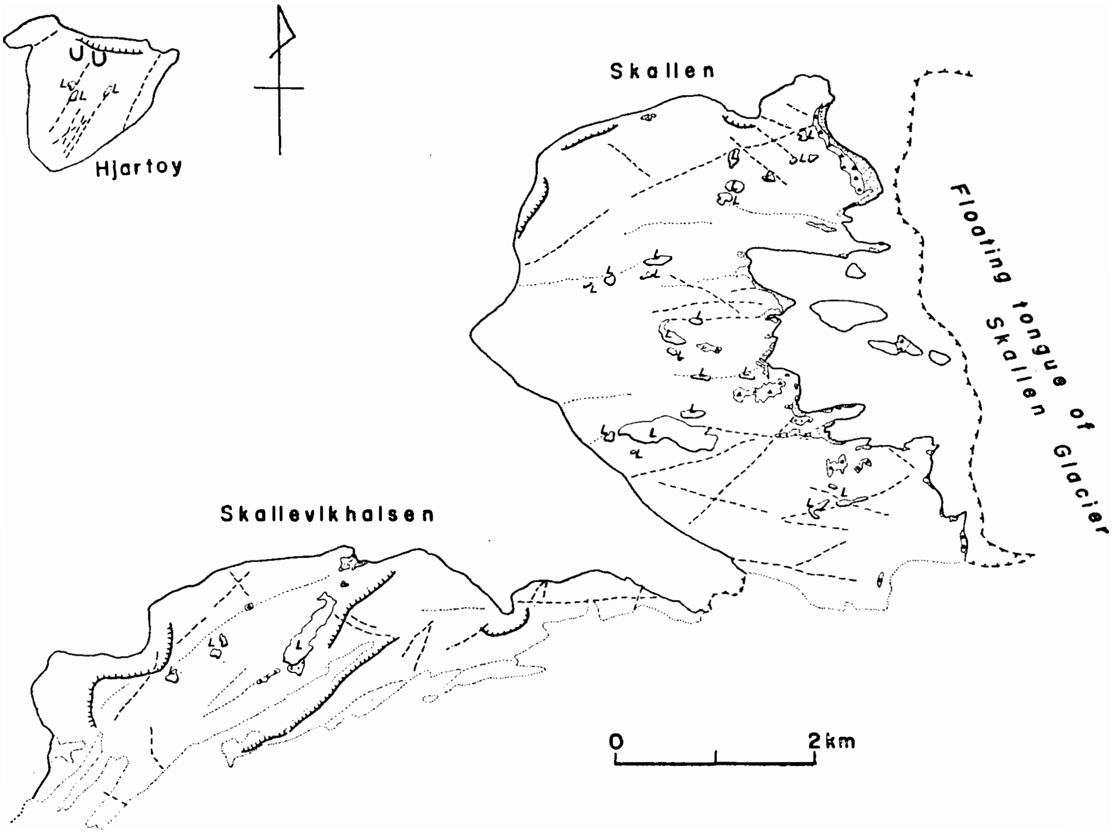

Fig. 12. Geomorphic map of the Skallen ice-free area (For legend, see Fig. 9).

sheet. A small-scale quarried surface is also found near the western end of the Lake Skallen Ôike. Its abraded stoss side and plucked lee side indicate that the ice flowed in the depression locally from east to west. Striations trending E-W and NE-SW were described by TATSUMI and KIKUCHI (1959a). But these directions appear to represent local ice flow. Curved grooves, with overhanging sidewalls (TATSUMI and KIKUCHI, 1959a) in some places, is inferred to be gouged by plastically moulded ice. These features are considered to be formed by a wet-based glacier by some researchers (SUGDEN and JOHN, 1976). However, no other evidence has been found in this area.

Preservation of glacially scoured surfaces is good in this district as a whole. Frost shattering and exfoliation can be seen in some places. However, the depth of exfoliated crust amounts at most to a few centimetres on flat surface. The weathering of bedrock in the area north of the depression containing the Lake Skallen Ôike has slightly advanced in comparison with that in the southern area, suggesting that the ice sheet retreated from the northern area a little earlier than from the southern one.

On the other hand, stone stripes and stone circles are developed well on thin morainic deposits in some places. In this district much snow accumulates in winter and melts away in summer. Abundance of meltwater appears to be favourable for the growing of sorted patterned ground in this district.

Raised beaches are distributed along the east coast, but hardly found on the west coast. Raised beach deposits are composed mostly of sand and gravel derived from morainic detritus, containing marine silt and clay in places. Shell fragments of a very small amount were found in the raised beach deposit at the northeastern tip (Osiage Beach) of the district, but well-preserved fossil molluscan shells to be dated have not been collected yet. OMOTO (1977) distinguished 15 levels of raised beaches and step landforms below 32.3 m. However, it seems to the present author that the levels above probably 15 m and possibly 10 m high have no clear evidence of marine origin in the southern area (south of the depression including the Lake Skallen Ôike). In the northern area, evidence of marine origin for levels above 20 m is also uncertain.

The Skallevikhalsen district, 7 km long, 1.8 km wide and 290 m high, is separated from the Skallen district by the narrow Skallevika inlet which ends at the snout of a small outlet glacier. A characteristic feature is the stepped topography with precipitous slopes and rather flat surfaces, forming a stoss-and-lee topography affected by the E-W trending and southerly dipping gneissic structure. Calcareous rocks (marble and skarn) have been eroded more intensely than other rocks. Directions of glacial striae are generally NW-SE, perpendicular to the general trend of the topography. Two cirque-like depressions cutting the gneissic structure occur as small coves in the eastern and the western parts. Joints trending NW-SE and NE-SW are developed throughout the district. The eastern cirque-like depression is formed in the place where the joints are thick. But in general, erosion along joints is insignificant.

The mechanical weathering has been advanced a little in this district compared with that in the Skallen district. The Skallevikhalsen is considered to have emerged from the ice sheet a little earlier than the Skallen, probably because of the difference in elevation of these ditricts.

Raised beaches are found on the western coast below 22 m. On the northern coast, the landform created by marine agency is indistinct, but sand and gravel containing fragments of marine molluscan shell are distributed below 25 m in a small valley down from the Lake Dairi. Raised deltaic fan is formed below 5 m at the mouth of this valley. OMOTO (1977) distinguished 12 levels of raised beaches and step landforms below 23.9 m in this district.

Ice-free areas near the snout of the Shirase Glacier: Ice-free areas south of the Skallevikhalsen are small in area, and the information of their landforms is comparatively limited.

Glaciated surfaces with striae and grooves are generally better preserved in this area than in the northern part of the Sôya Coast, except in the Austhovde district where the bedrock has been weathered considerably to form small mushroom rocks leaving the scoured surface on the top.

The major topographic trend of depressions and ridges is fairly in accord with the directions of the gneissic structure, while the direction of glacial striae and grooves often differs from this topographic trend. Intersection of striae and grooves can be seen in many places. A groove 1.5 m wide and 2 m deep trending N40°W scooped out the flat surface with striae trending E-W in the Berrodden district. MORIWAKI (1976) described the intersection of older (N73°W) striae and younger (N45°W) striae and a groove in the Rundvågskollane district. Therefore, the former ice flow is inferred to have changed its direction from E-W to SE-NW during the retreat of the ice sheet.

The direction of striation on an exposed rock rising a little from the ice surface near the right bank of the Shirase Glacier is nearly perpendicular to the direction of a maximum slope of the surrounding ice sheet. This suggests that the ice flow changed its direction from SE-NW to the present NE-SW at a fairly later stage of ice shrinkage when the shrinking ice level came close to the present level. The striation found in the Austhovde district trends in the direction of SE-NW (MORIWAKI, 1976), while the general ice sheet surface around the Austhovde slopes northward at present. The Austhovde is situated in the Botnneset area some 40 km west of the Shirase Glacier. Therefore, there is a possibility that the ice sheet had flowed from southeast to northwest beyond the Shirase Glacier some time in the past when the ice sheet occupied most of Lützow-Holm Bay (YOSHIDA and MAE, 1978). The ice flow towards northeast by the shrank ice sheet at a later stage appears to be indicated by striation trending northeast in the Djupvikneset area (YOSHIDA, M., 1975). M. YOSHIDA (1975) also described holes a few decimetres in diameter aggregating on the bedrock in the Botnneset area. He thought that they were formed by marine agency (?). Judging from his photograph, however, they seem to be glacial potholes. Curved grooves in the Rundvågskollane (MORIWAKI, 1976) and these potholes suggest the possibility of glacial erosion by the wet-based ice sheet.

The Shirase Glacier has cut the Instekleppane rock exposure to form a vertical wall some 100 m high on the west side. The areal scouring appears to be replaced by selective linear erosion (SUGDEN and JOHN, 1976) in this district, as the retreat of ice proceeded. It is difficult to estimate, however, that the selective linear erosion took place alternately or contemporaneously with the areal scouring.

Raised marine landforms are poorly developed, and often difficult to be identified in this region. OMOTO (1977) distinguished 8 levels of raised beaches and steps below 23 m in the Rundvågshetta district. Raised beaches below 5 m are composed of sand and gravel with marine silt. And a beach ridge (raised?) runs parallel to the present shoreline. He also described the wave-cut platforms at the heights of 6.5–7, 11.3 and 23–24 m. However, it is not easy to discriminate a wave-cut platform from a glacial bench. A rock bench 30 to 40 m wide and about 10 m high is found on Strandnebba. Small rock benches covered with a small amount of sand and gravel occur at 5 m high on Berrodden and 20 m high on Fleinöya (island). But the evidence of marine agency is not clear. MORIWAKI (1976) noted an indistinct raised beach 4.5 m high and an ice-pushed ridge (NICHOLS, 1966) of gravel and sand 3 m high on Strandnebba. The latter includes fragments of shells and worm tubes, and seems to have been formed very recently.

4.2.2. *Summary of glacial landforms and some discussions*

Glacial landforms of the coastal ice-free areas can be summarized as follows:

1) The coastal ice-free areas consist of low-lying hills and hill-like mountains and were subjected to glacial erosion by the ice sheet some time in the past throughout the areas. However, traces of preglacial landforms are recognized as accordant heights of summits, erosional flat or undulating surfaces, and piedmont benchland. The piedmont benchland with elevations of 40 to 50 m is often designated as strand-

flat (KOAZE, 1964), though it shows no specific relationship to marine or glacial erosion. This fact suggests that the amount of erosion by areal scouring of the ice sheet is not large enough to modify extensively the preglacial relief in this region, though there is a discrepancy in views on the total amount of denudation over the Antarctic Continent.

2) Areal scouring of the ice sheet, as is well known, acts on bedrock very selectively, and produces relief of a relatively small scale subsequent to geologic structure and other physical properties of gneisses. Shallow depressions and low ridges extending in the direction perpendicular or oblique to the former direction of ice sheet flow were formed along the banding or foliation of gneisses. They show a typical stoss-and-lee topography when the bandings dip toward the upstream side, but such characteristics are rarely seen where the dip is vertical or inclines toward the downstream side. Massive parts of gneisses often shape flat surfaces or mammilated peaks. Linear depressions were carved along joints and fractures where joints are densely spaced. Most of these depressions are shallow, but some of them are deep and narrow. Rather wide basins occur at intersecting places of two or more joint systems of different orientation. Some of them have cirque forms. Several cirques were evidently modified by cirque erosion after emerging from the ice sheet.

The general directions of the former ice flow inferred from glacial striae and grooves are almost coincident with those inferred from a larger-scale topography, and are estimated as E-W to SE-NW, roughly perpendicular to the present coastline, though crossing of striae indicating the change of flow direction is found in some places within limits of general flow directions. Some grooves and quarried surfaces, however, show local directions of ice flow considerably different from the general flow direction. There is a possibility that a part of a basal layer of the ice sheet flowed in a fairly different direction from that of the overlying ice layer. One example of large-scale appears to be the case of the Langhovde. Striae indicating the sheet flow from east to west are found on the bedrock very close to the surface of the Langhovde Glacier flowing toward north. This suggests that the former ice sheet flowed toward west, overriding the Langhovde ice-free areas, while the lower part of the ice flowed northward along the depression beneath the Langhovde Glacier, though more field evidence would be necessary for detailed discussion.

3) Glacial abrasion is thought to be ineffective beneath the polar glaciers (BOULTON, 1974), because the basal sliding of ice on bedrock is hard to occur due to low ice temperature below the pressure-melting point. Glacial striae and grooves are regarded as one of the two strongest proofs that a region has been glaciated by a wet-based glacier (BLOOM, 1978). These small-scale features are widely distributed in the ice-free areas along the Sôya Coast, suggesting that the wet-based ice sheet has abraded this region. Especially, curved striation and a groove with over-hanging side wall may support this inferrence. However, the field evidence other than striation and a groove for the wet-based state of the former ice sheet has not been obtained yet. It is also difficult to judge whether the present sheet flow of ice is wet-based in its marginal part where the ice becomes thin. MAE and NARUSE (1978) pointed out the possible basal sliding beneath the ice sheet of the Shirase drainage basin 200 km inland from the coast. However, phenomena indicating basal melting of the present

ice sheet are not known along the coastal zone.

4) On the other hand, an ice stream in the ice sheet has cut a glacial trough of a large-scale at a limited place. The down-cutting of troughs leaving intervening slopes or flat surfaces unmodified or areally scoured is often described as selective linear erosion. However, the glacial erosion has proceeded to form precipitous cliffs, often irrespective of the local geologic structure. Many of ice streams are likely to be formed along preglacial depressions of the region, though they are not always situated in the places where long-continued preglacial valleys can be assumed. Conspicuous cliffs such as Skjegget in the Skarvsnes and Mt. Tyôtô in the Langhovde districts are side walls of asymmetric U-shaped valleys. They were possibly formed by ice streams of which the one side was confined by a local rise of bedrock and the other side by rather sluggish ice sheet, as seen in the cases of the present Shirase Glacier and the Honnör Glacier. OMOTO (1977) considered that a trough wall of the one side had been removed by the glacial erosion. Such possibility cannot be completely denied at present. However, the areal scouring seems to be difficult to erode away a trough wall of the one side. And, at least in Skjegget, the geomorphic evidence indicating that the selective linear erosion removed a trough wall of the one side has not been obtained.

5) It is difficult to know when the terminal of the ice sheet having exerted glacial erosion to the area receded to the present position. Radiocarbon dates of organic remains deposited *in situ* and contained in raised beach deposit unmodified by glacial erosion indicate that the ice sheet had receded from main ice-free areas such as Langhovde, Skarvsnes, and the Ongul Islands prior to 30000 years B.P. at least.

It is also difficult to judge whether there was a stagnant stage or readvance of the ice sheet during shrinkage. But formation of cirque glaciers in the Langhovde district probably took place during a stagnant stage. The fact that the weathering in a distant part from the margin of the ice sheet has fairly progressed than in a close part to the margin in many ice-free areas suggests the stepped retreat of the ice sheet margin, though no distinct terminal moraines can be found.

6) The ice-free areas have been subjected to frost-heaving congelifraction and wind abrasion of bedrocks and cryoturbation of unconsolidated materials under a literally periglacial environment, after the ice receded from the region. However, these agencies are weak in general, and the modification of landforms by periglacial processes is small in quantity. In general, weathering of bedrock has more progressed in the northern part of the Sôya Coast than in the southern part except in very small-scale ice-free areas, suggesting that the ice sheet had receded at least a little earlier from the northern part than from the southern part.

The development of patterned ground is not much remarkable, because of less severe climate in the case of non-sorted patterned ground compared with that in the McMurdo Sound region (BERG and BLACK, 1966), and because of fewer freeze-thaw cycles in the case of sorted patterned ground compared with those in lower latitude regions. Difference in degree of development of sorted pattern-ground in the Prince Olav and Prince Harald Coasts appears to be caused by several factors such as the amount of finer materials in deposits and the thickness of deposits, among which the supply of meltwater seems to be the main factor.

Sorted patterned ground emerging from the snow field in the Langhovde district suggests that this patterned ground might have been formed some time in the past when the snow accumulation was less than at present.

4.2.3. Summary and discussions on raised beach topography

Distribution and characteristic features: Raised marine features are widely distributed in ice-free areas on the Prince Olav and the Sôya Coasts, indicating that the whole region had been submerged partially by the sea after deglaciation, and then has been elevated to some extent. Characteristics of raised marine features in this region are summarized as follows: 1) wave-cut bench and sea cliff develop only in very limited places, as wave action has been very weak, 2) most of raised beaches consist of sand and gravel derived from glacial drifts deposited in rather sheltered places as ground moraines, 3) beach deposits often contain fossiliferous sand and silt, 4) wave-washed features such as marine-boulder pavement develop better on the Prince Olav Coast than on the Sôya Coast, 5) raised beaches are often marked with stepped topography composed of berm-like steps and beach faces, indistinct beach ridges, or ice-pushed ridges, 6) raised beach topography has not been subjected to ice sheet erosion, and 7) pitted beaches are rarely found, and there is no beach

Table 3. Some characteristics of raised beaches on the Prince Olav and the Sôya Coasts.

Location	Maximum elevation above sea level (m)	Remarks
North coast of Skallevikhalsen	22	Gravelly sand deposited in a small valley
East coast of Skallen	8–10	Raised beaches composed of silty sand, associated with low sea cliff.
Northeast coast of Skallen	15	Boulderly morainic deposit mixed with marine sand containing shell fragments.
Lake Hunazoko, Skarvsnes	12	Raised beaches on a flank of Lake Hunazoko basin.
Northwest of Lake Suribati, Skarvsnes	16	Gravelly sand with many worm tubes in growth position, at a col separating Lake Suribati from the sea.
East of Lake Suribati	30?	Sand and gravel.
Kizahasi Beach, Skarvsnes	19	Raised beaches with many steps.
Southwest coast of Breidvågnipa	10	Pocket beach composed of boulderly deposit.
Northwest coast of Langhovde	20	Pocket beach, marine boulder pavement, containing shell fragments.
North coast of East Ongul Island	16	Raised beaches with several steps, containing shell fragments.
Northeast coast of West Ongul Island	15	Gravelly sand on a terrace-like surface.
North coast of Cape Hinode	27 35?	Thin boulderly deposit on a marine terrace, marine boulder pavement.
Northeast coast of Sinnan Rocks	31	Ice-pushed ridge composed of boulderly deposit, associated with rudimentary pitted beach.
Northwest coast of Akarui Point	10	Gravel with sand, mixed with shell fragments.

resting on ice slab (ice-foot), but perennial drift-snow ice covers raised beaches in some places (Table 3).

Wave action is very weak in littoral areas around the Antarctic Continent where fast ice and pack ice cover the surface of the sea almost all the year round. This is indicated by the fact that a shallow wave-cut bench or accumulation terrace is rarely found in the sea near the coast. It can be presumed that wave action might have been stronger some time in the past when the sea ice cover was small in quantity than at present or absent. However, raised wave-cut bench or sea cliff is generally difficult to be discriminated with ample evidence from a glaciated landform in this region. Time-scale during which wave action has worked might be too short for sculpturing hard bedrock to form distinct wave-cut bench and sea cliff. Only indistinct wave-cut bench and sea cliff are recognized in very limited places associated with sea-water worked sediments. Raised beach deposits were derived mainly from morainic deposits so that distinction between them is often difficult, though HAYASHI (1979) showed that flatness of pebbles was useful for discrimination of marine from morainic ones. Therefore, clear evidence for marine origin of deposit is organic remains in the deposit, in particular, which are associated with silty materials. It is corroborated by topographic features. Most of "raised beach" topography above the level of about 20 m high lack this most convincing evidence. Many of organic remains appear to have been settled *in situ*. However, fossil shells are or have been easily fragmented because the calcification in Antarctic water is weak.

Marine-boulder pavement must have been formed by wave-wash of a morainic deposit. It develops better on the Prince Olav Coast than on the Sôya Coast, probably reflecting the difference in intensity of wave action between the coast facing to the ocean and the coast more secluded from the ocean. Wave-wash seems too weak to form the typical marine-boulder pavement at present. Therefore, there is a possibility that the raised marine-boulder pavement was formed under the condition of less sea ice in the past than at present.

Many steps occur where raised beach deposits accumulated continuously on bedrock from the upper to the lower slopes near the strand-line. The steps must have been formed successively from the upper to the lower levels during the relative lowering of the sea level. It is difficult, however, to estimate the significance of standstills of the sea level during its lowering in these places. Stepped topography in other places, marked with ice-pushed ridges, wave-washed deposits, or benches in some places, seems to correspond a little more distinctively to standstills of the sea level.

Radiocarbon dates of fossil marine organisms and formation of raised beach: Radiocarbon datings have been conducted on organic remains contained in raised beach deposits. Materials used are shells of pelecypod (*Adamussium colbecki*; *Laternula elliptica*), worm tubes, tests of Foraminifera, and shell fragments. Coralline algae adhered on bedrock were also dated. Most of samples were collected from East and West Ongul Islands along the Sôya Coast, the Langhovde district, the Skarvsnes district, and one sample from Akarui Point of the Prince Olav Coast (Table 4).

Problems in considering the relationship between these dated samples and raised beachs are as follows: 1) habitats of dated animals were not always close to former sea levels, but in a wide range of water depth, 2) it was often difficult to collect samples at horizons of beach deposits favourable for the dating of raised beaches themselves, because organic remains are distributed very partially in the deposits, and 3) radiocarbon dates are sometimes problematical, especially in Antarctica.

The altitudes of occurrences and dates of organic remains, therefore, give the only minimum height of the former sea levels at approximate ages indicated by radiocarbon dates. Radiocarbon dating is based in principle on the assumption of constant radiocarbon activity over time. Uncertainty of this assumption increases when time becomes older. Therefore, old radiocarbon dates over 20000 years B.P. cannot be discussed precisely. OMOTO (1977) also pointed out the possibility of large counting error in the determination of radioactivity of old samples. This causes a wide range of radiocarbon ages. Living marine animals in the Antarctic waters are known to be dated as old ones, not modern, by radiocarbon method. OMOTO (1972) pointed out that sea water in Lützow-Holm Bay includes carbon of low ^{14}C concentration. Other possibilities to cause deficiency of ^{14}C may be the interference by ice against exchange of carbon dioxide between atmosphere and sea and ecosystem having a simple food chain. Apart from these causes, analysis of living marine organisms was attempted to obtain a reasonable correction value for radiocarbon datings (YOSHIDA and MORIWAKI, 1979). Samples were fish, sea urchin, sea snail and starfish (Table 4g). The values obtained are considerably uniform, the average being 1120 years old. STUIVER *et al.* (1976) suggested the correction factor of 850 to 1400 years for the radiocarbon dates of *Adamussium colbecki* collected from the McMurdo

Tables 4a–4f. The results of radiocarbon datings on marine organisms contained in the beach deposits.

Table 4a.

Date of sampling	Locality (East Ongul Island)	Elevation above sea level (m)	Sample	Age (yr.B.P.)
3 March 1960	Kai-no-hama Beach	3–4	*Adamussium colbecki**	3840±110
25 Jan. 1961	Kitami Beach	5–6	Fragments of molluscan shell**	over 30000
19 Jan. 1962	Kitami Beach	7–8	Fragments of molluscan shell***	25400±200
〃	〃	12	〃	34000^{+3000}_{-2000}
〃	Kai-no-hama Beach	9–10	〃	22800±1000
〃	〃	3–4	〃	29500^{+2400}_{-1800}
〃	Kitami Beach	7–8	Tests of Foraminifera***	31200^{+2500}_{-1900}

* Collected by Y. YOSHIDA.
** Collected by H. MEGURO.
*** Collected by K. FUJIWARA and H. KOAZE.

Remarks: Analysis was made by K. KIGOSHI, Gakushuin University, except in the case of TH-020. 5570 years is used as a half life of C. Yr.B.P. means years before 1950. 95% of N.B.S. standard oxalic acid was used as the reference. The error shows standard deviation calculated from statistical errors in counting of beta rays.

Table 4b.

Date of sampling	Locality	Elevation above sea level (m)	Sample*	Age (yr.B.P.)	Code No.
1 Feb. 1967	Kizahasi Beach, Skarvsnes	1.8	*Adamussium colbecki*	3600±100	GaK-2035
〃	〃	8	〃	4700±100	GaK-2034
2 Feb. 1967	Shore of an inlet, southwestern part of Skarvsnes	0.5	*Laternula elliptica*	3180±250	GaK-2039
3 Feb. 1967	Near shore of Lake Hunazoko, Skarvsnes	−23	〃	4190±100	GaK-2037
〃	〃	8	Fragments of molluscan shell	31600^{+2800}_{-2100}	GaK-2036
7 Oct. 1967	Shore of Lake Suribati, Skarvsnes	−30	Worm tubes	5640±130	GaK-2038
13 Feb. 1967	Northern part of East Ongul Island	16	Fragments of molluscan shell	5850±100	GaK-2032
1 March 1967	Mizukumi Stream, East Ongul Island	12	Fragments of molluscan shell	30700±2000	GaK-2033

* Collected by Y. YOSHIDA.

Table 4c.

Date of sampling	Locality	Elevation above sea level (m)	Sample	Age (yr.B.P.)
1970	East Ongul Is.	0.5	Coralline algae*	3340±90
〃	〃	1.5	Coralline algae*	3540±90
〃	〃	0.8	*Adamussium colbecki**	2040±90
〃	〃	−9	Shell of living sea urchin	150±80
10 Oct. 1967	Langhovde	2	Fragments of molluscan shell**	2000±220

* Collected by T. HOSHIAI. ** Collected by Y. YOSHIDA.

Table 4d.

Locality No.	Elevation above sea level (m)	Sample*	Age (yr.B.P.)	Code No.
Langhovde 03	5–6	*Laternula elliptica*	23830±910	GaK-4148
〃 04	1.5	*Adamussium colbecki*	4290±90	GaK-4151
〃 07	6	*Adamussium colbecki*	10250±210	GaK-4150
〃 08	6	*Laternula elliptica*	over 33400	GaK-4149

* Collected by K. MORIWAKI in summer of 1971–1972.

Table 4e.

Date of sampling	Locality	Elevation above sea level (m)	Sample	Age (yr.B.P.)	Code No.
1969	Near Lake Suribati	14	*Laternula elliptica**	6020±175	TH-020
1972	Simo-kama, Langhovde	1.5	*Laternula elliptica***	3840±90	GaK-4850

* Collected and analysed by K. OMOTO. ** Collected by T. ISHIKAWA.

Table 4f.

Date of sampling	Locality	Elevation above sea level (m)	Sample	Age (yr.B.P.)	Code No.
1974	Northern part of West Ongul Is.	below 8	*Adamussium colbecki* *	930±90	GaK-5832
Oct. 1974	South coast of Lake Suribati, Skarvsnes	below 6	Worm tubes*	6090±90	GaK-5840
Jan. 1975		15±5	Worm tubes**	7830±280	GaK-5837
	Kizahasi Beach, Skarvsnes	11	*Laternula elliptica* **	5580±180	GaK-5835
Feb. 1975	South coast of Osen, Skarvsnes	8	Worm tubes**	8370±270	GaK-5833
Jan. 1975		6	*Laternula elliptica* **	4430±90	GaK-5841
	Southernmost part of Skarvsnes	3		3370±120	GaK-5836
	Southwest coast of Lake Hunazoko	4		2540±160	GaK-5834
Feb. 1975	Akarui Point	10	Fragments of shell***	7730±110	GaK-5839

* Collected by K. Moriwaki. ** Collected by M. Hayashi.
*** Collected by Y. Yoshida and K. Moriwaki.
Reference: *Adamussium colbecki* collected at the mouth of Taylor Valley 0.8 m above sea level on 1972 by the present author shows age of 4360±110 years B.P.

Table 4g. Radiocarbon dates on living marine organisms.

Sample	Sampling depth (m)	Years before 1950	Code No.
Neoliuccinum eatoni dated on flesh	17~35	1190±90* ($\Delta^{14}=-137.9\pm9.2$‰, $\delta^{13}C=-19.1$‰)	Gak-6789a
Sample is same as GaK-6789 dated on shell	〃	1300±90	Gak-6789b
Ophionotus victoriae dated on flesh	92	1070±90* ($\Delta^{14}=-124.6\pm9.7$‰, $\delta^{13}C=-12.9$‰)	Gak-6790a
Sample is same as GaK-6790a dated on shell	〃	1210±100	Gak-6790b
Sterechinus neumayeri dated on flesh	17	1160±110* ($\Delta^{14}=-134.4\pm12.2$‰, $\delta^{13}C=-10.47$‰)	Gak-6791a
Sample is same as GaK-6790a dated on shell	〃	860±110	Gak-6791b
Trematomus berunacchii	15	1160±110* ($\Delta^{14}=-148.4\pm8.8$‰, $\delta^{13}C=-19.4$‰)	Gak-6792
Zoarcidae sp.	500	1010±110* ($\Delta^{14}=-118.2\pm12.5$‰, $\delta^{13}C=-21.6$‰)	Gak-6793

Samples were collected by Dr. T. Hoshiai from the sea near Syowa Station during October and December 1975. Datings were made by Dr. K. Kigoshi. Isotope fractionation correction was made on dates with asterisks.

Sound region. It is noteworthy that these values are fairly coincident with those obtained from the Lützow-Holm Bay region for the present. However, this indicates also that ages of organic remains may include uncertainty of the order of 1000 years in addition to the standard deviation of determination, even in younger ages.

In spite of the above shortcomings of radiocarbon datings, some inferences can be drawn from dates and raised beach topography. Abundance of samples makes up for these defects to some extent. Relationship between the radiocarbon dates of marine organic remains and the elevations of their occurrence is shown in Fig. 13 (The correction factor of 1120 years is not adopted). Radiocarbon dates ever obtained may be classified roughly into two age groups, the one belongs to the "post-Glacial" age, the dates falling between about 2000 and 10000 years B.P, and the other belongs to ages between 22000 and 34000 or more years B.P.

As to the younger group, the following discussions can be made. Dates of organic remains occurring at higher sites of raised beaches (or the highest raised beach ever dated) are concentrated around 6000 years B.P. In the McMurdo Sound region, the highest elevated beach is thought to have been formed 7000 years B.P. (NICHOLS,

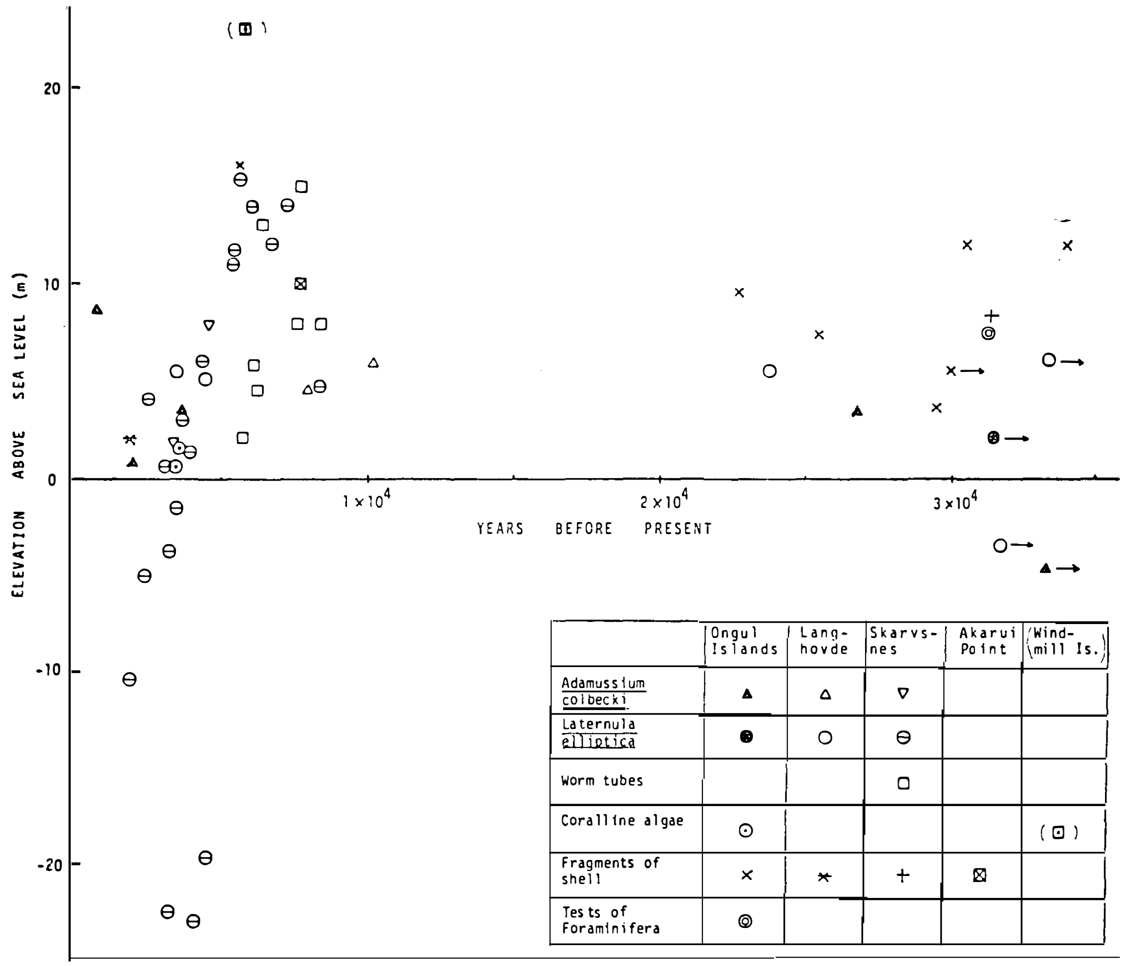

Fig. 13. Relationship between elevations of sampling sites and ^{14}C ages (The result from the Windmill Islands (CAMERON and GOLDTHWAITE, 1961) is included).

1966). A 6000 years old terrace was found at the elevation of 23 m on the Windmill Islands (CAMERON and GOLDTHWAIT, 1961). These facts seem to indicate that after deglaciation of the ice sheet from coastal areas, the rate of sea-level rise had been greater than or at least equal to the rate of crustal uplift prior to 6000 years B.P. in these regions. OMOTO (1977) thought that both rates had been almost equal. (If the correction factor of 1000 years is applied, then the dates are reduced by 1000 years from measured values.)

Crustal uplift on the Sôya Coast after 6000 years B.P. amounted to at least 16 m, if the sea level at 6000 B.P. was nearly the same as that at present. Judging from the freshness of the raised beach topography and from a wide range of habitat of dated marine organisms, raised beaches at least up to 20 m high might have been formed during the last 6000 years. Therefore, the rate of uplift in the last 6000 years is estimated to be 2.7–3.3 mm/year. OMOTO (1977) estimated the rate of uplift to be 4 mm/year in the last 10000 years, assuming that the 10000 years old *Adamussium colbecki* collected from the Langhovde district at the level of 6 m above the present sea level (MORIWAKI, 1974) was deposited 10000 years before *in situ* at the beach of those days 36 m below the present sea level. The sea level of 10000 years ago is based on either SHEPARD (1963) or FAIRBRIDGE (1961) curve. This seems to indicate that the rate of uplift has decreased from a little more than 4 mm/year to less than 3.3 mm/year, though the uplift since 6000 years ago is not always clear in the Langhovde because of no adequate sample. OMOTO (1977) also inferred that the isostatic upheaval in East Ongul Island and the Langhovde district decreased between 1000 years B.P. and the present, and in the Skarvsnes district it ceased about 2000 years before, from the graph showing interrelationship among sea level, radiocarbon age and total upheaval of land. It is difficult to judge whether the uplift is in progress in this region or not. Lake Nurume in the Langhovde district is a small lake separated by very low (about 50 cm high) rock and debris threshold from the sea. The water of the lake shows a two-layered structure (SANO *et al.*, 1977), clearly derived from sea water. This indicates that the lake had been connected to the sea, then had been cut off from the sea very recently by crustal uplift. Therefore, the uplift in this region seems to continue up to the present.

No radiocarbon date has been obtained as to the raised beach on the Prince Olav Coast except at Akarui Point. Shell fragments obtained from the raised beach 10 m high were dated as 7700 years B.P. at Akarui Point. This suggests that the mode of uplift of the area may be similar to that on the Sôya Coast, though fewer but more distinct steps of raised beaches are found on the Prince Olav Coast than on the Sôya Coast where rather continuous crustal uplift has been taking place. Judging from the freshness and common characteristics such as elevations, boulderly deposits, and degree of development of topographic features of raised beaches and marine terraces, the "post-Glacial" step levels may be traced up to the level probably 25–27 m high and possibly 30–32 m high. Therefore, the rate of uplift can be a little higher on the Prince Olav Coast than on the Sôya Coast.

On the other hand, it is difficult to interpret pertinently the relation between the raised beach topography and the organic remains of the older age group, ranging from 22000 to 34000 years, at the present state of knowledge.

It is sometimes difficult to judge whether the raised beach topography has been subjected to glaciation of the readvancing ice sheet or not. Some "residual beaches" of the Eemian Interglacial is considered to have survived the glacial destruction during the Weichselian Glacial in the South Shetland Islands area (JOHN, 1972). In the Prince Olav and the Prince Harald Coasts region, however, the raised beaches have not been affected by the ice sheet glaciation, because 1) the raised beach topography is well-preserved, and is not covered with morainic or erratic detritus, and 2) beach deposits have not been disturbed by glacial movement as is indicated by the presence of *in situ* fossils. Therefore, it can be safely said that before 30000 years at the latest the deglaciation of the ice sheet took place in the coastal area of the Prince Olav and the Prince Harald Coasts, and since that time the exposed land areas having raised beaches of older ages have not been covered with the ice sheet. This conclusion seems to be supported by the fact that weathering of bedrock is considerably proceeding in this region compared with that in the Antarctic Peninsula region where the readvancing ice exerted glaciation on coastal areas during the Late-Würm Glaciation.

After the retreat of the ice sheet, the coastal areas were partly inundated by the sea, and the glacial marine deposits and beaches were formed in favourable places. The high stand of sea level relative to land around 30000 years ago seems to correspond to the so-called Mid-Würm Interstadial, though the height of the sea level at that time relative to the present sea level is still uncertain not only in Antarctica but all over the world. Topographic relationship between the post-Glacial and the older raised beaches on the Prince Olav and the Sôya Coasts is still problematical. FUJIWARA (1973) considered that the sea level relative to land had lowered successively from the mid-Würm sea level to the present one. But the lack of fossils with radiocarbon dates between 10000 and 22000 years before suggests that the sea level had lowered and stayed below the present sea level during that time.

HOLLIN (1962) postulated the highly positive regimen of the present and the past Antarctic ice sheets, because the wastage is performed mainly through calving of icebergs which is controlled not by climate but only by the altitude of sea level, and maintained that the fall of sea level would induce the readvancing of the ice sheet, resulting in thick ice cover over the coastal area. In the Prince Olav and the Sôya Coasts region, however, the present ice-free areas have not been covered with the thick readvancing ice sheet after 30000 years before, as stated above. SCOTT (1905) proposed a paradox that during a colder stage in Antarctica, the accumulation could be reduced considerably, and the ice sheet might have expanded not so much. Fluctuation of the Antarctic ice sheet, especially in much old times (Pliocene–early Pleistocene?), has been not always synchronous with that of the ice sheets in the Northern Hemisphere (WILSON, 1964; CALKIN *et al.*, 1970; DENTON *et al.*, 1970). In this regard, SCOTT's hypothesis which seems to be applicable better to the East Antarctic ice sheet than to the West Antarctic one should be reevaluated or tested.

Readvance of the ice sheet postulated by HOLLIN is based on the seaward advance of "grounding line" which may be effective to an ice shelf floating in the relatively shallow waters. Such situation would be highly possible in West Antarctica where low-lying bedrock makes the ice sheet susceptible to fluctuation of sea level and ice shelves fringe most parts of the coastline. In the Prince Olav and the Sôya Coasts region where no ice shelf exists the circumstances might be different from those in West Antarctica. The fall of sea level by about 100 m might not cause the vast thickening of the ice sheet. It is possible that outlet glaciers had advanced during the low stand of sea level, but evidence has not yet been obtained.

Rise of sea level during "post-Glacial", culminated in around 6000 years before, caused deposition or rework of marine sediments on formerly formed, unconsolidated deposits. The highest altitude of older sea level inferred from the raised beach topography is not always certain. Organic remains with older dates ever obtained occur at sites below the highest site of fossils with younger date. This suggests that altitudes of the older raised beaches are not so different from the younger ones. If it is assumed that the highest level identified by OMOTO (1977) is correct and is 30000 years old, and sea level at that time was 10 m below the present sea level, the rate of uplift between 30000 years and 6000 years ago would be about 1 mm per year at most. This estimation is naturally not well-grounded. However, the above argument shows that the uplift of the region during that time was stagnant or considerably slower than that of after 6000 years ago, in spite of the major ice retreat before more than 30000 years.

OMOTO (1976a, 1977) estimated the amount of uplift due to isostatic rebound in this region as 200 to 250 m, based on the estimation that the former ice sheet was 1250 m thick on the present coastal region. He also considered that this figure supported the view that the foraminiferal assemblage with date of 31200 years B.P. collected from the site 8 m high on East Ongul Island had been deposited on the sea bottom 100 m deep. An actual amount of crustal uplift due to isostatic rebound is difficult to know, because we can estimate the amount of uplift only from the raised marine features in this area at present. Even if the 250 m uplift has taken place since the major retreat of ice, this uplift cannot be connected directly to the foraminiferal assemblage of 30000 years old, because it is difficult to find any highly raised marine features corresponding to such uplift. HENDY *et al.* (1969) described a marine sediment containing organic remains with radiocarbon dates between 37000 and 32000 years B.P. near Cape Barne of Ross Island. They deduced that the sediment was deposited on the sea bottom 180 m deep during the Mid-Würm Interstadial, and was uplifted to the present location 30 m high by isostatic rebound due to shrinkage of the ice sheet. But this conclusion was denied by BRADY (1977) from the study of fossil diatom in the sediment.

On the other hand, KAMINUMA *et al.* (1979) showed that elastic uplift of the crust caused by 50 km retreat of the marginal zone of the ice sheet which has a radius 1700 km long and the maximum thickness of 4000 m could amount to several tens of metres. Apart from the mechanism, isostatic or elastic, the view that the crustal uplift indicated by raised marine features was caused mainly by rebound of the crust caused by ice shrinkage seems to be reasonable, taking into account the widely occurring phenomenon around the Antarctic Continent and the high rate of uplift on the tectonically stable continent. On the other hand, the existence of a deep continental shelf and the proximity of ice-free areas to the present ice sheet seem to support the view of smaller magnitude of the crustal uplift.

Many case studies in other parts of Antarctica naturally indicate that there are similarity and dissimilarity in the development of raised beaches among different regions, as well as in the opinions on the ages of formation and levels of raised marine features. For instance, terraces around 25 m high above the present sea level were suggested to have been formed in a considerably old time (not mentioned clearly, but it seems to the present author to mean the last Interglacial), and the beach 10 m high was considered to have been formed during the "post-Glacial" climatic optimum, on Inexpressible Island in the Ross Sea (CLARIDGE and CAMPBELL, 1966). Similarly, two raised beach levels 7.7 m and 3 m high have been assigned to the "last Interglacial" and the "post-Glacial", respectively, near Cape Hallet in the northern Victoria Land, on the basis of geomorphology and degree of soil formation (CAMPBELL and CLARIDGE, 1966), while a raised beach 20 m high in the McMurdo Sound region is considered to have been formed during the "post-Glacial" period. Marine sediments higher or older than the raised beach 20 m high in the Ross Sea region are still problematical, as mentioned above about the work by HENDY *et al.* (1969). SPEDEN (1962) identified two formations of marine sediments around the McMurdo Sound. Of them, the younger formation was considered to be the post-Glacial, while the older occurring at the height of 180 m to be the early Pleistocene. However, BULL and WEBB (1973) correlated the latter, at least some equivalent of them, to the Pliocene. DENTON *et al.* (1970) obtained a radiocarbon age of >49000 years B.P. from shells contained in raised marine beds located at 59–63 m above the present sea level and an age of >47000 years B.P. from shells in the other deposit located at 29–32 m above the present sea level on Cape Barne, Ross Island. The shells obtained from the same locality as the latter case were dated as 34800 years B.P. by HENDY *et al.* (1969), as mentioned before. As the marine beds are covered with glacial drift, DENTON *et al.* assigned them to the interval between Ross Sea Glaciations I and II. Cores obtained by the Dry Valley Drilling Project are providing much information on the glacial history of this region. But most of them are old in age and their relation to the raised beach topography is not yet clear.

In the Windmill Islands on the Budd Coast, raised beaches develop below 30 m, and coralline algae obtained from the beach 23 m high were dated to be 6040 years B.P. (CAMERON and GOLDTHWAIT, 1961). This may be the highest beach ever dated as post-Glacial in East Antarctica.

On the other hand, many raised beaches have been identified in the Antarctic Peninsula region. Of these, a raised beach 54 m high on King George Island is the highest which has been assigned to the post-Glacial, as mentioned above (JOHN, 1972). A large amount of crustal uplift may be caused by the ice shrinkage of a considerable scale compared with the ice mass which covers the greater part of the island at present (namely, the ice mass which covered the island and the surrounding island shelf was considerably large compared with the present ice mass on the island). However, the view that the residual beach up to 275 m high assigned to the last Interglacial was uplifted by the mechanism of "bulging" (compensational uplift which accompanied renewed ice-sheet growth and isostatic depression in the continental interior) should be reexamined.

As mentioned above, the "post-Glacial" raised beaches are widely distributed on the Antarctic Continent, though their altitudes differ in each area, reflecting the regional difference in topographic and crustal situations. However, problems of raised beach development in older ages still await further detailed studies.

Summarizing the above argument, the following geomorphic sequence concerning the raised beach topography is inferred. After the ice sheet had receded from the coastal region, some land areas emerged from the ice and were partly covered with sea water. Morainic materials were subjected to weak marine agency and were reworked to some extent to form glacial marine sediments, as the sea level rose up relative to land. Marine organisms lived and were deposited on glacial marine sediments around 30000 years B.P. Then, a relative upheaval of ice-free areas took place after 22000 years B.P., resulting in the formation of "older" raised beaches. Possibility that the fluctuation of the relative sea level occurred between 30000 and 22000 years B.P. (MORIWAKI, 1974) is not excluded, but the insufficient data prevent any definite conclusion to be reached. Some time before 10000 years ago, the sea level began to rise, corresponding probably to the eustatic rise of sea level during the "post-Glacial" period. On the other hand, the land mass which had emerged from the ice was uplifted by crustal rebound caused by unloading of ice. But the rate of uplift was less than the rate of sea level rise at least between 10000 and 6000 years B.P. The preexistent raised marine sediments were reworked weakly, and some clastics together with marine organisms were deposited, as the rise of sea level proceeded. The rise of sea level culminated at around 6000 years B.P. Then the uplift of land mass has left the "recent" raised beach topography below the level of about 20 m high on the Sôya Coast and 25 m high on the Prince Olav Coast. Older raised beaches have reached likely 35 m high at the maximum up to present. The rate of uplift since 6000 years B.P. is estimated to be at least 2.7 mm/year but probably 3.3 mm/year. The rate of uplift seems to have been much accelerated in the "post-Glacial" age. It is inferred from the nature of raised beach deposit that there was a period between 6000 years B.P. and the present when sea ice and snow were less than those at present.

5. Some Problems on Submarine Topography

5.1. Characteristics of the continental shelf in Antarctica

The most characteristic features of the continental shelf surrounding the Antarctic Continent are its great depth and highly irregular surface. The deep shelf, not only its edge but the greater part of it, has attracted attention for a long time, and several hypotheses have been proposed on its origin, such as, continental flexure (FAIRBRIDGE, 1952), glacial erosion by the ice sheet (SHEPARD, 1931), and isostatic subsidence caused by loading of the ice sheet (many authors). The last is thought to be the major cause, but the age of its occurrence has not been determined yet.

The irregular surface of the continental shelf is characterized by elongated depressions, circular or oval basins of various sizes, and rises of ledges and morainic deposits. The transverse (longitudinal to elongation of the shelf) depression of a large scale, extending more or less parallel to the coastline from the Davis Sea to the d'Urville Sea off East Antarctica, was considered to have been formed by a regional tectonic movement during the Quaternary (LISITZIN and ZHIVAGO, 1960), though its asymmetric cross profile with a steep slope on the landward side is thought to have been modified by glacial erosion (NICHOLS, 1966). However, this depression was found to be not so continuous but separated into several depressions. Most of the depressions might have been formed by glacial erosion. But the view that ancient tectonism might be responsible for the primary shaping of two transverse depressions east of 135°E is thought to be valid (GRINNELL, 1971). Submarine valleys extending from the outer edge of the shelf to the lower margin of the continental rise are also recognized by GRINNELL.

On the other hand, depressions of a smaller size in the inner part of the continental shelf are mostly longitudinal (transverse to elongation of the shelf) and more or less perpendicular to the coast line. Their features are similar to those of longitudinal depressions found in glaciated regions of the Northern Hemisphere (SHEPARD, 1931), though the former are generally much deeper. They are thought by many researchers to have been shaped by glacial sculpture of formerly extending ice sheet.

Conspicuous ridges and shoals are also distributed on the continental shelf. Several large banks were regarded as morainic heap (for instance, GOULD, 1940). However, glacial marine sediments cover the greater part of the ice shelf (HOUGH, 1950), and distinction between moraines formed at former glacier termini and deposits transported by icebergs is not always easy. The Pennel and Mawson Banks in the Ross Sea are inferred from the topography to be primarily the relief of basement rocks, contrary to the old views.

The continental shelf is also characterized in many places by an irregular inner and a smooth outer surfaces (LISITZIN and ZHIVAGO, 1960). The outer smooth shelf is often considered to be of accumulation origin (for instance, NEETHLING, 1972).

It seems most likely that nearly the whole area of the Antarctic continental shelf was once or several times covered with the ice sheet. HOLLIN (1962) discussed that isostatic subsidence of the continental shelf took place at the first glaciation of the whole Antarctic Continent and has been maintained since that time. And the shelf underwent the repeated ice cover induced by eustatic lowering of the sea level. Primary sculpturing of submarine troughs off the Wilson Piedmont Glacier in the McMurdo Sound region was considered to have taken place long before (BARRETT *et al.*, 1974), correlated to the stage of glacial erosion of the trunk "Dry Valleys" in the southern Victoria Land (DENTON *et al.*, 1970; CALKIN *et al.*, 1970). The glaciers from the plateau had not flowed through trunk valleys since the oldest glaciation in the Dry Valleys region. Therefore, the later glacier cover on preexistent submarine troughs, if it had occurred, might have been conducted by the grounded Ross Ice Shelf.

5.2. Characteristics of submarine topography off the Prince Olav and the Prince Harald Coasts

The foregoing outline of the Antarctic continental shelf topography includes many problems to be solved in the future, hence gives a kind of framework to the geomorphic study of the continental shelf in the Prince Olav and the Prince Harald Coasts region. Intensive surveys in Lützow-Holm Bay (FUJIWARA, 1971; OMOTO 1975, 1976a; MORIWAKI, 1975, 1979) revealed the detailed feature of the Antarctic continental shelf close to the ice sheet, though they cover a limited area at present. The following descriptions are based mainly on the results obtained by these studies.

Outline of the continental shelf: The margin of the continental shelf off the Prince Olav Coast extends from northeast to southwest in parallel with the coastline at a distance of 60–70 km. The width is larger off the Prince Olav Coast than off the Enderby Land, but narrower than the average of the Antarctic continental shelf (BELINSKAYA, 1968). The depth of the outer edge of the shelf is 350 to 450 m in general, but it decreases up to 250 m in some places. The outer edge seems to be indented a little in the north of the central part of Lützow-Holm Bay, increasing its depth up to 600 m. The same is recognized in Amundsen Bay, Enderby Land. To the west, the Gunnerus Bank (or ridge) extends to the north some 300 km from the northern tip of the Riiser-Larsen Peninsula, increasing its depth gradually from 400 to 2000 m. No apparent break in a cross section can be seen between 400 and 2000 m deep (Fig. 14).

The continental slope is divided into the upper and the lower slopes at a depth of about 2000 m. The upper is considerably steep and narrow. Submarine contour of the upper slope suggests that shallow submarine valleys cut the upper slope. The lower slope descends gently down to the depth of 3000 to 3500 m.

The shelf is divided topographically into two parts, as is the case with many other parts of the Antarctic. The uneven inner part is characterized by submarine troughs and basins, and the outer part by low relief of broad depressions and rises.

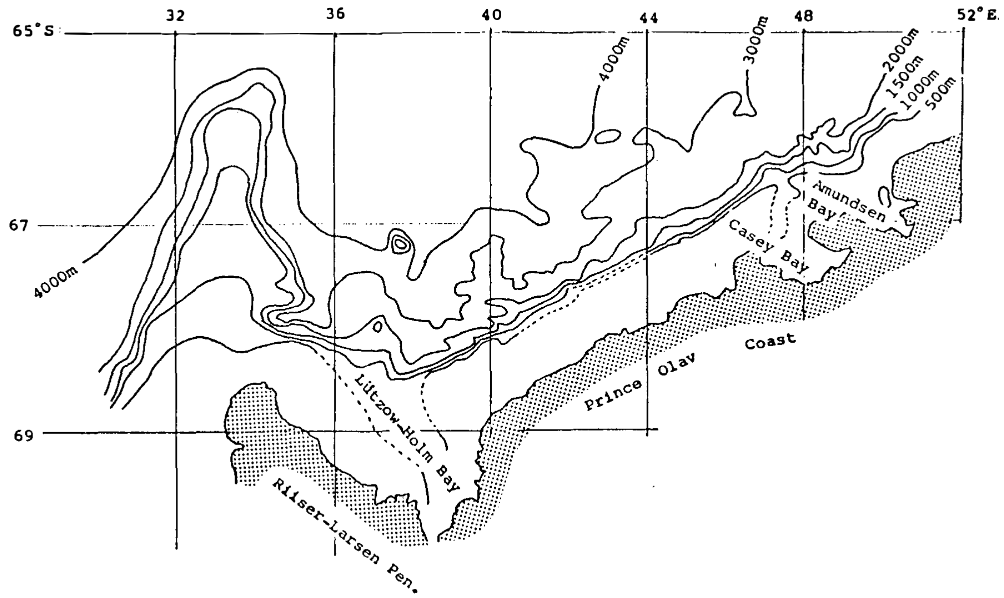

Fig. 14. Bathymetric chart of the area off the Prince Olav and the Prince Harald Coasts.

Submarine topography in the eastern Lützow-Holm Bay: Several deep submarine troughs and many small basins have been found along the Sôya Coast (FUJIWARA, 1971; OMOTO, 1976a). The troughs cut into the general shelf surface shallower than 400 m deep by 300 to 700 m. There are basins and sills in their longitudinal profiles, and their cross profiles are U-shaped, often asymmetrical. Therefore, the troughs were considered to be of glacial origin. They may be designated as "drowned fjord". The troughs are located in the immediate offing of outlet glaciers, and their directions of extension are NW-SE off the Honnör Glacier and the Telen Glacier, conformably with those of the former ice flow in ice-free areas. These troughs join a large trough situated in the central part of Lützow-Holm Bay (MORIWAKI, 1979). Another narrow trough runs in the N-S direction in the Ongul Strait, in parallel with the coast line. This trough extends from the Ongul Strait towards north for some 30 km, oblique to the coast line. To the south, it joins a trough extending NNW from a terminus of the Langhovde Glacier. The topographic features of the central trough are not known well. According to MORIWAKI (1979), the trough's width at the mouth of Lützow-Holm Bay reaches at least 20 km. A cross profile shows asymmetric valley walls which are steep and high in the east and gentle and low in the west. The eastern wall extends for more than 30 km in the direction of NNW without any break. The trough seems to extend towards the terminus of the Shirase Glacier (Fig. 15).

On the other hand, intensive sounding around the Ongul Islands (FUJIWARA, 1971; MORIWAKI, 1975, 1979) shows that the detailed submarine topography develops subsequently to the geological structure which can be inferred from the geology of an ice-free land area. This suggests sedimentation in this area has been very inactive since submergence. Drowned cirque-like features, small troughs, and stoss-and-lee topography are recognized. Many of small oval depressions and ledges suggest the same situation.

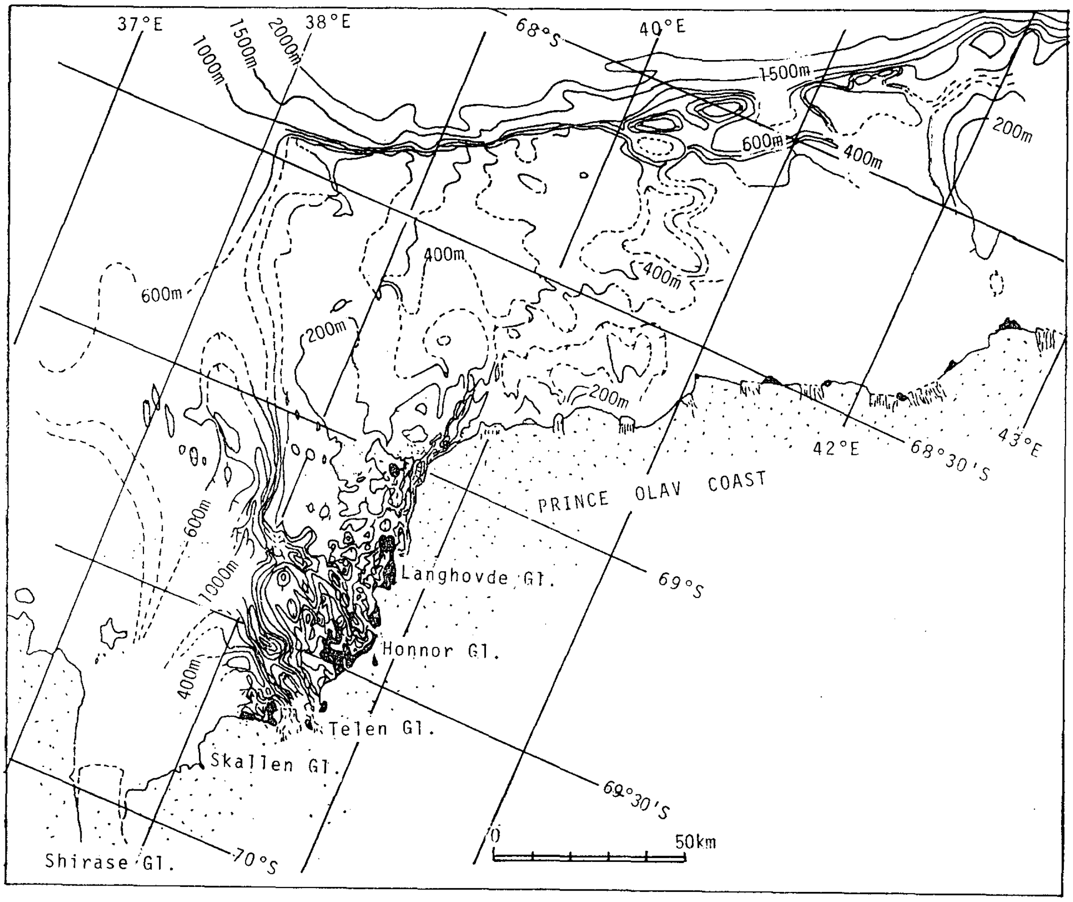

Fig. 15. Bathymetric chart of the eastern Lützow-Holm Bay and off the Prince Olav Coast (Compiled by K. MORIWAKI).

Information on submarine topography off the Prince Olav Coast and of the outer shelf is insufficient. A fairly large rise at the depth shallower than 300 m below sea level extends from the Ongul Islands to the edge of the shelf, trending NNW. To the east from this rise, there seem to exist broad depressions at the depth oı 400 m to 500 m. On the inner shelf west of 41°E meridian, there exist glacial troughs extending towards the present terminus of a glacier. But they are shallower than those in Lützow-Holm Bay. LISITZIN and ZHIVAGO (1960) described the lava mass on the continental shelf off the Prince Olav Coast around 40°E meridian. However, any evidence has not been obtained by JARE as yet.

Former ice sheet and submarine topography: Some problems of submarine topography development in this region are mentioned in relation to glacier fluctuation.

The "drowned fjord" was formed evidently by glacial sculpture of ice streams in the formerly expanded ice sheet (Fig. 16). Preglacial valleys might have been favourable for formation of ice streams, but overdeepening greatly modified the preexistent landform so as to make it difficult to identify the exact role of preglacial valleys. The confluent trough system in central Lützow-Holm Bay may be a feeble trace of a preglacial valley system, but no other evidence is found. In the Vincennes Bay area, the Vanderford submarine valley, the deepest drowned fjord in the world, is considered to have been originated in the preglacial valley (CAMERON, 1965). It

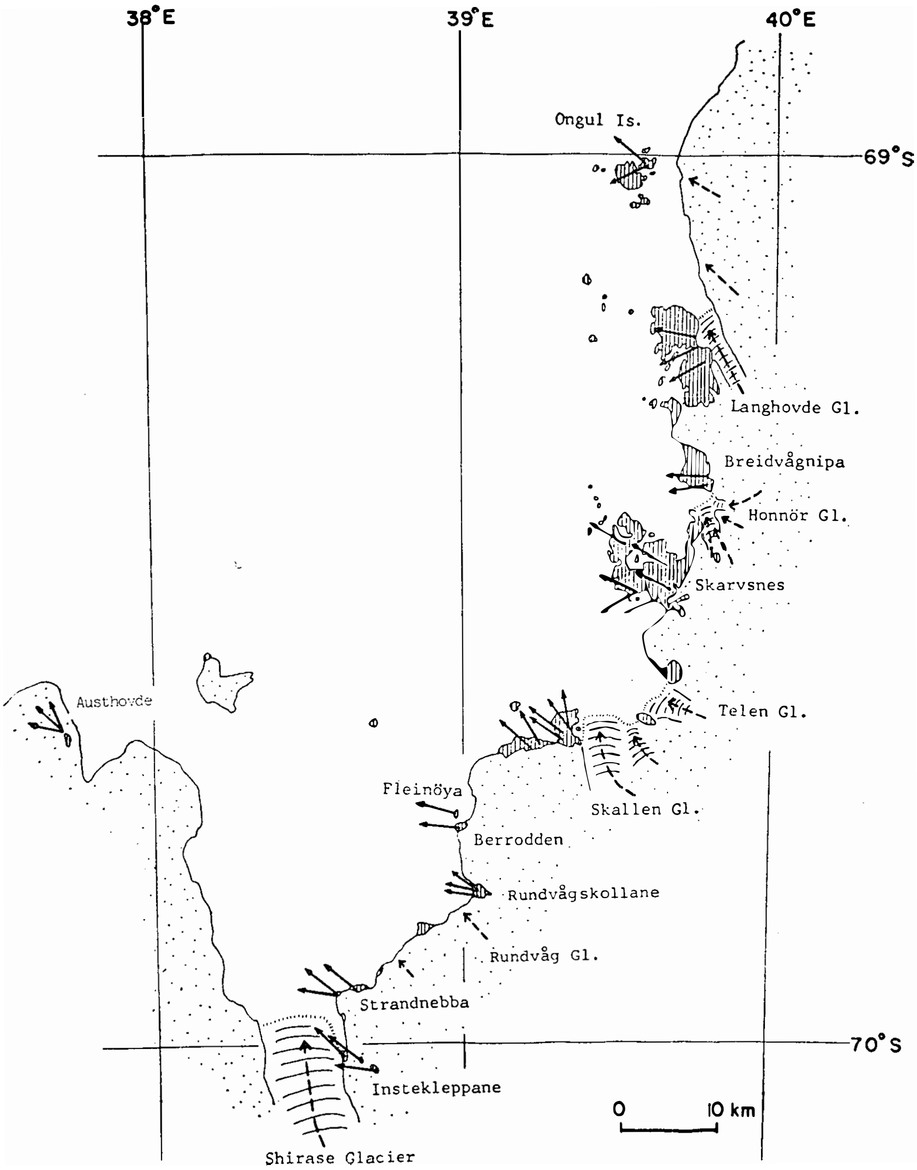

Fig. 16. Former ice movement indicated by glacial striation.

seems, however, positive evidence has not been obtained as yet.

On the other hand, FUJIWARA (1971) pointed out that most parts of low undulating sea bottom shallower than 300 m deep might represent the remnant of preglacial peneplain, though they had been modified by glacial agency to some extent. This view seems to be reasonable, judging from the topography of the adjacent ice-free areas where rather flat surfaces originated in preglacial landforms remain in some places and from the nature of ice sheet erosion which is rather protective for regional relief except in places where ice concentrates to form an ice stream.

It is not well known when and to what extent the former ice sheet covered the continental shelf. The outer smooth shelf surface has been ascribed to accumulation on the glacially scoured shelf of old times, as mentioned above, though FUJIWARA (1971) pointed out other possibility that the outer shelf had been a non-glaciated area during the extension of the former ice sheet. Geological structure of accumulation on the shelf has not been made clear. However, some broad depressions on the outer shelf off the Prince Olav Coast suggest that the continental shelf in this region had been covered with the ice sheet completely at its maximum extension, as in other cases in Antarctica. SHOJI and SATO (1959) suggested that well-sorted medium sand obtained from the bottom 700 m deep on the Gunnerus Bank off the Riiser-Larsen Peninsula could be originated in the relict of old shore sand. This is suggestive for consideration of the limit of the former ice sheet glaciation as well as the submarine valleys cutting into the continental slope, which remains for future study.

The approximate ice thickness at the present coast during the maximum extension of the ice sheet can be calculated by the formula,

$$h=4.7\sqrt{d} \text{ (HOLLIN, 1962)}, \qquad (1),$$

or

$$\left(\frac{h}{H}\right)^2+\left(\frac{x}{L}\right)^2=1 \text{ (VIALOV, 1958)}, \qquad (2),$$

assuming that the ice sheet had covered the continental shelf completely. In formula (1) h and d are the thickness of the ice at any point and the distance from the edge of the ice sheet, respectively. And in formula (2) H, L, x and h are the elevation at the centre of the ice sheet, the length between the centre and the edge of the ice sheet, the distance from the centre at any point, and the elevation of the ice sheet at any point, respectively. In either case, the estimated former ice thickness at the present coast is about 1000 m on the Prince Olav Coast and about 1400 m on the Sôya Coast. OMOTO (1976a) estimated the distance of ice advance at its maximum from the ice thickness at the coast inferred from the depth of drowned fjords. The method of estimating the ice thickness seems somewhat immature. However, his estimation of ice advance 75 km long and ice thickness 1250 m is reasonably consistent with the above-mentioned estimate. He also estimated the amount of isostatic subsidence in the coastal area by the maximum ice sheet to be 250 m. If the ice sheet disappears completely, isostatic uplift of the area will amount to 250 m, or 300 to 460 m, simply assuming that the depression of the earth's crust was one third of the ice thickness. However, a thick ice mass still exists adjacent to the present coastline. The zone of downward displacement by uniform loading was calculated to extend approximately 2 to $3\times l$ ($l=58$ km for the continent) beyond the edge of the load (BROTCHIE and SILVESTER, 1969). Moreover, isostatic uplift caused by unloading of the ice sheet from the continental shelf might have been conpensated by loading of invading sea water. Therefore, the amount of uplift at the present coast caused by shrinkage of the ice sheet margin may be considerably lower than 300 m as is indicated by the altitude of raised beaches.

The distribution of free-air gravity anomaly in Lützow-Holm Bay is shown in Fig. 17. The free-air anomaly indicates anomaly from isostatic equilibrium of the crust as the first approximation. Therefore, distribution of free-air anomaly is expected to express to some extent the state of glacial isostatic condition, though anomaly alone does not give direct evidence. HARADA *et al.* (1964) mentioned the isostatic recovery in this region from gravity survey, but they did not express a defi-

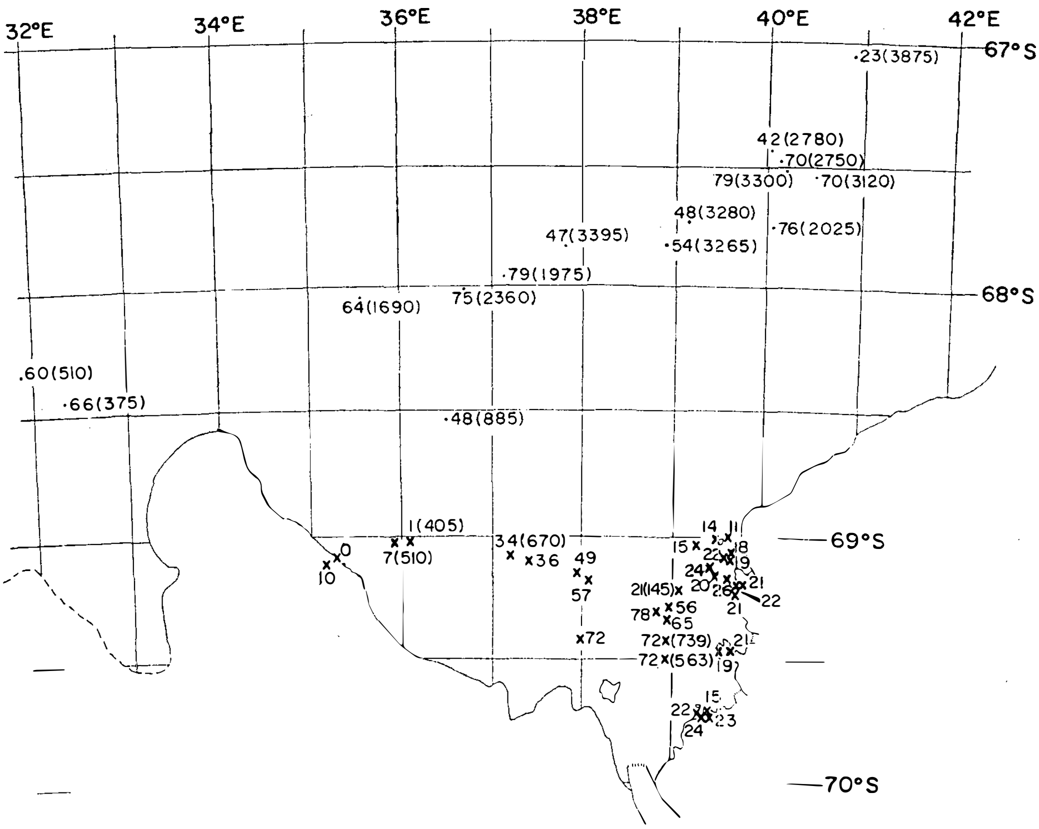

Fig. 17. Distribution of free air gravity anomaly in the vicinity of Lützow-Holm Bay. (Data: HARADA et al., 1964, ÔURA, 1965, and measurement by the author. A dot represents positive value in milligal, and a cross represents negative value. A numeral in parentheses indicates depth of water in metre.)

nite opinion. From Fig. 17, the following can be pointed out: 1) positive anomaly is distributed on the continental slope, 2) in contrast, considerable negative anomaly is found on the continental shelf, 3) large negative values seem to relate closely to the occurrence of deep submarine troughs (ÔURA, 1965), and 4) negative anomaly of about 20 milligals is found on the ice-free areas on the Sôya Coast, and the values increase slightly towards south. From these facts the anomaly can be explained in some degree by the effect of combination of the mass deficiency due to excavation of deep troughs and the delay of isostatic recovery since ice disappeared from the shelf area.

Time of the maximum glaciation and time and extent of later readvances are not known in this region. Major ice retreat took place before 30000 years B.P. at the latest, and since that time most of ice-free areas have not been covered with the ice sheet, as stated above. On the other hand, deep and shallow submarine troughs and basins having distinct, glacially sculptured features suggest that a period of marine sedimentation on them has not been very long. The fact that the ice sheet has not covered the presently ice-free areas since 30000 years ago does not always mean the ice has not occupied the drowned fjord since that time. It is possible that the outlet glaciers expanded to some extent induced by eustatic fall of sea level in the late-Würm Glacial and occupied the drowned fjords, leaving the ice-free areas as nunakols. A slow rate of the crustal uplift in this region before the "post-Glacial" seems to suggest this possibility. In any case, however, the evidence is insufficient for detailed discussion. The study on foraminiferal assemblages in sediment cores obtained from bottoms of submarine troughs suggests that the bottom sediment of the Ongul Strait trough was eroded to some extent (KATO and TAI, 1979). The study of bottom sediments will provide a clue to the future discussion of the above problems.

Geological structure and submarine topography: Relationship between the submarine topography and the recent (LISITZIN and ZHIVAGO, 1960; VORONOV, 1964) and/or ancient (GRINNELL, 1971) tectonism is difficult to discuss on the basis of direct evidence in this region. However, the following can be pointed out along this line. Lützow-Holm Bay is situated between the ENE-WSW trending Prince Olav Coast and the NNW-SSE trending Riiser-Larsen Peninsula. The outer edge of the continental shelf running straight is cut abruptly by the line connecting the Shirase Glacier and the east coast of the Riiser-Larsen Peninsula. The broad central trough in Lützow-Holm Bay develops along this line, and seems to continue to the large scale depression beneath the Shirase Glacier. Its asymmetrical valley wall of the east side extends considerably straight. These facts indicate this trough was formed along a structural line of a large scale. Geology of the Riiser-Larsen Peninsula and the Gunnerus Bank is scarcely known at present. But a block dredged from the Gunnerus Bank was assigned to Tertiary sedimentary rock (NIINO, 1958). This suggests that there may exist a sedimentary basin near the Gunnerus Bank, and geological structure west of Lützow-Holm Bay considerably differs from that east of Lützow-Holm Bay. Therefore, embayment of Lützow-Holm Bay may represent the large scale structural or tectonic line dividing two tectonically different structural provinces. Similar situation, though in a smaller scale, seems to exist in Amundsen Bay in the Enderby Land to the east (YOSHIDA *et al.*, 1964).

6. Conclusion

Summarizing the foregoing descriptions and discussions, the following remarks on geomorphic development of the surveyed region are presented.

In the inland area, the Yamato Mountains had been once buried completely by the ice sheet and subjected to its glacial agency. Glacial erosion by the ice sheet was selective, subsequent to the geological structure. Trends of extension of ridges and spurs are mostly concordant with trends of gneissic foliation and joints. Low undulating surfaces supposed to be originated in the remnant of preglacial erosional surface were subjected to areal scouring to some extent, forming glacially smoothed surface. Selective linear erosion sculptured troughs on previously existed depressions. Mountain glaciation which preceded the "flood" of the ice sheet made cirques on the Yamato Mountains. Some of these cirques were modified only slightly by ice sheet erosion, having been protected probably by topographic obstacles to ice flow. This stage of the ice sheet is designated as the "Yamato Glacial Stage".

After that stage, the fall of the ice sheet surface took place and the mountains began to emerge from the ice. The ice surface lowered by at least 400 m on the west side of the mountains and in a lesser amount on the east side; thenceforth, the fall of ice surface came to a standstill for a certain period of time. During this period, a cirque glacier was formed by nourishment by drifted snow on a lee-side slope between upper and lower gentle slopes in the Fukushima Massif, and the glacier exerted cirque erosion a little on the gentle slope. This standstill is designated as the "Fukushima Glacial Stage".

Thenceforth, the ice surface began to lower again and reached the present level. The total fall of the ice surface is at least 400 m on the east side and 700–800 m on the west side. Progress of the fall of ice surface interrupted the nourishment by drifted snow on cirque glaciers at high levels, which resulted in the wastage of ice masses in cirques. The present cirque glaciers, which are being nourished by drifted snow, show various states of regimen, according to their locations relative to the ice sheet surface and against the wind direction, and exerted slight cirque erosion to the mountains. Thrusting up of the lower part of the ice sheet caused by obstruction of the mountains and surrounding shallow subglacial bedrock and ablation of ice surface brought subglacial moraines and englacial meteorite fragments on to the ice surface to form a wide moraine field. This stage is designated as the "Meteorite Glacial Stage".

In the Yamato Mountains area, patterned ground has been formed on the morainic detritus deposited on the mountain slopes by the periglacial process. The degree of development of patterned ground differs in each massif. This difference may reflect to some extent the difference in elapsed time of each massif since deglaciation.

It is difficult to estimate the time scale of the above glacial stages. A similar sequence of glacial events has been found in the mountainous areas of the Queen Maud Land. Its time scale is also not known. But the correlation between the mountainous areas is necessary for understanding the ice sheet characteristics in this area of the past and the present; thus it is one of the main subjects for future research.

In the coastal area, the ice sheet extended probably to the outer edge of the continental shelf at its maximum expansion. The present ice-free areas and the continental shelf were subjected to the glacial agency during that time. Crossed striations on the ice-free bedrock indicate the possibility of readvance of the ice sheet. But decisive evidence has not been obtained as yet.

Areal scouring by the ice sheet exterted selective erosion on crystalline bedrocks, shaping various kinds of glacial landforms but leaving traces of preglacial relief in some places. Basins and shallow linear depressions were formed along gneissic foliation or banding and joint systems. Typical stoss-and-lee topography is seen in places where dips of gneissic banding and joints are appropriate for scouring and plucking. Mammilated peaks and rather flat surfaces were formed on the massive rocks. Curved grooves, grooves with overhanging walls and pothole-like features found in places suggest that erosion was conducted by the wet-based ice sheet.

On the other hand, selective linear erosion by ice streams in the ice sheet formed large glacial troughs. Upstreams of deep troughs on the continental shelf are still concealed by the present outlet glaciers. Some trough walls have a remarkably asymmetric cross profile, probably being controlled by original topography and an ice stream confined between bedrock rise and sheet flow. Some troughs might have been formed along conspicious structural lines of geology. In particular, a trough in the central part of Lützow-Holm Bay seems to have been formed along the structural line dividing two tectonic provinces of the Prince Olav Coast and the Riiser-Larsen Peninsula.

Major retreat of the ice sheet from the region took place prior to 30000 years B.P. Since then, most of the present ice-free areas have not been covered with the ice sheet. During the retreat of the ice sheet, small cirque glaciers were formed in some depressions of the ice-free areas. These glaciers eroded bedrock slightly. The retreat of the ice sheet might have been interrupted for some time in this period.

After the retreat of the ice sheet, the ice-free areas were inundated partially by the sea around 30000 years B.P. Glacial marine sediments derived mainly from ground moraine detritus were deposited together with marine organisms. After 22000 years B.P. the sea level lowered relative to land, and raised marine features emerged from the sea. Then, the sea level rose again, probably due to the post-Glacial eustatic change in sea level, and culminated around 6000 years B.P. After that, the raised beach topography was formed successively by the crustal uplift of the region at levels lower than 20 m high on the Sôya Coast and 30 m high on the Prince Olav Coast. Older raised beach topography higher than 20 m on the Sôya Coast is distributed below the level of 35 m high. The difference in altitudes between the older and the younger raised beach levels is at most 15 m. This suggests the rate of uplift prior to 6000 or more years B.P. was considerably smaller than that after 6000 years B.P.

Finally, it is pointed out that a period of a little warmer climate or less snow accumulation and sea ice than those at present might have existed in this area.

Acknowledgements

The author is greatly indebted to Professor T. YOSHIKAWA, Department of Geography of University of Tokyo, who guided the author into the study of the Antarctic geography and has given invaluable advice. The author also would like to express his sincere thanks to Professor T. NAGATA, Director of the National Institute of Polar Research, for his encouragement. Finally, he gives his hearty gratitude to many senior members and colleagues of the Japanese Antarctic Research Expeditions. Without their encouragement and cooperation he could not have continued the present research.

References

ADIE, R. J. (1964): Sea-level changes in the Scotia Arc and Graham Land. Antarctic Geology, ed. by R. J. ADIE. Amsterdam, North-Holland, 27–32.

AGETA, Y. (1971a): Nankyoku no shitsuryô shûshi to kongo no mondai (Mass balance studies in Antarctica and the future problem). Seppyo (J. Jpn. Soc. Snow Ice), **33**, 205–214.

AGETA, Y. (1971b): Higashi Nankyoku Mizuho Kôgen fukin no kishô jôtai ni tsuite (Some aspects of the weather conditions in the vicinity of the Mizuho Plateau, East Antarctica). Nankyoku Shiryô (Antarct. Rec.), **41**, 42–61.

AGETA, Y. and NARUSE, R. (1971): Measurements of ice flow around Skallen Rock, south of Syowa Station, Antarctica. Nankyoku Shiryô (Antarct. Rec.), **42**, 61–64.

ANDO, H. (1971): Dai-10-ji Nankyoku Chiiki Kansokutai nairiku chôsa gaihô (Preliminary report of the oversnow traverse of the 10th Japanese Antarctic Research Expedition in 1969–1970). Nankyoku Shiryô (Antarct. Rec.), **39**, 39–45.

ARMSTRONG, T., ROBERTS, B. and SWITHINBANK, C. (1966): Illustrated Glossary of Snow and Ice. Cambridge, University Printing House, 60 p.

AUTENBOER, T. VAN (1964): The geomorphology and glacial geology of the Sør-Rondane, Dronning Maud Land. Antarctic Geology, ed. by R. J. ADIE. Amsterdam, North-Holland, 81–103.

BARDIN, V. I. (1972): Moraines of Antarctica. Antarctic Geology and Geophysics, ed. by R. J. ADIE. Oslo, Universitetsforlaget, 663–667.

BARDIN, V. I. (1977): Lithology of East Antarctic moraines and some problems of Cenozoic history. Paper presented to the Symposium on Antarctic Geology and Geophysics, Madison, 1977.

BARRETT, P. J., CHRISTOFFEL, D. A. and NORTHEY, D. J. (1974): Seismic profiles across the extension of Wright Valley into McMurdo Sound. Antarct. J. U.S., **9**, 138–140.

BELINSKAYA, L. A. (1968): Geomorfologiya, osadki i vzvesi shel'fa i materikovogo sklona Antarktidy (Geomorphology, sediments and suspended material of the shelf and contiental slope of Antarctica). Tr. Sov. Antarkt. Eksped., **38**, 102–110.

BERG, T. E. and BLACK, R. F. (1966): Preliminary measurements of growth of nonsorted polygons, Victoria Land, Antarctica. Antarctic Soils and Soil Forming Processes, ed. by J. C. F. TEDROW. Washington, D.C., American Geophysical Union, 61–108.

BLOOM, A. L. (1978): Geomorphology; A Systematic Analysis of Late Cenozoic Landforms. Englewood Cliffs, Prentice-Hall, 510 p.

BOULTON, G. S. (1974): Processes and patterns of glacial erosion. Glacial Geomorphology, ed. by D. R. COATES. New York, State University of New York, 41–87.

BRADY, H. T. (1977): Late Neogene history of Taylor and Wright Valleys and McMurdo Sound, derived from diatom stratigraphy and palaeoecology of DVDP cores. Paper presented at the Symposium on Antarctic Geology and Geophysics, Madison, 1977.

BROTCHIE, J. F. and SILVESTER, R. (1969): On crustal flexure. J. Geophys. Res., **74**, 5240–5252.

BUDD, W., JENSSEN, D. and RADOK, U. (1970): The extent of basal melting in Antarctica. Polarforschung, **6**(39), 293–306.

BULL, C. and WEBB, P. N. (1973): Some recent developments in the investigation of the glacial history and glaciology of Antarctica. ICSU SCAR Conference on Quaternary Studies, Canberra, 1972, ed. by E. M. VAN ZINDEREN BAKKER. Cape Town, A. A. Balkema, 55–84 (Palaeoecology of Africa & of the Surrounding Islands & Antarctica, Vol. 8).

CALKIN, P. E., BEHLING, R. E. and BULL, C. (1970): Glacial history of Wright Valley, southern Victoria Land, Antarctica. Antarct. J. U.S., **5**, 22–27.

CAMERON, R. L. (1965): The Vanderford submarine valley, Vincennes Bay, Antarctica. Geology and Paleontology of the Antarctic, ed. by J. B. HADLEY. Washington, D.C., American Geophysical Union, 211–216.

CAMERON, R. L. and GOLDTHWAIT, R. P. (1961): The US-IGY contribution to Antarctic glaciology. Symposium on Antarctic Glaciology, Gentbrugge, IASH, 7–13 (IASH Publ., No. 55).

CAMPBELL, I. B. and CLARIDGE, G. G. C. (1966): Evidence of former sea levels near Cape Hallet, Antarctica. N. Z. J. Sci., **9**, 776–781.

CHRISTENSEN, L. (1939): Recent reconnaissance flights in the Antarctic. Geogr. J., **94**, 192–203.

CLARIDGE, G. G. C. and CAMPBELL, L. B. (1966): The raised beaches at Inexpressible Island, Antarctica. N. Z. J. Sci., **9**, 889–900.

COTTON, C. A. (1947): Climatic Accidents in Landscape-making. Christchurch, Whitcombe & Tombs, 354 p.

DENTON, G. H., ARMSTRONG, R. L. and STUIVER, M. (1970): Late Cenozoic glaciation in Antarctica: The record in the McMurdo Sound region. Antarct. J. U.S., **5**, 15–21.

DORT, W., Jr. and DORT, D. S. (1971): Sodium sulfate deposits in Antarctica. Mod. Geol., **1**, 97–117.

DREWRY, D. J. (1975): Initiation and growth of the East Antarctic ice sheet. J. Geol. Soc. London, **131**, 255–273.

EMBLETON, C. and KING, C. A. M. (1968): Glacial and Periglacial Geomorphology. London, Edward Arnold, 608 p.

FAIRBRIDGE, R. W. (1952): The geology of the Antarctic. Antarctic Today, ed. by F. A. SIMPSON. Wellington, New Zealand Antarctic Society, 56–101.

FAIRBRIDGE, R. W. (1961): Eustatic changes in sea level. Phys. Chem. Earth, **4**, 99–185.

FLINT, R. F. (1971): Glacial and Quaternary Geology. New York, John Wiley, 892 p.

FOSSUM, B. A., MAISEY, G. H. and TORSEN, H. O. (1977): Marine geophysical research in the Weddell Sea during the Norwegian Antarctic Research Expedition 1976/77. Paper presented at the Symposium on Antarctic Geology and Geophysics, Madison, 1977.

FUJIWARA, K. (1964): Preliminary report on the morphology of the inland ice sheet of the Mizuho Plateau, East Antarctica. Nankyoku Shiryô (Antarct. Rec.), **23**, 1–11.

FUJIWARA, K. (1971): Ryutsuo-Horumu Wan ni okeru sokushin to hyôshoku tairikudana no chikei (Soundings and submarine topography of the glaciated continental shelf in Lützow-Holm Bay, East Antarctica). Nankyoku Shiryô (Antarct. Rec.), **41**, 81–103.

FUJIWARA, K. (1973): Higashi Onguru Tô Mizukumi Zawa no ryûki teisen no chikei to shûhyôga chikei (The landforms of the Mizukumi Zawa near Syowa Station, East Antarctica). Nankyoku Shiryô (Antarct. Rec.), **46**, 44–66.

FUJIWARA, K. and YOSHIDA, Y. (1972): Ryutsuo-Horumu Wan tôgan no hyôga ryûdô sokutei (Ice flow measurements on the east coast of Lützow-Holm Bay, Antarctica). Nankyoku Shiryô (Antarct. Rec.), **44**, 79–92.

GLAVNOE UPRAVLENIE GEODEZII I KARTOGRAFII MG SSSR (1966): Atlas Antarktiki. Moskva, Sovetskaya Antarkticheskaya Ekspeditsiya.

GOULD, L. M. (1940): Glaciers of Antarctica. Proc. Am. Philos. Soc., **82**, 835–876.

GRINNELL, D. V. (1971): Physiography of the continental margin of Antarctica from 125°E to 150°E. Antarct. J. U.S., **6**, 164–165.

HANSEN, H. E. (1946): Antarctica, from Lat. 68°40′ to 70°20′S. and from Long. 36°50′ to 40° 20′E., scale 1: 250,000. Oslo, Norges Geogr. Opmåling.

HARADA, Y., MURATA, I., SAITO, T. and FUJII, Y. (1964): Report on the gravity measurements by the Japanese Antarctic Research Expedition, 1957–62. Antarctic Geology, ed. by R. J. ADIE. Amsterdam, North-Holland, 528–531.

HATANO, S. (1961): Some geomorphic studies of East Ongul Island; Landform classification and morphometry. Nankyoku Shiryô (Antarct. Rec.), **11**, 17–26.

HAYASHI, M. (1977): Shirase Hyôga no ryûdôryô sokutei (Measurement of flow rate of the Shirase Glacier). Nihon Nankyoku Chiiki Kansokutai Dai-16-ji Tai Hôkoku (1974–1976). Tokyo, Natl Inst. Polar Res., 85.

HAYASHI, M. (1979): A study on lithology and shape characteristics of moraine and marine pebbles in the Lützow-Holm Bay region. Mem. Natl Inst. Polar Res., Spec. Issue, **14**, 221–229.

HEINE, A. J. (1963): Ice breakout around the southern end of Ross Island, Antarctica. N. Z. J. Geol. Geophys., **6**, 395–401.

HENDY, C., NEALL, V. and WILSON, A. (1969): Recent marine deposits from Cape Barne, McMurdo Sound, Antarctica. N. Z. J. Geol. Geophys., **12**, 707–712.

HOLLIN, J. T. (1962): On the glacial history of Antarctica. J. Glaciol., **4**, 173–195.

HOUGH, J. L. (1950): Pleistocene lithology of Antarctic ocean-bottom sediments. J. Geol., **58**, 254–260.

ISHIDA, T. (1962): Yamato Sanmyaku ryokô no toji ni okeru jinkô jishin tansa (Seismic observation of the Yamato Mountains traversing trip). Nankyoku Shiryô (Antarct. Rec.), **14**, 36–43.

ISHIDA, T. (1970): Glaciological research on the inland traverse of the 8th Japanese Antarctic Research Expedition. International Symposium on Antarctic Glaciological Exploration (ISAGE). Gentbrugge, IASH, 130–138 (IASH Publ., No. 86).

ISHIDA, T. ed. (1978): Glaciological studies in Mizuho Plateau, East Antarctica, 1969–1975. Mem. Natl Inst. Polar Res., Spec. Issue, **7**, 274 p.

ISHIKAWA, T. (1974): Nankyoku Ryutsuo-Horumu Wan Ranguhobude chiiki no chishitsu (Geology of Langhovde, Lützow-Holm Bay, East Antarctica). Nankyoku Shiryô (Antarct. Rec.), **51**, 1–17.

JOHN, B. S. (1972): Evidence from the South Shetland Islands towards a glacial history of West Antarctica. Polar Geomorphology, ed. by R. J. PRICE and D. E. SUGDEN. London, Institute of British Geographers, 75–92.

KAGAMI, H. (1964): A description of bottom sediments in Lützow-Holmbukta. Antarctic Geology, ed. by R. J. ADIE. Amsterdam, North-Holland, 703–709.

KAMINUMA, K., SATO, R., SUZUKI, Y. and MATSUURA, M. (1979): Uplift of the Antarctic Continent in elastic deformation. Mem. Natl Inst. Polar Res., Spec. Issue, **14**, 70–82.

KAPITSA, A. P. (1964): The nature of central Antarctica, recent studies by Soviet scientists. Antarctic Report of USARP, Dec. 1964, 2–13.

KATO, M. and TAI, Y. (1979): Foraminifera from the eastern part of Lützow-Holm Bay, Antarctica. Mem. Natl Inst. Polar Res., Spec. Issue, **14**, 210–220.

KENNETT, J. P. (1978): Cainozoic evolution of circumantarctic palaeoceanography. Antarctic Glacial History and World Palaeoenvironment, ed. by E. M. van ZINDEREN BAKKER. Rotterdam, A. A. Balkema, 41–56.

KIZAKI, K. (1965): Geology and petrography of the Yamato Sanmyaku, East Antarctica. JARE Sci. Rep., Ser. C, **3**, 27 p.

KIZAKI, K. (1979): Nankyoku chigaku no tenbô (Review of Antarctic earth science). Chikyu (The Earth Monthly), **12**, 883–888.

KLEBELSBERG, R. V. (1948): Handbuch der Gletscherkunde und Glazialgeologie. Erster Band. Wien, Springer, 403 p.

KOAZE, T. (1964): Syowa Kiti fukin, rogan chiiki no chikei (The landform of the northern part of Prince Harald Coast, East Antarctica). Nankyoku Shiryô (Antarct. Rec.), **20**, 61–74.

KOBAYASHI, S. (1979): Some features of the turbulent transfer on the bare ice field near the Yamato Mountains, East Antarctica. Mem. Natl Inst. Polar Res., Spec. Issue, **12**, 9–18.

KOBAYASHI, S. and NARUSE, R. (1975): Meteorological data. JARE Data Rep., **28**, 83–113.

KOGAN, A. L. (1969): Relief of the glacier bed and geological description of the bedrock along the Molodezhnaya–Pole of Relative Inaccessibility traverse. Sov. Antarct. Exped. Inf. Bull., **7**, 140–143.

LEMASURIER, W. E. (1972): Volcanic record of Antarctic glacial history: Implications with regard to Cenozoic sea level. Polar Geomorphology, ed. by R. J. PRICE and D. E. SUGDEN. London, Institute of British Geographers, 59–74.

LISITZIN, A. P. and ZHIVAGO, A. V. (1960): Marine geological work of the Soviet Antarctic Expedition, 1955–1957. Deep-Sea Res., **6**, 77–87.

MAE, S. (1978): The bedrock topography deduced from multiple radar echoes observed in the Mizuho Plateau, East Antarctica. Nankyoku Shiryô (Antarct. Rec.), **61**, 23–31.

MAE, S. and NARUSE, R. (1978): Possible causes of ice sheet thinning in the Mizuho Plateau. Nature, **273**, 291–292.

MAEGOYA, T., NOHDA, S. and HAYASE, I. (1968): Rb-Sr dating of the gneissic rocks from the east coast of Lützow-Holm Bay, Antarctica. Mem. Fac. Sci., Kyoto Univ., Ser. Geol. Mineral., **35**, 131–138.

MARGOLIS, S. V. and KENNETT, J. P. (1971): Cenozoic paleoglacial history of Antartica recorded in subantarctic deep-sea cores. Am. J. Sci., **271**, 1–36.

MARKOV, K. K., BARDIN, V. I., LEBEDEV, V. L., ORLOV, A. I. and SUETOVA, I. A. (1970): The Geography of Antarctica, tr. by N. KANER. Jerusalem, Wiener Bindery, 370 p.

MCLEOD, I. R. (1964): The saline lakes of the Vestfold Hills, Princess Elizabeth Land. Antarctic Geology, ed. by R. J. ADIE. Amsterdam, North-Holland, 65–72.

MEGURO, H., YOSHIDA, Y., UCHIO, T., KIGOSHI, K. and SUGAWARA, K. (1964): Quaternary marine sediments and their geological dates with reference to the geomorphology of Kronprins Olav Kyst. Antarctic Geology, ed. by R. J. ADIE. Amsterdam, North-Holland, 73–80.

MERCER, J. H. (1968): The discontinuous glacio-eustatic fall in Tertiary sea level. Palaeogeogr., Palaeoclimatol., Palaeoecol., **5** (spec. issue), 77–85.

MERCER, J. H. (1972): Some observations on the glacial geology of the Beardmore Glacier area. Antarctic Geology and Geophysics, ed. by R. J. ADIE. Oslo, Universitetsforlaget, 427–433.

MINISTRY OF EDUCATION (1962): Nankyoku Kansoku Dai-5-ji Ettôtai Hôkoku. Tokyo, 202 p.

MORITA, Y. (1961): The effect of sea ice upon the climate of Syowa Base. Nankyoku Shiryô (Antarct. Rec.), **11**, 98–102.

MORITA, Y., MURAKOSHI, N. and NISHIBORI, E. (1961): An experimental observation of the evaporation from snow surface. Nankyoku Shiryô (Antarct. Rec.), **11**, 103–108.

MORIWAKI, K. (1974): Ryutsuo-Horumu Wan tôgan no ryûki teisen to kai kaseki no ^{14}C nendai (Radiocarbon datings of fossil shells on raised beaches on the east coast of Lützow-Holm Bay, East Antarctica). Nankyoku Shiryô (Antarct. Rec.), **48**, 82–90.

MORIWAKI, K. (1975): Syowa Kiti fukin no kaitei chikei—Dai-15-ji ettôtai chiri bumon hôkoku, 1 (Submarine topography near Syowa Station, Antarctica). Nankyoku Shiryô (Antarct. Rec.), **54**, 101–115.

MORIWAKI, K. (1976): Syowa Kiti fukin no rogan chiiki no chikei to tairikuhyô enpenbu no chigakuteki kansatsu (Glacio-geomorphological observations in and around ice-free areas in the vicinity of Syowa Station, Antarctica). Nankyoku Shiryô (Antarct. Rec.), **57**, 24–55.

MORIWAKI, K. (1979): Submarine topography of the central part of Lützow-Holm Bay and around Ongul Islands, Antarctica. Mem. Natl Inst. Polar Res., Spec. Issue, **14**, 195–209.

MORIWAKI, K. (1980): Glacio-geomorphological observations in Cape Omega on the Prince Olav Coast, East Antarctica. Nankyoku Shiryô (Antarct. Rec.), **70**, 1–14.

MURAKOSHI, N. (1958): Dai-1-ji ettôtai kishô bumon hôkoku (Meteorological observations at the Syowa Base during the period from March, 1957, to February, 1958). Nankyoku Shiryô (Antarct. Rec.), **4**, 1–22.

MURAKOSHI, N. (1959): Dai-1-ji ettôtai kishô bumon hôkoku, 2 (Some synoptic problems relating to the abnormal weather at Syowa Base). Nankyoku Shiryô (Antarct. Rec.), **6**, 22–33.

MURAUCHI, S. and YOSHIDA, Y. (1959): Dai-2-ji nankyoku kansoku ni saishite okonatta kaihyô kansoku oyobi soreni kansuru 2, 3 no kôsatsu (Reports of sea ice observation of the Japanese Antarctic Research Expedition 2, 1957–58). Nankyoku Shiryô (Antarct. Rec.), **8**, 22–39.

MURAYAMA, H. (1977): Syowa Kiti fukin no rogan chiiki ni sonzai suru koshô no ippanteki seijô ni tsuite (General characteristics of the Antarctic lakes near Syowa Station). Nankyoku Shiryô (Antarct. Rec.), **58**, 43–62.

NAGATA, T. (1961): Glaciology. Japanese Contribution to the International Geophysical Year 1957/8 and International Geophysical Cooperation 1959, **3**, 99–101.

NAGATA, T. ed. (1975): Yamato Meteorites Collected in Antarctica in 1969. Mem. Natl Inst. Polar Res., Spec. Issue, **5**, 110 p.

NAGATA, T. (1978a): A possible mechanism of concentration of meteorites within the Meteorite Ice Field in Antarctica. Mem. Natl Inst. Polar Res., Spec. Issue, **8**, 70–92.

NAGATA, T. ed. (1978b): Proceedings of the Second Symposium on Yamato Meteorites. Mem.

Natl Inst. Polar Res., Spec. Issue, **8**, 267 p.

NAGATA, T. ed. (1979): Proceedings of the Third Symposium on Antarctic Meteorites. Mem. Natl Inst. Polar Res., Spec. Issue, **12**, 282 p.

NAGATA, T. and YOSHIDA, Y. (1962): Glaciology. Japanese Contribution to the International Geophysical Year and the International Geophysical Cooperation, **4**, 138–141.

NAKANO, T., KAJI, T. and HARADA, Y. (1960): Some information on the glacier and geology in the vicinity of Ongul Island. Nankyoku Shiryô (Antarct. Rec.), **10**, 32–36.

NAKAWO, M, AGETA, Y. and YOSHIMURA, A. (1978): Discharge of ice across the Sôya Coast. Mem. Natl Inst. Polar Res., Spec. Issue, **7**, 235–244.

NARUSE, R. (1975): Movement of the ice sheet observed by a triangulation chain. JARE Data Rep., **28**, 48–61.

NARUSE, R. (1978): Surface flow and strain on the ice sheet measured by a triangulation chain in Mizuho Plateau. Mem. Natl Inst. Polar Res., Spec. Issue, **7**, 198–226.

NARUSE, R. and YOKOYAMA, K. (1975): Position, elevation and ice thickness of stations. JARE Data Rep., **28**, 7–47.

NATIONAL INSTITUTE OF POLAR RESEARCH (1975): Nihon Nankyoku Chiiki Kansokutai Dai-15-ji Tai Hôkoku. Tokyo, 191 p.

NEETHLING, D. C. (1972): Submarine and subglacial morphology, Kronprinsesse Martha Kyst, Drounning Maud Land. Antarctic Geology and Geophysics, ed. by R. J. ADIE. Oslo, Universitetsforlaget, 705–711.

NICHOLS, R. L. (1961): Characteristics of beaches formed in polar climates. Am. J. Sci., **259**, 694–708.

NICHOLS, R. L. (1966): Geomorphology of Antarctica. Antarctic Soils and Soil Forming Processes, ed. by J. C. F. TEDROW. Washington, D.C., American Geophysical Union, 1–46.

NICHOLS, R. L. (1968): Coastal geomorphology, McMurdo Sound, Antarctica. J. Glaciol., **7**, 449–478.

NIINO, H. (1958): Nankyoku Purinsu Hararudo Kukku Misaki oki no teishitsu (On the bottom sediments off Cook Point, Prins Harald Kyst, Antarctica). Tokyo Suisan Daigaku Tokubetsu Kenkyû Hokoku (J. Tokyo Univ. Fisheries), **1**(3) (Spec. ed.), 250–257.

NISHIIZUMI, K. and ARNOLD, J. R. (1980): Ages of Antarctic meteorites. Paper presented to the Eleventh Lunar and Planetary Science Conference, Houston, 1980.

NOGAMI, M. (1977): Higashi Nankyoku Purinsu Orafu Kaigan no ryûki teisen ni tsuite (On the raised beaches along the Prince Olav Coast, East Antarctica). Nihon Dai-4-ki Gakkai Yokôshû, **6**, 10.

OMOTO, K. (1972): A preliminary report on modern carbon datings at Syowa Station and its neighbourhood, East Antarctica. Nankyoku Shiryô (Antarct. Rec.), **43**, 20–24.

OMOTO, K. (1975): Syowa Kiti nanpô tairiku hyôenbu no kaitei chikei (Submarine topography near the margin of the ice sheet south of Syowa Station). Nihon Chiri Gakkai Yokôshû, **8**, 68–69.

OMOTO, K. (1976a): Subglacial geomorphology of Mizuho Plateau and around Yamato Mountains, East Antarctica. Sci. Rep. Tohoku Univ., 7th Ser. (Geography), **26**, 47–99.

OMOTO, K. (1976b): Glacio-geomorphic study on submarine morphology east of Lützow-Holm Bay, East Antarctica. Sci. Rep. Tohoku Univ., 7th Ser. (Geography), **26**, 241–297.

OMOTO, K. (1977): Geomorphic development of the Sôya Coast, East Antarctica—Chronological interpretation of raised beaches based on levellings and radiocarbon datings. Sci. Rep. Tohoku Univ., 7th Ser. (Geography), **27**, 95–148.

ÔURA, H. (1965): Syowa Kiti yori nan'i 75 do made no ryôkô oyobi Kukku Misaki e no ryôkô ni okeru jûryoku sokutei ni tsuite (On the measurements of gravity on the courses from Syowa Station to 75°S and to Cook Point). Nankyoku Shiryô (Antarct. Rec.), **25**, 86–107.

PÉWÉ, T. L. (1960): Multiple glaciation in the McMurdo Sound region, Antarctica: A progress report. J. Geol., **68**, 498–514.

PRICE, R. J. (1973): Glacial and Fluvioglacial Landforms. Edinburgh, Oliver & Bond, 242 p.

ROBERTSON, J. D., BENTLEY, C. R., CLOUGH, J. W. and GREISCHAR, L. L. (1977): Sea bottom topography and crustal structure below the Ross Ice Shelf, Antarctica. Paper presented at the Symposium on Antarctic Geology and Geophysics, Madison, 1977.

ROBIN, G. de Q., SWITHINBANK, C. W. M. and SMITH, B. M. E. (1970): Radio echo exploration of the Antarctic ice sheet. International Symposium on Antarctic Glaciological Exploration (ISAGE). Gentbrugge, IASH, 97–115 (IASH Publ., No. 86).

RONCA, L. B. (1964): Minimum length of time of frigid conditions in Antarctica as determined by thermoluminescence. Am. J. Sci., **262**, 767–781.

RUTFORD, R. H. (1972): Glacial geomorphology of the Ellswarth Mountains. Antarctic Geology and Geophysics, ed. by R. J. ADIE. Oslo, Universitetsforlaget, 225–232.

SANO, M., NAKAI, N. and TORII, T. (1977): Nurume Ike no 2-sô kôzô ni tsuite (On the two-layer structure in Lake Nurume, Antarctica). Nankyoku Shiryô (Antarct. Rec.), **58**, 63–68.

SATO, T. (1964): Submarine topography and several bottom sediments around Prince Olav and Prince Harald Coasts, Antarctica. Nankyoku Shiryô (Antarct. Rec.), **21**, 1–11.

SATO, T. (1969): Kaitei Chikeigaku (Submarine Geomorphology). Tokyo, Maruzen, 191 p.

SATO, T., IWABUCHI, Y., IIJIMA, A., OINUMA, K. and KOBAYASHI, K. (1965): Geological research on the bottom sediments sampled by the fifth Japanese Antarctic Research Expedition. JARE Sci. Rep., Ser. C, **4**, 42 p.

SCHWERDTFEGER, W. (1970): The climate of the Antarctic. Climates of the Polar Regions, ed. by S. ORVIG. Amsterdam, Elsevier, 253–355 (World Survey of Climatology, Vol. 14).

SCOTT, R. F. (1905): Results of the National Antarctic Expedition, I. Geogr. J., **25**, 353–373.

SHEPARD, F. P. (1931): Glacial troughs of the continental shelves. J. Geol., **39**, 345–360.

SHEPARD, F. P. (1963): Submarine Geology. 2nd ed. New York, Harper & Row, 267 p.

SHIMIZU, H., YOSHIMURA, A., NARUSE, R. and YOKOYAMA, K. (1978a): Morphological feature of the ice sheet in Mizuho Plateau. Mem. Natl Inst. Polar Res., Spec. Issue, **7**, 14–25.

SHIMIZU, H., WATANABE, O., KOBAYASHI, S., YAMADA, T., NARUSE, R. and AGETA, Y. (1978b): Glaciological aspects and mass budget of the ice sheet in Mizuho Plateau. Mem. Natl Inst. Polar Res., Spec. Issue, **7**, 264–274.

SHIRAISHI, K. (1977): Geology and petrography of the northern Yamato Mountains, East Antarctica. Mem. Natl Inst. Polar Res., Ser. C, **12**, 33 p.

SHIRAISHI, K. and KIZAKI, K. (1979): Yamato Sanmyaku no chishitsu (Geology of the Yamato Mountains). Chikyu (The Earth Monthly), **12**, 928–937.

SHIRAISHI, K., NARUSE, R. and KUSUNOKI, K. (1976): Collection of Yamato meteorites, Antarctica, in December 1973. Nankyoku Shiryô (Antarct. Rec.), **55**, 49–60.

SHOJI, D. and SATO, T. (1959): Nankyoku no teishitsu—sono ryûdo bunseki—(On the bottom sediment dredged by the second Japanese Antarctic Exploration). Nankyoku Shiryô (Antarct. Rec.), **7**, 39–49.

SIEGEL, F. R. and DORT, W., Jr. (1969): Mirabilite and associated seal bones, southern Victoria Land, Antarctica. Antarct. J. U.S., **4**, 173.

SPEDEN, I. G. (1962): Fossiliferous Quaternary marine deposits in the McMurdo Sound region, Antarctica. N. Z. J. Geol. Geophys., **5**, 746–777.

STUIVER, M., DENTON, G. H. and BORNS, H. W., Jr. (1976): Carbon-14 dates of *Adamussium colbecki* (Mollusca) in marine deposits at New Harbor, Taylor Valley. Antarct. J. U.S., **11**, 86–88.

SUGDEN, D. E. and JOHN, B. S. (1976): Glaciers and Landscape, a Geomorphological Approach. London, Edward Arnold, 376 p.

TATSUMI, T. and KIKUCHI, T. (1959a): Nankyoku Syowa Kiti fukin no chigakuteki kansatsu (1) (Report of geomorphological and geological studies of the wintering team (1957–58) of the first Japanese Antarctic Research Expedition, Part 1). Nankyoku Shiryô (Antarct. Rec.), **7**, 1–16.

TATSUMI, T. and KIKUCHI, T. (1959b): Nankyoku Syowa Kiti fukin no chigakuteki kansatsu (2) (Report of geomorphological and geological studies of the wintering team (1957–58) of the first Japanese Antarctic Research Expediton, Part 2). Nankyoku Shiryô (Antarct. Rec.), **8**, 1–21.

TATSUMI, T., KIKUCHI, T. and KIZAKI, K. (1964): Geology of the region around Lützow-Holmbukta and the "Yamato Mountains" (Dronning Fabiolafjella). Antarctic Geology, ed. by R. J. ADIE. Amsterdam, North-Holland, 293–303.

TAYLOR, G. (1922): The Physiography of the McMurdo Sound and Granite Harbour Region. London, Harrison, 246 p.

TORII, T. and YAMAGATA, N. (1973): Oashisu (Oasis). Nankyoku, ed. by K. KUSUNOKI *et al.* Tokyo, Kyoritsu Shuppan, 282–330.

TRAIL, D. S. (1964): The glacial geology of the Prince Charles Mountains. Antarctic Geology, ed. by R. J. ADIE. Amsterdam, North-Holland, 143–151.

UCHIO, T. (1966): Nankyoku Syowa Kiti fukin no kaiyô chishitsugaku tenbô (Critical review of the marine geological results around the Syowa Station obtained during the first series of the Japanese Antarctic Research Expeditions). Nankyoku Shiryô (Antarct. Rec.), **27**, 78–87.

VIALOV, S. S. (1958): Regularities of glacial shields movement and the theory of plastic viscous flow. Physics of the Movement of the Ice. Gentbrugge, 266–275 (IASH Publ., No. 47).

VORONOV, P. S. (1964): Tectonics and neotectonics of Antarctica. Antarctic Geology, ed. by R. J. ADIE. Amsterdam, North-Holland, 692–700.

WATANABE, O. and YOSHIMURA, A. (1972): Higashi Nankyoku Tairiku, Sandâkokku Nunatâku gun ni okeru chigaku chôsa ni tsuite (Preliminary geo-scientific researches on Sandercock Nunataks, Enderby Land, East Antarctica). Nankyoku Shiryô (Antarct. Rec.), **45**, 47–65.

WATKINS, N. D., KEANY, J., LEDBETTER, M. T. and HUANG, T. C. (1974): Antarctic glacial history from analyses of ice-rafted deposits in marine sediments: New model and initial test. Science, **186**, 533–536.

WATKINS, N. D., LEDBETTER, M. T. and HUANG, T. C. (1977): Antarctic glacial history using spatial and temporal variations of ice rafted debris in abyssal sediments of the Southern Ocean. Paper presented at the Symposium on Antarctic Geology and Geophysics, Madison, 1977.

WEBB, P. N. and WRENN, J. H. (1977): Late Cenozoic micropaleontology and biostratigraphy of eastern Taylor Valley, Antarctica. Paper presented at the Symposium on Antarctic Geology and Geophysics, Madison, 1977.

WILSON, A. T. (1964): Origin of ice ages: An ice shelf theory for Pleistocene glaciation. Nature, **201**, 147–149.

WRENN, J. H. and WEBB, P. N. (1977): Physiographic analysis and interpretation of the Ferrar Glacier–Victoria Valley area, Antarctica. Paper presented at the Symposium on Antarctic Geology and Geophysics, Madison, 1977.

YANAI, K. (1976): 1974-nen no nankyoku-san Yamato inseki no tansa to saishû (Search and collection of Yamato meteorites, Antarctica, in October and November 1974). Nankyoku Shiryô (Antarct. Rec.), **56**, 70–81.

YOKOYAMA, K. (1976): Geomorphological and glaciological survey of the Minami-Yamato Nunataks and the Kabuto Nunatak, East Antarctica. Nankyoku Shiryô (Antarct. Rec.), **56**, 14–19.

YOSHIDA, M. (1975): Geology of the region around Botnneset, East Antarctica. Mem. Natl Inst. Polar Res., Ser. C, **8**, 43 p.

YOSHIDA, M. (1979): Tectonics and metamorphism of the region around Lützow-Holmbukta, East Antarctica. Mem. Natl Inst. Polar Res., Spec. Issue, **14**, 28–40.

YOSHIDA, M., ANDO, H., OMOTO, K., NARUSE, R. and AGETA, Y. (1971): Discovery of meteorites near Yamato Mountains, East Antarctica. Nankyoku Shiryô (Antarct. Rec.), **39**, 62–65.

YOSHIDA, Y. (1969): Echo-sounding of ice-covered waters from surface of ice. Nankyoku Shiryô (Antarct. Rec.), **34**, 14–22.

YOSHIDA, Y. (1970a): Glacio-geological survey on the Kronprins Olav Kyst, Antarctica (Extended abstract). International Symposium on Antarctic Glaciological Exploration (ISAGE). Gentbrugge, IASH, 461–463 (IASH Publ., No. 86).

YOSHIDA, Y. (1970b): Higashi Nankyoku Purinsu Orafu Kaigan no ryûki teisen to enko (Raised beaches and saline lakes on the Prince Olav Coast, East Antarctica). Gendai no Chirigaku. Tokyo, Kokon Shoin, 93–118.

YOSHIDA, Y. (1972): On meltwater channels in the marginal zone of the ice sheet along Lützow-Holm Bay, Antarctica. Chiri Kagaku (Geogr. Sciences), **16**, 19–24.

YOSHIDA, Y. (1973): Rogan no chikei to hyôshô no hendô (Geomorphology of the ice-free areas and fluctuation of the ice sheet). Nankyoku, ed. by KUSUNOKI *et al.* Tokyo, Kyoritsu Shuppan, 237–281.

YOSHIDA, Y. (1975): Dai-16-ji Nankyoku Chiiki Kansokutai natsutai (1974–1975) hôkoku (Report of the summer party of the 16th Japanese Antarctic Research Expetidion in 1974–1975). Nankyoku Shiryô (Antarct. Rec.), **53**, 141–179.

YOSHIDA, Y. (1976): Studies in physical geography of Prince Harald and Prince Olav Coasts, Antarctica. Paper presented at the 23rd International Geographical Congress, Moscow, 1976.

YOSHIDA, Y. and FUJIWARA, K. (1963): Yamato Sanmyaku no chikei (Geomorphology of the Yamato (Queen Fabiola) Mountains). Nankyoku Shiryô (Antarct. Rec.), **18**, 1–26.

YOSHIDA, Y. and MAE, S. (1978): Some information on topographic features and the characteristics of the ice sheet around the Yamato Mountains. Mem. Natl Inst. Polar Res., Spec. Issue, **8**, 93–100.

YOSHIDA, Y. and MORIWAKI, K. (1979): Some consideration on elevated coastal features and their dates around Syowa Station, Antarctica. Mem. Natl Inst. Polar Res., Spec. Issue, **13**, 220–226.

YOSHIDA, Y., MURAKOSHI, N. and YATA, A. (1962): Dai-4-ji Nankyoku Chiiki Kansokutai ettôtai no chôsa ryokô ni okeru kishô kansoku ni tsuite (On the climatological observations in survey trips of the wintering party JARE 4, 1960). Nankyoku Shiryô (Antarct. Rec.), **15**, 12–24.

YOSHIDA, Y., MURAUCHI, S. and FUJIWARA, K. (1964): Submarine topography off the Kronprins Olav Kyst. Antarctic Geology, ed. by R. J. ADIE. Amsterdam, North-Holland, 710–714.

YOSHIDA, Y., TORII, T., YUSA, Y., NAKAYA, A. and MORIWAKI, K. (1975): A limnological study of some lakes in the Antarctic. Quaternary Studies, ed. by R. P. SUGGATE and M. M. CRESSWELL. Wellington, The Royal Society of New Zealand, 311–320.

YOSHIKAWA, T. and TOYA, H. (1957): Dai-1-ji Nankyoku Chiiki Kansokutai chiri bumon hôkoku (Report on geomorphological results of the Japanese Antarctic Research Expedition, 1956–57). Nankyoku Shiryô (Antarct. Rec.), **1**, 1–13.

(Manuscript received May 1, 1982;
Revised manuscript received September 30, 1982)