NATURE ENVIRONMENT MAP

DRONNING MAUD LAND

1:150 000

Editors: S. Elvevold & Y. Ohta

H.U. SVERDRUPFJELLA

NORSK·POLARINSTITUTT

NATURE ENVIRONMENT MAP
H.U. SVERDRUPFJELLA
DRONNING MAUD LAND, EAST ANTARCTICA
1:150 000

Editors: S. Elvevold & Y. Ohta

Contributors to the map:
Topography: T. Eiken, B. Lytskjold & B.Å. Luktvasslimo
Geology: Y. Ohta, G.H. Grantham, P.B. Groenewald, A.B. Moyes, A. Hjelle
& T.S. Winsnes
Glaciology: E. Isaksson, K. Melvold & J.O. Hagen
Biology: L. Sømme & F. Mehlum

Contributors to the explanatory text:
Y. Ohta, S. Elvevold, J.O. Hagen, K. Melvold, T. Eiken, B. Lytskjold,
E. Isaksson & H. Anschütz

The Norwegian Polar Institute is Norway's main institution for research, monitoring and topographic mapping in Norwegian polar regions. The Institute also advises Norwegian authorities on matters concerning polar environmental management.

Norsk Polarinstitutt er Norges sentralinstitusjon for kartlegging, miljøovervåking og forvaltningsrettet forskning i Arktis og Antarktis. Instituttet er faglig og strategisk rådgiver i miljøvernsaker i disse områdene og har forvaltningsmyndighet i norsk del av Antarktis.

Contents

Nature Environment map
H.U. Sverdrupfjella,
Dronning Maud Land, East Antarctica, 1:150 000

S. Elvevold & Y. Ohta, editors

Abstract

The area covered by the map extends from 71° 55' to 72° 50' S and 0° 35' W to 1° 45' E in western Dronning Maud Land and includes H.U. Sverdrupfjella and Gburekfjella. The ice topography of the map area is divided into three zones: the grounded ice sheet of lower altitudes (lower than 1400 m a.s.l.), the NE-SW trending zone of exposed nunataks which forms H.U. Sverdrupfjella, and the polar plateau ice at elevations higher than 2200 m a.s.l. For the most part, the area has a positive accumulation of snow, 0.05-0.40 m water equivalent/y. The nunatak zone acts as a dam to the plateau ice and the outlet glacier, Jutulstraumen, is the main drainage with an ice-flow velocity of 442 m/y measured 80 km upstream from the grounding line, and transporting ca. 16 km³ ice/y. Sveabreen, in the northeast, transports a small amount of ice. The ice thickness is variable, reaching a surface elevation of 1500 m. The bedrock surface is approximately at sea level, but is higher where the bedrock consists of massive igneous rock. Two deep valleys, as deep as 1900 m b.s.l., occur along the western side of Jutulstraumen. They follow a major fault zone, the Jutulstraumen-Pencksökket Rift Zone, which separates two geological terranes.

H.U. Sverdrupfjella is dominated by various high-grade metasupracrustal rocks, the Sverdrupfjella supersuite. These rocks have a polyphase tectonothermal history with two orogenic episodes, the first in late Mesoproterozoic and the second in late Neoproterozoic/Cambrian times. The main fabric (foliation) and the high-grade metamorphic mineral assemblages are the result of intense reworking during the Neoproterozoic. The Sverdrupfjella supersuite is subdivided into three lithological complexes. These are, from lower to structurally higher levels, the Jutulrøra gneiss complex, the Fuglefjellet gneiss complex and the Rootshorga gneiss complex. The Jutulrøra gneiss complex crops out in the western region and contains amphibolite facies, quartzo-feldspathic and banded gneisses and minor mafic rocks. The Fuglefjellet gneiss complex consists of upper amphibolite facies, banded gneisses and marble and calc-silicates. The eastern region, underlain by the Rootshorga gneiss complex, comprises upper amphibolite to granulite facies gneisses interlayered by numerous granitoids, mafic rocks and migmatitic leucosomes.

Various igneous rocks intruded the Sverdrupfjella supersuite during Neoproterozoic to Mesozoic times. The oldest granitic rocks (1140-1130 Ma) record the formation of an extensive volcanic arc. This was followed by 1070 Ma tabular granitoids (Sveabreen migmatitic granites). Several generations of deformed mafic intrusions are distinguished. The majority of the mafic rocks occur as lenses, boudins and concordant layers; dykes occur rarely. The common mafic rocks pre-date the deformation and are intrusive to the oldest pre-tectonic granite. Late- to post-tectonic magmatism, reflected by plutons, sheets and dykes of monzogranitic composition (Brattskarvet suite), is widespread. The alkaline syenites at Straumsvola and Tvora intruded along the eastern border of the Jutulstraumen-Pencksökket Rift Zone. Igneous layering and zoned distribution of different lithologies are distinct in both intrusions. The alkaline syenites and the dolerite were emplaced roughly simultaneously, around 170-180 Ma.

Mesoscopic structures record five deformation phases (D_1-D_5). D_1 and D_2 are represented by isoclinal, rootless and recumbent folds. The penetrative fabric (S_2) is related to E- to SE-plunging tight to isoclinal folds and dominant top-to-NW transport. Several thrust faults, which occur in the eastern region, transported high-grade rocks westwards onto lower grade rocks during D_2. Regional warping and local NE-SW-trending open folds led to localized fabric transposition defined by biotite (D_3). Finally, NE-striking normal faults and joints (D_5) are consistent with the break-up of Gondwana at approximately 170 Ma.

The first high-grade metamorphism is constrained at ca. 1035 Ma, whereas the second high-grade recrystallization and metamorphic zircon overgrowths took place during the Pan-African orogeny at ca. 540 Ma.

S. Elvevold, E. Isaksson & H. Anschütz, Norwegian Polar Institute, Fram Centre, NO-9296 Tromsø; Y. Ohta, Norwegian Polar Institute, c/o Natural History Museum, Po-box 1172 Blindern, NO-0318 Oslo; J.O. Hagen, Department of Geosciences, Po-box 1147 Blindern, NO-1316 Oslo; T. Eiken, Department of Geosciences, Po-box 1147 Blindern, NO-1316 Oslo; K. Melvold & B. Lytskjold, Norwegian Water Resources and Energy Directorate, Po-box 5091 Majorstuen, NO-0301 Oslo.

Introduction

This map covers the area between 71° 55' - 72° 50' S and 0° 35' W - 1° 45' E in western Dronning Maud Land (DML), which covers part of the topographic map sheets G5, G-6, H5 and H6. The area mapped is located about 200-300 km south of the ice front of Kronprinsesse Märtha Kyst.

Published locations of birds and invertebrates are marked on the map by courtesy of Professor L. Sømme, University of Oslo (Strandtmann & Sømme 1977). The species are described in the explanatory text on the adjacent map sheets (Ohta 1999). The present explanatory text contains physical information, such as topography, glaciology and geology. Minor adjustments to the map, which was printed in 1996, are indicated in the explanatory text below.

Geomorphology

Topography

H.U. Sverdrupfjella forms part of the coastal mountain chain along the Indian Ocean side of East Antarctica, which extends in a general NE-SW direction for some 100 km along the eastern side of the major glacier, Jutulstraumen. H.U. Sverdrupfjella is separated from adjacent mountain ranges by two large outlet glaciers, Svea-

breen in the north-east and upper Jutulstraumen in the south-west.

The north-western half of the map area, Nilsevidda and Hellehallet, is a flat ice field at 900-1600 m a.s.l., formed by a grounded ice sheet. In the south-east, it rises from the flat ice, via H.U. Sverdrupfjella, to ca. 2000 m, the Wegenerisen ice plateau. Nine valley glaciers flowing in a NW-SE direction divide H.U. Sverdrupfjella into ten groups of nunataks. The highest peak, 2885 m, is Hamartind (72° 33' S and 0° 40' E).

H.U. Sverdrupfjella forms the boundary zone that separates the grounded ice sheet and the inland plateau ice sheet. The altitude drops from more than 2400 m on the plateau to 1200 m a.s.l. in the foothills of the nunatak ranges. The high velocity of ice flow in narrow glaciers creates a high density of crevasses in the nunatak zone, and small moraine tails are locally seen. Major crevasse zones and moraines are visible on the satellite image maps.

History of discovery and topographical survey

The first aeroplane observation of the coast and the interior was made by H. Riiser-Larsen during the Norwegian expedition in 1929-30. In 1930-31, G. Isachsen and H. Riiser-Larsen mapped the area along Prinsesse Ragnhild Kyst from the air (Aagaard 1934). Norway proposed the terri-

torial claim of Dronning Maud Land (DML) in 1939 (Hansen 1949), but the claim has been frozen since 1961 when the International Antarctic Treaty became effective.

The first extensive topographical mapping was performed by an expedition sent out by Lars Christian in 1936-37. Its factory ship, Thorshavn, used an aeroplane for aerial photography. The expedition took more than 2200 photographs of the coastline and mountains in eastern DML, and as far east as 81° E. Eleven maps on the scale of 1:250 000 and one 1:500 000 map were published in 1946. The maps were compiled solely from the aerial photographs and aeroplane navigation, and no ground control was made.

Die Deutsche Antarktische Expedition in 1938-39 carried out aerial photography of DML and produced a sketch map of the mountains between 12° W and 21° E. The Neuschwabenland Expedition, led by Alfred Ritscher, first sighted mountains between 4° 15' W and 14° 45' E, in 1939, including Mt. Heddeon, which is now called Brattskarvet, and the Paulsen Mts, probably including Vendeholten and Vendehö in the northeastern part of H.U. Sverdrupfjella.

The first systematic and modern mapping of the mountainous area was performed by the Norwegian-British-Swedish Antarctic Expedition in 1949-52. The field work took place from the wintering base, Maudheim, at 71° S, 11° W. A triangulation network was established and aerial photographs were taken during the austral seasons of 1950-51 and 1951-52 (Roer 1954) in the area between 5° W and 2° E. The results of this expedition, with additional British measurements on Heimefrontfjella (1963-65) and Norwegian data on Vestfjella (1967-68), yielded the basis for the preparation of 12 map sheets on the scale of 1:250 000.

The Norwegian Antarctic Expeditions of 1956-60 continued the topographical work from Fimbulheimen. During the austral summer of 1957-58, aerial photographs were taken to extend the coverage, and 20 map sheets on the scale of 1:250 000 were compiled with Belgian reference measurements and supplementary photographs. These were published in 1961 by the Norwegian Polar Institute.

In the late 1980s, South African Antarctic expeditions made a new topographical survey of the area and, based on the aerial photographs taken by the German Antarctic expeditions in 1986-87, new maps on the scale of 1:50 000 were compiled and published by the South African Scientific Centre for Antarctic Research in 1993.

During later Norwegian expeditions in 1984-85 (Eiken & Svendsen 1985), 1989-90 (Eiken et al. 1990) and 1992-93 (Barstad et al. 1997), the main topographical efforts concentrated on a better geodetic reference network and the production of satellite image maps. Norwegian topographers made new measurements of reference points using GPS. The satellite images were rectified using new control points, and satellite image maps of the present area were published in 1991 (eastern part) and 1997 (western part). These images, with additional topographical data, are valuable complementary sources of information, especially in the snow- and ice-covered areas where topographical map information is very scarce. Ten satellite image maps have been produced on the scale of 1:250 000 and one map on the scale of 1:100 000. The map used for the present geological map was made from a digitized map based on the 1:250 000 scale series.

Glaciology and climate

General statement

Antarctica is almost completely covered by ice. Only 2.4 % of the rock mass is exposed above the ice, mainly on the Antarctic Peninsula and mountain ridges and nunatak areas, which usually appear 100-300 km from the ice shelf edge. The ice thickness has been mapped by airborne radio-echo sounding and surface seismic shooting methods. The average ice thickness is 2160 m, maximum 4770 m. The total ice volume is about 30 x 106 km³, which is more than 90 % of the world's volume of ice. 86.5 % of the ice volume is in East Antarctica, 11.5 % in West Antarctica, including the Antarctic Peninsula, and the remaining 2 % constitute the major ice shelves in the Ross and Weddel seas.
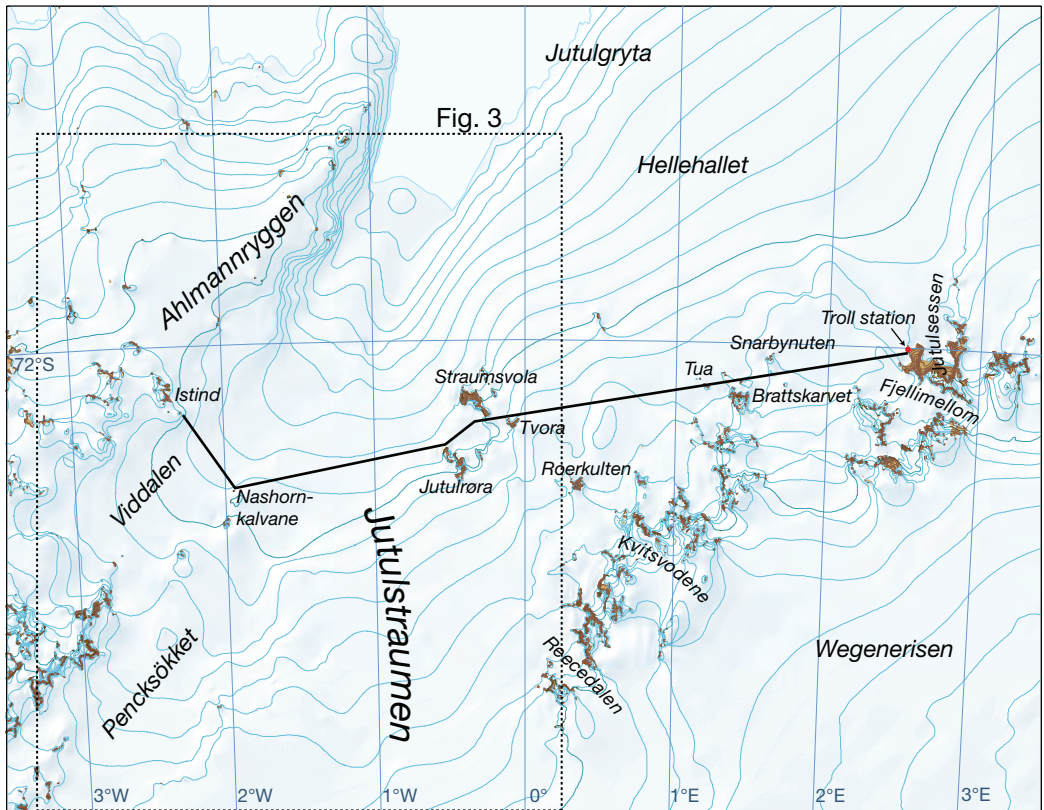
Fig. 1. Location map. Ice from the plateau drains mainly through Jutulstraumen. In the south, Jutulstraumen flows north-westwards, but at about 72° 30' S, where the glacier merges with Pencksökket, the flow turns north-east. The black line shows the radio-echo sounding profile line (Fig. 2). The dotted box marks Figure 3.

The ice occurs in three zones:

1) The coastal zone. Floating ice shelves or outlet glaciers and ice streams beyond the grounding line, which represent about 60 % of the Antarctic coastline.

2) The mountain ridge zone. This zone is usually a 200-300 km wide area inland from the grounding line and from ca. 100 m to >2500 m a.s.l.

3) The interior. This zone comprises the main ice sheet, with many ice domes.

Ice conditions

The first glaciological reconnaissance of the area was performed by the Norwegian-British-Swedish Antarctic Expedition in 1949-52 (e.g. Swithinbank 1959). The next glaciological studies were carried out from the Norwegian station during 1956-60 in connection with the International Geophysical Year 1958-59 (Lunde 1965). From the late 1960s to the beginning of the 1970s, South African, Belgian and Norwegian glaciologists studied Jutulstraumen, Viddalen and Ahlmannryggen. The work included ice thickness measurements by gravimeter and radio-echo sounding, ice velocity, firn temperature and snow stratigraphy (e.g. Bredell 1973, Decleir & Van Autenboer 1982, Gjessing 1972, Wolmarans 1982).

The Jutulstraumen drainage basin (Fig. 1) has been one of the main areas of interest for Norwegian glaciological research in Antarctica since 1990. The studies have focused on mass balance, snow accumulation distribution, ice thickness and ice dynamics (Høydal 1996, Melvold 1999, Rolstad et al. 2000, Melvold & Rolstad 2000). The interior of DML has been one of the large, unexplored parts of the Antarctic continent until relatively recently. Some parts of the area farther inland were investigated under the South Pole Queen Maud Land Traverses in the 1960s (Picciotto et al. 1971). During the EU-funded

project "European Project for Ice Coring in Antarctica (EPICA)" (1996-2006) (EPICA community members 2006) several pre-site surveys were performed in the area, and data on snow accumulation, snow chemistry and meteorology were collected in areas south of the present map area mainly by Nordic and German groups (e.g. Isaksson et al. 1999, Van den Broeke et al. 1999, Hofstede et al. 2004, Karlöf et al. 2005, Oerter et al. 1999, 2000). During IPY (International Polar Year) 2007-09 the Norwegian-American Scientific Traverse of East Antarctica completed two overland traverses in East Antarctica; 1) from the Norwegian Troll Station to the United States South Pole Station in 2007-08; and 2) a return traverse by a different route in 2008-09 rolled in to Troll Station. Among the scientific aims were to investigate climate and establish spatial and temporal variability in snow accumulation over this area of Antarctica to understand its impact on sea level change (Anschütz et al. 2009, Müller at al. 2010).

The areas around H.U. Sverdrupfjella are typical for the mountain ridge zone. Small outlet glaciers between the nunatak groups (e.g. Reecedalen, Kvitsvodene and Rogstadbreen) drain areas of the ice sheet that are several hundred km² in extent. H.U. Sverdrupfjella as a whole acts as a dam to the inland ice, forcing the main part of the areas to the south to be drained through the larger outlet glacier, Jutulstraumen, to the west. Some ice also drains through Sveabreen, but only areas around 2500 m a.s.l. The significant drop in altitude, more than 1000 m over a distance of 50-70 km, leads to increased ice-flow velocity, forming crevasse zones on the small outlet glaciers. The ice-flow directions can easily be detected from satellite images. Locally the ice flow is directed by the mountain ridges, and blue-ice fields with negative balance also initiate local ice flow into cirques.

To the north of the mountain range, the Nilsevidda and Hellehallet ice fields slope gently north-north-westwards from about 1200-1400 m a.s.l. close to the mountain range to about 100 m a.s.l. at the grounding line on the Fimbulisen ice shelf (Jutulgryta). Hellehallet is characterized by a series of large, rounded steps and terraces on the surface (e.g. Lunde 1961, Swithinbank 1957). Several areas of blue ice exist on the lower part of these slopes close to the grounding line (e.g. Orheim & Lucchitta 1987) and ice movement is sluggish, amounting to some tens of metres/year (T. Eiken, pers. comm.). The distance from the mountain range to the grounding line of Fimbulisen is 80-120 km.

Snow accumulation pattern

As a part of the EPICA pre-site survey the snow accumulation rate has been measured within the map area using oxygen isotope analysis, electrical conductivity and radioactive isotope measurements in 10-30 m deep firn cores. The isotope and electrical conductivity stratigraphy were used to calculate the annual net accumulation rate for the last 15 to 30 years and the mean annual net accumulation was deduced from detection of radioactive layers and volcanic eruption layers in the cores (Van den Broeke et al. 1999, Isaksson et al. 1999, Isaksson & Melvold, 2002). The accumulation and climate history for the past 250 years has been determined from a 100 m deep ice core from a coastal location (Kaczmarska et al. 2004, Divine et al. 2009). In addition, ground-penetrating radar surveys were carried out to obtain the spatial accumulation pattern (Richardson & Holmlund 1999, Eisen et al. 2005, Rotschky et al. 2004). Firn drilling and core analysis have also been carried out in Jutulstraumen, Fimbulisen and to the east of the map area by Norwegian scientists (Melvold 1999, Melvold et al. 1998, Orheim et al. 1986). The plateau south of the map area have been extensively investigated as part of the EPICA project where also one of the deep ice cores where drilled (EPICA community members, 2006) by Swedish (Richardson & Holmlund 1999), German (Oerter et al. 1999) and Norwegian (Isaksson et al. 1999) scientists. The spatial accumulation pattern for a large part of DML including the map area has been compiled by Rotschky et al. (2007).

Surface net accumulation generally decreases with the distance from the coast, varying from about 0.40 m water eq./y in the coastal region, less than 0.16-0.10 m water eq./y just south of the nunatak area, and less than 0.05 m water eq./y in the interior at ca. 3500 m a.s.l. (Rotschky et al. 2007, Anschütz et al. 2009). This overall decreasing trend is expected since precipitation (accumulation) is a function of distance from the

moisture source (decreasing as the distance from the coast increases), temperature and elevation (decreasing with decreasing temperature and increasing elevation). The topographic changes from the flat ice shelf areas to the steeper mountain ridge zones cause orographic uplift and adiabatic cooling of humid air, leading to a zone of increased precipitation. The strong katabatic winds redistribute the snow, resulting in accumulation and wind erosion, so that locally the snow accumulation can differ widely from the general trend. By modelling with a simple katabatic-wind model, Van den Broeke et al. (1999) showed that both erosion and deposition expected on Nilsevidda relate to the surface topography. Erosion will occur over ridges, leading to a lower accumulation rate, whereas leeward sides and small depressions and glacier basins will have a relatively high accumulation due to deposition of drifting snow.

The snow density at 1000 m a.s.l. is usually ca. 0.4 g/cm³ in the upper 1 m, increasing to ca. 0.5 g/cm³ at 5 m and ca. 0.6 g/cm³ at 15 m depth. In the coastal high accumulation region the firn-ice transition occur around 60 m (Kaczmarska et al. 2006). The temperature in the snow pack at a depth of 15 m is very close to the mean air temperature at the surface. At 10 m depth, the amplitude of the annual temperature wave at the snow surface is reduced to approximately 5 % of its surface temperature. The 10 m temperature is thus a reasonable measure of the annual mean surface temperature. In drill holes in Jutulstraumen, the 10 m-depth temperature at 710 m a.s.l. was close to -20°C and at 1100 m a.s.l. close to -23°C (Melvold 1999). On the polar plateau, the temperature decreases to -30°C at 2400 m a.s.l. and -51.3°C at 3450 m a.s.l. (Van den Broeke et al. 1999).

Surface features of the snow and ice
The surface of the snow, except in blue-ice areas, is sculptured and modified by strong wind, often at gale force. Wind erosion of hardened snow commonly forms sharp-crested, irregular ridges, known as sastrugi, which may rise as high as one metre above the general snow level, although they seldom exceed 0.3 m. Other features are dunes, which range from ca. 1.5 to 30 m in length and 1.0 to 8 m in width, but their height seldom

exceeds 0.3 m. Dunes have been observed to advance as much as 3.5 m in one hour (Wolmarans et al. 1982). Wind scoops, basin-like depressions caused by funnelling wind action, occur around most nunataks. Most of these wind scoops have floors of blue ice. Away from the nunataks, NNW-directed eddies often scour out elongated to oval depressions in the snow.

Net accumulation takes place over most parts of the map area, but in the Jutulgryta area to the north, radiation is sufficiently high to cause significant amounts of melting. Negative surface mass balance seems to be caused by sublimation in some inland areas of blue ice. These are generally located on the northern and north-western sides of the large nunataks such as Straumsvola and Brattskarvet.

Blue-ice fields are typically developed on the lee side of the mountains or nunataks where strong katabatic winds blow away most of the snow and local climatic conditions result in net ablation. Some studies of ice flow and mass balance have been done in the blue-ice fields close to Troll Station in Jutulsessen, and an annual net ablation of 10 to 15 cm/y was measured in the period from 1990 to 1993 (Hagen 1997, Thomsen & Hagen 1997). More detailed studies on the formation and surface energy balance have been carried out in similar blue-ice areas in the western part of Dronning Maud Land, in Heimefrontfjella (e.g. Jonsson 1990).

Blue-ice areas have been observed in the Jutulgryta area, north of the area mapped, at an elevation of 150 m a.s.l. The grounding line there has been regularly monitored by recent Norwegian Antarctic Expeditions (NARE) (e.g. Winther et al. 1996). These blue-ice areas continuously experience below-surface melting in summer, generally while the air and the snow- and ice-surface temperatures are below freezing. The subsurface melting is mainly explained by the interaction between the snow, ice and water matrix and the near-surface radiation and energy balance (Liston et al. 1999).

Ice thickness and bedrock topography
Only limited data on ice thickness exist from the map area. However, the ice thickness and bed topography along a profile from Istind to Jutulsessen (Fig. 1) provide important information.
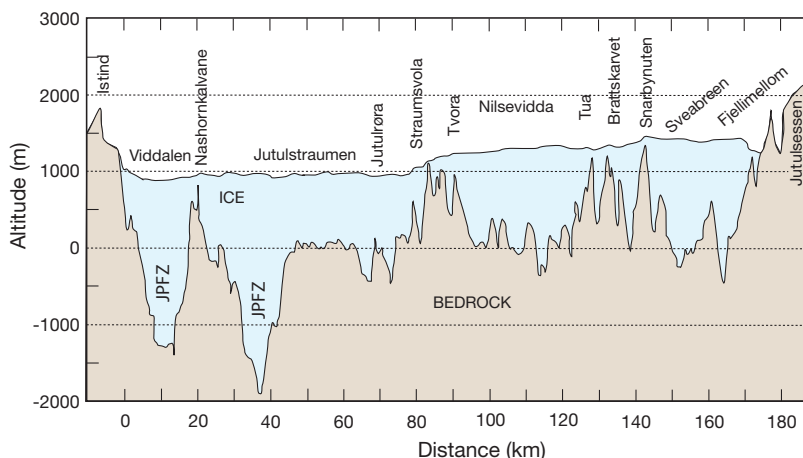
Fig. 2. Bedrock and ice surface profile along a 190 km long line from Istind to the Jutulsessen nunataks. Positions of outlet glaciers and ice fields are marked along the profile and vertical names indicate areas where the profile line passes close to the nunataks (Näslund 1998, Melvold & Rolstad 2000).

The profile (Fig. 2) is based on airborne radio-echo sounding data carried out during the NARE (1996-97) (Melvold & Rolstad 2000, Näslund 1998) and shows the cross-section of Viddalen and Jutulstraumen, Nilsevidda and Sveabreen.

The bedrock surface is below or close to sea level along significant parts of the profile, inferring a pre-glaciation and/or continental ice sheet erosion surface. The bedrock surface beneath Viddalen, on the west side of Nashornkalvane, is symmetrical and more U-shaped than suggested by Decleir & Van Autenboer (1982). It is 1300 m b.s.l. and the maximum depth of 1900 m b.s.l. (ice thickness 2850 m) occurs 17 km east of the Nashornkalvane nunatak. The deepest trenches occur where the profile traverses the Jutulstraumen-Pencksökket Trough, which is a major subglacial valley that can be traced southwestwards through Pencksökket as far as 6° W. The U-shaped valley has a more than 5 km wide bottom, caused by ice flow, probably along a distinctly soft rock zone, perhaps a shear zone. This valley, together with the deepest valley along the eastern side of Neshornkalvane, must be formed in close relationship with the JPRZ, which is a major plate boundary fault between the South African and East Antarctic plates.

In the central and eastern parts of the Jutulstraumen profile, the ice thickness is relatively constant, ca. 1000 m, and the bedrock floor is close to sea level. The ice thickness increases to about 1300 m close to Jutulrøra and is generally somewhat thicker than that inferred from gravimetrical data along the same profile (Decleir & Van Autenboer 1982). Variable ice thickness and bedrock elevation is found when the radio-echo sounding profile runs close to Jutulrøra, Straumsvola and the Tvora nunataks. Further to the east, across Nilsevidda, the ice thickness increases to about 1000-1200 m and the bedrock floor is close to sea level. As the profile passes close to the Brattskarvet and Snarbynuten nunataks, the ice thickness and bedrock floor show large variations along the profile. The bedrock surface is high around the massive, resistant igneous bodies of the Straumsvola and Tvora syenites and the Brattskarvet granitoids. Increased ice thickness occurs beneath the Sveabreen and Fjellimellom outlet glaciers where the bedrock floor is close to, or below, sea level. The E-W trending geological structures from the east turn to the NE-SW trend of H.U. Sverdrupfjella around these glaciers. A NW-SE trending structural break, in the same trend as the direction of the upper Jutulstraumen which limits the southern end of H.U. Sverdrupfjella, is inferred along these glaciers.

The profile (Fig. 2) shows examples of extreme topography, for instance from the top of the Nashornkalvane nunatak (ca. 1020 m a.s.l.) to the trough beneath Jutulstraumen (ca. 1900 b.s.l.), where the elevation difference is 2920 m
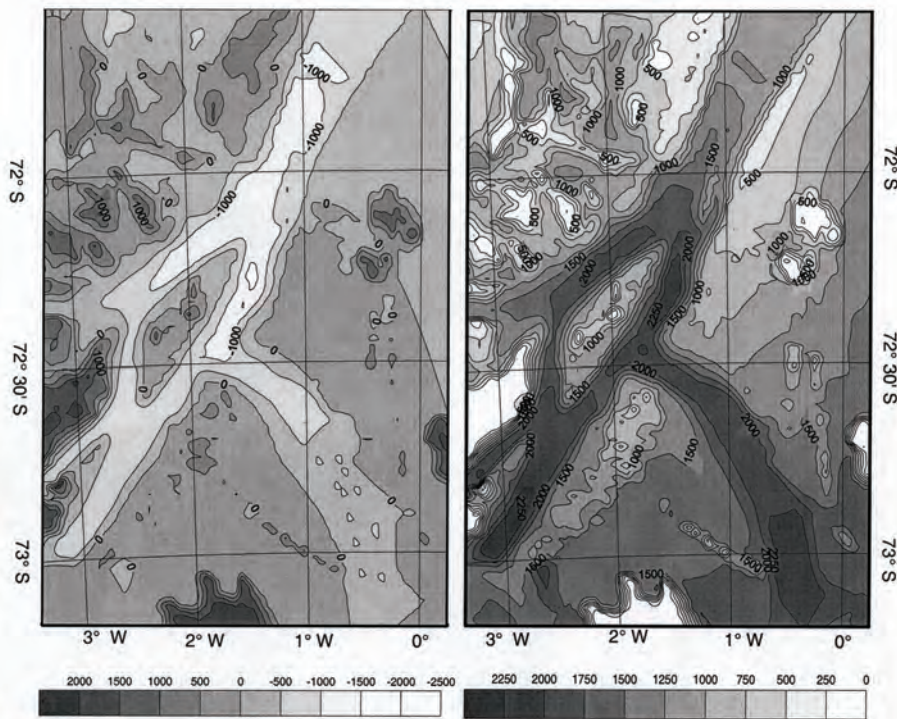
10

Fig. 3. Bedrock topography (left) and ice thickness (right) grids. The contour interval is 500 m (modified from Melvold & Rolstad 2000). The map area is shown in Fig. 1.

over a horizontal distance of 17 km, resulting in a slope gradient of ca. 10º. Similar gradients are found at several places in connection with the deep valley systems in Antarctica.

The bedrock floor and topography beneath Jutulstraumen can be studied on a regional scale in Fig. 3 (Melvold & Rolstad 2000). The general elevation of the nunataks and the subglacial topography decreases northwards. Several deep, subglacial valleys cut through relatively flat platforms and high peaks. The outlet glaciers, Jutulstraumen, Pencksökket and Viddalen, occupy this large valley system. The ice in the deep valleys is, in general, approximately 1000-1500 m thicker than in the surrounding areas. In addition to the well-known Pencksökket depression, a prominent depression with a WNW-ESE orientation at about 72° 30' S and 1° 30' W is obvious in Jutulstraumen from the bed topography grid. The depression is 800 m deep relative to the adjacent subglacial terrain, but its south-eastward extent is unknown.

Geology

History of geological studies

The first geological descriptions of H.U. Sverdrupfjella were presented by Roer (1954), Roots (1953, 1969) and Giaever (1954). The Soviet Antarctic Expedition during the 1950s and 1960s made reconnaissance visits to the area, and metamorphic rocks in the north-eastern part of the map area were described by Ravich & Solov'ev (1966). South African geologists started their investigations of the area in the late 1960s. Their early activity was concentrated in the area west of Jutulstraumen (Neethling 1969, Wolmarans & Kent 1982). Hjelle and Winsnes, members of the Norwegian Antarctic Expedition in 1971-72 (Hjelle 1974), carried out the first comprehensive geological field work on H.U. Sverdrupfjella and established the principal lithostratigraphy of the area.

From the mid 1980s, South African scientists (Grantham et al. 1988, Moyes & Barton 1990, Allen 1991, Grantham 1992, Groenewald 1995) performed extensive mapping and isotopic work on H.U. Sverdrupfjella. More recent work by

South African geologists includes tectonothermal studies on H.U. Sverdrupfjella (Board et al. 2005) and Gjelsvikfjella (Bisnath & Frimmel 2005, Bisnath et al. 2006).

Geological outline
The coastal mountain ranges of Dronning Maud Land are exposed subparallel to and about 200-250 km from the edge of the East Antarctic Ice Sheet. Geological structures display NE-SW trends in H.U. Sverdrupfjella (west of 1° 45' E) and show E-W trends in Gjelsvikfjella and Mühlig-Hofmannfjella east of 1° 45' E (Ohta et al. 1990). Ice cover makes it difficult to observe any geological break between these two areas. Transverse faults have been inferred along Sveabreen in the north-east and upper Jutulstraumen in the south-west based on radio-echo sounding on the ice fields (Fig. 3).

H.U. Sverdrupfjella forms the north-eastern part of a high-grade metamorphic terrane, known as the Maud Belt. The nunataks and mountains cover ca. 6000 km². and are located to the east of the Jutulstraumen-Pencksökket Rift Zone (JPRZ). Achaean rocks of the Grunehogna Province are exposed west of the JPRZ. The Maud Belt has been interpreted as a juvenile island arc that was tectonically juxtaposed along the margin of the Grunehogna Province at the end of the Mesoproterozoic (Arndt et al. 1991, Grantham et al. 1995, Groenewald et al. 1995). The orogenic belt was subsequently modified by metamorphism, deformation and magmatism during the Cambrian Pan-African orogeny (Moyes & Barton 1990, Moyes et al. 1993b, Board et al. 2005, Bisnath et al. 2006). Jurassic alkaline intrusions occur along the eastern margin of the JPFZ (Groenewald et al. 1991; Grantham & Hunter 1991).

The map area is divided into two NE-SW trending geographical regions.

The western region includes the nunataks at Gburekfjella and Nilsevidda, and extends to the western foothills of H.U. Sverdrupfjella. The area is dominated by amphibolite facies supracrustals and subordinate granitic gneisses, which are intruded by Early Palaeozoic granite and pegmatite sheets and by Jurassic alkaline igneous rocks. The eastern region includes most of the H.U. Sverdrupfjella nunataks, and consists of partially retrogressed granulite facies granitic

gneisses and migmatites. The Early Palaeozoic Brattskarvet granitoid batholith occurs in the north-eastern part of the region. There is also a structural difference in that the main eastern outcrops are characterized by an east to south-east dipping series of thrusts subparallel to the general layering and foliation, whereas the bedrock in the west is complexly folded in a variety of orientations.

The north-westward transport on the south-east dipping thrusts suggests that the high-grade rocks in the east have been thrust over the lower-grade rocks to the west. The rocks were metamorphosed during the Late Mesoproterozoic orogenic event at ca. 1035 Ma and were subsequently overprinted during the Early Palaeozoic thermal event at ca. 540 Ma (Board et al. 2005). Numerous igneous rocks have intruded the supracrustal units at various times during the Mesoproterozoic to Middle Mesozoic, the latest intrusions being concentrated along the eastern margin of the JPFZ.

Description of the lithostratigraphic units
Sverdrupfjella supersuite
The metasupracrustal rocks of the area are grouped into the Sverdrupfjella supersuite, and are subdivided into three complexes. These are, in structurally descending order:
i) Rootshorga gneiss complex, ii) Fuglefjellet gneiss complex, and iii) Jutulrøra gneiss complex.

Jutulrøra gneiss complex (13, 14) (numbers in parentheses correspond to the map legend)
The Jutulrøra gneiss complex, the structurally lowest exposed unit, is found in the western region and its thickness is estimated to be ca. 2.5 km. It is interpreted as a volcanic arc sequence of interlayered calc-alkaline volcanic and clastic sedimentary rocks on the basis of geochemical and field characteristics (Groenewald et al. 1995).

The Jutulrøra gneiss complex comprises two lithological units, grey quartzo-feldspathic gneiss (13) and banded gneiss (14). Both units are intruded by Mesoproterozoic and Early Palaeozoic granitoids and Mesozoic alkaline rocks. Dolerite dykes are present throughout the region.

Fig. 4. Grey quartzo-feldspathic gneisses of the Jutulrøra gneiss complex, with concordant amphibolite lenses (ca. 2 m thick), at Jutulrøra.

The grey quartzo-feldspathic gneiss (13) (map name: quartz-feldspar gneiss) are leucocratic to mesocratic, felsic to intermediate rocks (Fig. 4) which are medium grained, equigranular and homogeneous with respect to mineralogy. Typical assemblages comprise Qtz + Pl + Kfs + Hbl + Bt + Ep (mineral abbreviations are after Kretz 1983) with accessory apatite, zircon, titanite and allanite. The rocks have a tonalitic to granodioritic composition with plagioclase (An_{26-30}) > K-feldspar. The foliation is defined by hornblende and is partially overgrown by biotite and epidote. Concordant layering, local existence of large plagioclase porphyries/porphyroclasts, twinning characteristics, and the coherency of their chemistry suggest an extrusive volcanic origin (Grantham 1992). Zones containing dark schlieren are interpreted as autoliths of agglomeratic rocks. A west-to-east decrease in the epidote content reflects an eastward increase of metamorphic grade.

Banded gneiss (14) is interlayered with the quartzo-feldspathic gneiss, and is distinguished from the latter by its strong compositional, metre-scale layering (Fig. 5). The layers consist of felsic gneiss, amphibolite and subordinate Mg-rich mafic rocks, calcareous and semipelitic gneiss. The main constituents of mafic layers are hornblende and plagioclase (An_{25-35}) with subordinate Di, Grt, Qtz, Bt, Spn, Chl, Ep and Cum. The rocks display a fine- to medium-grained, granoblastic texture. The foliation is defined by hornblende, which is locally overgrown by biotite.

Diopside appears as inclusions in hornblende, whereas garnet occurs around poikiloblastic hornblende. Bulk rock compositions show a wide range of SiO_2 (47 to 56 %) and MgO (5 to 14 %), showing no coherent linear variation (Grantham 1992).

Felsic layers are fine- to medium-grained, granoblastic quartzo-feldspathic rocks. Planar fabrics are defined by garnet, hornblende and biotite with relict aluminosilicates. SiO_2 contents range from 58 to 81 % and no coherent linear variation of oxides vs. SiO_2 are seen (Grantham 1992). The interlayered concordant character and siliceous nature of the rocks suggest either a volcanoclastic or a sedimentary origin.

Mg-rich mafic bands are pale green and display a weak schistose texture defined by Tlc-Ath-Phl. Relict olivine and diopside are observed in some samples. The chemical compositions suggest ultramafic precursors (Grantham 1992). The interlayered concordant occurrence of the layers indicates that the rocks may represent Mg-rich lavas.

The mineralogy of the rare calc-silicate rocks includes quartz, diopside, grossular garnet and plagioclase with subordinate amounts of hornblende and titanite. The high quartz and SiO_2 contents with diopside and plagioclase and high Ni (ca. 300 ppm) and Cr (1200 ppm) suggest a highly silicified mafic-ultramafic protolith (Grantham 1992).

Fuglefjellet gneiss complex (15, 16)
The Fuglefjellet gneiss complex consists mainly

of banded gneisses (15) (the map legend, "pelitic gneisses", is in error), marbles and calc-silicates (16). The thickness of the Fuglefjellet gneiss complex is estimated to be less than 4 km. The banded gneisses in Jutulrøra and in the Fuglefjellet gneiss complex are interlayered with carbonate rocks, suggesting a lithological relation between the two complexes. An open synform, which involves both complexes, can be inferred around Roerkulten, while an open antiform with a moderately S-plunging axis is assumed in the nunatak area west of Knattebrauta (Hjelle 1974). These fold structures suggest that the Fuglefjellet gneiss complex structurally overlies the Jutulrøra gneiss complex.

The eastern boundary of the Fuglefjellet gneiss complex is inferred to be a thrust fault. Eastward-dipping, NE-SW trending thrusts are exposed at Kvithovden, Gordonnuten and Skarsnuten. The thrusts cut the open fold structures of the Fuglefjellet gneiss complex.

The banded gneisses (15) of the Fuglefjellet gneiss complex are similar to the banded gneisses of the Jutulrøra gneiss complex and include quartzo-feldspathic, mafic and Mg-rich gneisses. The quartzo-feldspathic gneisses exhibit a medium-grained granoblastic texture, and contain Pl, Kfs, Bt, Hbl, Cpx and Grt, and accessory epidote, titanite and opaque minerals. Ep-Hbl symplectites are developed along the boundary between clinopyroxene and hornblende, whereas Zo-Qtz symplectites occur between garnet and plagioclase. The bulk chemical composition is tonalitic (Grantham 1992). The presence of deformed conglomerates and quartz-rich lithologies suggest a sedimentary origin for parts of the quartzo-feldspathic sequence.

The mafic gneisses are medium-grained, granoblastic and contain the mineral assemblages Hbl + Bt + Pl ($>An_{23}$) $\pm$ Cpx $\pm$ Grt $\pm$ Ep $\pm$ Chl with accessory titanite, magnetite, calcite, apatite, tourmaline and opaques. The Mg-rich mafic gneisses contain serpentinized Ol, Amp, Phl, Tlc and Mgs. The gneissosity is defined by phlogopite and late serpentine. The Mg-rich rocks are interpreted to have an ultramafic igneous origin, based on their high Cr and Ni contents (Grantham 1992).

Marbles and calc-silicates (16) display a weakly foliated, fine- to coarse-grained granoblastic texture. Marbles contain the mineral assemblages Cal + Dol + Tr $\pm$ Tlc $\pm$ Phl $\pm$ Mtc $\pm$ Brc $\pm$ Srp. Serpentine and brucite aggregates are formed after olivine. The mineralogy of the calc-silicate rocks is Di + Hbl + Pl + Qtz $\pm$ Sph $\pm$ Sca $\pm$ Kfs $\pm$ Cal $\pm$ Bt $\pm$ Ep/Zo $\pm$ Chl. Centimetre- scale banding is defined by alternating diopside- and hornblende-rich layers. Various subordinate hydrous minerals indicate later hydrous retrogression under high-CO_2 conditions. The carbonaceous and calc-silicate rocks are most probably of calcareous arenite origin, suggested by low Cr, Ni and Cu and high SiO_2 (Grantham 1992).

Fig. 6. Quartzo-feldspathic gneiss of the Rootshorga gneiss complex in the lower part, and the Sveabreen megacrystic granite forming a tower above. A shear zone is developed along the boundary. Tua nunataks.

Rootshorga gneiss complex (17, 18, 19)
The Rootshorga gneiss complex dominates the outcrops along the entire length of eastern H.U. Sverdrupfjella. The complex strikes NE-SW and shows moderately SE-dipping foliations. The predominantly felsic to intermediate complex includes pelitic gneisses, minor amphibolites and ultramafic rocks. The Sveabreen megacrystic granites are tectonically interlayered with the gneisses (Fig. 6), and both complexes are intruded by the Brattskarvet granitoid batholith and a variety of granitic dykes in the north-eastern part. High-Al pelitic gneisses interlayered with semi-pelitic and quartzo-feldspathic gneisses (Fig. 7), and the presence of metaconglomerates and interlayered quartzite, indicate a sedimentary protolith for the Rootshorga gneiss complex. Segments containing blastoporphyritic texture are likely to have volcanic protoliths. A back-arc basin is envisaged for the depositional environment of the Fuglefjellet and Rootshorga gneiss complexes based on their field appearance and geochemistry.

The pelitic gneisses (17) are medium grained, banded and commonly display a migmatitic

Fig. 7. Banded gneisses of the Rootshorga gneiss complex on the north-western spur of the 2608 m peak at middle Isingen. The lower part consists of intermediate and quartzo-feldspathic gneisses and the upper part comprises semipelitic and pelitic gneisses. Two generations of granitic veins are recognizable.
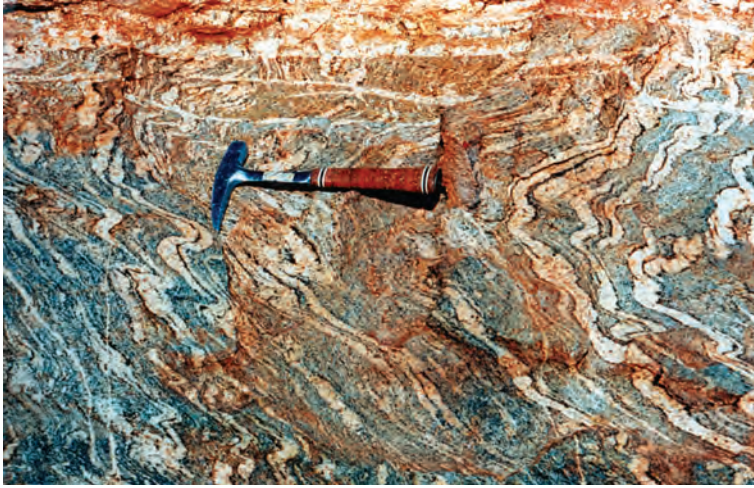
15

Fig. 8. Migmatite with ptygmatic veins in the semipelitic gneiss of the Rootshorga gneiss complex on the 1540 m peak of Heddenberget, northern Rømlingane.

structure (Fig. 8); they contain mafic and melanocratic lenses. Typical assemblages comprise Qtz, Pl and Kfs (each 0-30 modal %), Bt (<45 %), Sil, Crd and Grt (each 0-20 %), Hbl (<5 %) and Mag (<3 %). Two types of high-alumina rocks are recognized: 1) Sil-Grt-Crd bearing migmatite, and 2) Bt-rich rocks without aluminosilicates. Pelitic gneisses from western Salknappen are fine- to medium-grained, strongly foliated rocks with the assemblages Bt + Qtz + Kfs ± Grt ± Sil ± Ms and secondary chlorite.

Within unit 17, there are gneisses that show gradational variations to the pelitic and quartzofeldspathic gneisses. These rocks are medium- to coarse-grained, equigranular, with porphyroblasts of plagioclase and garnet, and they contain hornblende with inclusions of relict augite, hypersthene and plagioclase (An_{32}). Their bulk compositions range from tonalitic to dioritic.

The quartzo-feldspathic gneisses (18) comprise quartz, feldspar and small amounts of biotite and garnet. Preservation of cross-bedding and metaconglomerate indicates a sedimentary origin for the unit. The rocks are granular to granoblastic, with highly strained ribbons and polygonized leaves of quartz. Abundant microcline is rarely perthitic. Garnet commonly forms atolls or thin, irregularly curved stringers. Quartz-rich gneisses are distinctly banded with local magnetite-rich laminae. The protoliths are interpreted to be alternating, immature arenites and argillites.
Metamafic and calc-silicate gneisses are found as

inclusions in units 17 and 18.
Mafic rocks, which occur as layers and enclaves in other lithologies, are present as garnet pyroxenites, garnet amphibolites and biotite-hornblende schists. An eclogitic mafic relic is present on peak 1515 m, north-east of the Brattskarvet granitoid batholith. Biotite, hornblende, plagioclase and garnet are the major mineral phases. Clinopyroxene occurs as inclusions in hornblende. Calc-silicate gneisses occur as small concordant lenses. Typical calc-silicate assemblages comprise i) Ves + Di + Grt + An, ii) Zo + Czo + Di + Scp + An + Grt + Sph, and iii) Adr + Di + Ep + Mag + Hem.

The felsic orthogneisses (19) are pre-tectonic metagranitoid rocks that occur as conformable units, irregularly shaped bodies and deformed layers/veins and leucosomes. The orthogneisses occur on Rømlingane and Vendeholten as strongly deformed, discontinuous, irregular and tabular bodies, 30-100 m thick. A thick unit occurs along a thrust from Oppkuven in middle Rootshorge to Høyhamaren in Hamaren and the southernmost nunatak of Rootshorge in southern H.U. Sverdrupfjella. These rocks are mesocratic (mafic index ca. 30 %) monzogranite. The presence of dark-grey andesine porphyroclasts and xenocrystic garnets (<5 vol. %) is notable. Mafic enclaves are amphibolites and biotite-garnet gneisses (<20 vol %). The granitoids exhibit well-developed foliation, locally as a composite S-C fabric or double cleavage. No pre-existing

Fig. 9. A transecting contact of the Römlingane granite to the banded gneiss of the Rootshorga gneiss complex. Both are crosscut by a meta-olivine gabbronorite (left), which encloses gneiss enclaves. Near the 1590 peak of south-western Vendeholten.

foliation has been seen in any of the enclaves. These granites locally transgress the gneissosity of the surrounding gneisses (Fig. 9). The orthogneisses are characterized by normally zoned plagioclase, microperthitic K-feldspar porphyroclasts and myrmekite developed around the porphyroclastic feldspars. Biotite occurs predominantly in pressure shadows. Two types of biotite have been distinguished, dark-brown xenocrysts and secondary pale-brown subidiomorphic flakes. Garnet is a common accessory, both as irregular relicts and small idiomorphic grains in plagioclase. Rare hornblende is associated with biotite. Tourmaline (schorlitic), zoisite, allanite, magnetite and titanite are accessories.

Early granitoid rocks

The early granitoid rocks are highly deformed and gneissose, but are recognized as intrusives based on transgressive contacts, xenolithic enclaves, microstructures and geochemistry.

Sveabreen migmatitic granite (10) is widely distributed in the eastern region. It occurs as tabular units, 0.1-3 km thick and up to 30 km in strike length. The contacts are mostly thrust bounded, although at some localities there appear to be gradational contacts with the country rock gneisses. Mylonites occur parallel to the lower contacts at Tverrnipa, northern Vendehø, and a large-scale, low-angle discordance along this contact is recognized on western Vendeholten. The presence

of two intersecting cleavages and folded foliations suggests that this granite pre-dates all major deformation phases. Harris et al. (1995) obtained a U-Pb zircon age of 1127 ± 12 Ma for the Sveabreen granite, whereas Rb-Sr isotopic dates yielded 1028 ± 94 Ma (Groenewald et al. 1995). The Sveabreen migmatitic granite is medium- to coarse-grained and strongly foliated. It is mostly monzogranitic and has S-type characteristics; it locally carries sillimanite and garnet. Carlsbad twins and rare composite porphyroclasts of plagioclase and K-feldspar support an igneous origin. Finely polygonized K-feldspar and irregular to leaf-like quartz aggregates constitute the matrix. Mafic constituents (<15 %) are mainly fine flakes of biotite along foliations. Accessories are garnet, zircon, allanite and apatite.

Four main compositional variants are present: i) leucocratic, very coarse-grained, megacrystic granite, ii) darker porphyroclastic granite, iii) melanocratic granodiorite-diorite, and iv) equigranular leucogranite.

The western granitoid rocks (9) are exposed at Jutulrøra, Brekkerista and Roerkulten, and occur as sheet-like, concordant and subconcordant intrusions within the Jutulrøra gneiss complex. Cross-cutting relationships are locally preserved (Fig. 10). The granitoids at Jutulrøra and Roerkulten are pale pink, equigranular and medium grained, with ca. 25 % quartz and K-feldspar > plagioclase. Poikilitic hornblende and

17

Fig. 10. A sheet of the Jutulrøra granite, western granitoid rocks (9), intruding the banded gneiss of the Jutulrøra gneiss complex. E-W face of southern Jutulrøra, ca. 400 m height.

biotite define the foliation. Late biotite cross-cuts the fabric and replaces hornblende. Accessory minerals include zircon, apatite, garnet and allanite. Titanite is common in the Jutulrøra granite, whereas other opaques occur in the Roerkulten granite.

The Brekkerista granitoid is darker orange-pink and contains 25 % quartz and K-feldspar > plagioclase (An_{20}). K-feldspar forms rodded porphyroclasts (Fig. 11). Biotite defines a weak foliation and is locally altered to chlorite. Accessory minerals include titanite, allanite, apatite and zircon. Rare muscovite and epidote are present as late retrograde phases.

The major and trace element analyses are given in Grantham (1992). All three granites display a combination of I-, S- and A-type characteristics, which suggests a cogenetic origin and a hybrid nature of their sources. Partial melt models suggest that the Jutulrøra granite and the Roerkulten granite may be derived from a source similar to the quartzo-feldspathic gneisses of the Jutulrøra gneiss complex (Grantham 1992).

Early mafic rocks

The majority of the mafic bodies occur as lenses, boudins, concordant layers and rare dykes. They pre-date the deformation and are intrusive into the oldest pre-tectonic granite. Older metamafic rocks of the western region are present as amphibolites and grey diorites. The rocks are typically garnet-free and contain Hbl + Pl, consistent with medium-grade metamorphism. The grey diorites form narrow dykes in the banded gneisses of the Jutulrøra gneiss complex and the Brekkerista granite. Their granoblastic texture indicates metamorphism subsequent to intrusion. The mineral assemblage is Pl + Hbl + Qtz + Bt + Ep + Zo. Two generations of mapable early amphibolites are present.

Older amphibolites, A-1, (12) pre-date all recognizable deformation; they have a planar foliation parallel to S_1, but crosscut the compositional layering of surrounding gneisses. These older amphibolites occur in Vendeholten, Salknappen and Isingen. Some are isolated, apparently conformable boudins and lenses in the gneisses, implying a supracrustal origin, whilst others form arrays that suggest that the protolith was a crosscutting dyke. The rocks consist predominantly of hornblende, plagioclase and biotite. Garnet, quartz and clinopyroxene are present in some rocks, but all three of these phases never coexist. Relict clinopyroxene is commonly included in hornblende. Corundum and tourmaline occur as inclusions in garnet.

Younger amphibolites, A-2 and A-3, (11) are also pre-tectonic; they occur as narrow dykes and post-date the Sveabreen megacrystic granites and other early granitoids. They crosscut the lithological banding and S_1 foliation of surrounding gneisses (Fig. 12). The rocks have

18

Fig. 11. L-tectonite in the Brekkerista granite, western granitoid rocks (9). Brekkerista. Pencil = 10 cm.
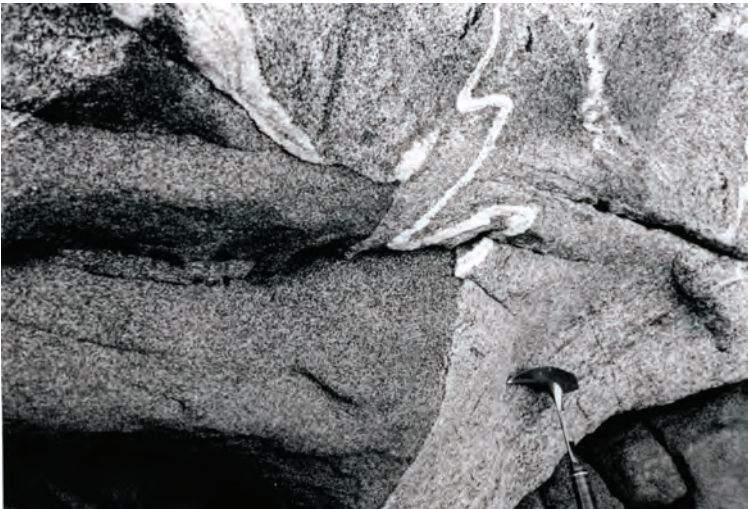

Fig. 12. In the left part of the photograph, an amphibolite (A-2) transects a leucosome of western granitoid (9). Roerkulten.

abundant hornblende and biotite, and fewer relict anhydrous minerals. The hornblende defines a penetrative fabric. Plagioclase, An_{26-38}, is polygonized and primary characteristics are present as strong normal and oscillatory zoning in some large grains. The amphibolites may have been emplaced during a period of general decompression and rehydration, synchronous to intense deformation.

Early metamafic rocks of the eastern region are typically found as boudins, which partly preserve primary igneous assemblages and textures in their cores (Hjelle 1974, Groenewald & Hunter 1991, Groenewald 1995). The mafic rocks occur in the paragneisses and, to a lesser extent,

in orthogneisses. They display a large range of mineral and chemical compositions. The cores of boudins are typically isotropic and comprise early metamorphic assemblages and textures. The non-mapable pre- and synorogenic metamafic rocks are divided into four groups (Groenewald 1993, Groenewald & Hunter 1991):

i) Tholeiitic, low-alkaline garnet-pyroxenites

Numerous lenses and boudins of pre-tectonic, tholeiitic metamafic rocks occur on the 1515 m nunatak, Tua, Brattskarvet, Vendeholten, Isingen and Kvitskarvet in northern H.U. Sverdrupfjella and at numerous localities in southern H.U. Sverdrupfjella. They are characterized by Grt-Cpx-Qtz-Rt ± Pl/Opx relict assemblages, replaced to

Fig. 13. Folded boudins of garnet pyroxenite in the semipelitic gneisses of the Rootshorga gneiss complex on the 1540 m peak of Heddenberget, northern Rømlingane. Pencil = 15 cm.

varying degrees by kelyphitic/symplectic Opx-An-Ilm or Hbl-Pl-Ilm intergrowths. The clinopyroxene has exsolved oligoclase, indicating an original omphacite content. A dyke-like lens of this rock is crosscut by the Rømlingane granite in nunataks in the western part of Vendeholten (Fig. 13). An eclogite facies for the early metamorphism has been proposed by Groenewald (1993, 1995). Board et al. (2005) reported relicts of eclogite facies garnet-omphacite assemblages within strain-protected mafic boudins.

ii) Metanorite

Metanorite occurs on peak 1540 m of southwestern Vendeholten as scattered boudins which consist mainly of metamorphic orthopyroxene and plagioclase. Biotite, plagioclase (strongly zoned, An_{28-45}), and rutile inclusions are aligned within the host hypersthene poikiloblasts, indicating a prograde penetrative foliation before the formation of hypersthene. Polygonized aggregates of pale-coloured Mg-hornblende are a local major phase and cummingtonite and/or anthophyllite occur as minor interstitial phases. Opx-Pl symplectite with sparse Grt-Cpx relicts between hypersthene poikiloblasts indicate that these rocks were emplaced before the major tectonothermal events.

iii) Garnet peridotite

More than 30 lenses of mafic and ultramafic rocks occur at Salknappen and Isingen, in the Rootshorga gneiss complex, close to an inferred tectonic break to the Fuglefjellet gneiss complex to the west. The metamorphic assemblages and apparent crosscutting relationships of the alignments of lenses relative to the local lithological layering suggest pre- or early- syntectonic emplacement. Primary phases in the core of these boudins are hornblende-olivine-pyroxene. Olivine and orthopyroxene are commonly euhedral to subhedral and equigranular. Some Mg-hornblende grains contain olivine and orthopyroxene inclusions, while other grains are idiomorphic and apparently coexist with olivine and orthopyroxene. Initial clusters of green-brown, subhedral spinel and Mg-garnet are surrounded by radiating pale-green spinel lamellae. The largest body at Salknappen has pyroxenitic and dunitic, 0.2-2 m thick, cumulate layers. The coarse-grained equigranular texture of the lenses suggests that they are tectonic fragments of a larger fractionated body. Amphibolites, consisting mainly of plagioclase, garnet, hornblende and biotite, represent re-equilibration during metamorphism.

iv) Olivine gabbronorites

Olivine gabbronorite occurs in the south-western Vendeholten as a wedge-like fragment of a dyke, 300 m in length. The margins of the body are strongly amphibolitized and foliated. Coarse-grained poikilitic, relict textures have been retained in the core and consist of olivine, bronzite and plagioclase enclosed by large augite. Spinel inclusions occur in the primary minerals. Coronas of Opx-Cpx-Grt are formed between olivine and plagioclase, and Cpx-Grt intergrowth is formed between hypersthene and plagioclase (Groenewald 1995). Primary plagioclase, An_{50}, is clouded by aluminosilicate needles. Primary ortho- and clinopyroxenes are also clouded, probably by garnet or spinel.

The subdivision of the metamafic rocks is supported by the differences in their geochemistry (Grantham 1992). All of these rocks display a general subalkaline nature. Strongly amphibolitized rocks plot in the alkaline field, while garnet pyroxenites are strongly subalkaline. The garnet pyroxenites, garnet peridotites, olivine gabbronorites, metanorites and amphibolites are all tholeiitic. The olivine gabbronorites show a tendency to enter the calc-alkaline field, due to amphibolitization. Estimation of the tectonic setting from chemical parameters is only provisional. The garnet pyroxenites plot in the field of low-K tholeiites, while all other rocks plot in ocean-floor basalts.

Late granitoid intrusions

The late intrusions are post-tectonic (with respect to the main D_2 fabric) and are mostly granitic in composition. They include the "Dalmatian-type" granite, the Brattskarvet granitoid suite, abundant granitic and pegmatitic dykes, and lamprophyre.

The "Dalmatian granite" occurs as sheet-like bodies, up to 10 m thick, which intrude the Jutulrøra and Fuglefjellet gneiss complexes and the Sveabreen migmatitic granite. The granite sheets are exposed from Jutulrøra to Roerkulten and Robinheia (not shown on the map). The best exposures are at Brekkerista, Dvergen and Fuglefjellet. The granite is typically equigranular and medium grained, and is composed of quartz (33 %), microcline (32 %), plagioclase (28 %), muscovite (4 %) and biotite (2 %) (Grantham et al. 1991). Accessory phases include apatite, magnetite and zircon. Two varieties have been recognized, i) a type with tourmaline-bearing nodules, and ii) a type with magnetite phenocrysts. The tourmaline nodules have a melanocratic (tourmaline-bearing) core and a leucocratic rim, and are up to 10 cm in diameter. The leucocratic rim is characterized by muscovite. The granite has dominantly S-type characteristics. Rb-Sr whole-rock isotopic data yielded an age of ca. 470 Ma for the emplacement of the granite (Grantham et al. 1991).

The Brattskarvet granitoid suite (7) intrudes the felsic paragneisses of the Rootshorga gneiss complex in the northern extremity of H.U. Sver-

drupfjella. The massif covers an area of ca. 100 km². The northern contact is a steep fault dipping ca. 80° towards the east. The granitoids are strongly foliated along their margins, whereas a weak foliation is present within the batholith. The intrusive contact at peak 1515 m, northeastern Brattskarvet, is sharp and dips at a low angle to SE. The contact is also exposed at the south-eastern extremity of the batholith, where it dips 25° NW. The main fabric in the surrounding gneisses is conformable with the outline of the intrusion. Granitic dykes derived from the batholith are common along the contact. Alkaline mafic enclaves and dykes occur within the batholith. On the north face of Brattskarvet, layering within the batholith is defined by grain-size and mafic-index differences. A weak alignment of biotite and schlieren relicts is also present. Local weak layering/foliations, augen gneiss texture, agmatitic amphibolites and xenoliths of country rocks suggest that the batholith has a funnel-like structure. The lack of any penetrative foliation in the Brattskarvet granitoids indicates that these intrusions post-dated all significant deformation. The Rb-Sr whole-rock data indicate an intrusive age of 517 ± 15 Ma and Sm-Nd whole-rock data an age of 522 ± 120 Ma. Rb-Sr mineralogical data (biotite, alkali-feldspar, sphene) yielded ages of 482-465 Ma (Moyes et al. 1993).

The Brattskarvet granitoid suite comprises monzonite, quartz monzogranite and monzogranite, and melasyenitic dykes. Monzonites and quartz monzonites are most abundant in the north-western part of the batholith. Gradational transitions between these lithologies suggest large-scale stratification. The uppermost part of the monzonite is irregularly banded, and the layering is defined by alternating coarse- and fine-grained units up to a metre thick. Coarse-grained microcline, <50 modal %, is perthitic in less quartz-rich rocks. Early oligoclase, ca. 40 %, is overgrown by albite. Primary clinopyroxene, <10 %, is augite. Poikilitic, dark arfvedsonite encloses idiomorphic augite. Primary biotite is green lepidomelane, while secondary grains are brown in colour. Titanite and allanite are accessory phases, the latter contains apatite inclusions. The crystallization order in the monzonite is apatite→ feldspars→ magnetite→ titanite→ augite→ allanite.

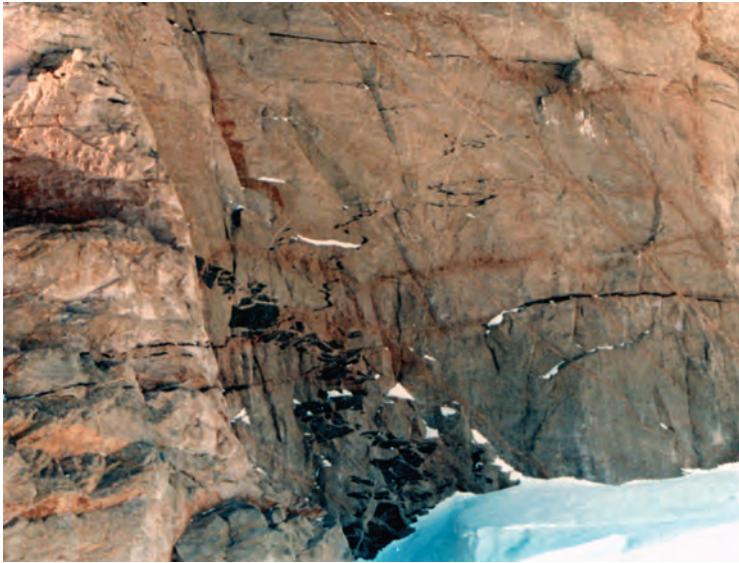The monzogranite consists of coarse perthitic

Fig. 14. Melasyenite dyke; dark fragments in the lower-middle part on the south-western spur of Brattskarvet, adjacent to Brattskarvbrekka. The height of the cliff is ca. 50 m.

K-feldspar (40-55 %), oligoclase (25-35 %), quartz (15-24 %) and biotite (3-5 %), with accessory titanite, apatite, magnetite, fluorite and allanite. Pale-green muscovite and hematite are also present. A few relicts of aegirine augite and arfvedsonite occur in the uppermost part. Biotite is secondary. Leuco-monzogranite makes up the south-western and eastern parts of the batholith. In eastern outcrops, the monzogranites are characterized by variations in the presence or absence of perthite and from medium to coarse grained.
The rocks of the Brattskarvet granitoid suite have characteristics of A-type granites, though some plot in the field of I + S type on the FeO*/MgO vs 1000xGa/Al diagram (Grantham 1992).

Dyke rocks (8)
Melasyenitic intrusives are melanocratic alkaline rocks which occur in the central and northern parts of the batholith (Fig. 14). These rocks appear as continuous units with locally gradational margins against the host monzonite, as rows of disrupted fragmental segments, and as scattered enclaves with reacted margins to poorly defined schlieren. The rocks consist of perthitic K-feldspar (25-60 %), sodic augite or aegirine augite (22-30 %), green amphiboles (4-36 %) and biotite (3-30 %); plagioclase is generally absent, and minor quartz occurs in a few specimens. The mafic minerals are inclusions in feldspar. Secondary actinolite has overgrown the augite and arfved-

sonite and large, interstitial, anhedral aggregates. Apatite and titanite are minor constituents (1-5 %), and allanite, zircon and monazite are accessories. Catapleite ($(Na,Ca)_2ZrSi_3O_9.2H_2O$) and aenigmatite ($Na_2Fe_5TiSi_6O_{20}$) have been provisionally identified.

Alkaline mafic dykes occur in the country gneisses and contain K-feldspar, aegirine-augite, arfvedsonite, titanite and apatite which form the same textures as in the melasyenites, except for the presence of more plagioclase (An_{50}, up to 10 modal %). Felsic veins crosscutting the dykes are isoclinally folded within the dyke and strong foliation parallel to the margins is axial planar to the folds. The veins are less intensely folded when they extend out into the country rocks, and they are affected only by D_3.

A lamprophyre dyke at Vendeholten consists of 2-15 cm aggregates of aegirine-augite and richterite/arfvedsonite in a fine-grained groundmass. A similar dyke occurs in the Rømlingane as part of composite dykes. This rock has a fine- to medium-grained, Kfs-Bt-Di groundmass containing up to 1.5 cm clots of diopside and arfvedsonite or katophorite. The dyke is strongly foliated parallel to its length, possibly due to flow rather than deformation. These dyke rocks are closely similar to the melasyenite occurring within the Brattskarvet batholith.

Xenolithic enclaves are locally abundant in the quartz-monzonites and monzogranites near the northern contacts. The pelitic, semipelitic and quartzo-feldspathic gneiss xenoliths have the same isotopic age range (ca. 950-1180 Ma) as the Rootshorga gneiss complex (Moyes et al. 1993). Autolithic enclaves are commonly present in the monzonites on the north-western spur of Brattskarvet and are mostly fine-grained monzonites. Schlieren are abundant in the transition zone from quartz-monzonites to monzonites. They are petrographically similar to the host rocks.

Alkaline intrusive rocks
The alkaline intrusive rocks in the western region represent the youngest magmatic pulse in the area. Two roughly circular alkaline intrusions cut the Jutulrøra gneiss complex near the eastern margin of the Jutulstraumen-Pencksökket Rift Zone (JPRZ). The contact metamorphic zone is up to ca. 400 m wide. Numerous dykes of alkaline granites, lamprophyres and trachytes intrude the syenite bodies and adjacent country rocks.

Straumsvola syenites (2, 3, 4)
The Straumsvola intrusion is ca. 5 km in diameter and is concentrically zoned with three lithological varieties (2, 3 and 4). The margin shows a clear intrusive relationship with a ca. 50 m zone of vein and dyke intrusions. Contact metamor-

phism of the host gneisses has locally resulted in partial melting and formation of migmatites.

Coarse-grained syenite (2) forms the outermost zone. Anhedral perthitic feldspar (<5 mm) with plagioclase rims and nepheline (2-3 mm) are the main mineral phases in the syenite. Hornblende, biotite and aegirine-augite occur interstitially in the matrix. Preferred alignment of the perthite defines a planar fabric concordant to the margin of the body.

Mesocratic syenite (3) occurs as ca. 5 m wide, near vertical, ring dyke-like, discordant bodies which divide the coarse-grained syenite (2) into outer and inner zones. The mesocratic syenite is coarse grained and has a significantly higher content of mafic and opaque minerals than the other two varieties of the Straumsvola intrusion. Perthite laths define a margin-parallel, planar fabric with interstitial nepheline, clinopyroxene, biotite and opaques. Leucocratic segregation lenses appear perpendicular to the margins. Aegirine is rimmed by vermicular amphiboles and locally radiates from opaque minerals.

The central syenite (4) is strongly layered and forms a saucer-like body with layers dipping 0-20°, sometimes approaching 30° (Harris & Grantham 1993). The layers are rhythmic and

Fig. 15. Layering in the Straumsvola syenite, showing an upward increase in mafic constituents. Southern ridge of Straumsvola.

graded in normal and reverse fashion (Fig. 15), though some show no grading. Some layers near the base have a spotted appearance with roughly spherical clots of mafic minerals, suggesting liquid immiscibility. The layering is defined by strongly oriented perthite laths which have plagioclase rims. Nepheline and amphiboles fill the interspaces, the former commonly being included in mafic minerals and partially replaced by cancrinite. Arfvedsonite-katophorite locally replaces clinopyroxene (aegirine-augite). Biotite partially replaces amphiboles. These textures are interpreted as indicating that perthite and nepheline accumulated on the magma chamber floor trapping intercumulus liquid from which mafic and accessory minerals later crystallized (Grantham 1992). No mafic cumulate phase has been seen. The common replacement of aegirine by amphiboles and amphibole by biotite suggests an enrichment of volatiles, whereas the replacement of nepheline by cancrinite implies the presence of CO_2 in these fluids.

A discordant mafic lens occurs in the coarse-grained syenite near the top of Straumsvola. The boundary is not sharp, and spherical structures of ultramafic composition are seen near the margin of the lenses. These rocks are dominated by nepheline, aegirine, amphiboles and biotite. Coarse-grained clinopyroxene consists of homogeneous rims and poikilitic cores with inclusions of biotite and amphibole. Pink eudialyte occurs on the joint surfaces near the margin. Syenitic pegmatite fills cavities in north-eastern Straumsvola, and these contain crystals of natrolite, acmite, leucite and some adular and zircon.

The coarse-grained syenite and the central layered syenite are typically alkaline with high Al_2O_3, Na_2O, K_2O and FeO/FeO+MgO and low CaO and MgO (Grantham 1992). Specimens of the central layered syenite have wide ranges of SiO_2 (53-65 %), Al_2O_3 (17-22 %), CaO (0.16-1.7 %), Na_2O (5-9 %) and K_2O (5-9 %), whereas those of the coarse-grained syenite do not vary greatly. The mesocratic syenites are distinguished by higher TiO_2, P_2O_5, MgO and CaO and lower Al_2O_3 than the others.

The differences in the chemistry of the three lithologies do not appear to show a continuum, which suggests that they are discrete intrusions with distinctive chemistry. The central layered syenite is interpreted to have formed from fractionation of K-feldspar and nepheline with intercumulus liquid from which amphiboles crystallized, believed to have originated by liquid immiscibility (Grantham 1992). The mesocratic syenite is an integral part of the central layered syenite, and the coarse-grained syenite is the oldest part.

Tvora syenites (5, 6)

The Tvora syenite intrusion is ca. 2-3 km in diameter. Two lithological facies have been recognized, an outer mesocratic syenite with a locally melanocratic part (5) and an inner leucocratic syenite (6). Xenoliths of mesocratic syenite (5) occur in the leucocratic type (6), which thus intruded the former. The two syenites have a similar modal composition. The feldspar of the mesocratic syenite is dark coloured.

The outer mesocratic syenite (5) is coarse grained, homogeneous, and contains varying proportions of perthite, amphibole, clinopyroxene, biotite and opaques, and accessory titanite, apatite, quartz and olivine (fayalitic). Quartz and fayalite occur in the melanocratic parts, but never in contact with one another. Plagioclase appears locally in the melanocratic part. Fine-grained mafic minerals occur interstitially among larger anhedral grains of feldspar. The aegirine-augite content is higher in the melanocratic rocks, whereas the main mafic phase of the mesocratic rocks is amphibole, which contains aegirine-augite inclusions. The minerals commonly show irregular grain boundaries, suggesting later recrystallization.

The inner leucocratic syenite (6) forms a plug-like intrusion. K-feldspar show schiller luminescence in a ca. 5 m wide zone along the contact to the outer mesocratic syenite. The average modal compositions are >65 % perthite, <10 % plagioclase (An_{20-40}), ca. 10 % clinopyroxene + hornblende + olivine and ca. 5 % magnetite. These rocks contain ca. 4 % normative nepheline. The mafic constituents have similar modal amounts to the mesocratic rocks.

These syenites differ in composition from the

Straumsvola syenites, having more SiO_2, FeO and CaO and less K_2O and Na_2O (Grantham 1992). If the Straumsvola and Tvora syenites had a common magmatic origin, an SiO_2 oversaturated phase would have had to be added to the former to form the latter.

Associated alkaline dykes
More than 2/3 of the younger mafic dykes on H.U. Sverdrupfjella were observed on Jutulrøra, Straumsvola and Tvora. The width of the dykes varies from 0.5 to 50 m, the majority being between 0.5 and 2 m. These dykes are subvertical and strike roughly N-S, the majority NNW-SSE. The dykes intruding the syenitic intrusions have trachytic textures with strongly aligned feldspar crystals. Some rocks have euhedral feldspar in an extremely fine-grained groundmass. One dyke is typically lamprophyric, containing biotite and clinopyroxene phenocrysts in a fine-grained groundmass. Feldspars in the rock are commonly perthite. Some rocks contain secondary catapleite, a variety of eudialyte.

Most dyke rocks have intermediate SiO_2 contents and high Al_2O_3, Na_2O and K_2O, while one has ca. 47 % and another 71 % SiO_2 (Grantham 1992). They are characterized by high Th, Zr, Nb and La, indicating late-stage residual liquids, and low Sc, V, Cu and Ni. The one with 47 % SiO_2 has high Sr and Ba. To generate intermediate to high SiO_2 contents by fractionation requires a high temperature to cross the thermal divide from undersaturated to granitic systems. High

SiO_2 contents may also result from assimilation of siliceous country rocks (Grantham 1992).

Mesozoic dolerite
Early reports of these dykes (Ravich & Solov'ev 1966, Hjelle 1974) mentioned only their presence and orientation. Grantham & Hunter (1991) reported the orientation of these dykes and discussed their relation with the fragmentation of Gondwana. Harris et al. (1990, 1991, 1995) presented chemical studies suggesting their consanguinity with the Jurassic basalt magmatism in Kirwanveggen and Heimefrontfjella to the west, and compared them with the Karoo dolerites in South Africa.

Dolerite dykes in the western region occur as xenoliths within and as dykes intruding the alkaline rocks at Straumsvola and Tvora (Fig. 16), suggesting that their emplacement was broadly synchronous with the latter, i.e. ca. 180-170 Ma (Grantham & Hunter 1991, Grantham 1992, Harris et al. 1995). They have NNW-SSE and NE-SW strikes with steep to vertical dips, and are less than 30 m wide. The wider dykes consist of medium- to coarse-grained, equigranular, locally porphyritic dolerite, whereas narrow dykes are fine-grained porphyritic and locally have amygdales filled by calcite and prehnite. The major minerals are phenocrystic olivine, ortho- and clinopyroxene, plagioclase and magnetite-ilmenite, and the subophitic groundmass consists of clinopyroxene, plagioclase and magnetite-il-
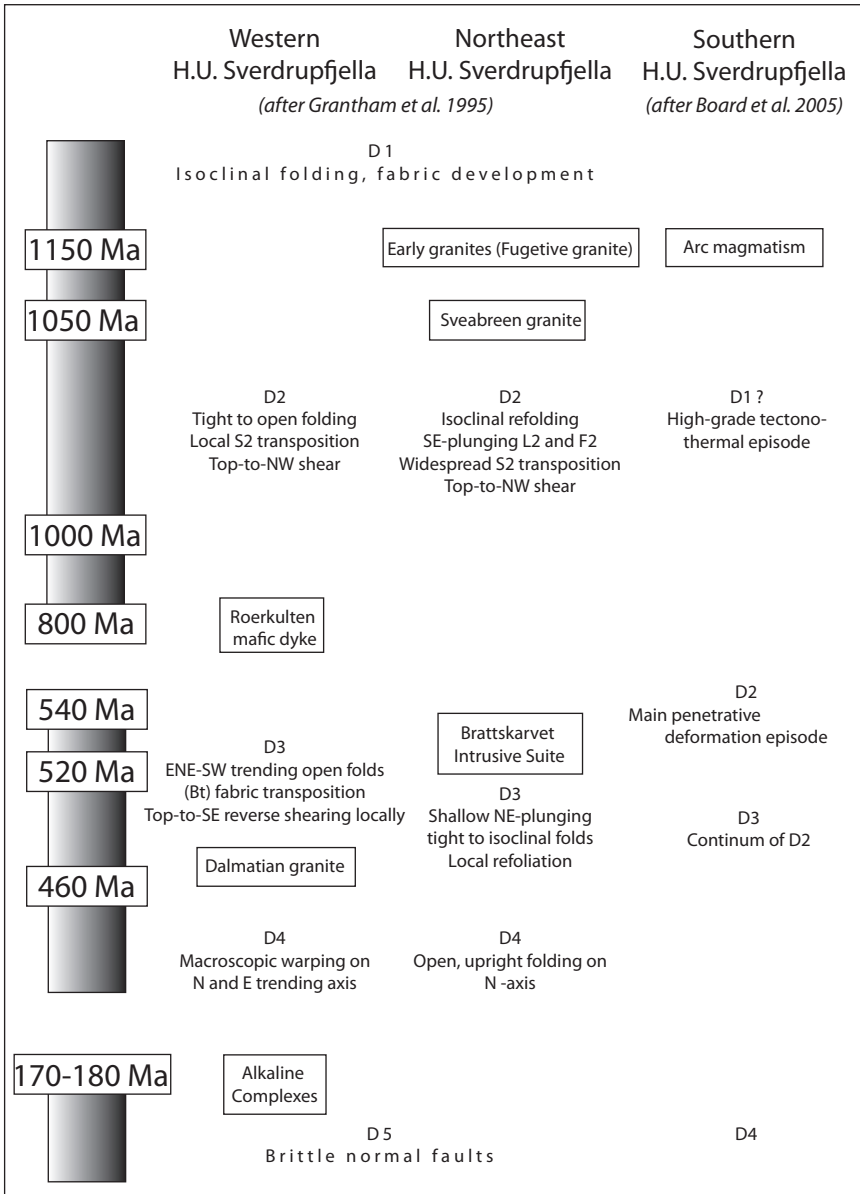
Western
H.U. Sverdrupfjella
(after Grantham et al. 1995)

Northeast
H.U. Sverdrupfjella

Southern
H.U. Sverdrupfjella
(after Board et al. 2005)

D 1
Isoclinal folding, fabric development

1150 Ma

Early granites (Fugetive granite)

Arc magmatism

1050 Ma

Sveabreen granite

D2
Tight to open folding
Local S2 transposition
Top-to-NW shear

D2
Isoclinal refolding
SE-plunging L2 and F2
Widespread S2 transposition
Top-to-NW shear

D1 ?
High-grade tectono-
thermal episode

1000 Ma

800 Ma

Roerkulten
mafic dyke

540 Ma

D2
Main penetrative
deformation episode

520 Ma

Brattskarvet
Intrusive Suite

D3
ENE-SW trending open folds
(Bt) fabric transposition
Top-to-SE reverse shearing locally

D3
Shallow NE-plunging
tight to isoclinal folds
Local refoliation

D3
Continum of D2

460 Ma

Dalmatian granite

D4
Macroscopic warping on
N and E trending axis

D4
Open, upright folding on
N -axis

170-180 Ma

Alkaline
Complexes

D 5
Brittle normal faults

D4

Fig. 17 Summary of the structural geology of H.U. Sverdrupfjella.

menite, locally with glass. Olivine, partially serpentinized and included in clinopyroxene, forms phenocrysts in all specimens, and clinopyroxene phenocrysts occur in approximately 2/3 of the specimens. Plagioclase phenocrysts, in about 1/3 of the specimens, locally show skeletal or radial intergrowth with clinopyroxene. The skeletal intergrowth and amygdales indicate rapid crystallization under low pressure <0.5 kb. Some rocks in Tvora are strongly altered and recrystallized. These tholeiitic rocks (Fig. 39a) are divided chemically into high-MgO (MgO >10 %), and low-MgO (MgO <10 %) varieties (Harris et al. 1990). The high-MgO rocks are characterized by olivine fractionation, whereas the low-MgO rocks crystallized from the residual magma after

36 % olivine fractionation from the high-MgO rock magma. Fractionation of clinopyroxene also occurred, but there was no plagioclase fractionation. The SiO_2 content ranges from 45 to 54 %. Later alteration significantly affected the concentrations of various trace elements.

The orientations of the dykes are consistent with the interpreted E-W extensional stress field during the separation of Antarctica from South Africa.

Approximately 30 dolerite dykes have been mapped in the eastern region (Groenewald 1995). They post-date all the deformation in this region, and vary from narrow aphanitic veins to vertically layered, coarse-grained dykes aligned in a northerly direction. The only aphanitic dyke occurs near the south-eastern extremity of Brattskarvet. It consists of irregular, finely devitrified glass which surrounds abundant, subhedral, bronzite phenocrysts (no other dolerite in the region contains orthopyroxene phenocrysts). A few rounded grains of olivine are present. Fine-grained dolerites mostly have clinopyroxene ± olivine set in either a very fine-grained palagonite-plagioclase groundmass, or in subophitic intergrowths with plagioclase. Coarse-grained, fractionated dykes at Doublet and Rømlingane show vertical layering with olivine and plagioclase enriched layers. The widest, a ca. 50 m wide intrusion at Salknappen, shows fractionation within the dyke that formed olivine gabbroic intersertal to orthocumulate textures in the vertical layers. Groenewald (1995) noted that clear vertical layering within this intrusion is defined by at least two periods of intrusion as well as internal differentiation.

Amygdales occur in the margins of many dykes. Those in the narrow dykes are spherical, while some in the margins of stratified dykes show elongation in a vertical direction. They are filled with natrolite, chalcedonic quartz and/or calcite. The dolerites range in SiO_2 from 44 to 53 %. MgO contents are generally high (9-15 %) (Groenewald 1995). Decreases of FeO* and MgO, and increases of Na_2O and K_2O, with increasing SiO_2 reflect olivine fractionation, whilst the scatter of CaO may indicate that augite was not a major liquidus phase.

The Salknappen intrusions and some narrow dykes are alkaline, while the remainder fall in the subalkaline field. The subalkaline rocks are tholeiitic on the AFM. All rocks plot in the within plate-volcanic arc field of the 2Nb-Zr/4-Y diagram (Groenewald 1995). A possibly deep source is suggested by relatively high Cr and Ni contents (Groenewald 1995). The Salknappen dolerites show different Cr/Ni ratios from the remainder, indicating at least two different primary magmas.

The dolerites of H.U. Sverdrupfjella are continental tholeiites, similar in chemistry and age (172 ±10 Ma) to those in Kirwanveggen, Heimefrontfjella and Vestfjella to the west (Faure et al. 1972, 1979). The picritic dolerites of H.U. Sverdrupfjella are significantly similar to the picritic basalts of the Karoo Supergroup in the northern Lebombo area of South Africa (Bristow 1984a, 1984b, Cox et al. 1979, 1984, Duncan et al. 1984, Groenewald 1995), which are considered to be derived from a mantle plume during the break-up of Gondwana (e.g. Sweeney et al. 1994).

Geological structures

General statement

The Sverdrupfjella supersuite and the various igneous rocks reveal a complex evolution with two tectonothermal episodes, the first in late Mesoproterozoic (1200-900 Ma) and the second in late Neoproterozoic/Cambrian times (Grantham & Hunter 1991, Groenewald et al. 1995, Grantham et al. 1995, Board et al. 2005).

Grantham et al. (1995) and Groenewald et al. (1995) proposed five phases of deformation (D_1-D_5) for the rocks of H.U. Sverdrupfjella (Fig. 17). The polyphase deformation involves two early phases (D_1 and D_2) of recumbent and isoclinal folding. The penetrative, regional fabric (S_2) is related to a top-to-NW shear sense. The D_2 deformation was followed by less intense upright folding (D_3) and normal and reverse faulting (D_4 to D_5). The main structures (D_1 and D_2) were assigned a Mesoproterozoic age by Grantham et al. (1995) on the assumption that the Sveabreen granitic gneiss (1028 ± 94 Ma; Moyes & Barton 1990) was emplaced post-D_1 and syntectonically with respect to D_2. Recently, the regional fabric has been reinterpreted as being Pan-African in age, based on U-Pb SHRIMP dating of mona-
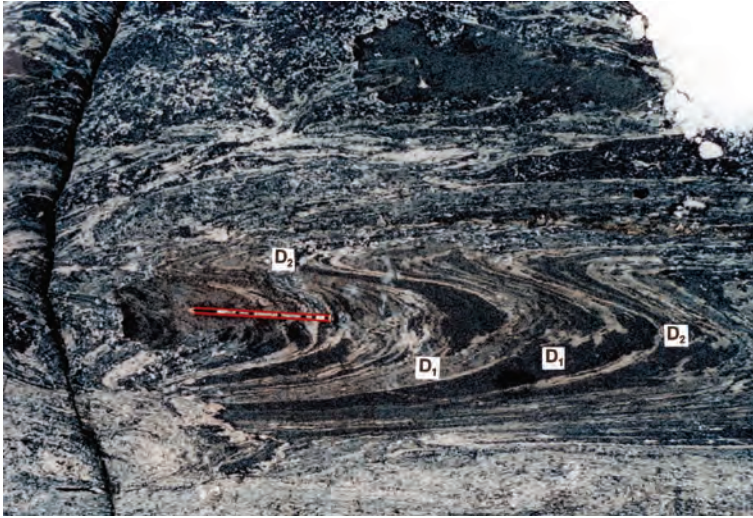
Fig. 18. Rootless, intrafolial, interference folds in the Rootshorga gneiss complex at northern Vendeholten. Pencil = 17 cm.

zite inclusions in fabric-forming mineral phases (Board et al. 2005). A strong Pan-African tectonothermal overprint has also been documented for Gjelsvikfjella (Bisnath et al. 2006).

Deformation episodes

Primary compositional layering (S_0) is present in the Jutulrøra and Fuglefjellet gneiss complexes as alternating calcareous and quartzo-feldspathic layers. Isoclinal, rootless, recumbent folding (D_1) of these layers resulted in axial planes verging toward W-NW. In the eastern region, marble and quartzite layers are interpreted as primary layering (S_0). The D_1 deformation resulted in transposition of primary layering, which is seen as numerous rootless intrafolial folds in the gneisses

(Fig. 18) and the Sveabreen megacrystic granites. The most intense deformation is D_2, which is characterized by the development of a regionally pervasive fabric (S_2), a SE-plunging mineral lineation and SE-dipping ductile shear zones with a top-to-NW shear sense. Open to tight folds associated with thrust faulting are observed on Fuglefjellet and Kvitkjølen. An angular discordance between the D_1 and D_2 structures has been observed on the northern part of Fuglefjellet. The younger amphibolites at Roerkulten cut the S1 fabric and display a D_2 foliation defined by aligned biotite.

In the eastern region, the D_2 deformation encompassed several phases of progressive refolding (Fig. 18). The principal strain direction is parallel
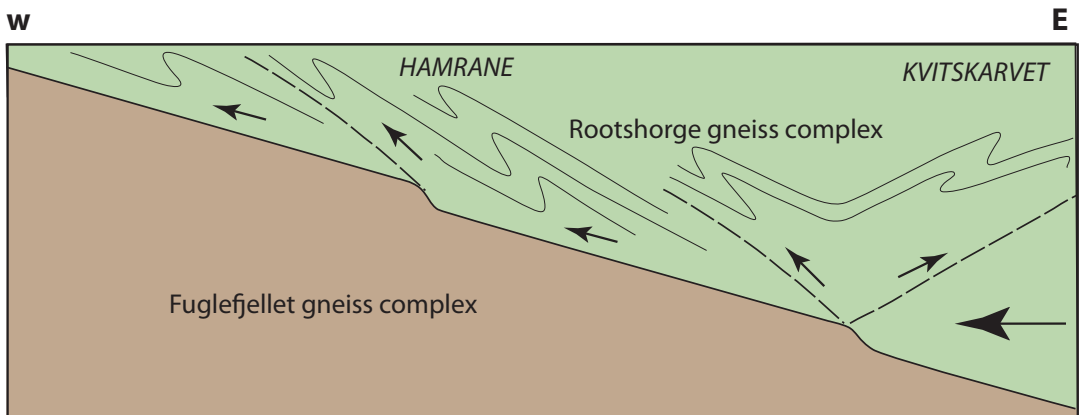

Fig. 19. A cartoon of structural reversal between Hamrane and Kvitskarvet (ca. 10 km apart), showing a pop-up structure in the south-eastern part of H.U. Sverdrupfjella (Groenewald 1995).

Fig. 20. Frontal ramps of the thrust fault at the top of Kvitkjolen.

to the regionally pervasive lineation, L_2, which plunges to the south-east in the western region. Shear-sense indicators suggest a north-westward transport, except in the Kvitskarvet area in the south-easternmost part of H.U. Sverdrupfjella, where movement was to the south-east (Fig. 19). The thrust and nappe structures are the dominant D_1 and D_2 structures in the eastern region and they indicate major north-west verging compression or crustal shortening.

The thrusts range from semi-conformable zones of ductile shear to multiple duplex structures (Fig. 20). Multi-duplex ramps occur on south-western Salknappen and south-eastern Isingen. Major thrust zones are also present near Robinheia and Hamrane. The movement along these thrusts is top-to-W or NW along east-south-eastward dipping thrust planes. The obliquity of lineations on some thrust planes suggests transpressive shear with a sinistral component. At Kvitskarvet, local reversal in the sense of shear suggests a pop-up structure involving back-thrusting to the east as a subsidiary of predominant north-westward thrust and nappe systems (Fig. 19). Early high-pressure parageneses are preserved better in this area than elsewhere. This south-eastern subarea is separated from the north-west by the Sveabreen megacrystic granites, where extensive mylonites are present.

Mylonites, up to 100 m thick, occur along the upper and lower contacts of the western Sveabreen megacrystic granite at Tverrnipa and Vendeholten. Intensely sheared textures and the pres-

ence of sillimanite in paraluminous parts of the granite suggest that the granite was emplaced synchronous to major decompression. The absence of mylonites along parts of the contact indicates that this emplacement was not solely tectonic. The mylonites may mark one of the major decollements which brought the high-grade Rootshorga gneiss complex above the moderately metamorphosed rocks in the middle and western regions. The presence of high-pressure rocks in the eastern region supports this interpretation (Groenewald 1995).

D_3 structures in the western region are present as steep faults and open folds. The latter display steeply NW-dipping axial planes and horizontal to gentle SE-plunging fold axes (Fig. 21). An axial planar foliation (S_3) is defined by aligned biotite. Steep axial planes were observed at Roerkulten, Brekkerista and Tvora with NW dips and E-NE vergence. In the eastern region, D_3 is characterized by locally tight, upright, mesoscale folds that generally plunge north-east at 10-25°. The later stage of D_3 is characterized by brittle faulting and brecciation. At Isingen and Vendeholten, a 30 m wide, brecciated fault, with a poorly estimated displacement of around 100-200 m, is a D_3 structure.

D_4 structures are open, regional structures without mesoscopic expressions in the western regions. Layered structures dipping south-south-westwards at Jutulrøra and north-north-eastwards at Straumsvola and Roerkulten may represent large-scale, open warping or dome-basin struc-
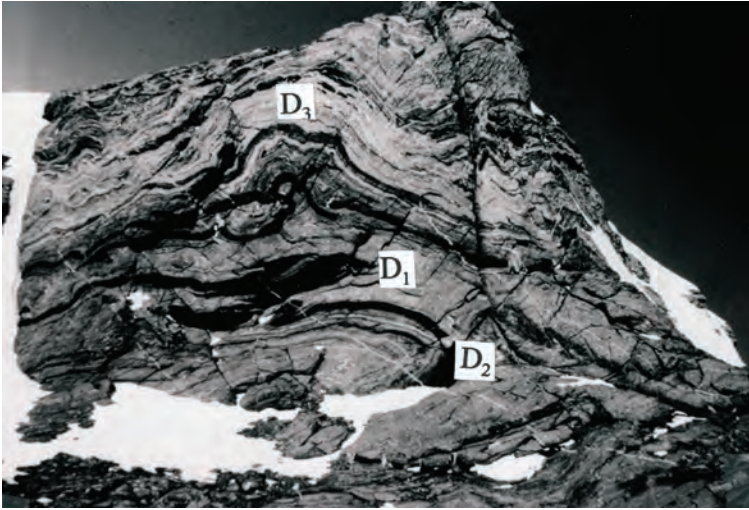
29

Fig. 21. Three phases of folding, D_1 and D_2 recumbent isoclinal folds and a D_3 open upright fold, at Jutulrøra.

tures. Mesoscale, open cylindrical folds occur at a few localities in the eastern region. These affect post-D_3 granite sills at Isingen and pre-date the Mesozoic dolerite dykes. At Vendeholten, these folds plunge 1-10° east, whereas in the Salknappen area they plunge 5-15 south.

D_5 structures are present as normal faults and joints striking N-S and NE-SW. This deformation is related to the Mesozoic break-up of Gondwana. Steep NE-striking faults are common at Jutulrøra. Chloritization of mafic minerals, saussuritisation of plagioclase and sericitisation of K-feldspar occur along the fracture surfaces. The faults may be related to the formation of major faulting along the Jutulstraumen-Pencksökket Rift Zone. NE-striking joints with steep, vertical dips are consistent with the N-S separation movement during the Gondwana fragmentation (Grantham & Hunter 1991). Spaced cleavages cutting the Tvora alkaline rocks show the same trend. Small shear zones observed in the granitoid rocks at Roerkulten, Jutulrøra and Brekkerista show both sinistral and dextral movement senses. The youngest deformation in the eastern region is block faulting. A large N-S striking, vertical fault was identified on the northern face of Brattskarvet. Vertical, NNE-striking extensional joints filled by stilbite veins are present at Brattskarvet and on the western side of Sveabreen.

Metamorphism

Metamorphic stages

Petrological investigations have indicated a complex multi-stage (M_1-M_3) metamorphic evolution of the metamorphic rocks at H.U. Sverdrupfjella (Groenewald & Hunter 1991, Grantham et al. 1995, Board et al. 2005). The majority of the gneisses in the eastern area comprise upper amphibolite and/or granulite facies mineral assemblages, whereas the western area is characterized by amphibolite facies. The difference in metamorphic grade between the eastern and western parts of H.U. Sverdrupfjella is interpreted to be the result of thrusting, with the higher-grade eastern area being thrust over the lower-grade western area during D_2 (Grantham et al. 1995).

Mineral assemblages in felsic and intermediate gneisses, carbonates and calc-silicate rocks provide only limited constraints on the metamorphic evolution. Metapelites and mafic rocks are, however, commonly more useful as monitors of metamorphic histories, and this is also the case at H.U. Sverdrupfjella.

The P-T evolution of the Sverdrupfjella supersuite is summarized in Figure 22. Relics of early metamorphic phases are found in the cores of strain-protected mafic pods and boudins in the eastern area. Groenewald & Hunter (1991) reported granulite facies Grt-Cpx-Opx-Pl-Qtz and Grt-Cpx-Qtz-Ilm assemblages preserved in mafic lenses (at Tua, Rømlingane, Vendeholten, Salknappen, Isingen and Kvitskarvet). Estimated P-T conditions for these M_1 assemblages are 9-11

kbar at 850°C (Groenewald & Hunter 1991). The second stage is a series of metamorphic events that is subdivided into M_{2a}, M_{2b} and M_{2c} and which defines a coherent clockwise P-T-T path. M_{2a} is represented by an early eclogite facies garnet-omphacite assemblage (Board et al. 2005), which is found in resistant tectonic mafic lenses and boudins. Geobarometry based on reintegrated omphacite composition has yielded a minimum pressure of 12.9-14.3 kbar for the eclogite facies metamorphism. Garnet-pyroxene assemblages have not been observed in the mafic rocks of the western area.

The high-pressure M_{2a} metamorphism was followed by isothermal decompression to amphibolite facies conditions (M_{2b}). The dominant mineral assemblages in H.U. Sverdrupfjella were formed during this event. In the mafic rocks, decompression from high pressures resulted in the formation of replacement textures such as kelyphite on paragasitic amphibole and plagioclase after garnet and omphacite, and clinopyroxene-plagioclase intergrowths after omphacite. Further equilibration during M_{2b} formed Hbl-Pl and Grt-Hbl-Pl bearing assemblages in mafic rocks. Geothermobarometry on mafic M_{2b} assemblages has yielded 680-760 and 9.4-11.3 kbar for the amphibolite facies metamorphism (Board et al. 2005). Typical M_2 mineral assemblages in metapelitic gneisses are Grt-Bt-Ky(Sil)-Pl-Kfs-Qtz-Gr, where the elongated laths of kyanite, biotite and graphite define the S_2 fabric (Board et al. 2005). Calculated P-T estimates for the metapelitic gneisses, based on garnet-biotite thermometry and garnet-kyanite-plagioclase-quartz barometry, are slightly lower than those estimated for the mafic rocks. According to Grantham et al. (1995), sillimanite is the stable aluminosilicate north of 72° 25' S, whereas kyanite is partly replaced by sillimanite to the south. The metapelitic gneisses show abundant evidence of partial melting.

M_{2c} is recognized largely by a planar fabric commonly defined by biotite aligned parallel to the D_3 axial plane. M_{2c} assemblages also include amphibolitized mafic intrusions which locally show no planar fabric. The youngest event, M_3, is characterized by low-grade retrograde metamorphism and involves hydrothermal alteration which resulted in chloritization of biotite and hornblende and saussuritisation of plagioclase (Grantham et al. 1995).
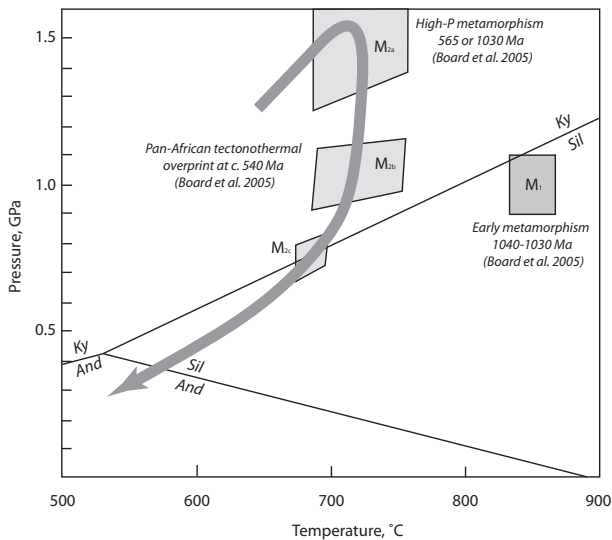

Fig. 22. P-T diagram showing the P-T evolution of the metasupracrustal rocks of H.U. Sverdrupfjella (modified after Board et al. 2005).

Tectonic interpretation

Geochronological data from H.U. Sverdrupfjella (Board et al. 2005), Gjelsvikfjella (Jacobs et al. 2003, Bisnath et al. 2006) and Heimefrontfjella (Arndt et al. 1991, Jacobs et al. 2003) have recorded two episodes of high-grade metamorphism in the Maud Belt, one towards the end of the Mesoproterozoic and again in late Neoproterozoic/Cambrian times. The late Mesoproterozoic metamorphism corresponds to the early-M_1 event. Relicts from the granulite facies M_1 metamorphism are, however, few and rare due to the intensity of the later Pan-African reworking. Similar relicts are also known from the eastern part of Gjelsvikfjella (Bisnath & Frimmel 2005).

The second metamorphic stage (M_2) involves near isothermal decompression from eclogite facies conditions (M_{2a}) to upper amphibolite facies (M_{2b}; 5-8 kbar, 600-790°C; Board et al. 2005). The decompression path is interpreted as the result of tectonic exhumation through thrusting. The M_2 metamorphism is related to Pan-African continent-continent collision between components of Gondwana. The main pulse of deformation, syntectonic magmatism and metamorphism, took place between 540 and 530 Ma (Board et al. 2005, Bisnath & Frimmel 2005). M3 was coeval with the emplacement of the alkaline syenites and dolerites, and is related to the Jurassic break-up of Gondwana.

Mineralization

No special attention has been paid to the mineralization in this area; however, the following observations were recorded.

Oxide and hydro-oxide concentrations

The heavy mineral fraction from the Tvora melanocratic syenite contains 10-20 vol. % of magnetite, while those of other rocks are from 0.1 to 2.0, average 0.6 vol. %. Scheelite was found in the detritus around Fuglefjellet. Secondary Cu minerals were found in the banded gneisses of the Jutulrøra gneiss complex on the south-eastern part of Jutulrøra, the northern slope of Roerkulten, Knattebrauta, between Gordonnuten and Skarsnuten, and the southern part of Fuglefjellet.

Radioactivity

Geiger counter measurements show 2-5 times higher radioactivity in the Brattskarvet granitoids than the other rocks.

Pegmatite minerals

Granitic pegmatite associated with the post-D_2 and pre-D_3 granitoid rocks contains K-feldspar (> 50 cm), very coarse graphic intergrowth of quartz and K-feldspar, biotite (<30 cm), green and black tourmaline (<9 cm long), magnetite (<1.5 cm diameter) and beryl (<3 cm long). A pink pegmatite in the eastern region contains very coarse-grained K-feldspar, tourmaline, magnetite, beryl, biotite and quartz. Fluorite is present as an accessory and vein mineral in the monzogranites of the Brattskarvet granitoids. Some pegmatites contain fluorite, celestine and magnetite crystals. The composite lamprophyre dykes at Vendeholten and Römlingane contain 1.5 cm large diopside, arfvedsonite and katophorite. Syenitic pegmatite which intrudes the Straumsvola alkaline syenite on the north-eastern part of Straumsvola contains large crystals of natrolite, acmite, leucite and some adular and zircon.

References

Aagaard, B., 1934: Fangst og forskning i Sydishavet. Vol. I-III, Gyldendal Norsk Forlag, Oslo.

Allen, A.R., 1991: The tectonic and metamorphic evolution of H.U. Sverdrupfjella, western Dronning Maud Land, Antarctica. In M.R.A. Thomson, J.A. Crame, & J.W. Thomson (eds): Geological evolution of Antarctica. Cambridge University Press, Cambridge, pp. 53-60.

Anschütz, H., Müller, K., Isaksson, E., McConnell, J. R., Fischer, H., Miller, H., Albert, M. & Winther, J-G., 2009: Revisiting sites of the South Pole Queen Maud Land Traverses in East Antarctica: Accumulation data from shallow firn cores. Journal of Geophysical Research, Vol. 114, No. D24, D24106. http://dx.doi.org/10.1029/2009JD012204

Arndt, N.T., Todt, W., Chauvel, C., Tapfer, M. & Weber, K., 1991: U-Pb zircon age and Nd isotopic composition of granitoids, charnockites and supracrustal rocks from Heimefrontfjella, Antarctica. Geologische Rundschau, 80/3, 759-777.

Barstad, B., Eiken, T., Finnes, R. & Øvstedal, O., 1997: Geodetic measurements in Dronning Maud Land. In J.-G. Winther (ed.): Report of the Norwegian Antarctic Research Expedition 1996/97. Norsk Polarinstitutt Meddelelser, 148, 67-75

Bisnath, A. & Frimmel, H.E., 2005: Metamorphic evolution of the Maud Belt: P-T-t path for high-grade gneisses in Gjelsvikfjella, Dronning Maud Land, East Antarctica. Journal of African Earth Sciences, 43, 505-54.

Bisnath, A., Frimmel, H.E., Armstrong, R.A. & Board, W.S., 2006: Tectono-thermal evolution of the Maud Belt: New SHRIMP U-Pb zircon data form Gjelsvikfjella, Dronning Maud Land, East Antarctica. Precambrian Research, 150, 95-121.

Board, W.S., Frimmel, H.E. & Armstrong, R.A., 2005: Pan-African tectonism in the western Maud Belt: P-T-t path for high-grade gneisses in the H.U. Sverdrupfjella, East Antarctica. Journal of Petrology, 46, 671-699.

Bredell, J.H., 1973: Ten-metre Snow Studies at Grunehogna Base, western Dronning Maud Land. South African Journal of Antarctic Research (3), 16-21.

Bristow, J.W., 1984a: Picritic rocks of the north Lebombo and southeast Zimbabwe. In A.J. Erlank (ed.): Petrogenesis of the volcanic rocks of the Karoo Province. Special Publication of the Geological Society of South Africa, 13, 105-123.

Bristow, J.W., 1984b: Nephelinites of the north Lebombo and south-east Zimbabwe. In A.J. Erlank (ed.): Petrogenesis of the volcanic rocks of the Karoo Province. Special Publication of the Geological Society of South Africa, 13, 87-104.

Cox, K.G., Macdonald, R. & Hornung, G., 1979: Geochemical and petrogenetic provinces in the Karroo Basalts of southern Africa. American Mineralogist, 52, 1451-1474.

Cox, K.G., Duncan, A.R., Bristow, J.W., Taylor, S.R. & Erlank, A.J., 1984: Petrogenesis of the basic rocks of the Lebombo. In A.J. Erlank (ed.): Petrogenesis of the volcanic rocks of the Karoo Province. Special Publication of the Geological Society of South Africa, 13, 149-170.

Decleir, H. & Van Autenboer, T., 1982: Gravity and magnetic anomalies across Jutulstraumen, a major geologic feature in Western Dronning Maud Land. In C. Craddock (ed.): Antarctic Geoscience, 941-948. University of Wisconsin Press.

Divine, D. V., Isaksson, E., Kaczmarska, M., Godtliebsen, F., Oerter, H., Schlosser, E., Johnsen, S. J., van den Broeke, M.R. & van de Wal, R. S. W, 2009: Tropical Pacific-high latitude south Atlantic teleconnections as seen in δ18O variability in Antarctic coastal ice cores. Journal of Geophysical Research, Vol. 114, No. D11, D11112 http://dx.doi.org/10.1029/2008JD010475

Divine, D.V., Koç, N., Isaksson, E., Nielsen, S., Crosta, X. & Godtliebsen, F., 2010: Holocene Antarctic climate variability from ice and marine sediment cores: Insights on ocean–atmosphere interaction. Quaternary Science Reviews Volume 29, 303-312.

Duncan, A.R., Erlank, A.J. & Marsh, J.S., 1984: Regional geochemistry of the Karoo igneous province. In A.J. Erlank (ed.): Petrogenesis of the volcanic rocks of the Karoo Province. Special Publication of the Geological Society of South Africa, 13, 355-388.

Eiken, T. & Svendsen, K., 1985: Geodetic, topographic measurements in Dronning Maud Land and on Bjørnøya. In O. Orheim (ed.): Report of the Norwegian Antarctic Research Expedition (NARE) 1984/85, Norsk Polarinstitutt Rapport serie, 22, 19-26.

Eiken, T., Luktvasslimo, B.Å. & Lytskjold, B., 1990: Geodetic measurements in Dronning Maud Land. In O. Orheim (ed.): Report of the Norwegian Antarctic Research Expedition 1989/90. Norsk Polarinstitutt Meddelelser, 113, 101-111.

EPICA community members, 2006: One-to-one coupling of glacial climate variability in Greenland and Antarctica. Nature, 444, 195-198.

Eisen, O., Rack, W., Nixdorf, U. & Wilhelms, F., 2005: Characteristics of accumulation rate in the vicinity of the EPICA deep-drilling site in Dronning Maud Land, Antarctica. Annals of Glaciology, 41, 41-46.

Faure, G., Hill, R., Jones, L.M. & Elliot, D.H., 1972: Isotope composition of strontium and silica content of Mesozoic basalt and dolerite from Antarctica. In R.J. Adie (ed.): Antarctic Geology and Geophysics. 617-624, Universitetsforlaget, Oslo.

Faure, G., Bowman, J.R. & Elliot, D.H., 1979: The initial 87Sr/86Sr ratios of the Kirwan volcanics of Dronning Maud Land: comparison with Kirkpatrick Basalt, Trans-Atlantic Mountains. Chemical Geology, 26, 77-90.

Giaever, J., 1954: The White Desert. The Official Account of the Norwegian-British-Swedish Antarctic Expedition. 304 pp. Chatto and Windus, London.

Gjessing, Y.T., 1972: Mass transport of Jutulstraumen ice stream in Dronning Maud Land. Norsk Polarinstitutt Årbok 1970, 227-232.

Grantham, G.H., 1992: Geological evolution of western H.U. Sverdrupfjella, Dronning Maud Land, Antarctica. PhD thesis (unpublished), University of Natal, Pietermaritzburg, South Africa, 322 pp.

Grantham, G.H. & Hunter, D.R., 1991: The timing and nature of faulting and jointing adjacent to the Pencksökket, western Dronning Maud Land, Antarctica. In M.R.A. Thomson, J.A. Crame & J.W. Thomson (eds.): Geological Evolution of Antarctica. 47-51. Cambridge University Press, Cambridge.

Grantham, G.H., Groenewald, P.B. & Hunter, D.R., 1988: Geology of the northern H.U. Sverdrupfjella, western Dronning Maud Land, and implications for Gondwana reconstructions. South African Journal of Antarctic Research, 18, 2-10.

Grantham, G.H., Moyes, A.B. & Hunter, D.R., 1991: The age, petrogenesis and emplacement of the Dalmatian Granite, H.U. Sverdrupfjella, Dronning Maud Land, Antarctica. Antarctic Science, 3, 197-204.

Grantham, G.H., Jackson, C., Moyes, A.B., Groenewald, P.B., Harris, P.D., Ferrar, G. & Krynauw, J.R., 1995: The tectonothermal evolution of the Kirwanveggen - H.U. Sverdrupfjella areas, Dronning Maud Land, Antarctica. Precambrian Research, 75, 209-229.

Groenewald, P.B., 1991: Metamorphic evolution of H.U. Sverdrupfjella in the context of Maudheim orogenic province, western Dronning Maud Land, Antarctica. Sixth International Symposium on Antarctic Earth Sciences 194. NIPR - Tokyo.

Groenewald, P.B., 1993: Eclogites in Dronning Maud Land, Antarctica: evidence of mid-crustal detachment during metamorphic inversion. Terra Nova - Abstract supplement, 5, p. 13.

Groenewald, P.B., 1995: The geology of northern H.U. Sverdrupfjella and its bearing on crustal evolution in Dronning Maud Land, Antarctica. PhD thesis, University of Natal, Pietermaritzburg, South Africa. 315 pp.

Groenewald, P.B. & Grantham, G.H., 1991: Geology of the H.U. Sverdrupfjella, Dronning Maud Land, Antarctica. Sixth International Symposium on Antarctic Earth Sciences. 195. NIPR - Tokyo.

Groenewald, P.B. & Hunter, D.R., 1991: Granulites of northern Sverdrupfjella, western Dronning Maud Land: metamorphic history from garnet pyroxene assemblages, coronas and rehydration reactions. In M.R.A. Thomson, J.A. Crame & J.W. Thomson (eds.): Geological Evolution of Antarctica. 61-66. Cambridge University Press, Cambridge.

Groenewald, P.B., Grantham, G.H. & Watkeys, M.K., 1991: Geological evidence for a Proterozoic to Mesozoic link between Dronning Maud Land, Antarctica, and south-eastern Africa. Journal Geological Society of London, 148, 1115 1123.

Groenewald, P.B., Moyes, A.B., Grantham, G.H. & Krynauw, J.R., 1995: East Antarctic crustal evolution: geological constraints and modelling in western Dronning Maud Land. Precambrian Research, 75, 231-250.

Hagen, J.O., 1997: Blue ice field studies in the Jutulsessen and Troll areas. In O. Orheim (ed.): Report on the Norwegian Antarctic Research Expedition NARE 1992/93. Norsk Polarinstitutt Meddelelser, 125

Hansen, H.E., 1949: Notes on the sighting of land in Antarctica between 17 west and 87 east. Manuscript at Norsk Polarinstitutt, 20 pp.

Harris, C., Marsh, J.S., Duncan, A.R. & Erlank, A.J., 1990: The petrogenesis of the Kirwan basalts of Dronning Maud Land, Antarctica. Journal of Petrology, 31, 341 370.

Harris, C., Watters, B.R. & Groenewald, P.B., 1991: Geochemistry of the Mesozoic regional basic dykes of western Dronning Maud Land, Antarctica. Contributions to Mineralogy and Petrology, 107, 100-111.

Harris, C. & Grantham, G.H., 1993: Geology and petrogenesis of the Straumsvola nepheline syenite complex, Dronning Maud Land, Antarctica. Geological Magazine, 130, 513-532.

Harris, P.D., Moyes, A.B., Fanning, C.M. & Armstrong, R.A., 1995: Zircon microprobe results from the Maudheim high-grade gneiss terrane, Western Dronning Maud Land, Antarctica. In J.M. Barton & T.E. Copperthwaite (eds.): Centennial Geocongress (1995) Extended Abstracts, Geological Society of South Africa, Johannesburg, Volume I, 240-243.

Hjelle, A., 1974: Some observations on the geology of the H.U. Sverdrupfjella, Dronning Maud Land, Antarctica. Norsk Polarinstitutt Årbok, 1972, 7 22.

Hofstede, C.M., van de Wal, R.S.W., Kaspers, K., van den Broeke, M.R., Karlöf, L., Winther, J-G., Isaksson, E., Lappegard, G., Mulvaney, R., Oerter, H. & Wilhelms, F., 2004: Firn accumulation records for the past 1000 years on the basis of dielectric profiling of six cores from Dronning Maud Land, Antarctica. Journal of Glaciology 50 (169), 279-291.

Høydal, Ø. A., 1996: A force-balance study of ice flow and basal conditions of Jutulstraumen, Antarctica. Journal of Glaciology, 42(142), 413-425.

Isaksson, E., van den Broeke, M.R., Winther, J-G., Karlöf, L. Pinglot, J.F. & Gundestrup, N., 1999: Accumulation and proxy-temperature variability in Dronning Maud Land, Antarctica, determined from shallow firn cores. Annals of Glaciology, 29, 17-22.

Isaksson, E. & Melvold, K., 2002: Trends and patterns in the recent accumulation and proxy-temperature in coastal Dronnning Maud Land, Antarctica - interpretations from shallow ice cores. Annals of Glaciology 35, 175-180.

Jacobs, J., Bauer, W. & Fanning, C.M., 2003: New age constraints for Grenville-age metamorphism in western central Dronning Maud Land (East Antarctica), and implications for the palaeogeography of Kalahari in Rodinia. International Journal of Earth Sciences (Geologische Rundschau), 92, 301-315.

Jonsson, S., 1990: Local climate and mass balance of a blue-ice area in Western Dronning Maud Land, Antarctica. Zeitschrift für Gletscherkunde und Glazialgeologie, 26(1), 11-29.

Kretz, R., 1983: Symbols for rock-forming minerals. American Mineralogist, 68, 277-279

Kaczmarska, M., Isaksson, E., Karlöf, K., Winther, J-G., Kohler, J., Godtliebsen, F., Ringstad Olsen, L., Hofstede, C M., van den Broeke, M.R., Van De Wal, R. S.W. & Gundestrup, N., 2004: Accumulation variability derived from an ice core from coastal Dronning Maud Land, Antarctica. Annals of Glaciology, 39, 339-345.

Kaczmarska, M., Isaksson, E., Karlöf, L., Brandt, O., Winther, J-G., van de Wal, R.S.W., van den Broeke, M.R. & Johnsen, S.J., 2006: Antarctic coastal climate interpreted from light transmittance in and ice core record. Antarctic Science 18 (2), 271-278. DOI:10.1017/S09544102006000319.

Karlöf, L., Isaksson, E., Winther, J-G., Gundestrup, N., Meijer, H.A.J., Mulvaney, R., Pourchet, M., Hofstede, C., Lappegard, G., Pettersson, R., van den Broeke, M.R. & van de Wal, R.S.W., 2005: Accumulation variability over a small area in east Dronning Maud Land, Antarctica, as determined from shallow firn cores and snow pits: some implications for ice-core records. Journal of Glaciology 51 (174), 343-352.

Liston, G.E., Winther, J-G., Bruland, O., Elvehøy, H. & Sand, K., 1999: Below-surface ice melt on the coastal Antarctic ice sheet. Journal of Glaciology, 45(150), 273-285.

Lunde, T., 1961: On the snow accumulation in Dronning Maud Land. Norsk Polarinstitutt Skrifter, 123, 48.

Lunde, T., 1965: On the firn temperatures and glacier flow in Dronning Maud Land. Norsk Polarinstitutt Årbok 1963, 7-24.

Melvold, K., 1999: Impact of recent climate on glacier mass balance studies on Kongsvegen, Svalbard and Jutulstraumen, Antarctica. Rapportserie i Naturgeografi. No. 13, PhD thesis, Department of Physical Geography, University of Oslo, Oslo.

Melvold, K., Hagen, J.O., Pinglot, J.F. & Gundestrup, N., 1998: Large spatial variation in accumulation rate in Jutulstraumen ice stream, Dronning Maud Land, Antarctica. Annals of Glaciology, 27, 231-238.

Melvold, K. & Rolstad, E.C., 2000: Subglacial topography of Jutulstraumen outlet glacier, East Antarctica, mapped from ground-penetrating radar, optical and interferometric synthetic aperture radar satellite data. Norwegian Journal of Geography, 54(4), 169-181.

Moyes, A.B. & Barton, J.M. Jr., 1990: A review of isotopic data from western Dronning Maud Land, Antarctica. Zeitblatt Geologie und Palaontologie, 1, 19 31.

Moyes, A.B., Groenewald, P.B. & Brown, R.W., 1993: Isotopic constraints on the age and origin of the Brattskarvet intrusive suite, Dronning Maud Land, Antarctica. Chemical Geology (Isotope Geoscience Section), 106, 453-466.

Müller, K., Sinisalo, A., Anschütz, H., Hamran, S-E., Hagen, J-O., McConnell, J. & Pasteris, D., 2010: An 860km surface mass-balance profile on the East Antarctic plateau derived by GPR, Annals of Glaciology, 51(55)

Neethling, D.C., 1969: Geology of the Ahlmann Ridge, western Queen Maud Land. In V.C. Bushnell & C. Craddock (eds.): Geological Maps of Antarctica. Antarctic Map Folio Series 12; PL. VII.

Näslund, J-O., 1998: Ice sheet, climate, and landscape interactions in Dronning Maud Land, Antarctica. Dissertation Series. No. 11, Department of Physical Geography, Stockholm University, Stockholm.

Oerter, H., Graf, W., Wilhelms, F., Minikin, A. & Miller, H., 1999: Accumulation studies on Amundsenisen, Dronning Maud Land, Antarctica, by means of tritium, dielectric profiling and stable-isotope measurements: First results from the 1995-96 and 1996-97 field seasons. Annals of Glaciology, 29, 1-9.

Ohta, Y., Tørudbakken, B. & Shiraishi, K., 1990: Geology of Gjelsvikfjella and Western Mühlig-Hofmannfjella, western Dronning Maud Land, East Antarctica. Polar Research, 8, 99-126.

Ohta, Y. (ed.) 1999: Nature environment map 1:100,000, Gjelsvikfjella and Western Mühlig–Hofmannfjella, Dronning Maud Land, Antarctica. Sheets 1 and 2. With explanatory text, 37 pp. Norsk Polarinstitutt Temakart 24.

Orheim, O., Gjessing, Y., Lunde, T., Repp, K., Wold, B., Clausen, B.H. & Liestøl, O., 1986: Oxygen isotopes and accumulation rates at Riiser-Larsenisen, Antarctica. Norsk Polarinstitutt Skrifter, 187, 33-47.

Orheim, O. & Lucchitta, B.K., 1987: Snow and ice studies by thematic mapper and multispectral scanner Landsat images. Annals of Glaciology, 9, 109-118.

Picciotto, E., Grozaz, G. & de Breuck, W., 1971: Accumulation on the South Pole Queen Maud Land Traverse, 1964-1968. Antartic Research Series, 16, 257-315.

Ravich, M.G. & Solov'ev, D.S., 1966: Geology and petrology of the mountains of central Queen Maud Land (eastern Antarctica). Transactions of the Scientific Research Institute of Arctic Geology, Ministry of Geology of the USSR, volume 141. (Translation (1969) Israel program for scientific translations, Jerusalem), 348 pp.

Richardson, C. & Holmlund, P., 1999: Spatial variability at shallow snow-layer depths in central Dronning Maud Land, East Antarctica. Annals of Glaciology, 29, 10-16.

Roer, N., 1954: Landmålerliv i Dronning Maud Land. Norsk Polarinstitutt Meddelelser, 78, 246-265.

Rolstad, C.E., Whillans, I.M., Hagen, J.O. & Isaksson, E., 2000: Large-scale force budget of an outlet-glacier: Jutulstraumen, Dronning Maud Land, East Antarctica. Annals of Glaciology, 30, 35-42.

Roots, E.F., 1953: Preliminary note on the geology of western Dronning Maud Land. Norsk Geologisk Tidsskrift, 32, 17 33.

Roots, E.F., 1969: Geology of western Queen Maud Land. Explanation of Plate VI, Folio 12, Antarctic map folio series, American Geographical Society, New York.

Rotschky, G., Holmlund, P., Isaksson, E., Mulvaney, R., Oerter, H., Van den Broeke, M.R. & Winther, J-G, 2007: A new surface accumulation map for western Dronning Maud Land, Antarctica, from interpolation of point measurements. Journal of Glaciology 53 (182), 385–398

Rotschky, G., Eisen, O., Nixdorf, U. & Oerter, H., 2004: Spatial distribution of surface mass balance on Amundsenisen plateau, Antarctica, derived from ice-penetrating radar studies. Annals of Glaciology, 39, 265-270.

Strandtmann, R.W. & Sømme, L., 1977: Prostigmate (Acari, Prostigmata) from Sverdrupfjella, Dronning Maud Land, with description of four new species. Norwegian Journal of Entomology, 24, 137-146.

Sweeney, R.J., Duncan, A.R. & Erlank, A.J., 1994: Geochemistry and petrogenesis of Central Lebombo Basalts of the Karoo Igneous province. Journal of Petrology, 35, 95-126.

Swithinbank, C., 1957: The morphology of the ice shelves of western Dronning Maud Land, Glaciology I (A). In C. Swithinbank (ed.) Norwegian-British-Swedish Antarctic Expedition, 1949-52. Scientific Results III, 3-36 + plates. Oslo, Norsk Polarinstitutt.

Swithinbank, C., 1959: The morphology of the inland ice sheet and nunatak areas of western Dronning Maud Land, Glaciology I (D). In C. Swithinbank (ed.): Norwegian-British-Swedish Antarctic Expedition 1949-52. Scientific Results, III, 99-116. Oslo, Norsk Polarinstitutt.

Thomsen, H.H. & Hagen, J.O., 1997: A stable isotope pilot study from blue ice areas of Jutulsessen, Dronning Maud Land, Antarctica. In O. Orheim (ed.): Report on the Norwegian Antarctic Research Expedition NARE 1992/93, Norsk Polarinstitutt Meddelelser, 125, 104-110.

Van den Broeke, M.R., Winther, J-G., Isaksson, E., Pinglot, J.F., Karlöf, L., Eiken, T. & Conrads, L., 1999: Climate variables along a traverse line in Dronning Maud Land, East Antarctica. Journal of Glaciology, 45(150), 295-302.

Winther, J-G., Elvehøy, H., Bøggild, C.E., Sand, K. & Liston, G., 1996: Melting, runoff and the formation of frozen lakes in a mixed snow and blue-ice field in Dronning Maud Land, Antarctica. Journal of Glaciology, 42(141), 271-278.

Wolmarans, L.G., 1982: Subglacial morphology of the Ahlmannryggen and Borgmassivet, western Dronning Maud Land. In C. Craddock (ed.): Antarctic Geoscience, 963-968. University of Wisconsin Press.

Wolmarans, L.G. & Kent, L.E., 1982: Geological investigations in Western Dronning Maud Land, Antarctica – a synthesis. South African Journal of Antarctic Research, Supplement 2, 93 pp.