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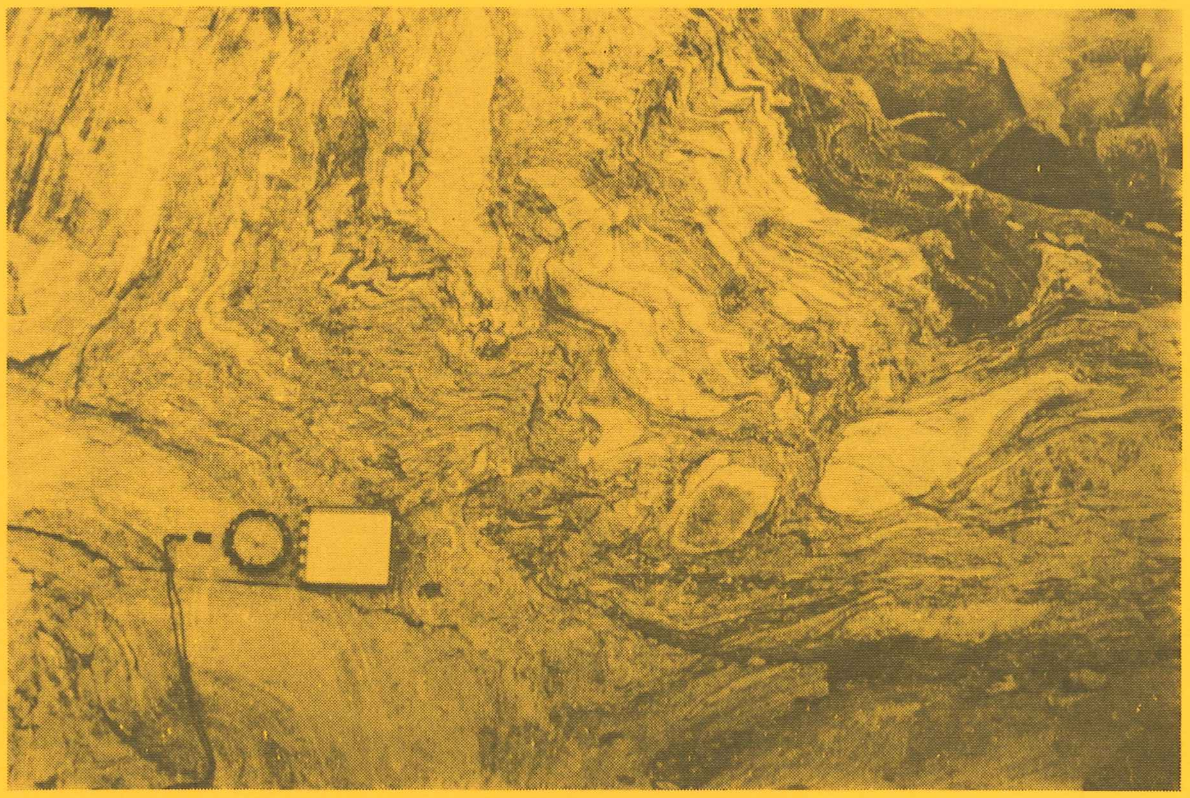

Basement - cover relationships in the Harkerbreen Group of the northern Ny Friesland Caledonides, Svalbard

Fredrik Hellman

Lund 1994 Examensarbete, 20 p Nr 58
Geologiska Institutionen, Lunds Universitet

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till Leif
Tack för all hjälp/. Fredrik

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Abstract: Mapping and isotopic work in the Instrumentberget area has shown that a granitic gneiss, previously referred to the Bangenhuk Formation, is unconformably overlain by a younger meta-sedimentary unit with a basal conglomerate horizon, previously included in the Rittervatnet Formation. In this paper it is suggested that the granitic gneiss, named the Instrumentberget Granitic Gneiss, occurs at a lower structural level than the Bangenhuk Formation and that this meta-sedimentary unit is the basal part of the Polhem Formation. Isotopic work on zircons from the Instrumentberget Granitic Gneiss and one boulder in the overlying basal conglomerate has shown that some of the conglomerate clasts are derived from the underlying Instrumentberget Granitic Gneiss the latter having a U-Pb zircon age of 1737±43 Ma. Evidence for this is three fold; firstly, the morphology of the zircons in one of the boulders is similar to zircons in the underlying granitic gneiss; secondly, a single U-Pb analysis from one conglomerate boulder falls right on the discordia line of the granitic gneiss; and thirdly, age analyses by the direct Pb-Pb evaporation method on single zircons from the granitic gneiss and the conglomerate boulder agree with the upper intercept age of the U-Pb analysis. A second conglomerate boulder has a quite different zircon morphology to the underlying granitic gneiss and a single U-Pb analysis falls significantly above the discordia line of the granitic gneiss, suggesting an exotic origin. Direct single zircon Pb-Pb evaporation analyses of this second boulder suggests that the Polhem Formation may be younger than 1000 Ma.

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Contents

Page

1. Introduction 1

2. Regional Geological Setting 3

2.1 The Hecla Hoek Succession in Ny Friesland 5
2.2 The Stubendorfbreen Supergroup (Lower Hecla Hoek) 6

3. Previous works in the Mosseldalen area 8

3.1 The Planetfjella Group 8
3.2 The Harkerbreen Group 9
3.2.1 The Harkerbreen Group conglomerates 10
3.3 Finnlandveggen Group 11
3.4 Structural setting 11

4. The Instrumentberget Area 12

4.1 The Instrumentberget Granitic Gneiss 13
4.2 The Meta-sediments of the Polhem Formation 17
4.2.1 The basal Mosselvatnet conglomerate 18
4.2.2 The intraformational conglomerates 21
4.2.3 The Quartzites of Polhem Formation 23
4.3 The depositional environment of the Polhem Formation 23

5. Isotopic age-determination work 24

5.1 Sample preparation and analytical methods 24
5.2. Morphology of the zircons 26
5.3 Results 26
5.3.1 The conventional U-Pb analyses 27
5.3.2 The Pb-Pb evaporation age determination analyses 28
5.4 Discussion of the isotopic results 31
5.4.1 The single zircon analyses of samples J92-010 and H92-29 31
5.4.2 The U-Pb analyses of samples J92-010 and H92-29 32
5.4.3 Isotopic analyses of sample H92-18 33

6. Summary and Conclusion 35

Acknowledgement 36

References 37

1. INTRODUCTION

This paper presents the results of new work on the conglomerate horizons within the Harkerbreen Group of the Lower Hecla Hoek of northern Ny Friesland, Svalbard (Fig 1 and 2). The area discussed is shown in Fig 2 and is located between Mosseldalen (latitude 79°53' N) and Femmilsjøen (latitude 79°47' N). The mapping and sample collections were made in the summer of 1992, during an expedition under a leadership of Prof. David Gee. The Harkerbreen conglomerate horizons have previously been described by Gayer & Wallis (1966), Harland et al. (1966) and Gayer (1969) as tilloid horizons. The new work includes (1) a detailed study of different conglomerate localities. (2) A new interpretation of the field relationships, structures and stratigraphy in the area of Instrumentberget on the south side of Mosseldalen. (3) An isotopic age determination study of zircons from two different conglomerate boulders and an underlying granitic gneiss to establish a maximum age of the conglomerte horizon. A basement-cover relationship is recognized, implying significant changes in the interpretation of the bedrock geology of Ny Friesland

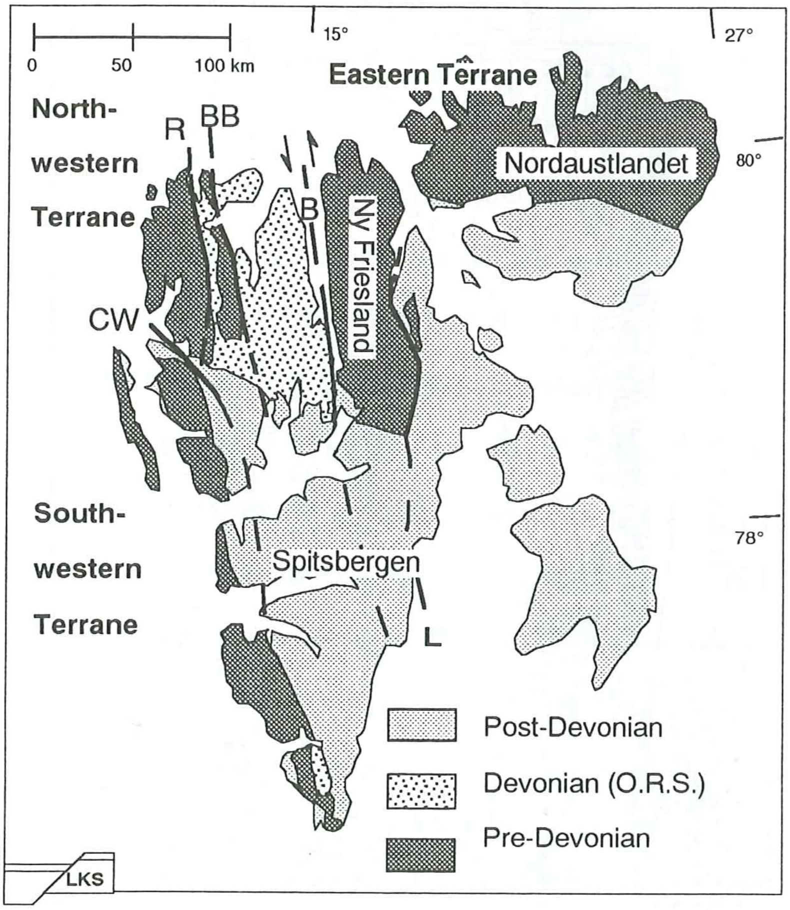

Fig 1. Geologic map over svalbards three different terranes and the major faults. B.= Billefjorden Fault; BB. Breibogen-Bockfjorden fault; L.= Lomfjorden Fault; R.= Raudfjorden Fault; C.W. = Central Western Fault; O.R.S.= Old Red Sandstone.

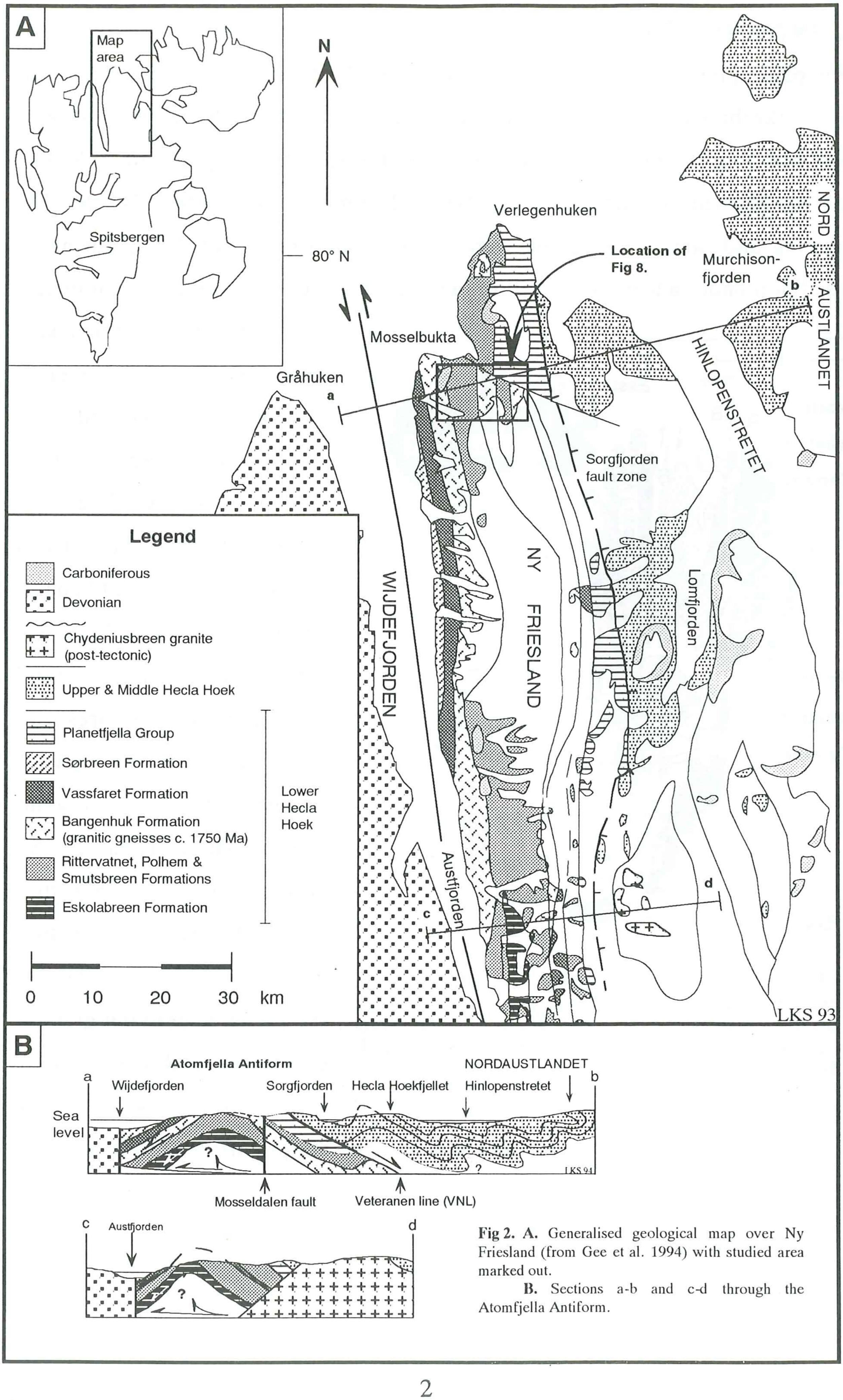

Fig 2. A. Generalised geological map over Ny Friesland (from Gee et al. 1994) with studied area marked out.

B. Sections a-b and c-d through the Atomfjella Antiform.

2. REGIONAL GEOLOGIC SETTING

The Svalbard's archipelago consists of at least three Caledonian terranes separated by major faults (Harland 1972; Harland 1985) and overlain by late Paleozoic and younger sediments. Devonian Old Red Sandstones (O.R.S.) occur in a major graben in north-central Svalbard and younger sedimentary rocks cover the central and the southeastern areas of the archipelago (Fig 1). The western Caledonian terranes show similarities with Ellesmere Island and northern Greenland; the eastern terrane, exposed in Ny Friesland and Nordaustlandet, shows considerable similarity with central eastern central Greenland (Harland 1972; Harland 1985; Harland & Wright 1979; Ohta et al. 1989). This has been inferred to imply that the eastern terrane was detached from Greenland during the Early Paleozoic and moved sinistrally northwards on a major transcurrent fault (Fig 3B; Harland 1985). During closing of the Iapetus Ocean the different terranes were accreted together and the present-day interrelationship were established. It has been suggested by Gee (1986) that the different terranes are a result of an orthogonal collision further to the south between the Laurentian and the Russo-Baltician plate and a northwards "escape" of fragments of the Laurentian plate (Fig 3A). The

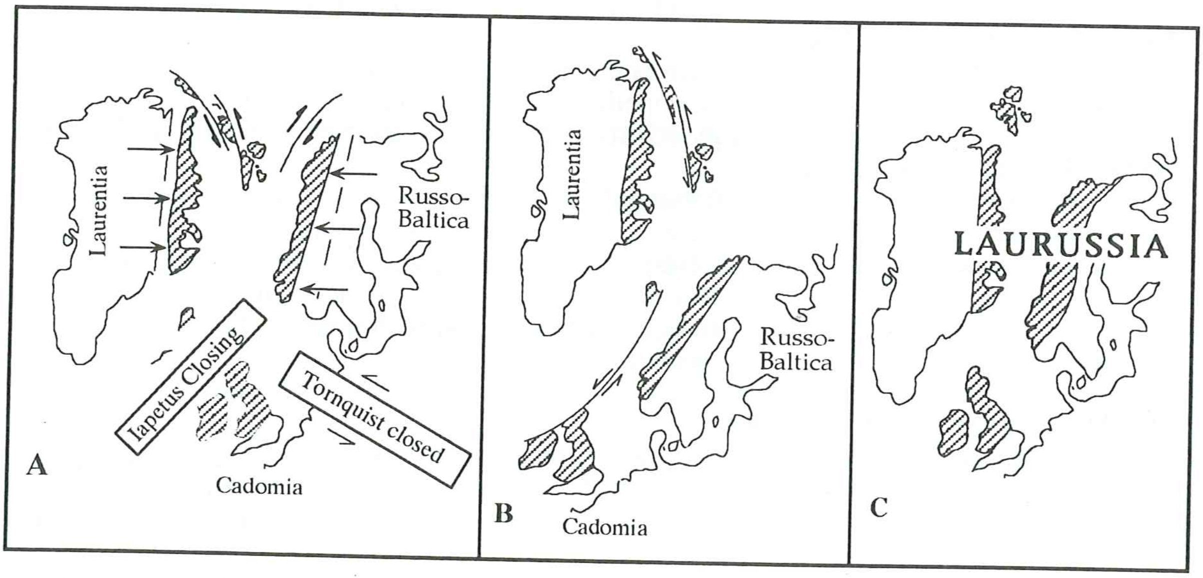

Fig 3 . Alternative interpretations of the accretion of Svalbard's different terranes during the closing of the Iapetus ocean. From Gee and Page (in press). A: The transpression results in an "escape" of the different terranes (Gee 1986). B: Sinestral shear in devonian time after the scandian collision (Harland 1985). C: Latest Devonian assembly of Caledonian terranes.

Table 1. The stratigraphy of the Hecla Hoek succession, based on Harland et al. (1985).

Hinlopenstretet Supergroup (Upper Hecla Hoek)

Oslobreen Group 1.2 Km	
Valhallfonna Fm	(limestone) *Arenig-Llanvirn*
Kirtonryggen Fm	(limestone and dolomite) *Arenig*
Tokammane Fm	(dolomite and sandstone) *Caerfai*
Polarisbreen Group 0.8 Km	
Drakoisen Fm	(shale)
Wilsonbreen Fm	(tillite) *Late Varangian*
Elbobreen Fm	(shale) *Early Varangian*

Lomfjorden Supergroup (Middle Hecla Hoek)

Akademikerbreen Group 2 km	
Backlundtoppen Fm	(dolomite and shale)
Draken Conglomerate Fm	*Early Sturtian*
Svanbergfjellet Fm	(limestone and dolomite)
Grusdievbreen Fm	(limestone)
Veteranen Group 3.8 Km	
Oxfordbreen Fm	(shales)
Glasgowbreen Fm	(graywacke and quartzite)
Kingbreen Fm	(quartzite and shale with graywacke and carbonate)
Kortbreen Fm	(quartzite and limestone)

Stubendorfbreen Supergroup (Lower Hecla Hoek)

Planetfjella Group 4.3 km	
Vildadalen Fm	(semipelite, psammite and quartzite)
Flåen Fm	(semipelite, psammite and quartzite with acid pyroclastics)
Harkerbreen Group 4.1 km	
Sørbreen Fm	(quartzite and amphibolite)
Vassfaret Fm	(semipelite, psammite and amphibolite)
Bangenhuk Fm	(feldspathite and amphibolite)
Rittervatnet Fm	(psammite,pelite, semipelite, dolomite, **conglomerate** and amphibolite)
Polhem Fm	(quartzite and amphibolite)
Finnlandveggen Group 2.7 km	
Smutsbreen Fm	(semipelite and marble)
Eskolabreen Fm	(feldspathite, semipelite and amphibolite)

early geological research on Svalbard's eastern terrane by Blomstrand (1864), Nordenskiöld (1863), De Geer (1909) and Nathorst (1910) separated an unmetamorphosed succession "the Heclahook Formation" (later the Middle and Upper Hecla Hoek succession) from an underlying "basement" of granitic gneisses and schists. Kulling (1934) extended the understanding of the stratigraphy of the Hecla Hoek succession with identification of Cambrian, Vendian and underlying Riphean strata. Harland et al. (1992) do not recognize basement and claim that the Hecla Hoek succession represents an 18 km thick more or less continuous sedimentary succession, without major stratigraphic breaks (Table 1). Other authors (Krasil'scikov 1973; 1979) have supported the previous interpretation (e.g., Nordenskiöld 1866) that at least the lower part (the Atomfjella series) of the Lower Hecla Hoek is a basement complex to the Caledonian succession. Recent work (Gee et al 1992) has demonstrated the presence of Early Proterozoic basement and has interpreted the lower part of the Hecla Hoek to be a stack of thrust sheets (Fig 2A and B) containing intercalations of basement and cover (Gee et al. 1994). U/Pb zircon ages of 1700-1800 Ma from granites in the Harkerbreen Group (Table 1) imply that significant segments of Early Proterozoic basement are incorporated in the lower part of the Hecla Hoek (Gee et al. 1992).

2.1 The Hecla Hoek Succession in Ny Friesland.

From the work of a number of Cambridge expeditions to Ny Friesland, a succession of Hecla Hoek rocks was outlined by Harland and Wilson (1956), Harland et al. (1966) and Harland et al. (1992). They divided the succession (Table 1) into three major units (1) the Upper Hecla Hoek or Hinlopenstretet Supergroup, (2) the Middle Hecla Hoek or Lomfjorden Supergroup and (3) the Lower Hecla Hoek or Stubendorfbreen Supergroup. The Upper Hecla Hoek consists of Cambrian to Ordovician sandstones, limestones and dolomites with Vendian tilloids and shales at the base. The Middle Hecla Hoek is build up of dolomites and limestones in the upper part, underlain by quartzites, shales and greywackes and the Lower Hecla Hoek is composed of a variety of meta-sedimentary and meta-igneous rocks, generally in high amphibolite facies, but decreasing in grade upwards to the contact with the Middle Hecla Hoek. The middle and upper supergroups reflect shallow marine conditions of deposition during the Neoproterozoic or Early Paleozoic, with subaerial exposure at intervals throughout and few lateral facies variations, thus suggesting a stable shelf environment (Harland et al 1992). The lower supergroup contains a wide range of highly deformed meta-sedimentary rocks apparently deposited in both shallow and deep marine environments.

2.2 The Stubendorfbreen Supergroup (Lower Hecla Hoek)

The Lower Hecla Hoek rocks are exposed within a major N-trending fold, the Atomfjella Antiform of western Ny Friesland (Fig 2B; Harland 1959). They are all highly tectonized and metamorphosed in amphibolite facies and demonstrate three or more phases of homoaxial folding (Gayer 1969; Manby 1990). In the eastern limb of the Atomfjella Antiform the Lower Hecla Hoek is overlain by the Middle Hecla Hoek. The contact zone has been interpreted as a major fault (Nathorst 1910; Manby 1990), or as a sedimentary transition, disturbed by minor faulting (Harland et al. 1966). Harland et al. (1992) referred to this contact zone as the Veteranen line (VNL). The Lower Hecla Hoek in the area studied here has previously been described by Gayer & Wallis (1966), and Gayer (1969) presented a map over the Mosselbukta-Femmilsjøen area (Fig 4).

Fig 4 The geologic map of the Mosseldalen area after Gayer (1969) with his interpretation of the location of the Atomfjella Antiform (AA). Section 2-2 in figure 6.

3. PREVIOUS WORK IN THE MOSSELDALEN AREA

Gayer and Wallis (1966) and Gayer (1969) divided the rocks in the Mosseldalen area (Fig 4) into two major units, the Planetfjella and the Harkerbreen Groups. Further south of this map area in Ny Friesland, the latter is underlain by a third unit, the Finnlandveggen Group (Harland et. al. 1966, 1992).

3.1 The Planetfjella Group

The Planetfjella Group represents the highest unit of the Lower Hecla Hoek (Table 1) and is composed of two lithological associations (Wallis 1969), the Vildadalen Formation, underlying the Veteranen Group (Middle Hecla Hoek) and the Flåen Formation, the latter overlying the Harkerbreen Group. The Vildadalen Formation consists of graded greywacke turbidites, orthoquartzites, impure psammites and, more rarely limestones and dolomites. The underlying Flåen Formation is mainly composed of semipelites with subordinate more massive units consisting of conspicuous pink feldspar megacrysts in a dark schistose matrix. The metamorphic grade in the Planetfjella Group varies from low greenschist facies at the VNL to high amphibolite grade at the base of the Flåen Formation. Rapid increase in metamorphic grade suggests normal faulting to be important in VNL (Gee et al 1994). The most notable differences between the Harkerbreen and Planetfjella Groups are the more pelitic to semi-pelite lithologies in the latter, always with the lack of amphibolites. The contact between the Harkerbreen Group and Planetfjella Group has been interpreted as a thrust fault, based on the evidence that the lower Flåen Formation lies on different units of the Harkerbreen Group (Gayer 1969; Gee et. al. 1994). Harland et al (1992) reject this interpretation and suggest that this contact is a transition zone, without any major tectonic movement.

3.2 The Harkerbreen Group

The Harkerbreen Group has been divided from top downwards into five mappable subunits, the Sørbreen, Vassfaret, Bangenhuk, Rittervatnet and Polhem Formations (Fig 5). The Sørbreen Formation is an unit at least 265 m thick, consisting of psammites (including quartzites) and foliated amphibolites; meta-acid-tuffs occur in the lower part (Gayer and Wallis 1966). This formation has a transitional lower boundary and has been interfolded with the underlying Vassfaret Formation. The latter consists of c. 600m of semi-pelites with bands of psammite and schistose amphibolites. Cross-bedding has been reported in this formation locally cut by amphibolites (Gee et al. 1974)

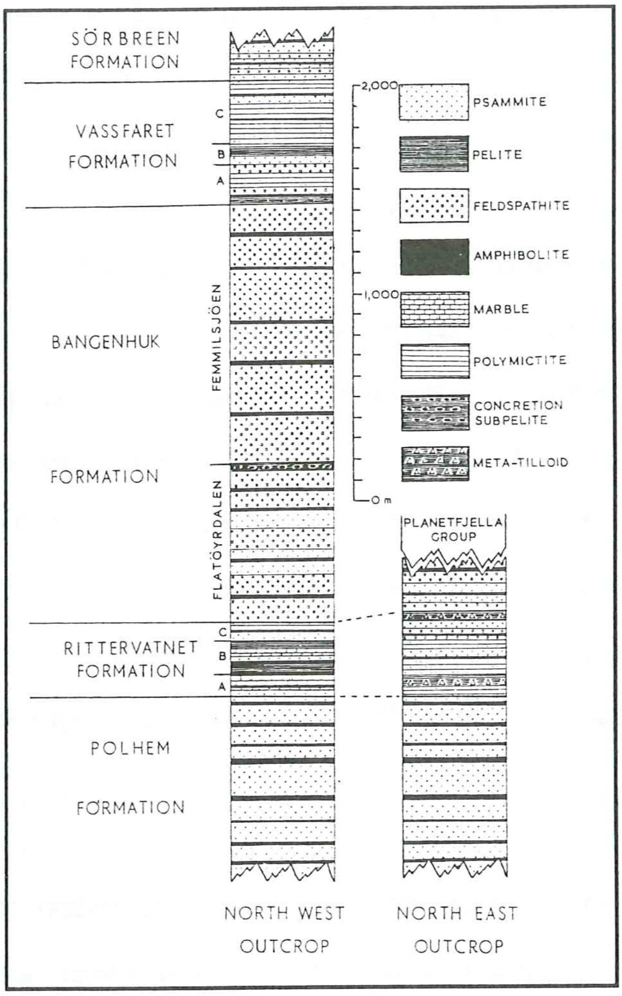

Fig 5. The stratigraphy of the Harkerbreen Group (Gayer and Wallis 1966).

The underlying Bangenhuk Formation has a distinct upper contact to the Vassfaret Formation which is frequently obscured by aplite veins and sheets (Gayer 1969). Gee et al. (1994) have described this contact to locally be cut by intrusive granites. The Bangenhuk Formation was further divided into the Flatøryrdalen Member, overlain by the Femmilsjøen Member. The c. 2000 m thick Bangenhuk Formation consists largely of foliated quartzo-feldspathitic gneisses and includes deformed granites (Gayer 1969; Gee et. al. 1992). Many authors have interpreted the Bangenhuk gneisses to be of acid volcanic origin (Gayer and Wallis 1966; Harland 1966, 1992; Manby 1990). Carlsson (1993) has geochemically analysed a variety of Bangenhuk Formation rocks and inferred that they are derived from A-type granites. Several U-Pb-age determinations of Bangenhuk granites have yielded ages of 1724-1766 Ma (Gee et. al. 1992, Johansson, unpublished data 1993).

The underlying Rittervatnet Formation (c. 350 m), subdivided into three divisions A, B and C, has been described to be a series of thin banded psammites, with interbanded feldspathic-psammites, schistose semipelites, graphitic schists, marbles and amphibolites, all strongly interfolded with both the overlying Bangenhuk Formation and the underlying Polhem Formation. Two conglomerate horizons were located by Gayer (1969) and included in the Rittervatnet Formation; these are the subject of this paper.

The lowest unit of the Harkerbreen Group as defined by Harland et al. (1992) is the Polhem Formation (c. 900 m), consisting of quartzites and coarsely banded and poorly foliated psammites together with schistose amphibolites. This formation is similar to the Sørbreen Formation. Harland et. al (1992) renamed the unit the Mosselbukta Formation, but in this study the original name is preferred.

3.2.1 The Harkerbreen Group conglomerates

Gayer and Wallis (1966) included the conglomerates into the Rittervatnet Formation and distinguished lower and an upper units (Fig 5). They were described to contain of clasts up to 2 m in diameter consisting of quartzites (10%), psammites (40%), fine-banded feldspathites (30%) and granites (20%). The matrix, a medium to coarse grained foliated biotite-feldspathite was said to constitute more than 50% of the rock. Gayer (1969) measured the strain in the clasts and showed that the largest strain ratio is in the quartzites (4.5:1) and the smallest in the granites (1.5:1). The lower conglomerate horizon (10-20 m) was described to occur within the lower part of the Rittervatnet Formation and to differ from the upper horizon in both the composition of the matrix and the phenoclasts. The matrix was said to be composed of a fine to medium grained biotite psammite, with a well developed schistosity, shown by the biotite fabric, the basal part (3-5 m) consisting of the lower conglomerate horizon containing of 80% amphibolite clasts and 20% vein quartz and granular psammite clasts. The proportion of vein quartz was described to increases gradually upwards, the upper 3-5 m of the unit being made up of 80% vein quartz clasts and 20% psammites and amphibolites in equal proportions.

3.3 Finnlandveggen Group

Finnlandveggen Group, the lowest unit of the Hecla Hoek succession of Ny Friesland outcrops in the core of the Atomfjella antiform in southern Ny Friesland. The upper part is composed of semipelite and marble and the lower part consists of quartzo-feldspathitic gneisses, psammites, marbles and amphibolites.

3.4 Structural Setting

Gayer (1969) considered the area from Mosselbukta to Femmilsjøen to be dominated by a major antiform, thought to be the northerly continuation of the

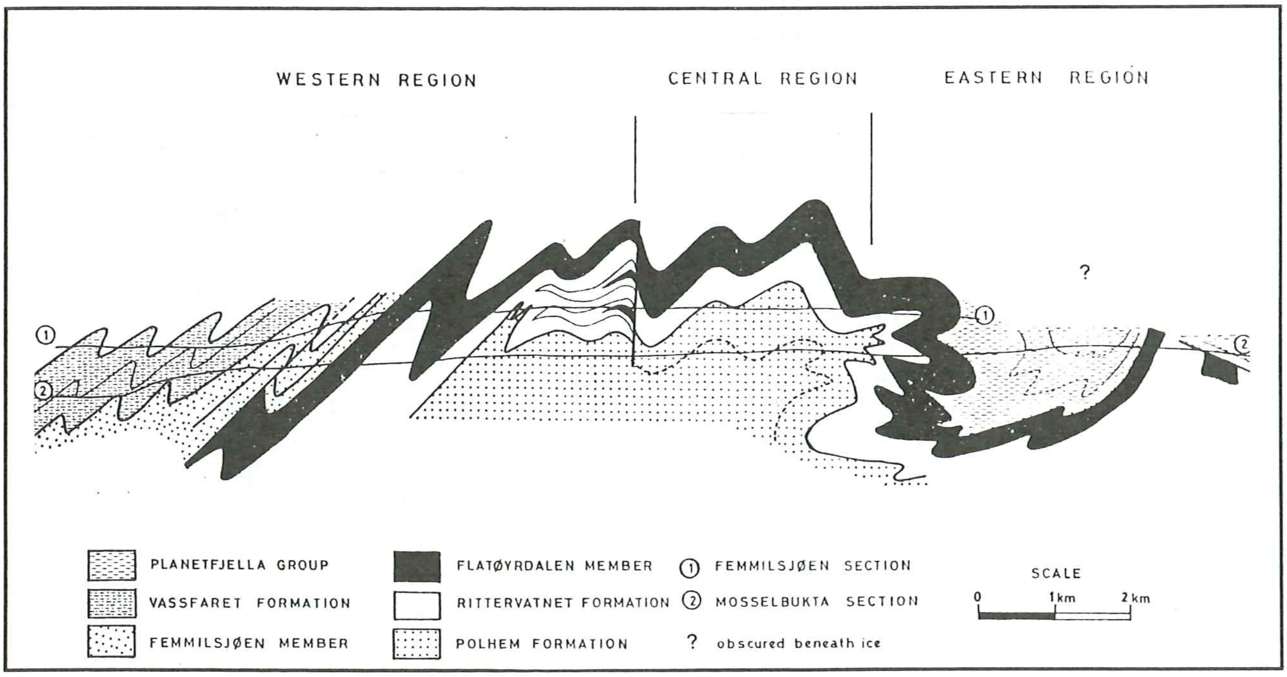

Fig 6. Generaliced section 2-2 (Fig 5) across the Atomfjella antiform. From Gayer (1969)

Atomfjella Antiform. He interpreted the fold axis to be located within the Polhem Formation (marked in Fig 4). The correlation between the western and the eastern limbs are problematic because the rocks of Rittervatnet Formation are different in east and west (Fig 5). In the west, divisions A and B occur, containing graphitic schists, marbles and quartzites; in the east, only division C is present, composed of psammites, feldspathites and two conglomerate horizons. Gayer and Wallis (1966) explained the difference in lithologes to be related to depositional facies changes. In addition to these problems of contrasts in the Rittervatnet Formation, the stratigraphy described by Gayer (1969) require a complicated refolding of the Bangenhuk and Rittervatnet Formations in the eastern limb of the antiform, with local inversion (Fig 6). The new work, described below requires revision of the geological map and structural interpretation.

4. THE INSTRUMENTBERGET AREA

The recent Swedarctic mapping in the area between Mosseldalen and Femmilsjøen interprets the main axis of the Atomfjella Antiform to be located east of the Polhem Formation (marked in Fig 7) and not as previously inferred (Fig 4). This structural interpretation implies that the stratigraphy is in the right order (Fig 8) and not upside down in a series of complicated folds in the eastern limb of the antiform. We conclude from the new mapping that the meta-sediments in the Instrumentberget area, previously referred to division C of the Rittervatnet Formation, are the basal part of the Polhem Formation; they overlie the granitic gneiss, earlier correlated with the Bangenhuk Formation (Fig 8). In this study, the lower granitic gneiss is called the Instrumentberget Granitic Gneiss, in order to separate it from the structurally higher Bangenhuk Formation gneisses; a close genetic relationship is still possible, but not confirmed. During the summer of

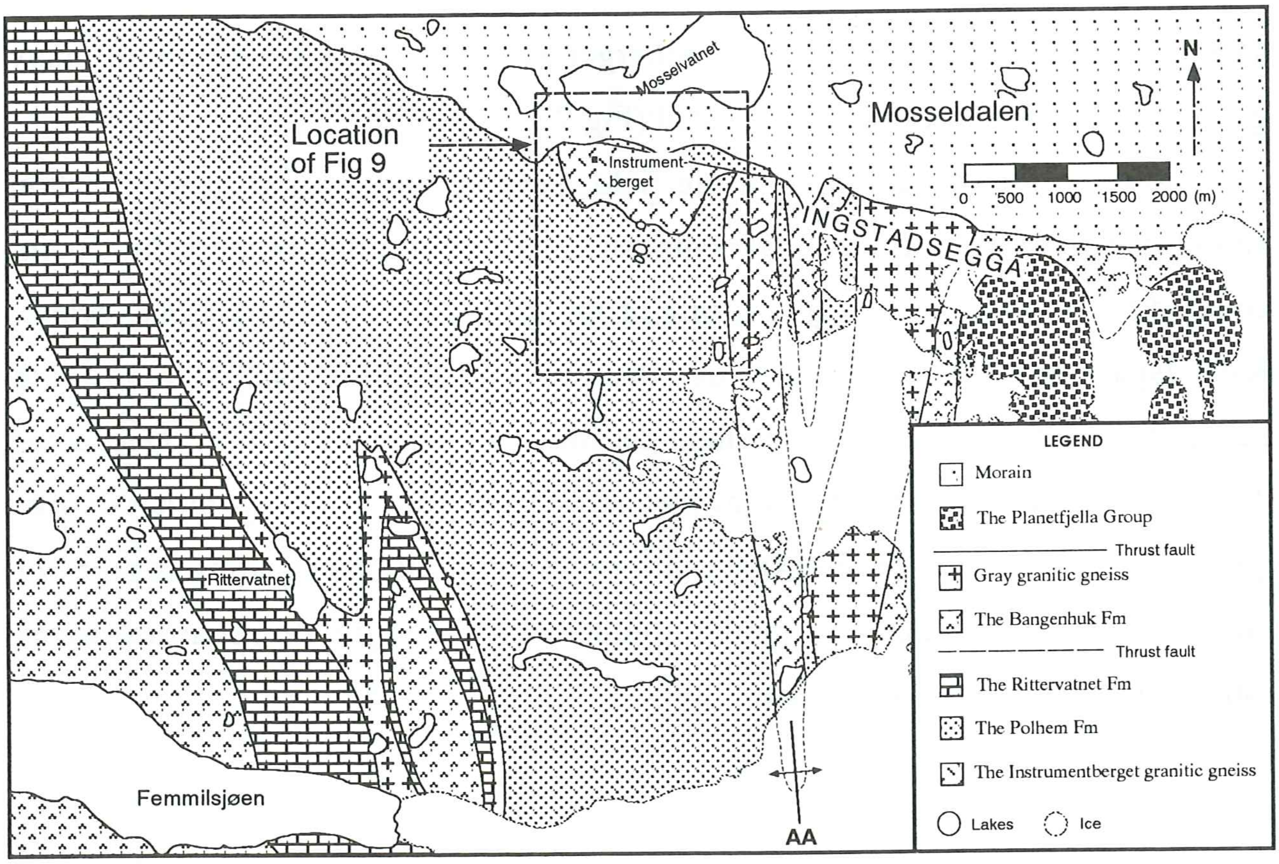

Fig 7. Geological map, based on the Swedartic mapping (pers. com P. Nilsson) in the Mosseldalen-Femmilsjøen area. The interpreted location of the main axis of the Atomfjella Antiform (AA) are marked in the figure.

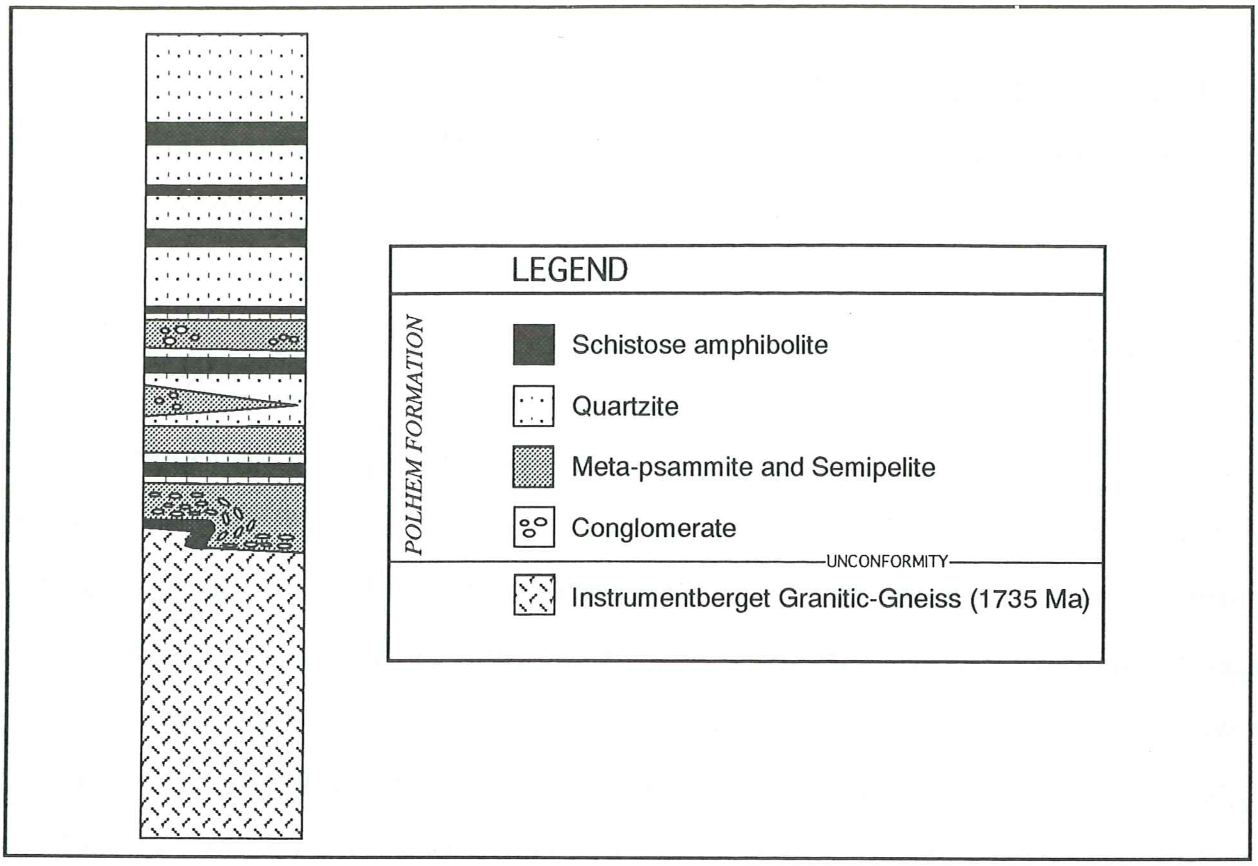

Fig 8. The stratigraphy in the Instrumentberget area.

1992, a small area around Instrumentberget was mapped in detail where many good conglomerate localities are located (Fig 9).

Rock units, distinguished in the field, have been examined in thin section and classified according to modal composition. The thin sections were first point counted (500 points or more). Because of the difficulties to separate plagioclase, K-feldspar and quartz, some of the thin sections were stained with hexanitrocoboltate, following a method described by Widmark (1979). Due to the deformation and amphibolite facies metamorphism, rock forming minerals like quartz and feldspar have segregated and formed compositional banding in many of the rocks, implying that the modal composition can only be regarded as an estimation of the true composition (Table 2).

Table 2. Modal composition of rocks in the Instrumentberget area (a = accessory).

	Sample nr	Quartz	K-feldspar	Plagioclase	Muscovite	Biotite	Chlorite	Hornblend	Garnet	Epidote	Opaque	Apatite	Karbonate
Instrumentberget granitic gneiss	H92-10	28%	30%	29%	6%	4%	2%						
Schistose amphibolite	H92-28	20%	12%	12%		16%		26%	1%	7%	6%		
Basal Conglomerate, Locality 1													
Psammite clast	H92-25	41%	14%	28%	2%	5%	3%		a	5%	a		
Psammite clast	H92-24	41%	12%	32%	2%	8%	1%			3%	1%		
Psammite clast	H92-19	43%	22%	21%	2%	7%	2%			a	a	a	
Psammite clast	H92-30a	95%	2%	2%	1%								
Matrix	H92-23	34%	2%	21%	a	36%	1%		1%	2%	a	1%	
Matrix	H92-27	51%	10%	15%	7%	10%			a	5%	a		
Granitic to granodioritic gneiss clast	H92-29	47%	13%	30%		9%				a	1%		
Granitic to granodioritic gneiss clast	H92-26	34%	7%	43%	2%	8%	1%		a	a	3%	a	
Homogenous granodioritic clast	H92-18	42%	6%	47%	3%		a		a			a	
Homogenous granodioritic clast	H92-20	34%	17%	14%	1%	12%	1%			19%	a	a	
Basal Conglomerate, Locality 3													
Homogenous granitic clast	H92-08	38%	21%	28%	6%	a	1%			3%	a	a	a

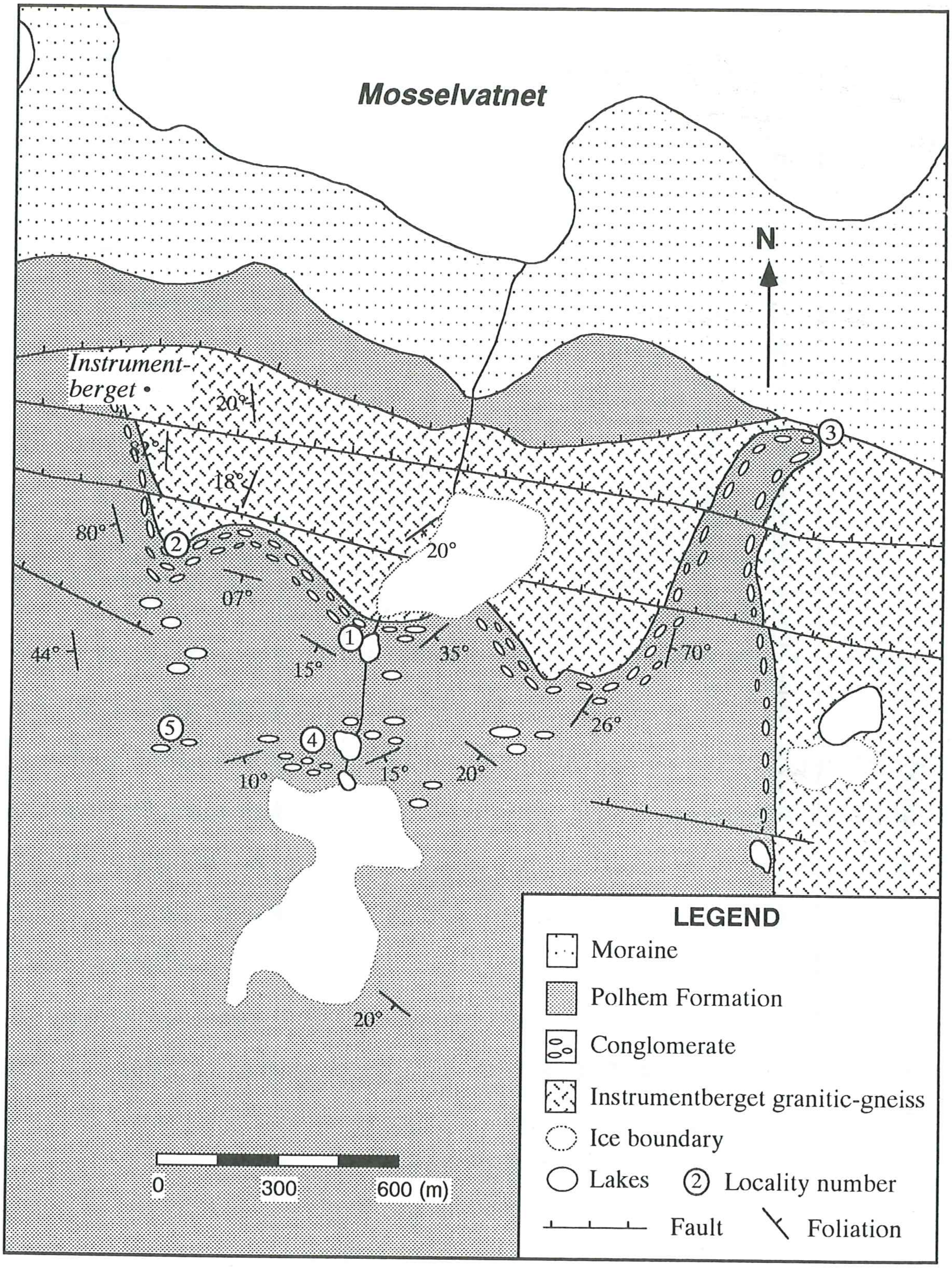

Fig 9. Geologic map over the Instrumentberget area, including the locality number referred to in the text.

4.1 The Instrumentberget Granitic Gneiss

This pink granitic gneiss represents the lowest stratigraphic unit in the Instrumentberget area (Fig 9). It is medium to fine grained and consists of quartz, plagioclase, K-feldspar with subordinate muscovite, biotite and secondary chlorite; accessories include apatite, magnetite and zircon. The gneissosity is defined by oriented brown biotite, muscovite, granular recrystallised quartz, plagioclase and K-feldspar. A penetrative, S-plunging (5-15°) stretching lineation in the quartz and feldspars is characteristic. The rock is completely recrystallised and no primary textures have been recognized. The plagioclase is highly sericitised in some domains and can be difficult to distinguish from K-feldspar, without staining. In other domains, the plagioclase has not been sericitised very much and exhibits albite twins.

4.2 The Meta-sediments of the Polhem Formation

The contact between the meta-sedimentary rocks of the Polhem Formation and the underlying Instrumentberget Granitic Gneiss is usually represented by an strongly schistose amphibolite, up to a few meters thick. Major constituents of this rock are hornblende, quartz, plagioclase, K-feldspar, biotite and epidote, with subordinate garnet and opaque minerals. A basal polymict conglomerate overlies the schistose amphibolite with a sharp contact; however, in a few localities, the conglomerate directly overlies the granitic gneiss. Often the contact is folded (Fig 11). The conglomerate sequence passes gradually up into a fine to medium grained granular schistose psammite, consisting of quartz (>50%), plagioclase, K-

Kuhi
IGG

Fig 10. Foliated psammite of the Polhem Formation.

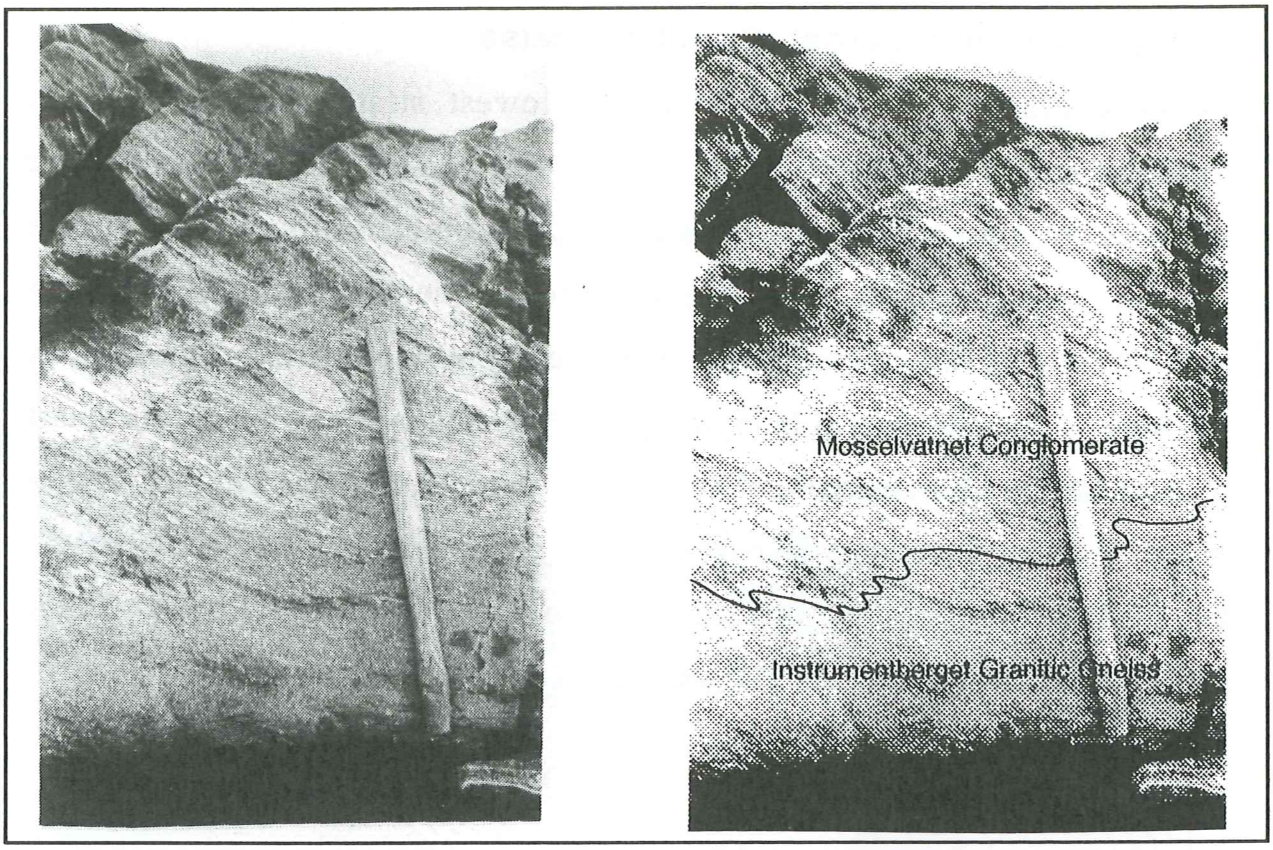

Fig 11. Folded contact between the Mosselvatnet Conglomerate and the Instrumentberget Granitic Gneiss.

feldspar, biotite, epidote and opaque minerals (Fig 10)15. The compositional banding reflects primary variations in the sediments, but the parallellism of the banding could be the result of the penetrative deformation. The psammite is interrupted by several schistose amphibolites, lying concordantly in the foliation. They range in thickness up to 20 metres and they are often boudined in a N-S direction. Further up in the sequence of meta-sediments, there occur various intraformational conglomerates. Gradually this impure lower part of the formation passes up into pure, but still foliated, quartzites and schistose amphibolites, typical for the Polhem Formation.

4.2.1 The basal Mosselvatnet Conglomerate.

Although most of the best conglomerate localities are on the top of the Instrumentberget, the most accessible exposure is located immediately south of Mosselvatnet at the base of the cliff; this name is therefore used for the basal conglomerate in this study. It is treated as a member of the Polhem Formation. The Mosselvatnet Conglomerate beds are up to 10 m in thickness and contain a variety of clasts, including psammites,vein quartz, granitic to granodioritic gneisses and more homogenous granites to granodiorites. The shortest axis of the clasts are up to 30 cm long; they are on average about 5 cm. Some more competent clasts have been boudined, with both N-S and E-W extension.

The *psammite clasts* are medium to fine grained, poorly lineated with a granular texture, and consisting mainly of quartz (40-95%), plagioclase, K-feldspar; minor biotite, epidote, chlorite and muscovite are present, with accessory opaques, garnet and limonite. Epidote may occur in significant proportions and constitute up to 20% of the rock. Larger plagioclase and quartz grains are often intergrown in an equidimensional matrix. The psammite clasts have been more ductile during deformation than the granitic to granodioritic gneisses and the more homogenous granodioritic clasts.

The *granitic to granodioritic gneiss clasts* consist of quartz, plagioclase, K-feldspar, biotite and epidote in excess of magnetite, chlorite and muscovite, with accessory garnet, apatite and zircon. The plagioclase usually shows albite twins. The gneissosity is shown by oriented brown biotite and bands of quartz and feldspar. In coarse to medium grained clasts, the K-feldspar often shows larger sutured grains in a mosaic textured recrystallized matrix of quartz and feldspar. Locally, the foliation in the clasts is discordant to the foliation in the matrix, suggesting that the clasts were deformed prior to the deposition of the conglomerate. These clasts resemble the underlying granitic gneiss, but they are often more foliated.

The *homogeneous granitic to granodioritic clasts* are coarse grained and show only a weakly oriented fabric. In general, they consist of quartz, plagioclase, K-feldspar and muscovite with accessory apatite, opaques and zircon. Some of the clasts contain small round garnets. The plagioclase is highly sericitised and often has inclusions of small muscovite grains. Both the quartz and K-feldspar forms large (up to 1 cm) sutured grains and enclose abundantly inclusions of epidote and muscovite.

The *conglomerate matrix* contains quartz, plagioclase, K-feldspar, biotite, muscovite and epidote with accessory garnet, apatite and opaques. The matrix is very similar in appearance to the psammite clasts, but usually has a higher biotite content and is more epidote rich. Within the matrix there are very thin pelitic layers (3-15 cm), containing mainly mica (40-60%) with biotite in excess of muscovite, quartz (<35%), plagioclase (<10%) and accessory garnet, opaques and epidote.

Described below are some typical localities where the basal conglomerates are well exposed.

Locality 1.

The conglomerate (Fig 12) occurs in an open gently S-plunging antiform on Instrumentberget (Fig 9) and is best preserved in the fold hinge, where it is up to 5 m thick. It is separated from the underlying granitic gneiss by a c. 2 m thick unit of schistose amphibolite with garnets up to 1 cm in diameter. The basal 2 m of the conglomerate is clast-supported; it becomes gradually matrix-supported upwards, the matrix constituting around 30% at the base and over 50% in the upper part of these basal beds. The clasts consist of granular psammite (75%), granitic to granodioritic gneisses (20%) and coarse grained more homogenous granitic to granodiorites (5%).

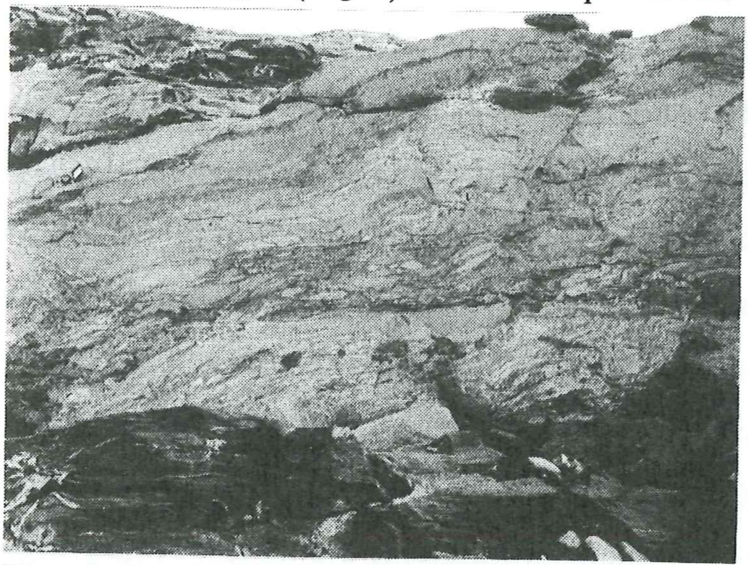

Fig 12. The Mosselvatnet Conglomerate at locality 1 (see Fig 9).

Locality 2.

The conglomerate described above can be followed westwards through a gentle synform into a close antiform at locality 2 (Fig 9), where the fold axial plane dips steeply and the western limb dips nearly vertically. The contact between the conglomerate and the schistose amphibolite at this locality is tightly folded (Fig 13) and the relationships to the granitic gneiss are mostly covered.

Locality 3.

The conglomerate occurs at this locality in a tight synform about 600 m south of Mosselvatnet. The fold has an axial plane dipping at a moderate angle eastwards. The contact to the under-lying schistose amphibolite is sharp and the latter is underlain by a granitic gneiss that is locally covered by moraine. The conglomerate sequence is 2-4 m thick and contains two main types of clasts, granitic to granodioritic gneisses (80%) and psammites (20%). The granitic to granodioritic gneiss clasts contain a small amount of carbonate. The conglomerate is matrix-supported and the clasts have a lensoid shape, where the shortest axes are up to 10 cm long. The matrix constitutes more than 60% of the rock and has a well developed compositional banding, consisting of beds dominated by epidote and biotite, alternating with bands largely composed of quartz and feldspar.

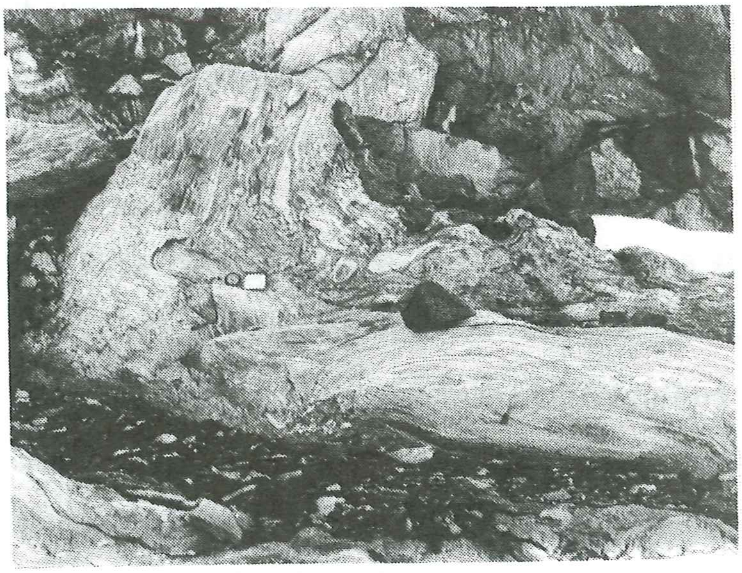

Fig 13. Folded Mosselvatnet Conglomerate at locality 2 (see Fig 9).

4.2.2 *The Intraformational Conglomerates*

Within the psammitic sequence overlying the Instrumentberget Granitic Gneiss, conglomerates occur at different levels. The contact between these intraformational conglomerates and the meta-sediments is not sharp. Several intraformational conglo-merates have been mapped and each of them persists over

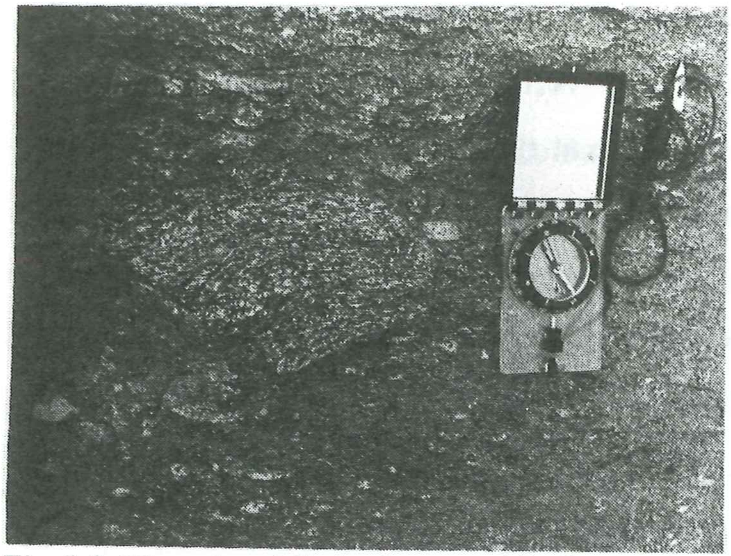

Fig 14. Discordant foliation in an amphibolite clast in the intraformational conglomerates at locality 4 (see Fig 9) horizon.

a small area before pinching out. The locality described by Gayer and Wallis (1966) where the conglomerate gradually changes clast content, from amphibolite domi-nated at the base to quartzite dominated at the top, has not been reinvestigated. The intraformational conglomerates are always matrix-supported, with clasts of psammite, granitic to granodioritic gneiss and amphibolite in a psammitic matrix. The amphibolite clasts are often foliated parallel to the foliation in the matrix, but in some cases the foliation is discordant implying a pre-depositional deformation (Fig 14).

Locality 4

This conglomerate sequence (c. 5m thick) contains clasts of quartzite (60%), psammite (15%), granitic to granodioritic gneiss, (20%) and amphibolite (5%). The sequence is more matrix supported than the basal unit. The shortest axes of the lensoid clasts are up to 30 cm, but are normally between 2-10 cm. The compositional banding in the matrix contains of up to 10 cm thick bands of epidote and biotite. The amphibolite pebbles are in general rich in epidote.

Locality 5

At locality 5, a thin (less than 3 m) conglomerate sequence contains clasts of foliated amphibolite (80%) and granitic gneiss (20%). The matrix is a fine to medium grained, quartz-biotite schist with large garnets. This sequence differs from the others in clast and matrix composition. Gayer and Wallis (1966) described a similar unit. The contact to the underlying psammites is very folded and the pebbles are best preserved in fold hinges.

4.2.3 The Quartzites of the Polhem Formation

The impure conglomerate bearing lower part of the Polhem Formation passes up into more pure meta-sandstones. These rocks consist of coarse banded, light quartzites with subordinate less pure psammites (Fig 10). The compositional banding varies up to 10 m in thickness and may have either sharp or gradational boundaries. The Polhem Formation quartzites contain, in general, quartz (60%) meta sandstones ? and seriticised plagioclase (20-30%) with subordinate muscovite, biotite, epidote and accessory apatite and zircon. The compositional banding is composed of thin bands of epidote and green biotite. In some localities, a tight folding exists, with the fold axes plunging gently to the south. They are interpreted to be parasitic folds to the major Atomfjella Antiform. Many of the schistose amphibolites have been boudinaged in N-S, due to the extension in the area.

4.3 The depositional environment of the Polhem Formation

The border between the Instrumentberget Granitic Gneiss and the Polhem Formation represents an unconformity. The Mosselvatnet Conglomerates, the basal part of the Polhem Formation, are interpreted to be deposited in a beach-shore face environment. During trangressions rapid sedimentation produced the impure sand-dominated sediments in the lower part of the Polhem Formation which include the intraformational conglomerates (Fig 8). These conglomerates can be deposited in many different ways, but one possible explanation is a nearby rise of the relif due to a fault from which larger blocks and pebbles were transported into the sandy sediments. The trangression continued with sedimentation of more quartz rich sand as the beach migrated landwards. A previous interpretation of the conglomerates as intraformational tilloids in the Rittervatnet Formation (Gayer and Wallis 1966; Harland et al. 1966) is not supported here.

5. ISOTOPE AGE-DETERMINATION WORK

As described above, the Mosselvatnet Conglomerate forms the lower member of the Polhem Formation that was deposited on the Instrumentberget Granitic Gneiss. Many of the granitic clasts in this basal conglomerate are lithologically similar to the underlying gneiss. It was therefore decided to carry out isotopic age-determination studies of zircons in both the gneiss and clasts in the basal conglomerate to see if they were comparable in age and test the basement-cover hypothesis. Two conglomerate boulders (Table 2) of different character were sampled at locality 1 (Fig 9), a coarse grained light homogenous granodioritic boulder (H92-18) and a medium grained strongly foliated dark grey granitic to granodioritic gneiss boulder (H92-29). The underlying Instrumentberget Granitic Gneiss (J92-010) was sampled c. 10 m below the contact to the conglomerates at the same locality.

5.1 Sample preparation and analytical methods

The zircons from the conglomerate boulders H92-18 and H92-29 were separated at the Department of Mineralogy and Petrology at the University of Lund. The two clasts were separated from all matrix and crushed down and sieved to a size less than 0.250 mm. The heavy minerals were separated using a Wilfley table. Magnetite was removed using a hand-magnet. Finally, the zircons were hand-picked under a binocular microscope. Preparation of the Instrumentberget Granitic Gneiss sample (J92-010) and the isotopic analyses of all three samples were made in Stockholm at the Laboratory of Isotope Geology of the Swedish Museum of Natural History. The U and the Pb isotope measurements were made on a Finnigan MAT 261 mass spectrometer.

Conventional U-Pb analyses: The chemical procedure followed standard procedures based on Krogh (1973), with zircon dissolution in HF + HNO_3 in microcapsules (Parresh 1987), spiking with ^{233}U - ^{235}U and ^{208}Pb tracers, separation of U and Pb by ion exchange in HBr, and further purification of U using ion exchange in HCl. The Pb analyses have been corrected for fractionation by 0.11±0.04% per mass unit for blank Pb by 0.065 ng and common Pb with $^{206}\mathrm{Pb}/^{204}\mathrm{Pb} = 15.57$, $^{207}\mathrm{Pb}/^{206}\mathrm{Pb} = 15.28$, $^{208}\mathrm{Pb}/^{204}\mathrm{Pb} = 35.2$ (Stacey and Kramers 1975, model at 1750 Ma). The U-Pb analyses as well as the intercept ages are given at 2 sigma confidence level and were calculated following the procedure of Ludwig (1991 a, b).

Single zircon analyses: The Pb evaporation analyses were made following the procedure described by Kober (1986, 1987) and Kröner and Todt (1988). The common-lead correction are made using the two-stage model developed by Stacey and Kramer (1975) and the ages are given in 1 sigma confidence level.

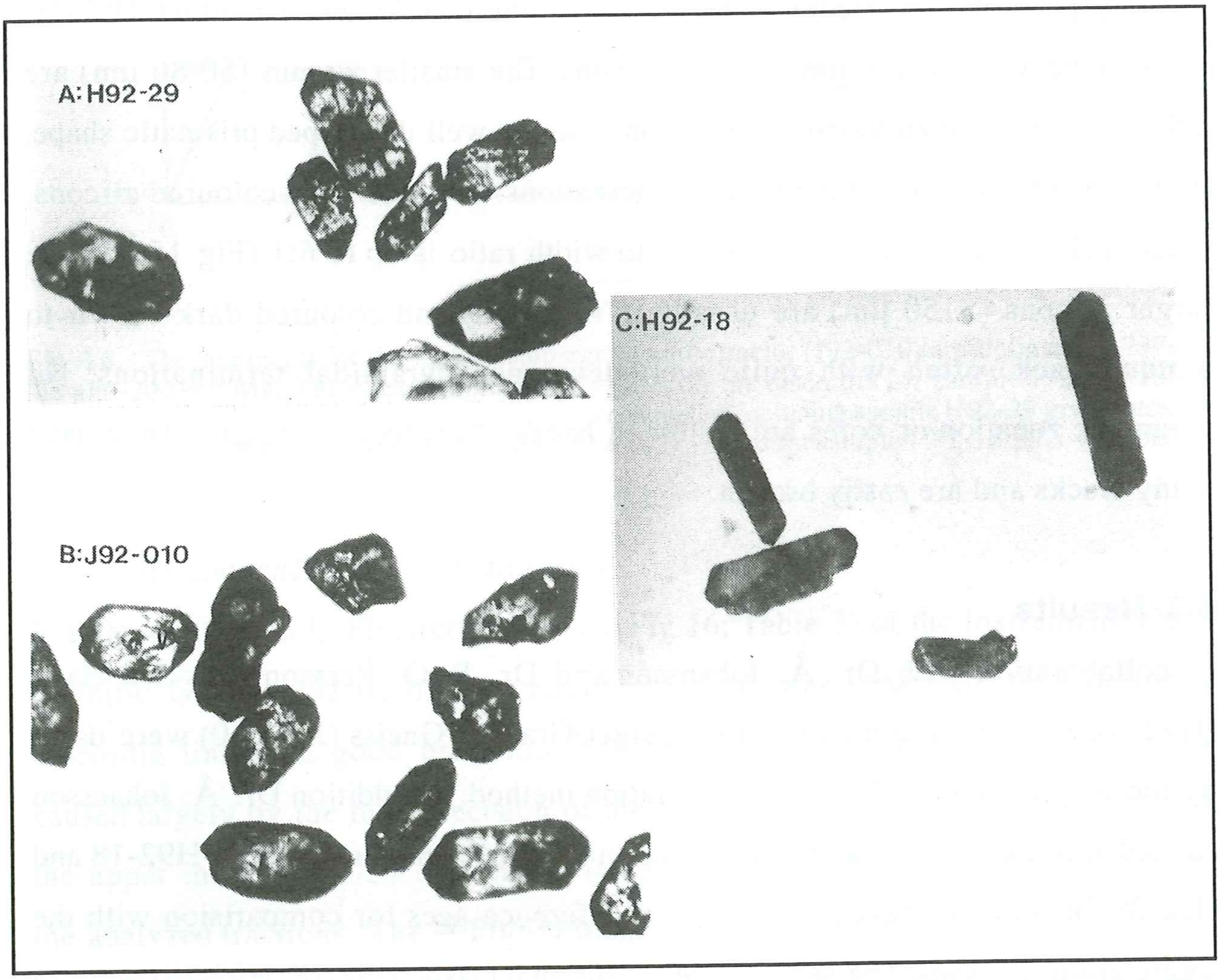

Fig 15. Zircons from three different samples at locality 1 (see Fig 9). These zircons are devided into type 1 (A and B) and type 2 (C) after their morphology. **A:** The Mosselvatnet Conglomerate boulder H92-29. **B:** The Instrumentbeget Granitic Gneiss sampled c. 10 m below the contact to the Mosselvatnet Conglomerate. **C:** The Mosselvatnet Conglomerate boulder H92-18.

5.2.Morphology of the zircons

Based on the morphology, the zircons in the three samples can be divided into two distinct types.

Type 1: This type of zircons is found in both the conglomerate boulder H92-29 and the Instrumentberget Granitic Gneiss (J92-010). The crystals have subhedral outline with well-developed, but quite rounded prismatic shape with a length to width ratio of approximatly 3:1. The color of the crystals is brown to pink and they are more or less transparent, with just a few visible fractures and inclusions (Fig 15 A and B). No zonation or cores were found in the backscatter electron image.

Type 2: These zircons, separated from the conglomerate boulder H92-18, are opaque, without any transparent domains. The smaller zircons (50-80 μm) are coloured light brown to dirty yellow and have a well developed prismatic shape, with less well developed pyramidal terminations. In some light-coloured zircons, black inclusion are visible. The length to width ratio is up to 6:1 (Fig 15 C). The larger zircons (>150 μm) are usually in fragments and coloured dark brown to almost black, often with quite well developed pyramidal terminations. No magmatic zonation or cores are visible in backscatter electron images. They have many cracks and are easily broken.

5.3 Results

In collaboration with Dr. Å. Johansson and Dr. P.-O. Persson, the two clasts (H92-18, H92-29) and the Instrumentberget Granitic Gneiss (J92-010) were dated by the single zircon $^{207}Pb/^{206}Pb$ evaporation method. In addition Dr. Å. Johansson carried out a conventional U-Pb zircon analysis of samples J92-010, H92-18 and H92-29, in order to have an independent reference ages for comparision with the single grain analyses. The sample sizes of the two clasts (H92-18, H92-29) were to small too allow more than one data point for each clast.

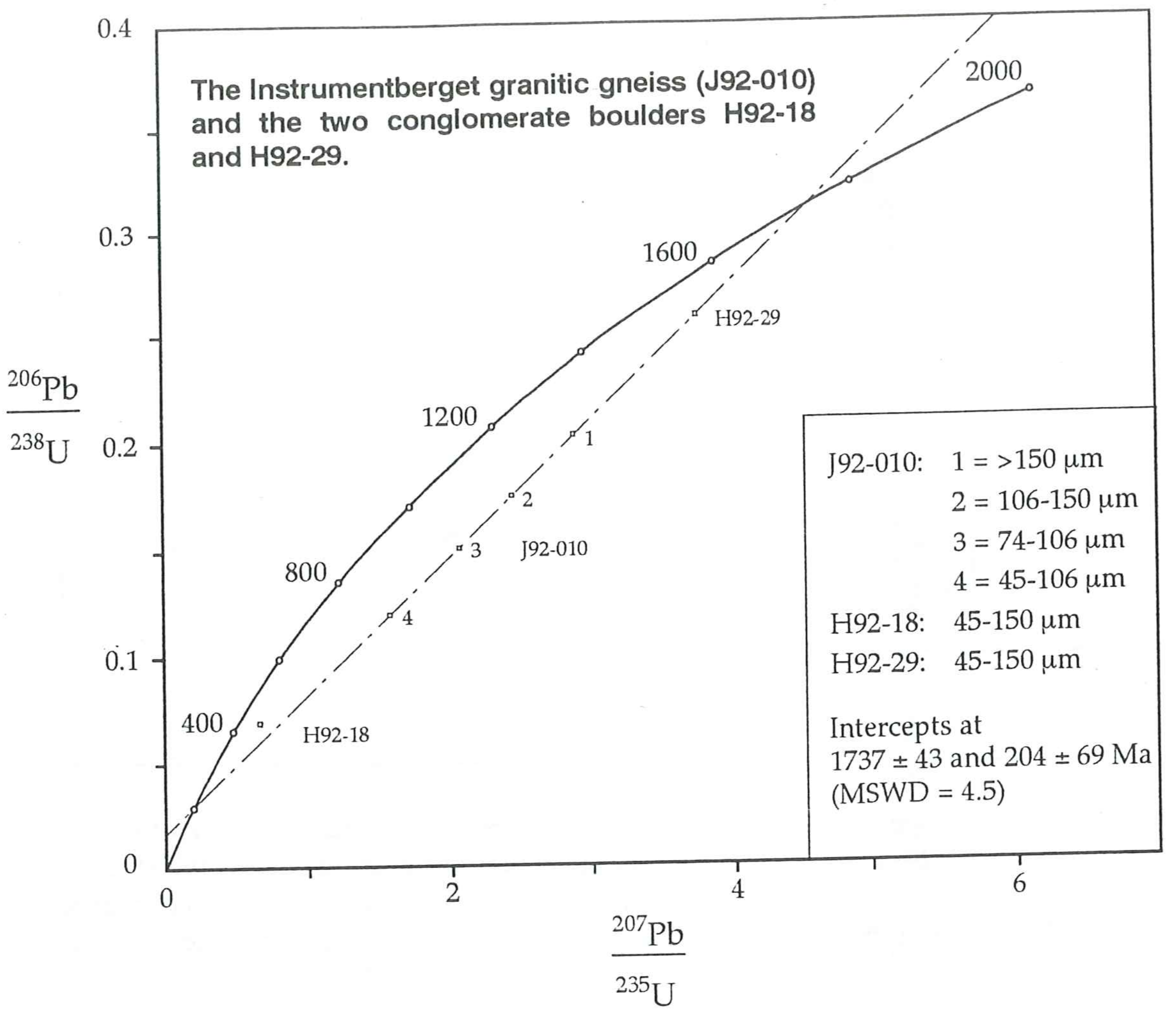

Fig 16. The intercepts for the Instrumentberget granitic gneiss (J92-010) are defined at 1737±43 Ma and 204±69 Ma. The clast sample H92-29 falls on the discordia for sample J92-010 and is interpreted to be of the same origin (see text); a recalculation including sample H92-29 gives intercept ages of 1735±14 Ma and 201±28 Ma (MSWD = 3.2). The clast sample H92-18 falls well off the discordia of sample J92-010.

5.3.1 The conventional U-Pb analyzes

For the multigrain U-Pb zircon analyses (Fig 16; Table 3) of the Instrumentberget Granitic Gneiss (J92-010), four zircon fractions was separated; they define a discordia line with good precision (MSWD=4.5). The slightly high MSWD is caused largely by the high precision of the data points and the high uncertainty of the upper intercept (1737 ±43 Ma) is mainly caused by the strong discordancy of the analyzed fractions. The $^{207}Pb/^{206}Pb$ ages of each fraction range from 1547 Ma to 1679 Ma. The single analysis of clast sample H92-29 plots less discordantly and right on the discordia line defined by the four data points from the Instrumentberget Granitic Gneiss; it yields a $^{207}Pb/^{206}Pb$ age of 1713 Ma. By contrast, the single U-Pb analysis of the zircons in clast sample H92-18 plots

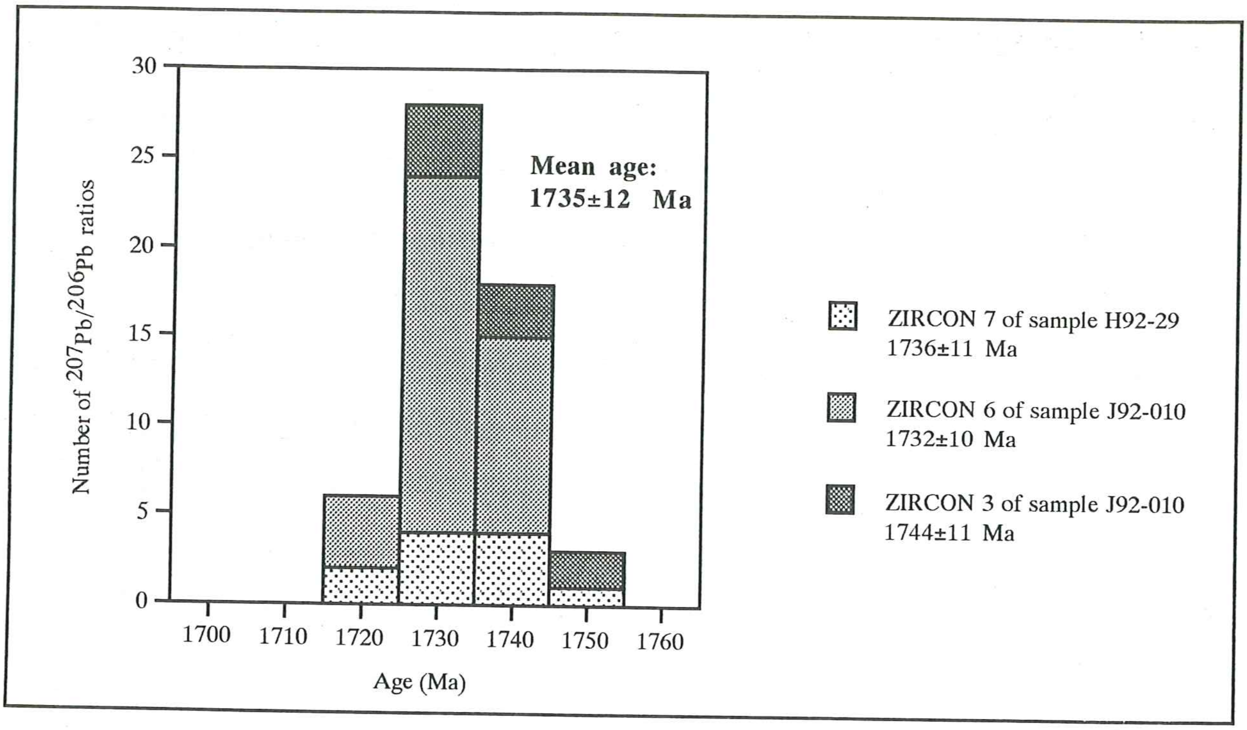

Fig 17. Distribution of radiogenic lead isotope ratios derived from evaporation of one zircon from conglomerate boulder (H92-29) and two zircons from the underlying granitic gneiss (J92-010). A common origin for the two samples is suggested, the mean age (1735±12) is calculated from all measured $^{207}Pb/^{206}Pb$ ratios of the three zircons and the age given for each zircon is the mean age of all measured ratios for that zircon.

significantly above this discordia line and gives a $^{207}Pb/^{206}Pb$ age of only 894 Ma. These zircons are extremely rich in uranium (3200 ppm) and common lead (52 ppm).

5.3.2 The Pb-Pb evaporation age determination analyses

Six individual zircon crystals from sample J92-010 gave $^{207}Pb/^{206}Pb$ evaporation ages ranging from 1550 to 1746 Ma (Table 4). Two of them provide maximum ages which agree with the upper intercept age (1737±43 Ma) of the Instrumentberget Granitic Gneiss (Fig 17). The only age obtained from sample H92-29 was 1736±11 Ma. Sample H92-18 yields ages from 734 to 933 Ma with low $^{206}Pb/^{204}Pb$ ratios which restrict their geological significance, but still are in agreement with the $^{207}Pb/^{206}Pb$ age of the convensional analysis (894 ± 9 Ma).

Table 3. U-Pb data for zircons from the granitic gneiss J92-010 and the two granitic boulders H92-18 and H92-29.

Sample	Fraction[1] (µm)	Weight (mg)	U (ppm)	Pb-rad (ppm)	Pb-com (ppm)	$\frac{^{206}\text{Pb}^{2}}{^{204}\text{Pb}}$	$\frac{^{207}\text{Pb}^{3}}{^{235}\text{U}}$	$\frac{^{206}\text{Pb}^{3}}{^{238}\text{U}}$	$\frac{^{207}\text{Pb}^{3}}{^{206}\text{Pb}}$	$\frac{^{207}\text{Pb}}{^{206}\text{Pb}}$ - age (Ma)	Error[4] (corr.)
J92-010	>150	0,95	470	100,8	1,9	2990	2,888±8	0,2033±2	0,1030±2	1679±4	0,573
	106-150	0,74	536	98,1	2,8	1970	2,420±10	0,1741±2	0,1009±4	1640±7	0,593
	74-106	1,02	547	87,0	2,8	1710	2,094±10	0,1519±2	0,1000±4	1624±8	0,608
	45-74	1,18	681	84,8	4,1	1170	1,590±11	0,1202±1	0,0960±6	1547±12	0,697
H92-18	45-250	0,81	3217	211	52	259	0,645±3	0,0680±1	0,0688±3	894±9	0,489
H92-29	45-250	0,89	460	124,5	0,7	9320	3,743±8	0,2587±2	0,1049±2	1713±3	0,524

note: 1. Non-magnetic at 1,6 A and 5 degrees side slope at Frantz Isodynamic Separator.

2. Corrected for fractionation (0,11 ± 0,04 % per AMU) and blank (0.02 ng U, 0.065 ng Pb).

3. The analytical errors are given at 2-sigma confidence level, ratio corrected for fractionation (0,11 ± 0,04 % per AMU) and blank (0.02 ng U, 0.065 ng Pb) and common Pb.

4. Error correction $^{207}\text{Pb} / ^{235}\text{Pb} - ^{206}\text{Pb} / ^{238}\text{U}$.

Table 4. Pb/Pb data from zircons using the Kober evaporation method of zircons from samples J92-010, H92-29 and H92-18. The ages are corrected for common lead and the analytical usertainty are given in 1σ.

Sample	Zircon	T (°C)	Number of ratios	$\frac{^{207}Pb}{^{206}Pb}$	$\frac{^{206}Pb}{^{204}Pb}$	Age (Ma)
J92-010	1	1300	-	No signal		
		1450	26	0,0995739	4123	1550±4
	2	1350	-	No signal		
		1450	18	0,1014138	8453	1618±4
	3	1450	-	No signal		
		1490	16	0,1048976	13560	1694±7
		1560	9	0,1072806	36300	1744±11
	4	1450	-	No signal		
		1650	18	0,1046584	7025	1673±12
	5	1400	-	No signal		
		1500	18	0,1048364	4620	1658±8
	6	1450	9	0,1036885	11461	1670±9
		1500	18	0,1059995	77732	1728±5
		1520	18	0,1063566	207745	1737±6
H92-29	7	1350	-	No signal		
		1550	9	0,1065369	60246	1736±11
H92-18	8	1400	9	0,1269261	232	734±213
		1450	17	0,1132913	336	933±145
		1500	9	0,0864615	690	784±96
	9	1360	9	0,0987978	478	880±100

5.4 Discussion of the isotopic results

The results of the isotopic analyses strongly suggest that the granitic gneiss boulder (H92-29) is derived from the underlying Instrumentberget Granitic Gneiss (J92-010). The first and perhaps the strongest evidence is that the single data point from the U-Pb analysis of the clast (H92-29) plots right on the discordia line of sample J92-010. Further indications for a close genetic relationship are the similarities in zircon morphology as well as the uranium and radiogenic lead content (Table 3). In addition, a common origin is supported by the fact that the single-zircon ^{207}Pb/^{206}Pb-age for H92-29 (1736 Ma) coincides with the upper intercept age (1737±43 Ma) of J92-010. Therefore, it may be justified to include the single data point from the U-Pb analysis of H92-29 in the calculation of the J92-010 intercept age. After recalculation, the upper intercept age decreases slightly to 1735 ± 14 Ma and the lower intercept decreases to 201 ± 28 Ma. The age uncertainties are strongly reduced and the MSWD becomes 3.2 (Fig 16). The significance of the upper intercept age for J92-010 is supported by the complementary single-zircon analyses of both sample J92-010 and H92-29. The mean ^{207}Pb/^{206}Pb-age of these analyses is 1735±12 Ma and is regarded as the true intrusion age of the granitic body (Fig 17).

5.4.1 The single zircon analyses of sample J92-010 and H92-29

The principal idea of single-zircon evaporation analyses is to sequentially evaporate different Pb-components at elevated temperatures from a single zircon grain embedded in a Re-filament. Already at lower temperatures, extraneous lead in cracks, and on surfaces as well as radiogenic lead sited within metamict domains is removed, while radiogenic lead localised in unaltered closed domains require significantly higher temperatures (c. 1500° C; Cocherie et al. 1991, Kober 1986; Kröner and Todt 1988) to evaporate.

In the single-zircon analyses of the gneiss sample (J92-010), zircons 1, 2, 4 and 5 give lower ^{207}Pb/^{206}Pb-ages than the upper intercept age, provided by the multi-grain U-Pb analyses (Table 4). The relatively low $^{20\overset{6}{7}}$Pb/$^{20\overset{4}{6}}$Pb ratio during these analyses suggests that at least a part of the radiogenic lead is coming from metamict domains. In zircons 1, 2 and 5 it was not possible to evaporate lead at temperatures over 1500° C; these ages are therefore considered to reflect discordant lead from metamict parts and thus lack geological significance. Zircon 4 did not provide a higher age even though a high evaporation temperature (around 1650° C) was used; this is explained by the fact that the lead was collected during the whole temperature interval between 1400° and 1650° C, resulting in a mixing of "pure" radiogenic lead released at higher temperatures with discordant lead released at lower temperatures. In contrast to the single-zircon analyses for zircons 1,2,4 and 5 from sample J92-010, zircons 3 and 6 as well as zircon 7 from sample H92-29 yield maximum ages (Fig 17) that agree well within the analytical uncertainty of the upper intercept from the U-Pb multigrain analysis. The analysis of zircon 6 at 1520° C and 1500° C gave the most precise ^{207}Pb/^{206}Pb ages with ^{206}Pb/^{204}Pb ratios of >74000; these ages are interpreted to correspond to the true zircon age. Finally, zircon 3 from sample J92-010 and zircon 7 from sample H92-29 yields analytically acceptable ^{207}Pb/^{206}Pb-ages of 1744±11 Ma and 1736±11 Ma, in good agreement with zircon 6, referred to above.

5.4.2 The U-Pb analyses of samples J92-010 and H92-29

All the data points of the U-Pb analyses plot dicordantly in the concordia diagram; various explanations are possible. Episodic lead-loss in zircons during a thermal metamorphic event is a well known mechanism, often argued for when discordant data points are discussed. If the lead loss occurred during a single thermal event, the age for this event is given by the lower intercept (201±31 Ma) in the concordia diagram (Wetherill 1956). ^{40}Ar/^{39}Ar ages of muscovite and hornblende from amphibolites in the Mosselbukta area provide evidence for Caledonian metamorphism and a cooling age (below 350° C) in Silurian time (about 415 Ma; Gee and Page in press) indicating that no regional metamorphic event occurred at

200 Ma in this area. Another explanation, postulated by Tilton (1960) is the continuous diffusion of lead from a zircon crystal at a rate governed by a diffusion constant that is dependent on the effective diameter of the crystal and the concentration gradient of lead in the zircon. This model results in a fictitious lower intercept age without any geolgical significance, but still can produce a straight line discordia with an upper intercept corresponding to the crystallisation age of the zircon. In the Ny Friesland analyses described here, it is impossible to say if the zircons have suffered continuous lead loss or not. A third interpretation of discordancy of U-Pb dates is the dilatancy model, explained by radiation damage due to alfa decay of U, leading to the formation of microcapillary channels that permit water to enter the zircons (Goldich and Mudrey 1972). The water is held tightly in the zircons until uplift and erosion reduce the pressure on the minerals, allowing the water to escape together with dissolved radiogenic Pb. This model suggests that the lower intercept at 201±31 indicates the time of uplift and erosion when the zircons begun to loose lead. A last possibility that can disturb the daughter to parent ratios and produce discordancy is chemical weathering (Stern et al. 1966). In recent lead loss due to chemical weathering, the zircons in the concordia diagram are supposed to lie on a straight line that starts from the discordia line and leads to the origo. In our analyzes we have not found any evidence for such recent lead loss.

5.3.3 Isotopic analyses of sample H92-18

The single U-Pb analysis of the boulder H92-18 plots significantly above the discordia line from the Instrumentberget Granitic Gneiss. This, together with a very different character and significantly higher radiogenic lead, common lead and uranium content of these zircons compared to zircons from the other two samples, strongly suggests that this conglomerate boulder is not derived from the underlying Instrumentberget Granitic Gneiss. Sample H92-18 yields ~900 Ma ages from the Pb-Pb evaporation analyses with low $^{206}Pb/^{204}Pb$ ratios (< 735). This means that it is likely that ^{207}Pb and ^{206}Pb exclusively is coming from metamict domains. The conventional U-Pb analysis show these zircons to be very ”dirty”, i.e. high in both uranium and common lead (and consequently yielding a low ^{206}Pb/^{204}Pb ratio), but gives a similar $^{20\overset{7}{6}}$Pb/$^{20\overset{6}{4}}$Pb age of 894 ± 9 Ma. This agreement makes a Grenvillian age of c. 900 for this boulder possible. Granites of similar age (~950 Ma) have been found on Nordaustlandet (Gee et al. in prep). If so, it would constrain the age of the Mosselvatnet conglomerate to be younger than 900 Ma.

6. SUMMARY AND CONCLUSIONS

The mapping in the Instrumentberget area (Fig 9) has shown the stratigraphy to be in the right order, with the Instrumentberget Granitic Gneiss unconformably overlain by a meta-sedimentary unit, the Polhem Formation. The basal part of this formation, unit including the Mosselvatnet conglomerates and impure meta-arkoses, has previously been referred to the Rittervatnet Formation, but we conclude from the new mapping that it is a member of the Polhem Formation. The isotopic work on the Instrumentberget Granitic Gneiss and clasts from the Mosselvatnet Conglomerate support the interpretation that the stratigraphy is the right way up,with strong indications that some of the clasts are derived from the Instrumentberget Granitic Gneiss. The crystallisation age (1735 Ma) of the Instrumentberget Granitic Gneiss is well demonstrated by both conventional multigrain U-Pb analyses (Fig 16) and single zircon Pb-Pb evaporation analyses (Fig 17). A Mesozoic or even younger age of the Polhem Formation is possible. This hypothesis is supported by single zircon Pb-Pb evaporation analyses of one conglomerate boulder that yields c. 900 Ma ages, with a high uncertainty caused by large common-lead correction, and one single U-Pb data point that gives a $^{207}Pb/^{206}Pb$ age of 894 Ma.

~~[illegible]~~ Paleozoic Middle Proterozoic

?

Future work in this area should include more isotopic age-determination studies of conglomerate clasts to test the hypothesis that it includes clasts of an younger generation than the Instrumentberget Granitic Gneiss. Another intresting problem to work out is the structural and stratigraphic relationships between the Instrumentberget Granitic Gneiss and the overlying granitic gneisses of Bangenhuk Formation, that are apperently of the same age.

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