Aeromagnetic and gravimetric interpretation of regional structural features in the Caledonides of West Finnmark and North Troms, northern Norway

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Interpretation of aeromagnetic and gravity data from Finnmark and North Troms has resulted in a new map of depth to Precambrian basement. The integrated interpretation of the data shows that the Seiland Igneous Province lies within a depression in the Caledonian allochthon and subjacent basement, to the southeast of which is an arcuate, elongate domal structure that includes the tectonic windows of Raipas Supergroup rocks. The Seiland plutonic rocks occur down to a depth of 7-8 km and define a gently NW-dipping disc-shaped body in the highest part of the Kalak Nappe Complex. The southwestern part of the Alta-Kvænangen window is essentially autochthonous; but some thrusting can be detected in the northeast. Further to the northeast, the Raipas rocks in the windows are more and more allochthonous and have yielded evidence of Scandian, greenschist-facies metamorphism. In the Repparfjord-Komagfjord area they define an antiformal structure of stacked thrust-sheets. The gravity and magnetic signatures of these rocks allow them to be followed northeastwards beneath the Kalak Nappe Complex, where they form a basement horse estimated to be at least 2.5 km thick. The two Raipas windows are situated within an elongate dome in the Precambrian basement, suggesting that gravitational sinking of the Seiland Igneous Province has contributed to local uplift around its rim. There can be few situations in an orogenic belt which involve greater gravitational disequilibrium in the upper crust than a 7 km-thick, dense, mafic igneous complex emplaced onto a continental margin. The negative buoyancy stress resulting from the emplacement was 20 MPa or more, depending on how much of the province had been removed by erosion. It is demonstrated that the calculated buoyancy stress was sufficient for the rocks to deform at geologically significant strain rates.

Several major fault zones can be identified within the Proterozoic basement. The NE-SW Vestfjorden-Vanna Fault extends out onto the continental shelf where a negative, residual gravity anomaly, offshore from Seiland, may represent a sedimentary basin of Late Palaeozoic age. On Finnmarksvidda, the Mierujavri-Sværholt Fault shows a complex system of strike-slip faulting with duplex development of Proterozoic age. This fault can be traced northeastwards, and is detected beneath the metasediments of the Gaissa Nappe where it deflects and truncates the Levajok Granulite Complex. Further northeast, the fault appears to link with a major, NE-SW fault on the shelf west of Nordkinn, which bears evidence of Late Palaeozoic and Mesozoic movements.

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Introduction

The survey area embraces a wide span of geological provinces. The Caledonian Orogen within northern Troms and western Finnmark is covered, in addition to the northwestern part of the passive continental margin in the Norwegian Sea and the northernmost part of the Precambrian basement on Finnmarksvidda. This paper, however, deals mainly with the structural relationships between the Precambrian basement and the Caledonian nappes. Major structural elements on the continental shelf as well as in the Precambrian basement on Finnmarksvidda will be included because they extend into, or beneath, the Caledonian Orogen.

The Caledonian allochthon of northern Norway is dominated by thick metasedimentary successions which range in age from Riphean to Silurian. Incorporated in the tectonostratigraphy, and particularly in the Kalak Nappe Complex (Fig. 1, Plate 2), are thrust-sheets of Archaean and Proterozoic gneisses, Early

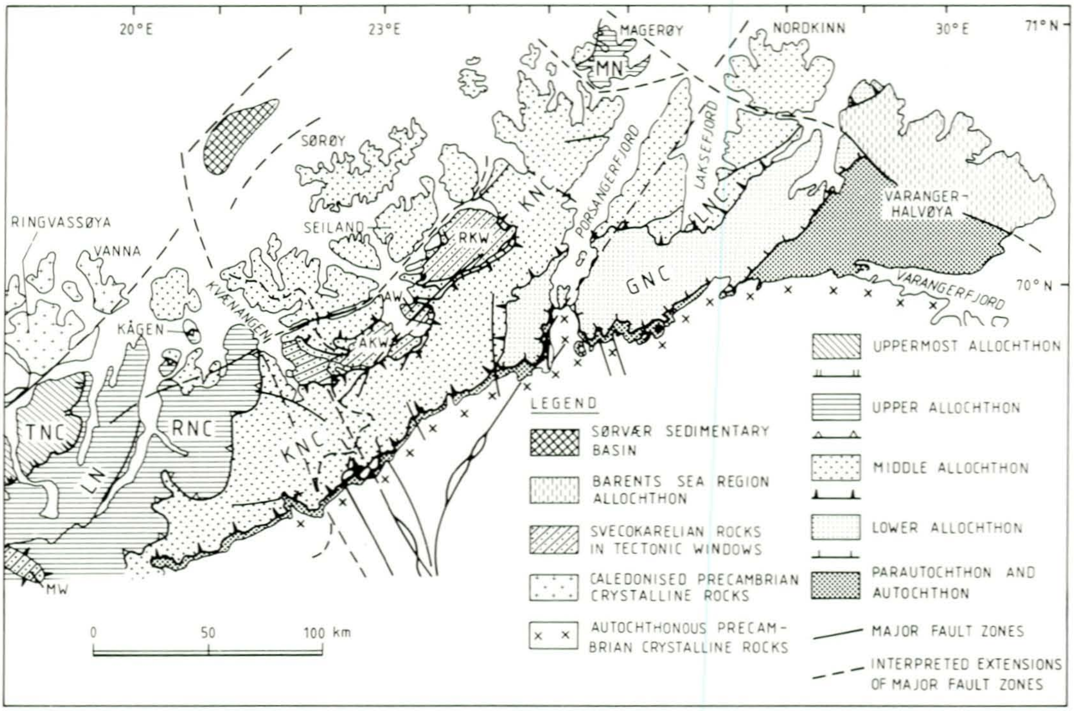


Fig. 1. Simplified tectonostratigraphy of the Caledonides of Finnmark and northern Troms. Modified from Roberts (1985). LNC — Laksefjord Nappe Complex; GNC — Gaissa Nappe Complex; KNC — Kalak Nappe Complex; MN — Magerøy Nappe; RNC — Reisa Nappe Complex; LN — Lyngen Nappe; TNC — Tromsø Nappe Complex. MW — Mauken window; AKW — Alta-Kvænangen window; AW — Altenes window; RKW — Repparfjord-Komagfjord window.

Proterozoic volcanosedimentary assemblages, and the ultramafic, gabbroic and alkaline rocks of the Seiland Igneous Province (SIP) (Roberts 1985). In northern Troms, volcanosedimentary rocks of Ordovician to probable Silurian age constitute the Reisa Nappe Complex (Zwaan 1988) and are tectonically overlain by the Lyngen Ophiolite.

Although generally considered synorogenic in relation to the early Caledonian 'Finnmarkian' polyphase deformation (Sturt & Ramsay 1965, Robins & Gardner 1975), the plutonic rocks of the SIP (also called Seiland Complex) have recently been the subject of reinterpretation following a study of basic dykes in SW Sørøy (Krill & Zwaan 1987). Rather than relating to mantle diapirism within a subduction zone (Ramsay 1973, Robins & Gardner 1975), the SIP magmatic rocks may have been generated in an extensional, rift-related situation (Krill & Zwaan 1987), and thus pre-date the main phase of Finnmarkian orogenesis. The Hasvik gabbro on Sørøy has yielded a Sm-Nd mineral isochron age of c. 700 Ma, while U-Pb zircon ages from minor granites which cut D2 structures in the Sørøy metasediments indicate intrusion at or before c. 800 Ma (Daly et al. 1990). Based on preliminary isotopic dating it has also been suggested (Krill et al. 1988) that the mafic-ultramafic Honningsvåg Complex on Magerøy (Robins 1987) (Plate 1) may possibly correlate with the Seiland rocks. While this suggestion is contentious (Andersen 1989), if upheld it would indicate that the original extent of these pre-Finnmarkian magmatic rocks was much greater than previously thought.

The mafic - ultramafic rocks of the allochthonous SIP are causing one of the most pronounced Bouguer gravity anomalies (+105 mGal) in northern Fennoscandia (Brooks 1970). According to Ramberg (1980, 1981), emplacement of dense rocks onto a less dense continental crust will cause gravitational instability accompanied by basement diapirism. One of the purposes of this contribution is to examine the possibility that basement diapirism may have been associated with thrust emplacement of the dense SIP. The density contrast (≈300 kg/m^3) and thickness (≈7 km) of the Seiland Complex are of the same order of magnitude as in the Semail Ophiolite of Oman which has been shown by Andrews-Speed & Johns (1985) to have caused basement diapirism. In the Nordland area of Norway, Ramberg (1980), Cooper & Bradshaw (1980) and Cooper (1985) have shown that gravity tectonics led to the development of mantled gneiss domes. The density contrast is significantly smaller in this case (70 kg/m^3) than in the Seiland region.

Regional geological setting

The Caledonian nappe succession

In a wider context of Scandinavian Caledonide tectonostratigraphy, most of the nappes and thrust-sheets recognised in Finnmark and northern Troms are part of the Lower and Middle Allochthons (Roberts & Gee 1985). The Upper Allochthon is represented in Troms in the form of the Reisa Nappe Complex, and in Finnmark by the Magerøy Nappe (Fig. 1). Details of the structural divisions are given in Ramsay et al. (1985), Roberts (1985), and Gayer et al. (1987). Here we outline only the main features of the principal nappe units.

Lying directly upon either the parautochthon or the autochthonous Dividal Group is the *Gaissa Nappe Complex* (GNC) (Roberts 1985), also informally termed the Gaissa Thrust Belt (Gayer et al. 1987). The GNC consists largely of fluvial and shallow-marine sediments of Vendian to Tremadoc age but also includes the Varangian tillites and a prominent unit of stromatolitic dolomites. Metamorphic grades, based on illite crystallinity studies (Bevins et al. 1986), are in anchizone, with highest grades occurring immediately beneath the basal thrust to the Kalak Nappe Complex (Rice et al. 1989 a,b).

Tectonic structural investigations in the GNC have revealed a complexity of fold-and-thrust deformation, with imbricate zones and duplex development (Townsend et al. 1986, Gayer et al. 1987). Based on attempted restoration of balanced cross-sections and branch-lines, a total displacement of some 150-165 km has been determined for the western trailing edge of the GNC, involving an internal shortening of up to 70% in some subunits. The thrust displacement vector for the Gaissa is towards ESE-E (Townsend et al. 1986).

East of Porsangerfjord the GNC is succeeded by the *Laksefjord Nappe Complex* (LNC) which is composed mainly of a sequence of clastic sediments of assumed latest Riphean to Cambrian age. A metamorphic grade of lowest greenschist facies is characteristic of these metasediments. Two phases of thrust displacement are known from the LNC; an early phase of ductile thrusting directed towards SSE, and a later more brittle phase with an ESE-E transport vector.

The *Kalak Nappe Complex* (KNC) (Fig. 1, Plate 2) is the most extensive and structurally complex of the Caledonian nappes, with a characteristic metasedimentary sequence of probable Late Proterozoic arkosic psammites, schists, metalimestones and turbiditic greywackes. Metamorphic mineral parageneses indicate a variation from greenschist to upper amphibolite facies, the highest grades generally occurring in the highest thrust-sheets. Lead isotope data from the Geitvann lead-copper mineralisation in the KNC of the Porsanger Peninsula have given a model age of about 940 Ma (Lindahl & Bjørlykke 1988). Isotopic dating (U-Pb, zircon, and Rb-Sr, whole-rock) of minor granitic bodies in two thrust-sheets in the KNC has indicated that the earliest deformation of these particular sediments occurred prior to c. 800 Ma (Daly et al. 1990).

An important element in the KNC is that of tectonic slices of older Precambrian crystalline rocks which in some places are unconformably overlain by the arkosic psammites (e.g. Ramsay & Sturt 1977). As in the case of the LNC, thrusting of the Kalak occurred in two main phases; an early and major ductile displacement, and a later more brittle thrusting (Gayer et al. 1987). Thrust displacement vectors are similar to those in the LNC.

Tectonostratigraphically above the KNC in the area considered here is the *Reisa Nappe Complex* (RNC) in Troms (Zwaan 1988) and the *Magerøy Nappe* (MN) in northern Finnmark (Andersen 1981). Metasediments dominate the successions and range in age from Ordovician to Early Silurian. In addition there are mafic/ultramafic rocks, including the Honningsvåg Complex on Magerøy, as well as syntectonic granitoids. The polyphase deformation in the RNC and MN is of Scandian (latest Silurian to earliest Devonian) age, the tectonothermal effects of which also extended down into the Kalak Nappe Complex (Sturt et al. 1967, Andersen et al. 1982, Dallmeyer 1988, Torsvik et al. in prep.).

In northwestern Troms, the *Lyngen Nappe* succeeds the RNC; this is composed of a basal gabbro/greenstone unit of ophiolitic affinity (now generally called the Lyngen Ophiolite) with an unconformably overlying volcano-sedimentary sequence containing Upper Ordovician and Silurian fossils (Minsaas & Sturt 1985, Andresen et al. 1985). The highest major tectonic unit in the region is that of the *Tromsø Nappe Complex* (TNC), a heterogeneous succession of schists, gneisses, amphibolites, anorthosites and local eclogites (Andersen et al. 1985).

The Seiland Igneous Province

Occupying a surface area of c. 100x70 km (Fig. 1) the SIP constitutes the largest magmatic province in the Norwegian Caledonides and forms part of the Sørøy-Seiland Nappe, the highest and most outboard, far-transported unit in the KNC. A large part of the southeastern margin of the SIP is bounded by the Vargsund Fault, a near-vertical structure of assumed Mesozoic age (Lippard & Roberts 1987).

Intrusive activity in the Seiland Complex evolved progressively from parental magmas of quartz-normative, low-K tholeiitic basalt through a calc-alkaline series to alkali-olivine basalt and peridotite (Robins & Gardner 1975). The final stages of magmatism are represented by an alkaline series of syenites, nepheline syenites and carbonatites (Sturt & Ramsay 1965, Robins & Gardner 1975). A minor but important component is that of basic dykes which intruded at several stages, ranging from pre-tectonic to clearly post-tectonic.

Numerous field studies have shown that the SIP magmas intruded during an important period of ongoing orogenesis, with regional fold and ductile mylonite development. This tectonism has usually been regarded as an integral part of the Finnmarkian orogenic event (e.g. Sturt et al. 1978) based on Rb-Sr isochron ages on several plutons and dykes in the SIP which range from c. 540 to 490 Ma. Recently, however, U-Pb ages of 531 ± 2 and 523 ± 2 Ma on zircons from late-phase nepheline syenite pegmatites from Seiland (Pedersen et al. 1989) strongly suggest that this syn-magmatic tectonism was of Middle Cambrian age or older, and not strictly equivalent to the 'Finnmarkian' as we know it from other parts of the Caledonides (cf. Roberts 1988). More zircon dates are required to help confirm this reinterpretation. In another interpretation of the SIP (Daly et al. 1990), two distinct episodes of magmatism are proposed, one synorogenic at or before c. 800 Ma and the other broadly 'inter-orogenic' in the period c.700 to c. 520 Ma.

The Raipas Windows and related rocks

Early Proterozoic (Karelian) rocks of the Raipas Supergroup occur in a series of tectonic windows penetrating the KNC (Fig. 1) — the Repparfjord-Komagfjord, Altenes and Alta-Kvænangen windows. In the Repparfjord-Komagfjord window a thin sequence of Vendian to Lower Cambrian sediments lies generally unconformably upon the Karelian rocks and below the basal Kalak Thrust (Pharaoh 1985). These sediments are correlative with those in the autochthonous Dividal Group. The c. 8 km-thick Karelian succession consists of greenstone lavas, tuffs and sediments, including stromatolitic dolomites (Pharaoh et al. 1983), cut by gabbros and serpentinised peridotites. The complete sequence of supracrustal and intrusive rocks was metamorphosed in greenschist facies at c.1840 Ma during the Svecokarelian orogeny (Pharaoh et al. 1982).

Opinions differ on the tectonic status of these Raipas windows; from autochthonous (Pharaoh et al. 1983) to allochthonous (Rhodes 1976, Gayer et al. 1987, Dallmeyer et al. 1988). The Vendian-Cambrian sediments which lie upon the Raipas rocks around and within the Komagfjord window are metamorphosed in lower greenschist facies and strongly folded and thrust-faulted — in sharp contrast to their diagenesis-grade Dividal Group correlatives which crop out just 40 km to the southeast beneath the GNC (Bevins et al. 1986). To the southwest, the grade of Caledonian metamorphism has decreased to high anchizone in the Altenes window (Rice et al 1989a). $^{40}Ar/^{39}Ar$ whole-rock ages on pelites have shown that this tectonothermal event is of Scandian generation and occurred c.425-420 Ma ago (Dallmeyer et al. 1988).

The origin of elongate doming connected with the windows has no unique solution. Possible causes include late-Caledonian open folding, basement diapirism, and the formation of an antiformal stack during thrust-sheet emplacement, or a combination of these. Townsend (1986) has shown that, in the Repparfjord-Komagfjord window, two Caledonian structural units may be recognised below the Kalak Thrust, a duplex structure in the west (Porsa Duplex) and a basement horse to the east. Together, these units constitute what has been termed the Komagfjord Antiformal Stack (Townsend 1986). Thus, a component of the present doming of the KNC was almost certainly imposed initially by the thrust-sheet stacking and duplex development in the form of a 'culmination' (Dahlström1970). As we will demonstrate below, basement diapirism most likely enhanced this thrust-related culmination.

Although detailed structural studies have not yet been carried out in the Altenes and Alta-Kvænangen windows, work by Fareth (1979) has shown that at least the western part of the Altenes Raipas and its Vendian cover sediments was strongly folded and imbricated in Caledonian time. Another feature of interest is that Raipas stratigraphies in the three windows differ quite appreciably, suggesting that juxtaposition may have been tectonically controlled (Gayer et al. 1987). Pratt (1989) has argued for thrusting of the Alta-Kvænangen Raipas above the Altenes, but the evidence for this is not compelling.

The Precambrian basement on Finnmarksvidda and in Troms

The Archaean to Early Proterozoic basement on Finnmarksvidda consists primarily of supracrustal sequences, generally referred to as greenstone belts, and of gneiss complexes (Siedlecka et al. 1985). The Kautokeino Greenstone Belt in the west is separated from the eastern Karasjok Greenstone Belt by the Jer'gul Gneiss Complex. Westernmost Finnmarksvidda is subdivided into the Raisædno Gneiss Complex and the Njallajåkka Complex, which is a minor greenstone belt. The Tanaelv Migmatite Complex and the Levajok Granulite Complex, the northern end of the Lapland Granulite Belt, have been thrusted from the east upon the Karasjok Greenstone Belt. Most of the rocks within the greenstone belts are of Early Proterozoic age. The Gåldenvarri and Vuomegielas Formations may be of Archaean age.

In western Troms, Precambrian crystalline basement is exposed on the islands Senja, Kvaløy, Ringvassøy and Vanna and in the Mauken window (Fig. 1) (Fareth 1981). This 'western gneiss region' comprises orthogneisses, migmatitic gneisses, amphibolites and some metasediments and metavolcanic rocks. Little isotopic dating is yet available from this region, but the indications are that the bulk of the gneisses pre-date the Svecofennian cycle (Andresen 1980). Caledonian thermal effects, recorded on Senja (Cumbest & Dallmeyer 1985), embrace both Early Ordovician and Early Devonian, post-metamorphic, $^{40}Ar/^{39}Ar$ mineral cooling ages. These Precambrian rocks are considered to belong to the Parautochthon of Caledonide tectonostratigraphy.

Geophysical data

Petrophysical data

Approximately 5700 rock samples, collected during geological mapping and the follow up of geophysical anomalies (Olesen & Solli 1985, Midtun 1988) and during a regional petrophysical sampling programme financed by Norsk Hydro and Statoil (Olesen 1988a) have been measured with respect to density, susceptibility and remanence. The measuring procedure is described by Olesen (1985) and Torsvik & Olesen (1988). Most of the samples are from Finnmark, and the sample locations are shown in Fig. 2. The data are stored in the national petrophysical database (Olesen & Sæther 1990) at NGU and the results for the main rock units are shown in Table 1. Q-values, the ratios of remanent to induced magnetisation, are reported rather than NRM intensity. The Q-value is not calculated if the susceptibility is below .00150 SI, since the accuracy of remanence measurements is poor for samples with low induced magnetisation and in aeromagnetic interpretation it is only necessary to quantify the remanence for ferromagnetic rock samples, i.e. susceptibilities above c.0.00150 SI.

The frequency distributions of the most common rock units are shown in Fig. 3. With the exception of Late Proterozoic - Early Palaeozoic mafic and ultramafic intrusive rocks and Archaean - Proterozoic gneisses and amphibolites within thrust-sheets, the Caledonides of northern Troms and Finnmark are practically non-magnetic. The frequency distributions in

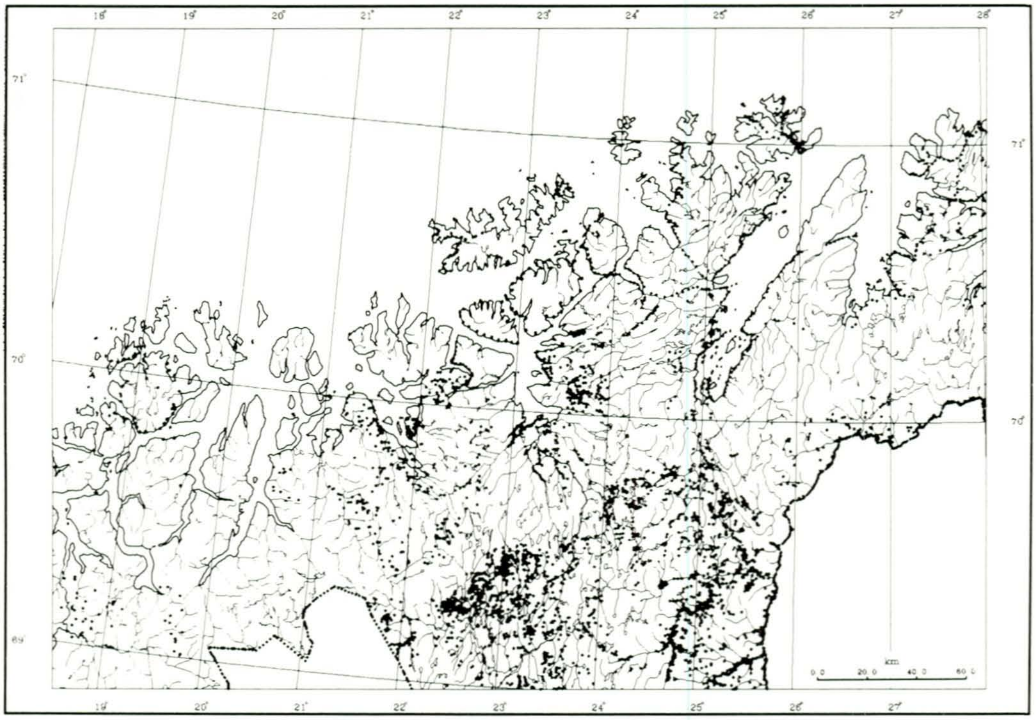


Fig. 2. Sample locations, 5,700 rock samples measured with respect to density, susceptibility and remanence.

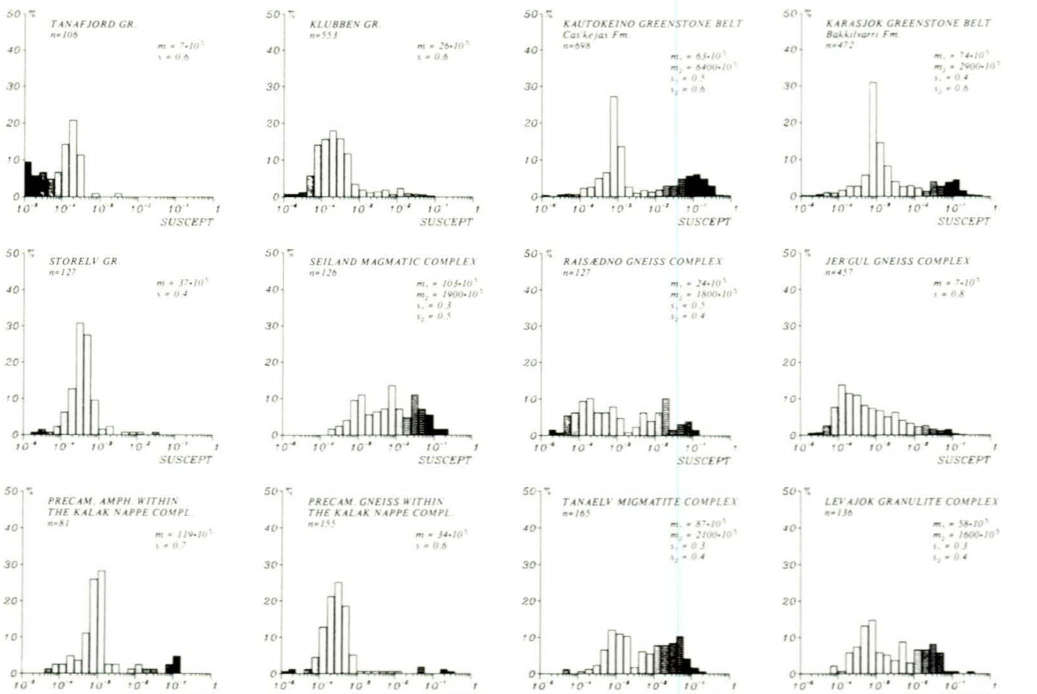


Fig. 3. Susceptibility spectra of major rock-units in northern Troms and western Finnmark. Logarithmic mean values (m) and standard deviations (s) expressed in decades are given in SI units. For rock units with both low- and high-magnetic fractions, mean value and standard deviation for both fractions are calculated.

ROCK UNIT		NO	DENSITY MIN	DENSITY MAX	DENSITY MEAN	DENSITY STD	SUSCEPTIBILITY MIN	SUSCEPTIBILITY MAX	SUSCEPTIBILITY MEAN	SUSCEPTIBILITY STD	NO	Q-VALUE MIN	Q-VALUE MAX	Q-VALUE MEAN	Q-VALUE STD
GAISSA NAPPE															
Tanafjord Gr. (metased.)	a	98	2584	2856	2699	73	.00001	.00207	.00006	.58					
Porsanger dolomite	a	14	2752	2849	2817	27	.00001	.00015	.00003	.36					
LAKSEFJORD NAPPE COMPL.															
All rock types	a	14	2535	2810	2693	81	.00001	.00050	.00013	.50					
KALAK NAPPE															
All rock types (except SIP)	a	965	2484	3224	2733	115	.00001	.35022	.00033	.64	59	.02	22.29	.29	.17
Klubben Gr. Metasandstone (psammites)	a	553	2484	3224	2692	73	.00001	.09544	.00026	.63	43	.02	4.66	.18	.17
	b	507							.00019	.40					
	c	46							.01117	.38					
Storely Gr. Metapelites	a	127	2613	3085	2753	80	.00001	.03190	.00037	.43	6	.41	12.21	1.41	.11
	b	123							.00033						
	c	4							.01233	.28					
Precamb. amphibolites within the Kalak Nappe Complex	a	81	2659	3187	2962	137	.00001	.15311	.00119	.74	5	.62	22.29	3.00	.14
	b	69							.00066	.39					
	c	12							.03634	.53					
Precamb. gneiss within the Kalak Nappe Complex	a	155	2563	3060	2739	111	.00001	.35022	.00034	.64	5	.10	.43	.22	.04
	b	145							.00025	.34					
	c	10							.03551	.67					
Eidsvågeid Gneiss Compl. All rock types	a	60	2613	3122	2783	122	.00002	.05743	.00035	.50	3	.38	.97	.54	.04
	b	57							.00029	.31					
	c	3							.01787	.36					
SEILAND MAGMATIC COMPL. Gabbro, peridotite, syenite	a	126	2593	3438	3025	185	.00021	.24627	.00621	.73	88	.07	20.53	.71	.38
	b	48							.00103	.28					
	c	78							.01880	.45					
MAGERØY NAPPE Metasediments	a	36	2602	2943	2777	61	.00001	.00303	.00031	.38					
	b	35							.00029	.34					
	c	1							.00303	.00					
HONNINGSVÅG COMPLEX (gabbro)	a	24	2684	3116	2937	94	.00034	.02427	.00193	.51	14	.26	104.83	1.55	.52
	b	15							.00088	.26					
	c	9							.00720	.24					
REISA NAPPE COMPL. All rock types	a	98	2578	3182	2795	121	.00002	.40981	.00042	.60					
	b	91							.00031	.36					
	c	7							.01899	.61					
KVÆNANGSTINDAN GABBRO	*a*	*14*	*2888*	*3072*	*2977*	*44*	*.00043*	.00880	.00070	.31					
LYNGEN NAPPE All rock types	a	4	2773	3074	2913	119	.00041	.14768	.00370	1.02					
TROMSØ NAPPE COMPL. All rock types	a	3	2775	2832	2798	24	.00054	.00155	.00089	.18					
RAIPAS SUPERGROUP															
Kvenvik Fm./Nussir Gr. Metavolcanics/-sediments	a	121	2578	3124	2912	113	.00001	.58309	.00149	.87	14	.04	1.71	.13	.14
	b	106							.00069	.37					
	c	21							.07410	.66					
Holmvatn Gr. Metavolcanics/-sediments	a	32	2636	3215	2812	133	.00013	.08867	.00056	.57	4	.20	.43	.28	.04
Saltvatn Gr. Quartzite	a	18	2621	2728	2669	29	.00005	.02628	.00135	.88	8	.05	.80	.23	.23
	b	11							.00033	.51					
	c	7							.01224	.30					
BASEMENT ROCKS IN TROMS All rock types	a	235	2512	3212	2746	152	.00001	.31511	.00107	.91	66	.03	59.83	1.82	.37
	b	166							.00034	.52					
	c	69							.01664	.42					
KAUTOKEINO GREENSTONE BELT All rock types	a	1290	2486	3390	2857	154	.00001	.84037	.00176	1.04	98	.02	14.74	.22	.14
Cas'kejas Fm. Metavolcanics/-sediments	a	698	2540	3390	2898	130	.00001	.84037	.00340	1.05	82	.03	7.11	.21	.16
	b	444							.00063	.37					
	c	254							.06378	.51					
KARASJOK GREENSTONE BELT All rock types	a	1130	2460	3400	2858	177	.00001	.92197	.00159	.98	230	.01	5.87	.27	.22
Bakkilvarri Fm. Metavolcanics/-sediments	a	472	2540	3400	2930	163	.00001	.51770	.00214	.84	102	.04	5.87	.28	.21
	b	336							.00074	.37					
	c	136							.02941	.57					
RAISÆDNO GNEISS COMPL. Gneiss, migmatite, amphibolite	a	127	2550	3070	2703	134	.00001	.15184	.00122	1.02	17	.08	4.34	.24	.16
	b	79							.00024	.49					
	c	48							.01786	.41					
JER'GUL GNEISS COMPL.	a	457	2520	3140	2688	107	.00001	.18790	.00067	.81	72	.01	37.70	.21	.23
	b	363							.00031	.49					
	c	94							.01259	.45					
TANAELV MIGMATITE COMPL.	a	165	2570	3636	2814	163	.00005	.24525	.00490	.79	69	.03	1.88	.24	.23
	b	76							.00087	.34					
	c	89							.02137	.41					
LEVAJOK GRANULITE COMPL.	a	136	2633	3256	2862	145	.00007	.26094	.00247	.81	22	.13	10.56	.61	.21
	b	77							.00058	.34					
	c	59							.01633	.41					

Table 1. Statistical data; density, susceptibility and Q-value for main rock-units in western Finnmark and northern Troms. The letters a, b and c denote total sample, low-magnetic fraction and high-magnetic fraction, respectively. Units are in SI. The standard deviation of susceptibility is expressed in decades. Susceptibility values have logarithmic mean values. An extended table is given in Olesen (1988a).

Fig. 3 show, however, that the Precambrian rocks are commonly highly magnetic. The bimodal distribution for the supracrustal rocks in the Kautokeino and Karasjok Greenstone Belts is clearly seen. The diagrams show one low paramagnetic component and one high ferromagnetic, the latter mostly caused by magnetite (usually more than 1 %). The distributions of the susceptibility for the igneous rocks and the gneisses are broad and unimodal, but these rocks also have a large proportion of highly magnetic lithologies.

ROCK DENSITIES

	N	Density 10^3 kg/m^3	Density adapted for the gravity model 10^3 kg/m^3
SEILAND IGNEOUS PROVINCE		3.01*	3.01
Gabbro, pyroxenites, syenites, etc.			
(Brooks 1970)	129	2.98	
(Chroston 1974)	114	3.02	
(Olesen 1988)	126	3.03	
LYNGEN OPHIOLITE			2.97
Gabbro (Chroston 1972)	7	3.05	
Kjosen Formation, amphibolite			
(Chroston 1972)	7	2.97	
Metasedimentary rocks			
(Chroston 1972)	20	2.75	2.71
HONNINGSVÅG COMPLEX			
Gabbro (Lønne & Sellevoll 1975)	42	2.93	2.91
Metasedimentary rocks			
(Lønne & Sellevoll 1975)	49	2.73	2.71
KALAK NAPPE COMPLEX (lower nappes)		2.73*	2.71
Klubben Group		2.69*	
(Brooks 1970)	15	2.65	
(Chroston 1974)	16	2.70	
(Olesen 1988)	553	2.69	
Eidsvågeid Gneiss Complex		2.77*	
(Brooks 1970)	16	2.73	
(Chroston 1974)	21	2.78	
(present study)	60	2.78	
PRECAMBRIAN BASEMENT		2.69*	2.71
Jergul Gneiss Complex			
(Olesen & Solli 1985)	131	2.67	
(Midtun 1988)	130	2.71	
(Olesen 1988)	457	2.69	
Ringvassøya, Kvaløya, gneiss			
(Chroston 1974)	71	2.71	
Ringvassøya, amphibolite			
(Chroston 1974)	9	3.00	3.00

* denotes weighted average

Table 2. Rock densities applied in the gravity modelling, western Finnmark and northern Troms.

Table 2 shows density data applied in the gravity modelling; an average is calculated from published data from the Seiland Province and adjacent areas (Brooks 1970, Chroston 1974, Olesen & Solli 1985, Midtun 1988, Olesen 1988a). The weighting is based on the number of measured samples in the different surveys, and non-representative samples are excluded. The average density of the basement (2.69 * 10^3kg/m^3) is lower than that of the rocks in the lower parts of the KNC (2.73 * 10^3 kg/m^3). The density of the SIP rocks is 3.01 * 10^3 kg/m^3. The density contrast between SIP and the surrounding rocks with an average density of 2.71 * 10^3 kg/m^3 is consequently about 300 kg/m^3. The density of the Karelian metavolcanites and metasediments overlying the Archaean gneiss basement has not been taken into account when computing the density of the Precambrian basement. The reason for this is that the volume of these rocks is small as compared with the SIP and the gneisses within the Precambrian basement.

Aeromagnetic data

The aeromagnetic measurements in this region were carried out in the period 1959-1972. Troms and the area northwest of the Caledonian front and to the east of Hammerfest in Finnmark were flown at constant altitudes above sea level of 800 m and 1500 m, respectively (Plate 1). The remaining areas, including Finnmarksvidda and the Kvænangen-Alta-Seiland region, were drape-flown at an altitude of 150 m. The line spacing was 500 m in the Karasjok region and 1 km in the remainder of Finnmark, with the exception of the Alta-Seiland region and Troms county which were flown with a spacing of 2 km. On the continental shelf, the line spacing was approximately 4 km with a flight altitude of 200 m (Aalstad 1970). Printed maps in the scale of 1:250,000 have been published (Nor. geol. unders. 1971, 1972a,b 1976, 1982a). Maps in the scale of 1:50,000 have been digitised into 500x500m grid cells and the Definite Geomagnetic Reference Field 1965 has been subtracted. The final map shown in Plate 1 is produced using the pseudo-relief technique (Kihle in prep.) with illumination from the east. This type of map enhances structural trends, lineations and contrasts not easily discernible in the conventional contour maps.

Gravity data

The present study is based on measurements from 11488 gravity stations (Table 3). During the field seasons 1978-1990, 4532 new gravity stations were established by NGU within western Finnmark and northern Troms using helicopter, snow scooter, car and boat for transportation. The complete Bouguer reduction (Mathisen 1976) of the gravity data has been computed using a rock density of 2670 kg/m^3. These gravity data and data from gravity stations from Norges geografiske oppmåling (1979a,b,c,d,e,f), Norges geografiske

Survey	Area	No of stations
Norges geol. unders. Gellein (1990a,b,c,d,e,f,g)	Northern Troms and western Finnmark	4532
Norges geografiske oppmåling (1979a,b,c,d,e,f)	Northern Troms and western Finnmark	1115
Norges geografiske oppmåling et al. (1971)	Offshore northern Troms and western Finnmark	1870
Oljedirektoratet/Amarok	Offshore Ringvassøya - Sørøya	2289
Brooks (1970)	Sørøya - Hammerfest	226
Chroston (1972, 1974)	Tromsø - Øksfjord	1220
Lønne & Sellevoll (1975)	Magerøya	163
S. Saxov	Bardu	73
Total		11488

Table 3. Gravity surveys compiled in the present study.

oppmåling et al. (1971), Brooks (1970), Chroston (1974), Brooks & Chroston (1974) and Lønne & Sellevoll (1975) are stored in the national gravity database at NGU. The International Standardization Net 1971 (I.G.S.N. 71) and the Gravity Formula 1980 for normal gravity have been used to refer the surveys to a common level. The variable areal distribution of the primary observations has been homogenised by extracting stations with a minimum spacing of 800 m from the original data-set. This reduced data-set (9805 stations) has been interpolated to a square grid of 1.5 x 1.5 km using the minimum curvature method (Briggs 1974, Swain 1976). The final grid is slightly smoothed using a 3 x 3 filter. Bouguer anomaly maps at scales of 1:500,000 (Olesen et al. 1990) and 1:250,000 (Gellein 1990a,b,c,d,e, f,g) are produced from this grid using the map production system by Kihle (in prep.). The maps are based on measurements taken both on land and at sea. The northwesternmost part of Plate 3 contains marine gravity measurements provided by the Norwegian Petroleum Directorate. The overall coverage of the area surveyed is about 1 station per 9 km^2. Plate 3 shows the residual gravity map after a regional field is subtracted. The computation of the regional field is described below. The contour interval of the residual field in Plate 3 is 2 mGal which is believed to be larger than the error of the Bouguer gravity data. The locations of the gravity stations are shown on the residual map and on an index map in Plate 3.

Methods of interpretation

Gravimetric interpretation

The Bouguer anomaly map is dominated by a paired long-wavelength anomaly trending parallel to the coast. The negative component is located on land and the positive component offshore. There are several techniques available for deriving a regional map, such as spectral factorisation, upward continuation or graphical smoothing. Each method may yield non-unique results (Gupta & Ramani 1980). For quantitative modelling purposes in Precambrian terrain, the residual based on the graphically produced regional field was shown by Gupta & Ramani (1980) to be most suitable. This method is empirical, clearly non-unique, but nevertheless superior since known surface geology and measured density values can be taken into account.

To estimate the regional gravity field, five Bouguer gravity profiles located perpendicular to the regional field were extracted from the database. On each profile the level of the regional field was visually estimated by considering the influence on the gravity field of localised geological features. The regional gravity map shown in Fig. 4 was obtained by interpolating between these profiles. The method is described in detail by Skilbrei (1988).

Model calculations have been carried out along two profiles which cross the SIP perpendicular to each other. Two such orthogonal profiles also help in determining the correct regional field. When computing the gravity response to a model, we have used a modified version (Hesselström 1987) of the computer programme by Enmark (1981). The basic model in the programme comprises 2 1/2 dimensional bodies, i.e. bodies of polygonal cross-section with the tails cut off in the strike direction.

Aeromagnetic interpretation

The aeromagnetic map is shown in Plate 1. A prominent pattern in northern Troms and Finnmark is that of anomalies continuing from the Precambrian basement in the southeast across the Caledonides to the offshore continental shelf. This pattern is mostly perpendicular to the main structural trend of the Caledonides. The Precambrian rocks causing these anomalies are exposed to the south of the Caledon-

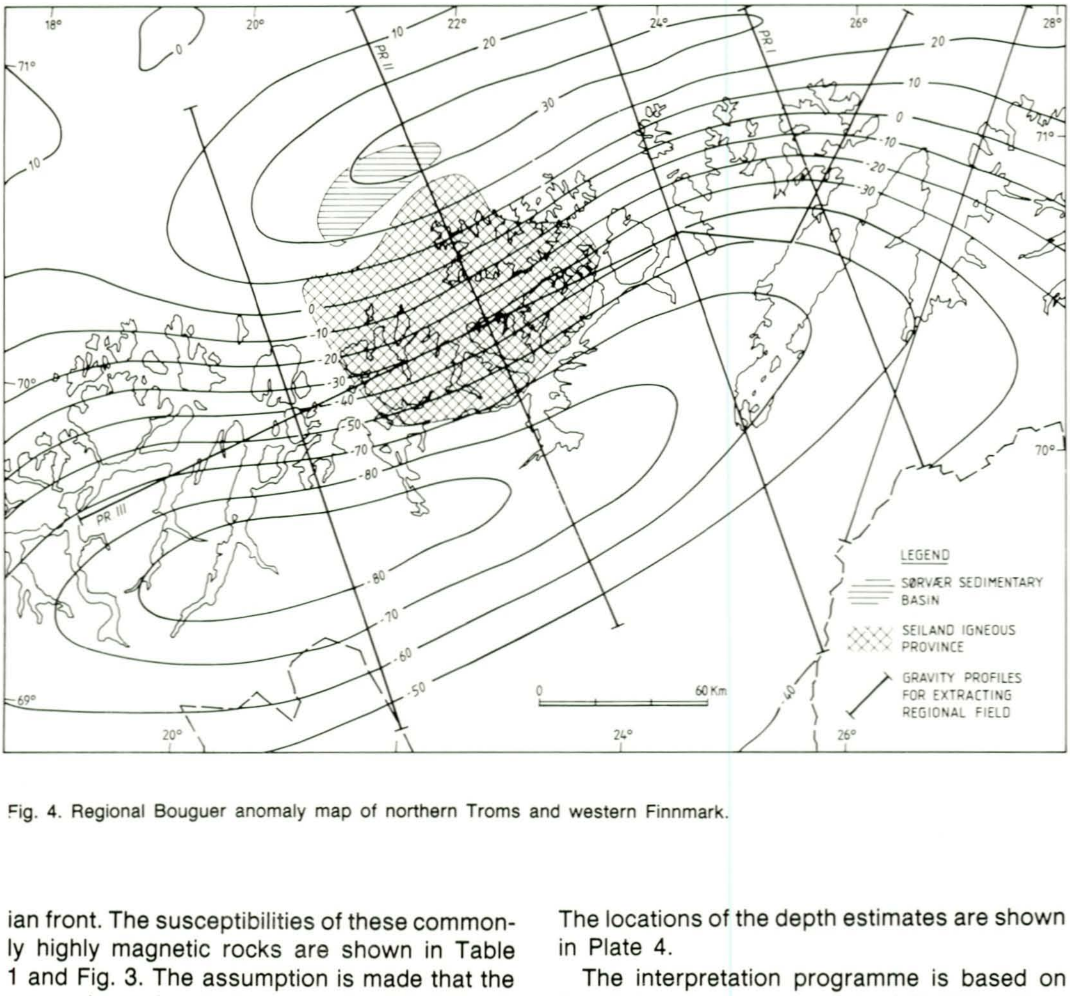


Fig. 4. Regional Bouguer anomaly map of northern Troms and western Finnmark.

ian front. The susceptibilities of these commonly highly magnetic rocks are shown in Table 1 and Fig. 3. The assumption is made that the magnetic rocks continue to the top of the Precambrian basement beneath the Caledonides and this is supported by geophysical interpretations of the basement area on Finnmarksvidda which show that the anomalies are caused mainly by the outcropping rocks (Olesen & Solli 1985). Thus, when extending this observation beneath the Caledonides, a depth to magnetic basement map of this region can be constructed. The interpretation has been made along profiles located perpendicular to the main trend of the magnetic anomalies. The equidistant profiles are extracted from an aeromagnetic database consisting of a 500 m by 500 m grid. The data interval along the profiles is 500 m. Previous interpretations made by Åm (1975) between the Alta-Kvænangen window and Finnmarksvidda, and by the Geological Surveys of Finland, Norway and Sweden (1986) in Troms, are incorporated. The locations of the depth estimates are shown in Plate 4.

The interpretation programme is based on the autocorrelation method of Phillips (1975, 1978, 1979) and the version used has been provided by Thorning (1982) of the Geological Survey of Greenland. The magnetic basement is defined as a two-dimensional surface (Fig. 5) constructed from a large number of very thin vertical 'dykes'. The method assumes that every one of these dykes extends to infinity in directions perpendicular to the profile, as well as vertically downwards. The upper termination of the dykes is the basement surface. This depth can vary from dyke to dyke. The dykes placed next to one another give the topography of the magnetic basement. It is further assumed that each dyke has a magnetisation which may differ from that of the adjoining dyke. The depth is estimated by passing a short window along the magnetic profile, estimating a depth for each position of the window. The width of the window var-

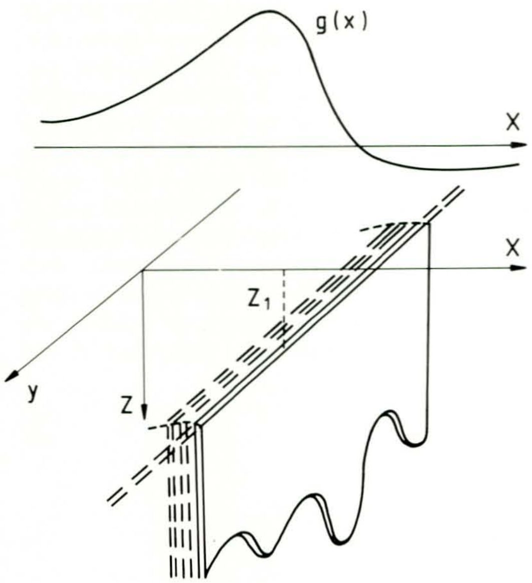


Fig. 5. The geometry for a thin sheet source, and its magnetic anomaly g(x) (after Phillips 1975). The source is of infinite extent in the ±y and +z directions. It terminates at a depth z_1 below the observation plane.

ies from 5 to 25 km depending on the wavelength of the anomalies. It has to be assumed that the anomaly within such a given window originates entirely from sources at a certain depth.

The depth expressions are:

$$(1)\quad Z=\frac{n\Delta x}{2}\sqrt{\frac{1}{1/\Phi_n-1}}$$

$$(2)\quad Z=\frac{\Delta x}{2}\sqrt{\frac{2n+1}{\Phi_n/\Phi_{n+1}-1}-n^2}$$

where Δx is the sampling interval, n is the number of intervals in the autocorrelation lag and Φ is the autocorrelation function.

The depth can be estimated from a value of the single autocorrelation at a single lag (1). In practice, the first lag (n=1) is used to estimate depth, while higher lags (n=2,3,4) are used to check the validity of the estimate. A second solution (2) can be expressed in terms of the autocorrelation at two successive lags. Sources at different depths can be separated using this formula, i.e. anomalies caused by deep and shallow bodies in the same profile.

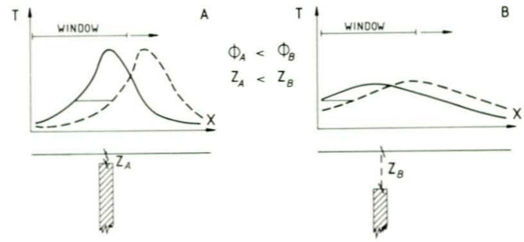


Fig. 6. Interpretation principle of the autocorrelation method. (A) A shallow magnetic dyke causes a narrow anomaly. The correlation of the waveform with itself is consequently small. (B) A deep-seated dyke causes a long-wavelength anomaly with a higher degree of autocorrelation. The depth is estimated by passing a short window along the magnetic profile and estimating a depth for each position of the window.

Fig. 6 illustrates the principles of the interpretation method. A shallow magnetic dyke causes a narrow anomaly. The correlation of the waveform with itself is consequently small. A deep-seated dyke causes a long-wavelength anomaly which will have a higher degree of autocorrelation.

Fig. 7 shows a residual E-W magnetic profile (A) across the Porsanger Peninsula where two anomalies interfere with each other. The profile (B) below the magnetic profile in Fig. 7 shows the depth synthesis of the profile. The depths are shown in kilometres below the observation plane. The lowermost curve (C) indicates the reliability of the depth estimates. The difference, in metres, between the first and the fourth lag is shown. If the difference is within the thresholds given by the dashed lines, the depth synthesis is acceptable. This is indicated by a continuous line for the depth estimate curve (B). It is worth noting that the depth estimate is accepted where the interference between the two anomalies is smallest, i.e. on the outer flanks of the anomalies. The depth to the magnetic basement is 2.7 km below sea level for both anomalies (3.5 km below the flight altitude). Interpretation of the same profile using the forward 2 1/2 dimensional programme gives the same depth (Profile III in Fig. 9). A weighted average of measured susceptibility values (fractions b and c in Table 1) for greenstones in the Kvenvik Formation/Nussir Group and the correlative Cas'kejas Formation (Siedlecka et al. 1985) within the Kautokeino Greenstone Belt has been applied in the interpretation.

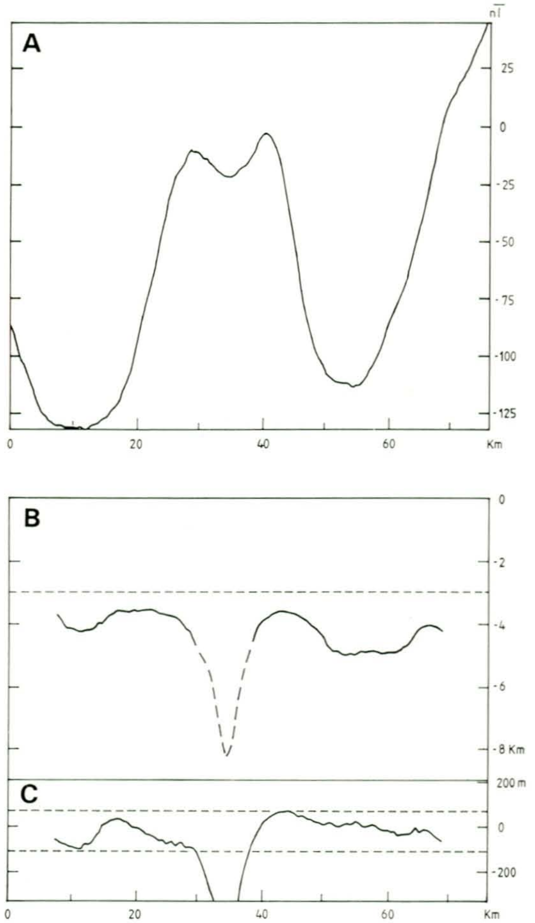


Fig. 7. (A) Residual magnetic profile across the Porsanger Peninsula. (B) Depth synthesis in kilometres of the residual field. (C) The reliability in metres of the depth estimates.

Fault zones occurring in the magnetic rock units, i.e. in the SIP and the Precambrian basement, are further interpreted from the aeromagnetic map. These fault zones are characterised by (Henkel 1975, 1984): 1. Linear discordances in the anomaly pattern. 2. Displacement of reference structures. 3. Linear magnetic gradients. 4. Discordant linear minima.

An interpretation of magnetic pattern and contacts is also included in Plate 4. Magnetic patterns are identified by the following criteria (Henkel 1975, 1984): 1. Banded pattern — representing continuous and parallel magnetic anomalies. 2. Dyke-like pattern — representing continuous and discordant magnetic anomalies. 3. Irregular anomaly pattern - areas lacking either of the previously mentioned features.

The analysis makes it possible to separate areas of supracrustal rocks from areas of plutonic rocks and gneisses in the Precambrian terrain. The interpretations have been made at the scale 1:100,000 on Finnmarksvidda, 1:250,000 in western Finnmark and 1:500,000 in the remaining areas. Aeromagnetic data along profiles have also been modelled using the same software package as used in the gravimetric interpretations described in the previous section.

Results

Regional gravity field

The obtained regional gravity map in Fig. 4 is dominated by the paired anomaly parallel to the coast. The negative component of this anomaly is aligned approximately along the axis of the highest elevations of the mountains in the area, while the maximum is located 30 km offshore. Brooks (1970) propounded three different hypotheses for the cause of the regional anomaly: 1. Zone of basic and ultramafic intrusions within the upper crust. 2. Granulitic facies lower crustal rocks approaching abnormally near to the surface. 3. Shallow depth to Moho under the coastal area. The paired ano maly is now thought to be caused mainly by Moho topography due to Airy-type isostatic compensation (Balling 1980, Dyrelius 1985), combined with thinning of the continental crust towards the continental shelf (Elo et al. 1989). Chroston et al. (1976) interpreted a crust thinning from 45 km to 38 km along a refraction seismic line from Storfjord in northern Troms to Dønnesfjord on Sørøy. The northernmost section of the 'Fennolora' profile (Lund 1987) crosses the survey area from North Cape to Kautokeino. Shotpoint H is located 10 km offshore to the west of North Cape. The Moho depth obtained from this shotpoint recording southwards is 45 km. At shotpoint G, 50 km south of Kautokeino, the corresponding depth is 46 km. These depth estimates do not support the existence of a root under the mountain chain in Finnmark. The calculated depths may, however, have errors of some kilometres due to the effect of the low-velocity layers in the upper crust (Lund 1987). The crustal models at shot points G and H show extremely high apparent velocities in the upper crust and several velocity reversals. Lateral density variations may therefore be partially responsible for the regional gravity field. From the 'Polar' profile (Luosto et al. 1989) the Moho is found at 47 km in the Karelian Province, 150 km south of Kautokeino, and rises to 40 km beneath the Lapland Granulite Belt 100 km southeast of Karasjok.

Smoothed depth to Precambrian basement map

The interpolated and contoured depth to Precambrian basement map of West Finnmark and North Troms is shown in Plate 4. The contour interval is 1 km. This map is based on depth to magnetic basement with additions from gravity interpretations in the Seiland, Porsanger Peninsula and Lyngen areas. The depth to the magnetic basement is quite shallow in the Målselv and Alta areas, consistent with the existence of basement windows in these regions. In the Porsanger-Laksefjord area the depth increases fairly gradually to the northwest at an angle of 5-6°.

Special problems arise in the SIP due to the presence of highly magnetic rocks within the Caledonian nappes. It is not possible to estimate the depth to a basement from aeromagnetic measurements in this area. This is also a problem, but not to the same extent, in the case of the smaller Lyngen and Kvænangstindan gabbros. In the Seiland area, other methods must be used to estimate the depth to the Precambrian basement. Another major problem relates to the difficulties arising from the probability that parts of the Raipas tectonic windows constitute thrust-sheets.

The depth to magnetic basement in the offshore areas is significantly greater than that obtained from drill-holes and reflection seismics. The likely reason for this is that the magnetic basement represents the Precambrian basement while the acoustic basement reflects the top of the Caledonian nappes. The magnetic anomalies selected for estimates of the basement in Plate 4 have, in most cases, an orientation perpendicular to the Caledonian trend, and are thought mostly to reflect Precambrian greenstones and amphibolites. Some of them may, however, also represent gneisses and intrusive rocks. Åm (1975) based his shallow magnetic basement in the Barents Sea mostly on small-amplitude anomalies which were thought to represent low-magnetic rocks within the Caledonides. Consequently, he arrived at a more shallow basement in this region as compared with the calculations and interpretations reported here.

Major basement faults

The importance of regional aeromagnetic and gravimetric anomaly patterns in detecting fault displacements is particularly great in areas dominated by extensive glacial deposits, such as Finnmarksvidda (Olesen & Solli 1985). The patterns are equally valuable to structural interpretations in this external part of the Caledonides where the non-magnetic, sandstone-dominated nappes form a relatively thin, allochthonous, 'transparent' skin above the variably magnetic Proterozoic basement (Plate 2).

Two of the most prominent, NE-SW trending faults in this region are the Vestfjorden-Vanna (VVF) and Mierujavri-Sværholt Faults (MSF) (Fig 8, Plate 4). The former, described by Forslund (1988), has a cumulative oblique displacement consisting mainly of a 4 km sinistral strike-slip component and a minimum of 2 km dip-slip. The vertical component is estimated from the presence of 2 km of non-magnetic Caledonian nappes to the southeast of the fault. The sinistral component can be detected from the displacement of magnetic and gravimetric anomalies. The offset of the gravity anomaly representing the Ringvassøya amphibolite is the most distinct evidence of an accumulated sinistral displacement.

The MSF extends from Mierujavri, 30 km north of Kautokeino, northeastwards through Masi, Iesjav'ri and Lakselv, and then further northeast, beneath the Caledonian nappes, to the Sværholt Peninsula (Fig. 8). At its southwestern termination it is truncated by the NNW-SSE trending faults interpreted by Olesen & Solli (1985). Berthelsen & Marker (1986) and Henkel (1987, 1990) include these faults in the regional Baltic-Bothnian megashear (BB) (Fig. 8) and Bothnian-Seiland shear zone, respectively. From the aeromagnetic data based on low-altitude measurements covering western Finnmarksvidda (Nor.geol. unders. 1988, Olesen et al. in prep.), a complex system of strike-slip contractional duplexes can be delineated along the MSF from Mierujavri to Sko-

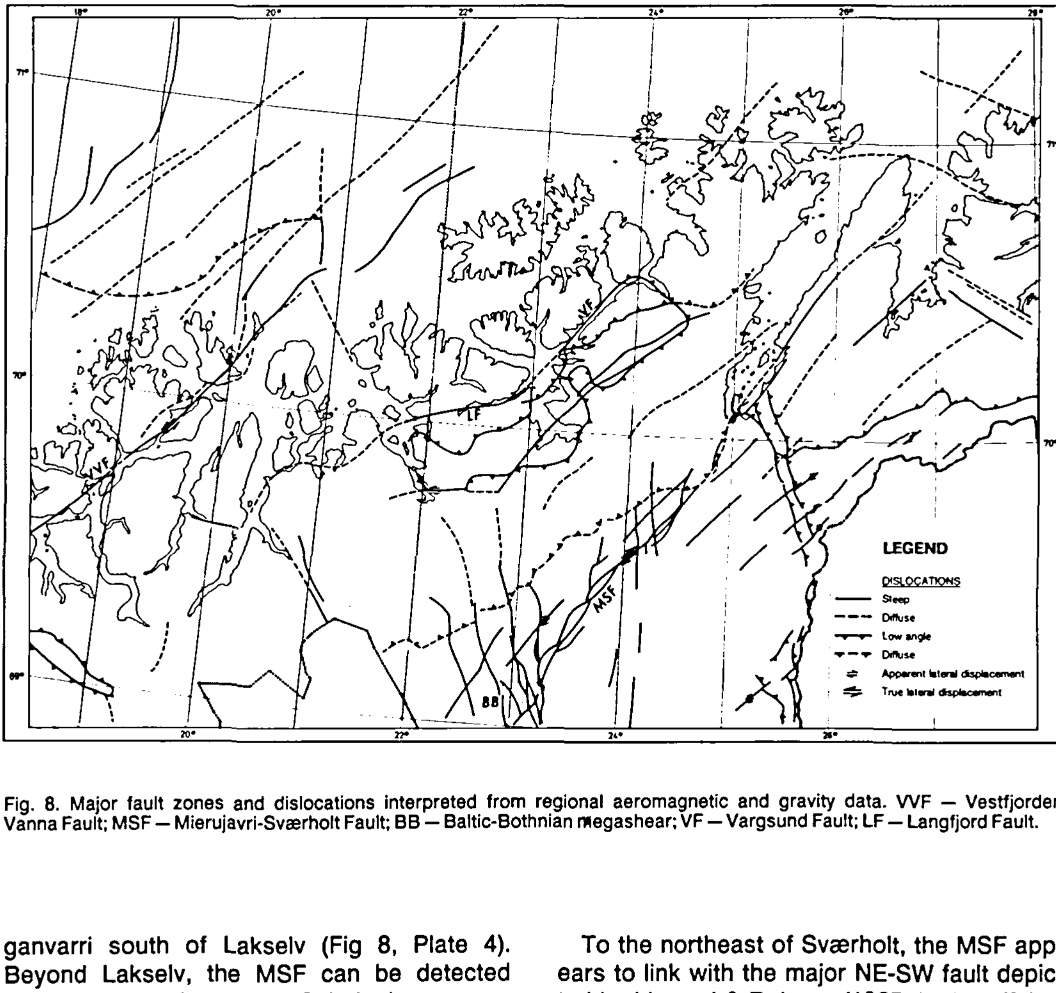


Fig. 8. Major fault zones and dislocations interpreted from regional aeromagnetic and gravity data. VVF — Vestfjorden-Vanna Fault; MSF — Mierujavri-Sværholt Fault; BB — Baltic-Bothnian megashear; VF — Vargsund Fault; LF — Langfjord Fault.

ganvarri south of Lakselv (Fig 8, Plate 4). Beyond Lakselv, the MSF can be detected beneath the various, thin, Caledonian nappes where it can be seen to deflect and truncate the Levajok Granulite Complex. This truncation is particularly pronounced on the aeromagnetic map (Plate 1), but can also be observed on the residual gravity map (Plate 3) and on the interpretation profile I (Fig. 9), and is considered to be of Proterozoic age coeval with, or just slightly younger than, the emplacement of the Early Proterozoic Levajok Granulite Complex. The Archaean to Early Proterozoic Gåldenvarri and Vuomegielas Formations, which have been correlated by Siedlecka et al. (1985), have been dextrally displaced approximately 20 km along the MSF. The 1815±24 Ma albite diabases in Masi (Krill et al. 1985) have intruded the MSF. Locally, younger deformation of these diabases can be observed. The diabases cause the characteristic magnetic anomalies along the MSF (Plate 1).

To the northeast of Sværholt, the MSF appears to link with the major NE-SW fault depicted by Lippard & Roberts (1987) in the offshore area just west of Nordkinn, which shows a cumulative downthrow to the northwest, from Late Palaeozoic and Mesozoic movements, of over 1.5 km. The Proterozoic MSF may have been reactivated and acted as a ramp during the Caledonian thrusting of the Munkavarri Imbricate Zone within the Gaissa Nappe Complex on the Sværholt Peninsula. On the Skoganvarre map-sheet, Townsend et al. (1989) mapped a post/late-Caledonian displacement which cuts the Gaissa Thrust, and a syn-sedimentary movement during the deposition of the Cambrian Dividal Group. In the Mierujavri-Lakselv area, a minimum downthrow of 1 km to the northwest has been recorded (Solli 1988), and there are also components of post-glacial movement. The 80 km-long post-glacial Stuoragurra Fault (Olesen 1988b) is situated within the MSF and comprises mainly reverse-fault movements. The MSF is, consequently, an exceptionally long-lived fault zone.

The BB, within the survey area, can be followed from Kautokeino to Kvænangen (Fig. 8). The fault complex is 30 km wide and consists of 3-4 major fault zones accompanied by duplex structures. In the Kvænangen area the BB is distorted by the Caledonian deformation in the Raipas windows. It seems likely that there is an offshore extension of the BB, but the exact location of these faults is difficult to determine because of the smoothing of the magnetic anomalies on the shelf.

The Raipas windows

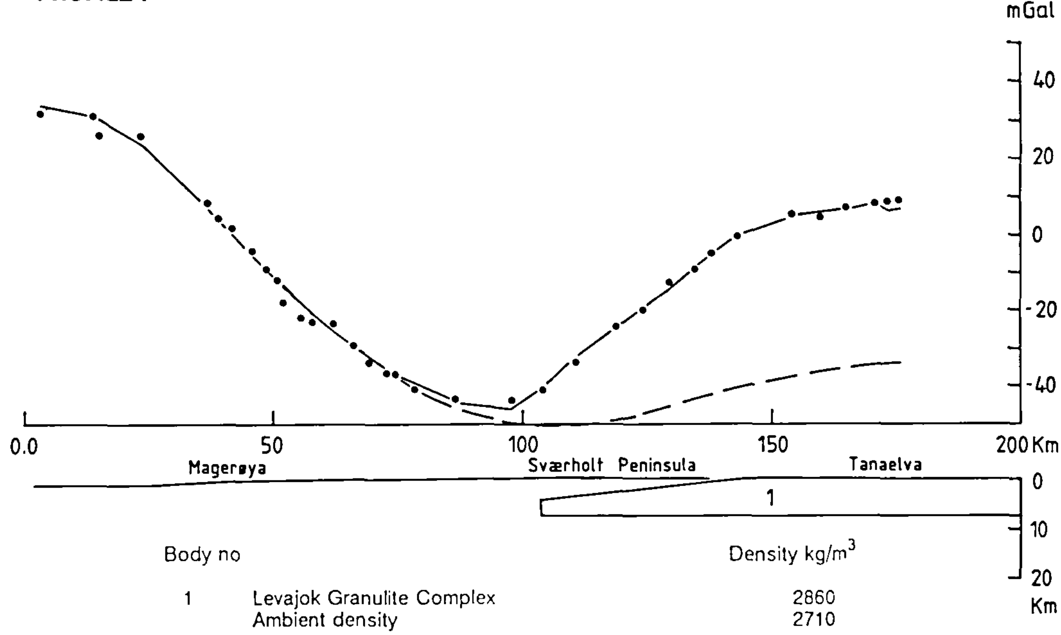
The precise status of the Raipas tectonic windows within the Caledonian nappe sequence, as outlined earlier, is a matter of debate. Geological and geophysical investigations have shown that in the extreme southwest, the Alta-Kvænangen window is autochthonous or parautochthonous (Åm 1975, Berg & Torske 1986), while to the northeast the western part of the Repparfjord-Komagfjord window is allochthonous (Gayer et al. 1987). The continuation of the Svecokarelian volcanosedimentary sequences within the Kautokeino Greenstone Belt from Finnmarksvidda into the Alta-Kvænangen window is clearly seen from the aeromagnetic map (Plate 1) as well as the residual gravity map (Plate 3). There may, however, be a displacement in the order of a few kilometres below and along the southeastern border of the Alta-Kvænangen window. This apparent displacement consists of a sinistral strike-slip and a vertical component (Åm 1975) of c. 5 and 2 km, respectively. The exact amount of detachment beneath the window is difficult to estimate but the southward transport does not appear to exceed a few tens of kilometres. The main reason for this assumption is seen in the very similar anomaly pattern (Plates 1 and 3) and stratigraphy within the window and in the area to the south (Zwaan & Gautier 1980, Vik 1981, Sandstad 1983, Solli 1984, Bergh & Torske 1986). The Vuomegielas Formation (Siedlecka et al. 1985) in the Iesjav'ri area on Finnmarksvidda can also be traced to the north, on the aeromagnetic and gravity maps (Plates 1 and 3), beneath the Caledonian nappes to the correlative Holmvatn Formation in the Komagfjord window. The trend of the anomaly, however, changes from N-S north of Iesjav'ri to NE-SW in the Komagfjord window.

The Altenes and Repparfjord-Komagfjord windows are characterised by NE-SW trending aeromagnetic and gravimetric anomalies reflecting the presence of thick greenstone units, but the stratigraphies in the two windows are not strictly comparable (c.f. Fareth 1979, Pharaoh et al. 1983). In the case of the Altenes window, the western half is strongly imbricated by steep, northwest-dipping, Caledonian thrust-faults. Although these appear to die out gradually southeastwards, the Dividal Group equivalent, the Rafsbotn Formation (Roberts & Fareth 1974), is itself dissected by Caledonian faults and strongly cleaved.

The Repparfjord-Komagfjord window, in Caledonian structural terms, constitutes the Komagfjord Antiformal Stack, comprising the Porsa Duplex and Rundfjell Basement Horse (Thrust Sheet) (Townsend 1986, Gayer et al. 1987), separated by a major fault zone. The imbrication in the west, in the Porsa Duplex, is somewhat similar to that seen in the western part of the Altenes window, suggesting that the latter segment, at least, is also allochthonous. The Porsa Duplex and Rundfjell Basement Horse may be represented by the large gravity and magnetic anomalies which can readily be followed northeastwards on to the Porsanger Peninsula (Plates 1 and 3, profile III in Fig. 9) beneath the KNC, moreso the Rundfjell segment. This latter unit must constitute a rock body of at least 2.5 km thickness, interpreted from the gravity anomaly. This may, in fact, correspond to the postulated Hatteras Basement Horse (HBH) of Gayer et al. (1987, Figs. 3 and 7) which is thought to have been accreted to the base of the KNC, and may overlie the Gaissa Nappe Complex. The depth down to this HBH thrust-sheet can be interpreted from aeromagnetic data (Fig. 7), and the thickness of the greenstones within the HBH can be interpreted from the gravity data (profile III in Fig. 9). The sum of the depth to the HBH thrust-sheet and the thickness of the thrust-sheet is a minimum estimate of the depth to the autochthonous Precambrian basement in this particular area.

Taking an overall view of the Raipas windows, the change of trend of the aeromagnetic anomalies from Kvænangen, where the trend is NW-SE to N-S, to Repparfjord, where it is NE-SW to locally ENE-WSW, is considered to be a result of Caledonian detachment.

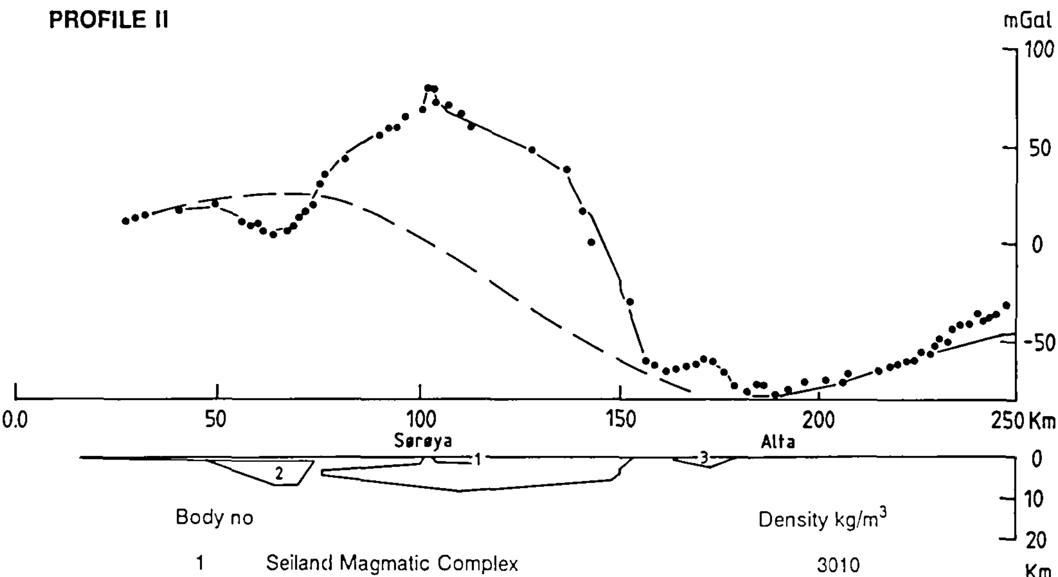
PROFILE I



Body no		Density kg/m³
1	Levajok Granulite Complex	2860
	Ambient density	2710

● Observed Bouguer anomaly
— — Regional gravity field
—— Calculated gravity effect

PROFILE II



Body no		Density kg/m³
1	Seiland Magmatic Complex	3010
2	*Palaeozoic sedimentary basin*	2550
3	Kvenvik Formation	2910
	Ambient density	2710

● Observed Bouguer anomaly
— — Regional gravity field
—— *Calculated gravity effect*

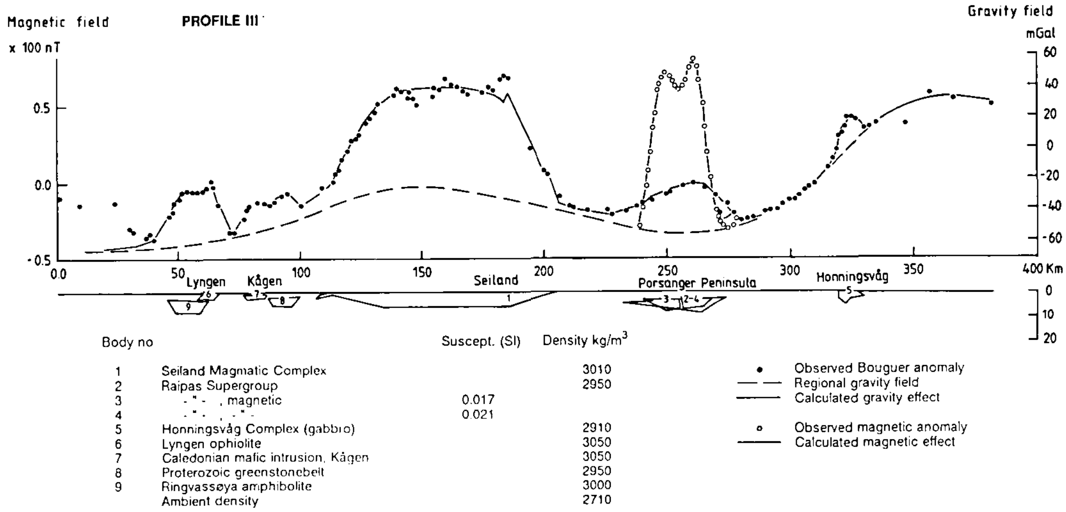


Fig. 9. Three gravity sections across the Caledonian Orogen in western Finnmark and northern Troms. Location of profiles I, II and III shown in Plate 3.

The southwestern parts of the Alta-Kvænangen window Raipas rocks are autochthonous or parautochthonous and one can detect and accept a greater degree of allochthoneity in moving, geographically, from southwest to northeast. Supportive evidence for this may be related to the fact that the grade of Caledonian (Scandian) metamorphism in cover sediments overlying the Raipas rocks around the windows, and correlated with the Dividal Group, decreases gradually from greenschist facies in the Repparfjord area to anchizone in the Alta-Kvænangen window (Rice et al. 1989b). One of the aeromagnetic anomalies within the eastern part of the Rundfjell Basement Horse is produced by a Svecokarelian dolerite. This intrusion is continuous from the Komagfjord window through the eastern part of the Altenes window and into the eastern part of the Alta-Kvænangen window. Since the latter window has to be autochthonous or parautochthonous, as discussed above, there was probably a significant component of clockwise rotation involved during the detachment of the Rundfjell Basement Horse to allow an incorporation of this unit into the base of the KNC at a comparatively late stage in the nappe transport.

An alternative to the interpretation proposed by Gayer et al (1987) based on balanced reconstructions is that the southeastern parts of the Raipas windows are parautochthonous, and that imbrication of the Gaissa Nappe took place to the northeast of its present position with the nappe being transported to the south by an anticlockwise rotation as proposed by Føyn (1967). In adopting this model the problem of having to allow the Gaissa Nappe to be transported beneath the Alta-Kvænangen window is avoided. Unfortunately, however, there is no minor structural evidence in the GNC to support this hypothesis.

Seiland Igneous Province and the adjacent offshore area

Plate 3 shows the residual gravity anomaly map from the Finnmark region. One of the most pronounced anomalies in northern Fennoscandia is a positive anomaly which largely coincides with the SIP. The amplitude of this anomaly is quite large: +75 to +105 mGal, depending on the choice of the regional field. The critical point in interpreting this anomaly is actually the determination of this regional field. The existence of a negative gravity anomaly to the northwest of the SIP, the most pronounced negative anomaly on the residual gravity anomaly map (Plate 3), complicates this process. A termination of the SIP at depth cannot explain the steep gradient offshore from Seiland. The most likely explanation is that a rock body exists which has a negative density contrast to the background density (2710 kg/m³) and that this body is situated immediately below the sea bottom. A less probable explanation involves the presence of an elongated body of background density surrounded on all four sides by rocks with a positive density contrast. When selecting the same amplitude for the regional field as for the remaining area in offshore West Finnmark, the regional field must be located above the negative anomaly as indicated in profile II in Fig. 9.

The most likely geological explanation for the negative gravity anomaly is either a sedimentary basin or a felsic intrusive body. The anomaly is elongated and aligned parallel to known sedimentary basins to the northwest, e.g. the Hammerfest Basin. It is also situated along the inferred seaward extension of the Vestfjorden-Vanna Fault, and a seismic profile (OD 2100/1973) crossing the negative anomaly does, in fact, indicate the presence of a sedimentary basin, which we propose being named the Sørvær Basin. The sediments in this basin may be of Palaeozoic age. Lower Carboniferous to Permian sediments occupy the area on a 1:3 mill. geological map compiled by Sigmond (in prep.). The quality of the seismic data, however, is poor due to strong sea bottom multiples and the depth of the Sørvær Basin cannot be determined. A density estimate of the sedimentary basin can be determined from well log data of Permian sediments from the petroleum exploration wells 7120/12-2 and 7120/12-4 provided by Norsk Hydro (M. Edvardsen pers. comm. 1989). These two wells are situated 40-50 km to the NNW and well 7120/12-2 penetrated the basement which consists of granitic gneisses (F. Riis pers. comm. 1990) at a depth of 4,600 m. The mean density of the sediments is $2.55*10^3$ kg/m^3 and gravity modelling of the anomaly using the resultant density contrast of 160 kg/m^3 gives a depth of 7 km. This estimate must be regarded as a maximum estimate since it is assumed that the basement comprises Klubben psammites and Precambrian gneisses. The presence of amphibolite or schists within the basement would increase the density contrast, and thus decrease the basin depth. Since the density estimate of the Sørvær Basin does not include Carboniferous sediments, there is an additional inaccuracy in the 7 km estimate. The model in profile II (Fig. 9) should therefore be regarded as a schematic tectonic model for the area offshore from Sørøya.

The maximum gradient of the residual gravity anomaly coincides with the borders of the SIP except to the northwest of Sørøy. The dense rocks of the SIP are therefore thought to cause the positive gravity anomaly. To the northwest, however, the SIP has to extend at depth beneath the metasedimentary rocks of the Kalak Nappe. The interpreted depth to the base of the SIP from the residual gravity anomaly is 7-8 km. A significant part of the crust in this area is consequently composed of mafic - ultramafic intrusive rocks. With the exception of its southern contact, the boundaries of the SIP are generally gently dipping (10-20°). In the south, the SIP is bounded by the Langfjord and Vargsund Faults, dipping at approximately 45° to the north. The shape of the SIP is consequently a gently northwestward inclined disc-shaped body tapering off gradually in all directions except to the south where it is cut by the Vargsund and Langfjord Faults. The northwestern part of the SIP therefore most likely represents the highest level within the complex. Local positive anomalies are superimposed on the main Seiland anomaly on profiles II and III (Fig. 9). These are mainly caused by density heterogeneities within the SIP. The peridotites, for example, have a mean density of 3164 kg/m^3 (n=11). There are also several, thick, intercalated units of psammite. On profile II (Fig. 9), the local positive anomaly is caused by the Breivikbotn gabbro on Sørøya. According to the gravity interpretations in profiles II and III there is no evidence for the Revsbotn Basement Horse as suggested by Gayer et al. (1987).

The distributions of the susceptibility for all major rock types within the SIP — gabbro, diabase, peridotite, hornblendite, syenite and carbonatite — are broad, usually more than two orders of magnitude. The high frequency anomalies occurring in an irregular pattern within the SIP are consequently caused by several of the intrusive rocks within the complex.

The tectonostratigraphic position of the SIP, within the KNC, requires the existence of a basal thrust. On the assumption that the Precambrian basement is situated directly beneath this thrust or just below an intervening thin thrust-sheet of metasediments (Gayer et al. 1987), the gravity depth interpretations can be incorporated into a map of depth to magnetic basement. Like the depth estimate on Porsanger Peninsula, this depth from gravity modelling is a minimum estimate for the depth to the Precambrian basement. A striking pattern emerges from this combined map of depth to Precambrian basement (Plate 4). The SIP is situated within a depression in the basement. An adjacent uprise curves around from the northeast via southeast to the southwest, suggesting that gravitational instability resulted in the local diapiric uplift following the thrust emplacement and subsidence of the SIP.

The 7-km thick igneous complex, situated in a high position in the nappe sequence, may have originally been a part of a more extensive complex of plutonic rocks. It is now preserved in the depression in the basement partly due to the gravitational subsidence following thrust-emplacement. The Raipas windows are situated within the flanking uprise, a feature which appears to support the idea that gravity tectonics has contributed to the development of these basement windows.

Gravity modelling across the Honningsvåg Complex on Magerøya using the density contrast 200 kg/m³ has indicated a depth of 6 km which is identical to the interpretation proposed by Lønne & Sellevoll (1975). This mafic intrusive complex was probably also detected on the 'Fennolora' profile, where a velocity as high as 6.4 km/s at a depth of 4 km has been recorded southwards across Magerøya from shotpoint H, 10 km to the west of North Cape (Lund 1987).

Lyngen-Kågen area

Gravity data from this area have earlier been interpreted by Chroston (1972, 1974). He pointed out the problem existing in the northern part of the Lyngen Peninsula where the gravity anomaly from the Lyngen Ophiolite is superimposed on the gravity anomaly arising from the extension of the Ringvassøya amphibolite at depth. Additional gravity readings have, however, made it possible to identify a short-wavelength and a long-wavelength anomaly in profile III. Using the density data from Chroston (1972, 1974), a model calculation has been made (profile III in Fig. 9). The depth extent of the Lyngen Ophiolite is 2-3 km. The Ringvassøya amphibolite extends from a depth of 2-3 km down to 5-7 km. These results have also been incorporated in the map of the depth to Precambrian basement (Plate 4).

A similar problem to that on the Lyngen Peninsula arises on the island of Kågen. To model the gravity anomaly in this area, an amphibolite, which must be an extension of the amphibolite on Finnmarksvidda, is included in the calculation. The upper surface of this body at 2 km depth is interpreted from depth to magnetic basement calculations. This body, however, cannot explain the existence of a neighbouring short-wavelength anomaly which, in the profile, is interpreted as a dense body within the Caledonian nappe sequence. The applied density is the same as for the SIP. The interpretation of this body is therefore rather schematic.

Strain caused by negative buoyancy stress

Consider a dense mafic complex B overlying a less dense basement A with densities ρ_B and ρ_A. The negative buoyancy stress on the overlying denser complex B of thickness h is (Molnar & Gray 1979, Andrews-Speed & Johns 1985):

$$\sigma_b = gh(\rho_A - \rho_B)$$

From this expression, the negative buoyancy stress on the SIP overlying less dense nappe metasediments and continental crust can be calculated. As previously shown the density contrast between the SIP and the Precambrian basement is 300 kg/m³. Applying the calculated thickness of 7 km yields a negative buoyancy stress of c. 20 MPa.

Handin & Carter (1980) have demonstrated that high-temperature creep is predominantly steady-state. The experimentally determined flow for practically all crustal and upper mantle rocks can be described by a power law relationship (Kirby 1983). The strain rate $d\varepsilon/dt$ can be related to the applied differential stress (σ) by:

$$d\varepsilon/dt = A \cdot \exp(-Q/RT) \cdot \sigma^n$$

where A is a material constant, Q is the activation energy for creep, R is the gas constant, T the absolute temperature and n is the stress exponent. The steady-state equation for semi-brittle creep of granite derived by Hansen & Carter (1983) is used to describe the deformation of basement, while a corresponding equation for diabase (Shelton & Tullis 1981) is applied to describe the deformation of the Seiland Complex. Diabase is chosen because the rocks of the SIP are predominantly mafic. Equivalent data for gabbro are unfortunately not available. Fig. 10 shows the calculated

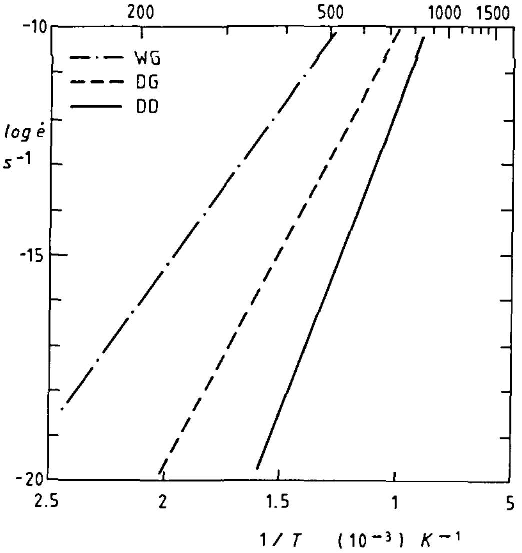


Fig. 10. Steady state strain rate in wet and dry granite and dry diabase as a function of temperature for deviatoric stress of 20 MPa. WG = wet granite (Hansen & Carter 1983); DG = dry granite (Hansen & Carter 1983); DD = dry diabase (Shelton & Tullis 1981).

steady-state strain rate in dry and wet granite as a function of temperature for the stress 20 MPa. In the temperature range 400°-500°C, dry granite deforms at rates of 10^{-15} to 10^{-13} s^{-1} when subjected to a stress of 20 MPa. However, the presence of water dramatically weakens granitoid rocks (Hansen & Carter 1983, Kirby 1983). Deep resistivity measurements (Nekut et al. 1977) and abundant geological evidence (Kirby 1983) suggest that aqueous fluids exist to a depth of at least 20 km in the continental crust. In a wet granite, strain rates of 10^{-15} to 10^{-13} s^{-1} can be achieved at a temperature of 150°C lower than for dry granite, i.e. at a temperature of 250°-350°C. If the ductile zone existed for 10 Ma before the temperature decreased due to erosion, creep rates of 10^{-13} s^{-1} will result in a strain of 30. An initial perturbation of 100 m in the plutonic complex basement interface would grow to a maximum of 3 km.

Discussion

The negative buoyancy stress resulting from the final, Scandian, thrust emplacement of the Seiland Complex is calculated to be 20 MPa. A temperature of approximately 350°C in a wet basement would be high enough to permit the gneisses in this basement to deform at geologically significant strain rates in response to the negative buoyancy stress. For a temperature gradient of 15°C/km a temperature of 350°C can be achieved at a depth of 20-25 km, which could represent the depth to the granitoid basement beneath the SIP shortly after the culmination of the Scandian orogeny. $^{40}Ar/^{39}Ar$ dating of white micas from both Raipas Supergroup phyllites and overlying Dividal Group equivalents in the Repparfjord-Komagfjord window have indicated Scandian (425-400 Ma) temperatures directly beneath the KNC basal thrust in excess of 300-350°C (Dallmeyer et al. 1988). In the lower parts of the KNC at Repparfjord, temperatures exceeded 350°C at c.428 Ma, whereas in parts of the SIP, higher up in the KNC, temperatures at 425 Ma had cooled through c.500°C (Dallmeyer 1988). Thus, comparatively high-temperature conditions would have persisted for some time within the SIP during and subsequent to its Scandian amphibolite-facies metamorphism and tectonic dismemberment and transport.

Temperatures in the mafic igneous complex may not have been sufficient for steady state strain to occur (Fig. 10). The complex itself may therefore have behaved in a more rigid manner during the post-metamorphic stage of actual subsidence, and the deformation of these rocks in this phase of rapid cooling could consequently have been of a more brittle character than that in the basement. There is, however, also some evidence of gravitational tectonics within the SIP. Ultramafic rocks are situated in the cores of deep synforms within paragneisses (Robins 1971).

The deformation which was necessary to reestablish gravity equilibrium could also have been partly brittle. Frictional sliding (Kirby 1983) is likely to have occurred along the Langfjord and Vargsund Faults. The observed fault between the Caledonian nappe sequence and the basement rocks south of Bergmark (Vik 1981) may also be of this type.

The aeromagnetic banded pattern representing the primary layering of the supracrustal sequences in the basement indicates that the Raipas basement windows are bent around the Seiland Igneous Province. This feature can be explained by a southeastward translation and possibly a rotation of the Raipas Supergroup rocks in front of the advancing and gradually subsiding plutonic complex. At the same time the Porsa Duplex was formed. However, the fact that a significantly smaller density contrast (70 kg/m³) could produce granitoid diapirism in the Nordland area (Ramberg 1980, Cooper & Bradshaw 1980, Cooper 1985) further to the south in the Caledonides, supports the hypothesis that the Raipas Windows were partly caused by diapirism. The doming around the windows is thus conceived as the combined effect of a developing duplex and antiformal stack, and subsequent diapiric uprise consequent upon gravitational instability. There can be few situations in an orogenic belt which involve greater gravitational instability than that of a major mafic plutonic complex, like the SIP, being thrust-emplaced onto an imbricating continental margin. In the future, a more complete understanding of the tectonic position of the Raipas Windows will hopefully be gained from a seismic reflection profile across the northernmost part of the windows.

Conclusions

1. From the aeromagnetic and gravimetric data, several major fault zones can be interpreted and traced within the Proterozoic basement. The NE-SW trending Vestfjorden-Vanna and Mierujavri-Sværholt Faults extend out onto the continental shelf. Truncation and dextral offset of the Levajok Granulite Complex by the Mierujavri-Sværholt Fault can be detected beneath the thin cap of metasediments composing the Gaissa Nappe Complex. This fault was probably active as a strike-slip structure during Svecokarelian thrusting of the granulite belt. The Vestfjorden-Vanna Fault has an apparent oblique displacement consisting of a 4 km sinistral strike-slip component and a minimum of 2 km vertical displacement.

2. A negative residual gravity anomaly offshore from Sørøya is caused by a shallow rock body with negative density contrast in relation to the surrounding bedrock. This body probably represents a sedimentary basin consisting of Late Palaeozoic sediments. This basin, which we propose being named the Sørvær Basin, is oriented parallel to and along the extension of the Vestfjorden-Vanna Fault.

3. When defining the regional gravity field above the negative Bouguer anomaly offshore from Sørøya, a depth of 7-8 km can be interpreted for the base of the Seiland Igneous Province. This oval, 70x100 km plutonic complex is a gently northwest-dipping disc-shaped body tapering off gradually in all directions except to the south where it is cut by the Vargsund and Langfjord Faults.

4. Interpretation of depth to magnetic basement combined with gravimetric interpretations show that the Seiland Complex is situated within a depression in the Caledonian allochthon and subjacent basement. Outside this depression the 'basement' describes an elongate arcuate domal structure. From experimental data on granite samples it can be shown that a wet granitoid basement would have responded to a load such as that of the SIP by deforming at geologically significant strain rates to form this semicircular elongate dome which includes the tectonic windows of Raipas rocks.

5. The Alta-Kvænangen window does not appear to be a part of the Caledonian nappe sequence as has recently been suggested, but occurs in its raised domal position as a consequence of gravitational instability during the Scandian orogeny. Some thrust detachment has been detected, however, in the northeastern part of this window. Northeast of Alta, the Raipas rocks show an increasing degree of allochthoneity, with evidence of greenschist-facies metamorphism of Scandian age. The westernmost Raipas rocks within the Repparfjord-Komagfjord window form an integral part of the Caledonian nappe sequence, defining an antiformal area of stacked thrust-sheets. The window doming was thus initiated at this stage of thrusting, just ahead of and beneath the KNC and the wedge of magmatic rocks of the SIP in the foreland-propagating and possibly rotating nappe sequence.

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Plate 1. Aeromagnetic anomaly map, pseudo-relief, western Finnmark and northern Troms.
500 x 500 m grid cells digitized from manually drawn contour maps.
Total intensity referred to DGRF-65.

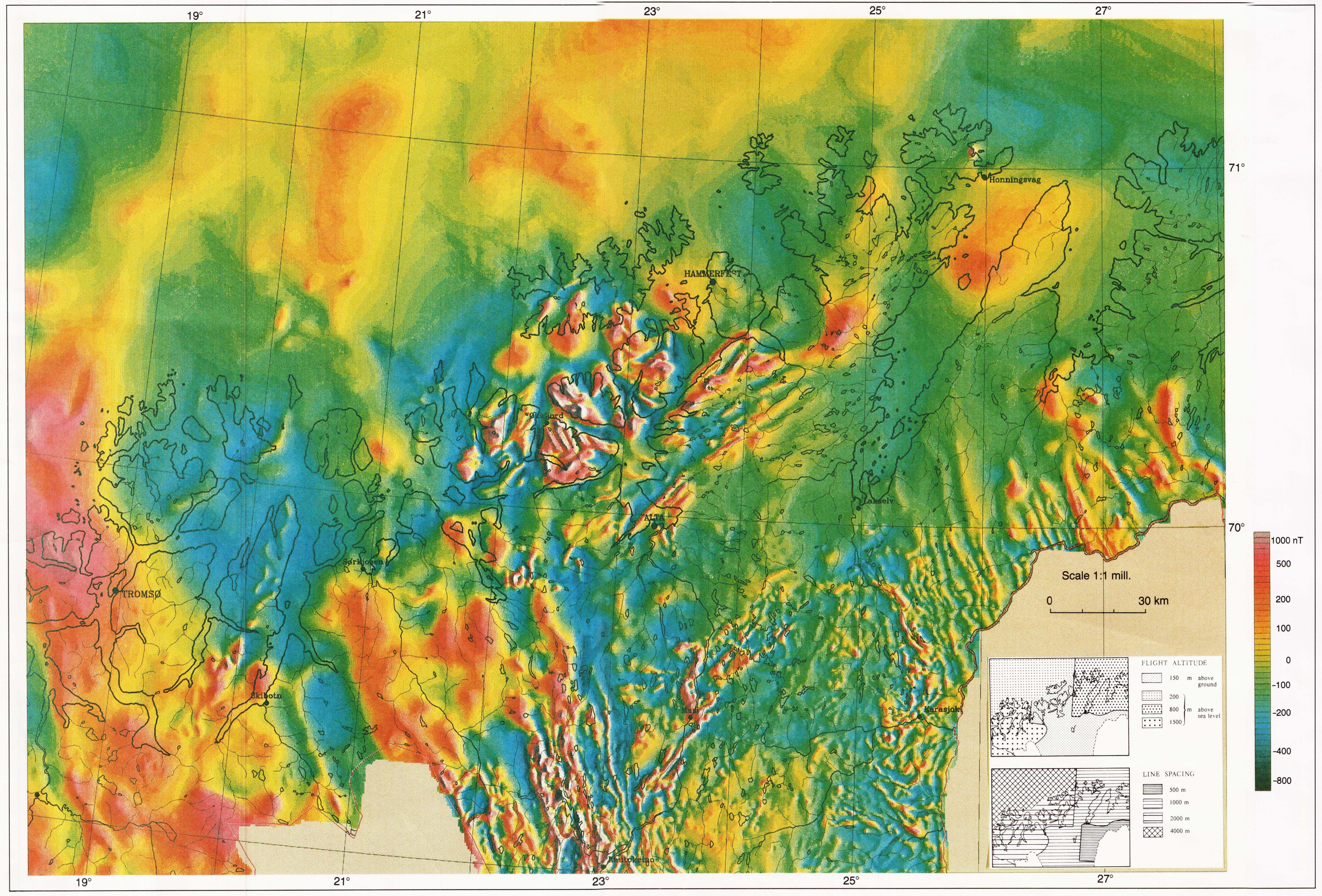


Plate 2. Geological map of western Finnmark and northern Troms.
Part of the 1:1 million bedrock geological map of Norway (Sigmond et al. 1984).

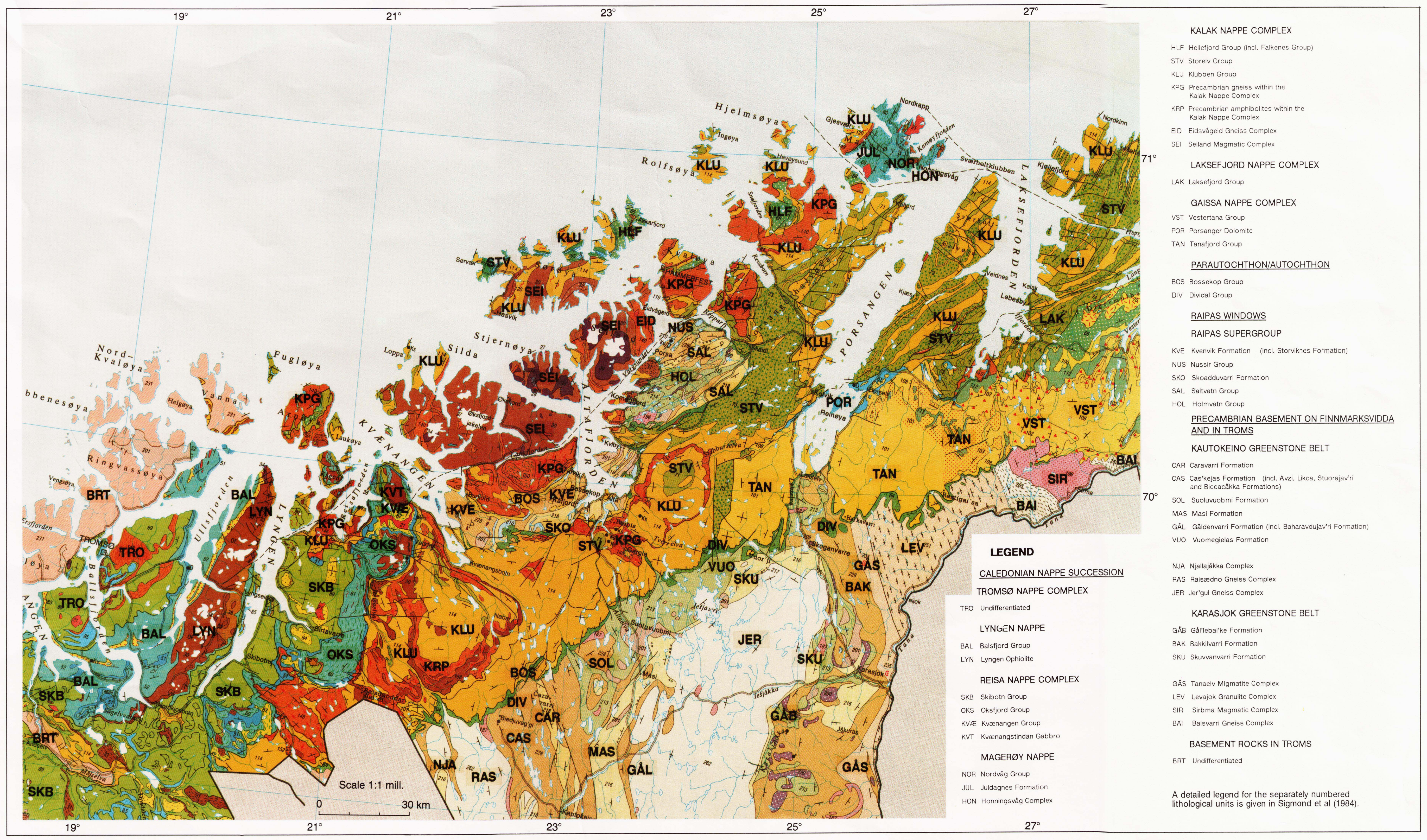


Plate 3: Residual gravity map of western Finnmark and northern Troms.
1500 x 1500 m grid cells interpolated from the Bouguer gravity data-set (I.G.S.N.71) using the minimum curvature method. Regional gravity field subtracted.
Distribution of the original observation sites shown in index map.

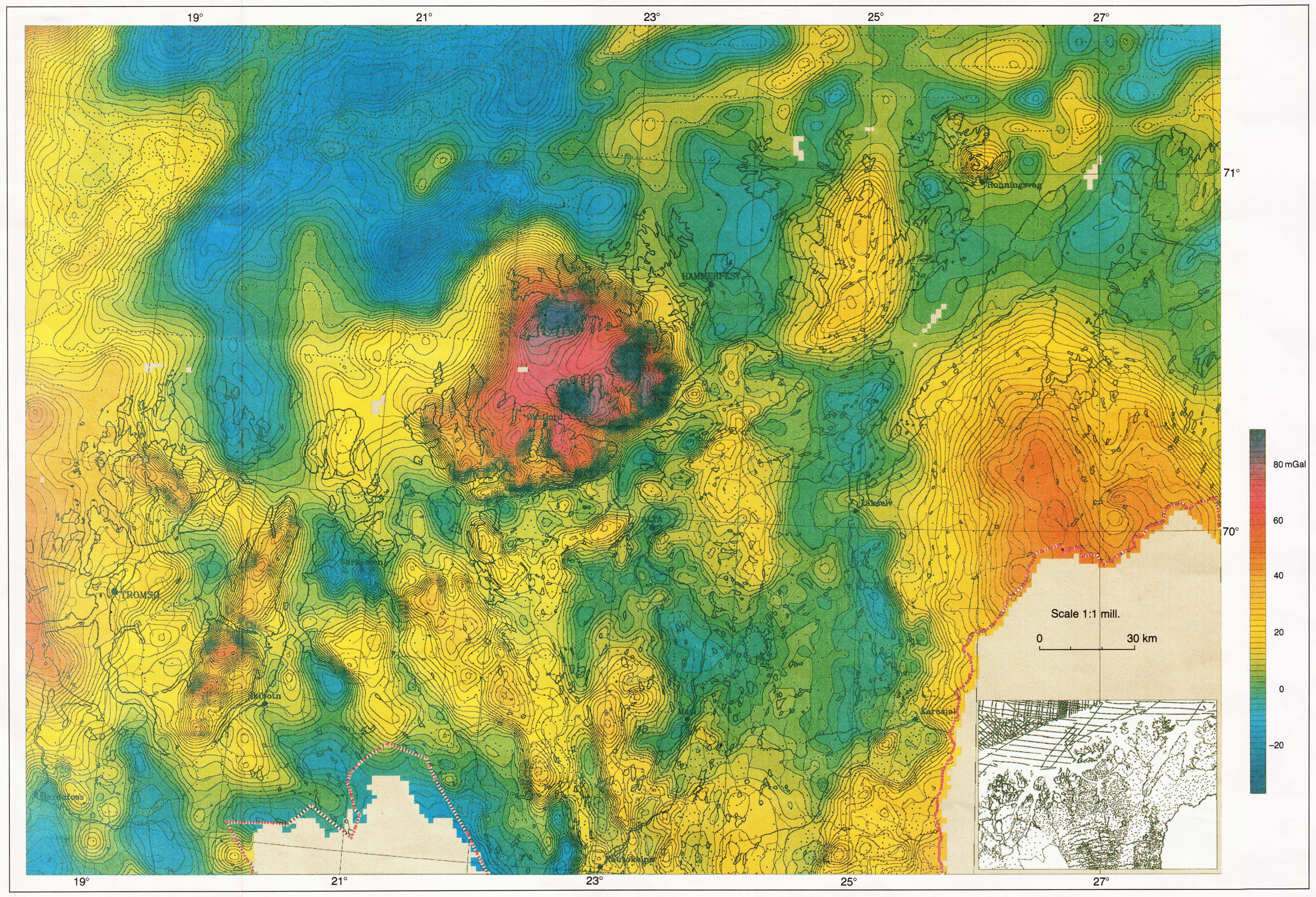
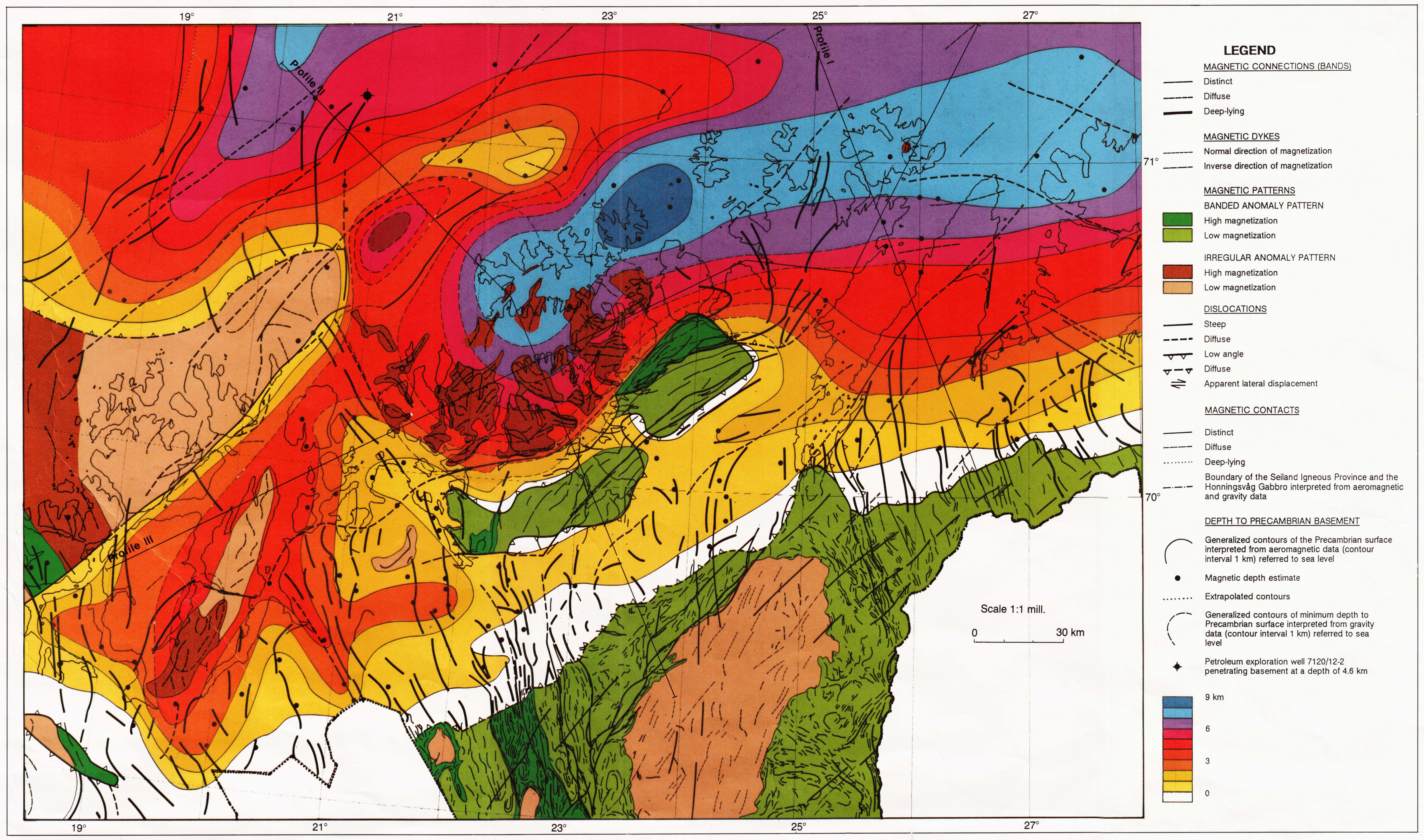


Plate 4. Geophysical interpretation map of western Finnmark and northern Troms.



Reprografi : Norges geologiske undersøkelse
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