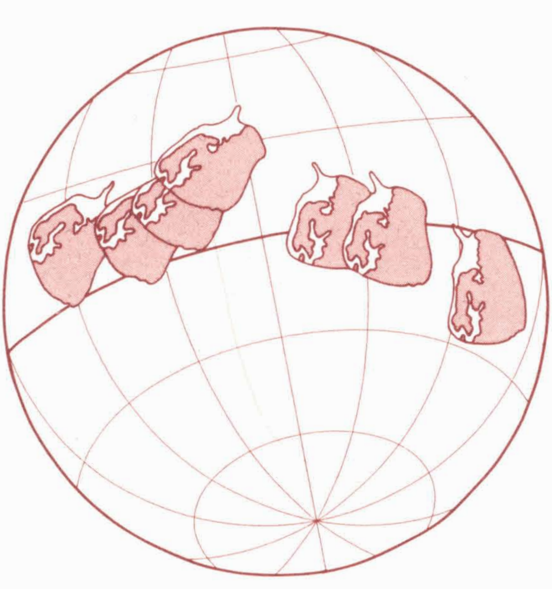
MULTICOMPONENT REMANENT MAGNETIZATIONS REFLECTING THE GEOLOGICAL EVOLUTION OF THE FENNOSCANDIAN SHIELD

- a palaeomagnetic study with emphasis on the Svecofennian orogeny



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Geological Survey of Finland
Espoo 1995

MULTICOMPONENT REMANENT MAGNETIZATIONS REFLECTING THE GEOLOGICAL EVOLUTION OF THE FENNOSCANDIAN SHIELD
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by

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ACADEMIC DISSERTATION

To be presented, with the permission of the Faculty of Science, University of Helsinki, for public criticism in Lecture Room I of the Department of Geology, on October 27th, 1995, at 12 noon

GEOLOGICAL SURVEY OF FINLAND
ESPOO 1995

Mertanen, Satu 1995. Multicomponent remanent magnetizations reflecting the geological evolution of the Fennoscandian Shield - a palaeomagnetic study with emphasis on the Svecofennian orogeny. *Geological Survey of Finland, Espoo.* 46 pages, 16 figures and 3 tables, with original articles (I - IV).

Palaeomagnetic studies in the Fennoscandian (Baltic) Shield reveal multistage magnetizations which are related to the tectonothermal evolution of the shield. The most extensive modifier of primary Archaean and Palaeoproterozoic magnetizations was the Svecofennian orogeny, at 1.9-1.87 Ga. The Svecofennian partial remagnetization is a pervasive overprint throughout the Karelian Province and in the formations studied in the Kola Province. Identification of this remagnetization provides a tool for studying the extent of Svecofennian metamorphism and reactivation of the crust. Based on thermal demagnetization and petrographic studies of Proterozoic mafic dykes and their host rocks, it is suggested that the overprint represents partial thermochemical remanent magnetization, acquired as a result of fluid flux channelled along the dykes. The overall remagnetization of the Archaean and Palaeoproterozoic crust suggests that further cause of the partial remagnetization was uplift and cooling following the Svecofennian orogeny.

New preliminary palaeomagnetic results from the 2.44 Ga old gabbronorite dykes in Russian Karelia, northeastern Karelian Province, are in agreement with those obtained from the nearby Koillismaa layered intrusions of similar age. The high unblocking temperature component of the gabbronorite dykes (Plat = -20.4°, Plong = 276.1°, A95 = 13.5°, N = 3 dykes) is regarded as primary, due to the absence of a similar component in younger crosscutting dykes. Comparative palaeopoles from the southeastern Karelian Province suggest that no major rotations have taken place within the province since 2.44 Ga. At 2.44 Ga the Fennoscandian Shield was located at a latitude of about 30°S.

Key words (GeoRef Thesaurus, AGI): paleomagnetism, remagnetization, hydrothermal alteration, dikes, layered intrusions, polar wandering, continental drift, Proterozoic, Neoproterozoic, Paleoproterozoic, Fennoscandia, Finland, Russia, Norway

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Cover: Drift history of the Fennoscandian Shield from 1930 Ma to 1270 Ma. The figure is compiled by using GMAP © of Trond Torsvik & Mark Smethurst (GSN) and MapDraw © of Matti Leino (GSF).

ISBN 951-690-601-X

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PREFACE

The thesis is based on the following four papers, which are referred to in the text by the Roman numerals:

I **Mertanen, S., Pesonen, L.J., Huhma H. and Leino, M.A.H., 1989.** Palaeomagnetism of the Early Proterozoic layered intrusions, northern Finland. Geological Survey of Finland, Bulletin 347, 40 pp.

II **Mertanen, S. and Pesonen, L.J., 1994.** Preliminary results of a palaeomagnetic and rock magnetic study of the Proterozoic Tsuomasvarri intrusions, northern Fennoscandia. Precambrian Research 69, 25-50.

III **Mertanen, S. and Pesonen, L.J., 1995.** Palaeomagnetic and rock magnetic investigations of the Sipoo Subjotnian quartz porphyry and diabase dykes, southern Fennoscandia. Physics of the Earth and Planetary Interiors 88, 145-175.

IV **Mertanen, S., Pesonen, L.J. and Huhma, H.** Palaeomagnetism and Sm-Nd ages of the Neoproterozoic diabase dykes in Laanila and Kautokeino, northern Fennoscandia. In: T.Brewer (ed.), Precambrian Crustal Evolution in the North Atlantic Region. Geological Society Special Publication (in press).

The contents of each paper are summarized below. The locations of the study areas are shown in Figure 1.

Paper I

This palaeomagnetic study of the 2.44 Ga old Koillismaa layered intrusions was the first to apply modern multicomponent analysing methods in Finland. Several distinct remanent magnetization components, all of them of geological importance could thus be isolated. In addition to the isolation of the ca. 2.44 Ga old primary magnetization, secondary components thought to have been acquired at ca. 2.1 Ga, 1.9 Ga and 1.75 Ga were separated. The results of field tests indicated that the primary remanence was acquired during tilting and break-up of the intrusions during or soon after emplacement and cooling of the intrusions. The dominant overprint is associated with the ca. 1.9-1.8 Ga Svecofennian orogeny. The 2.1 and 1.75 Ga ages for the other overprints were determined from their pole positions on the Apparent Polar Wander Path (APWP).

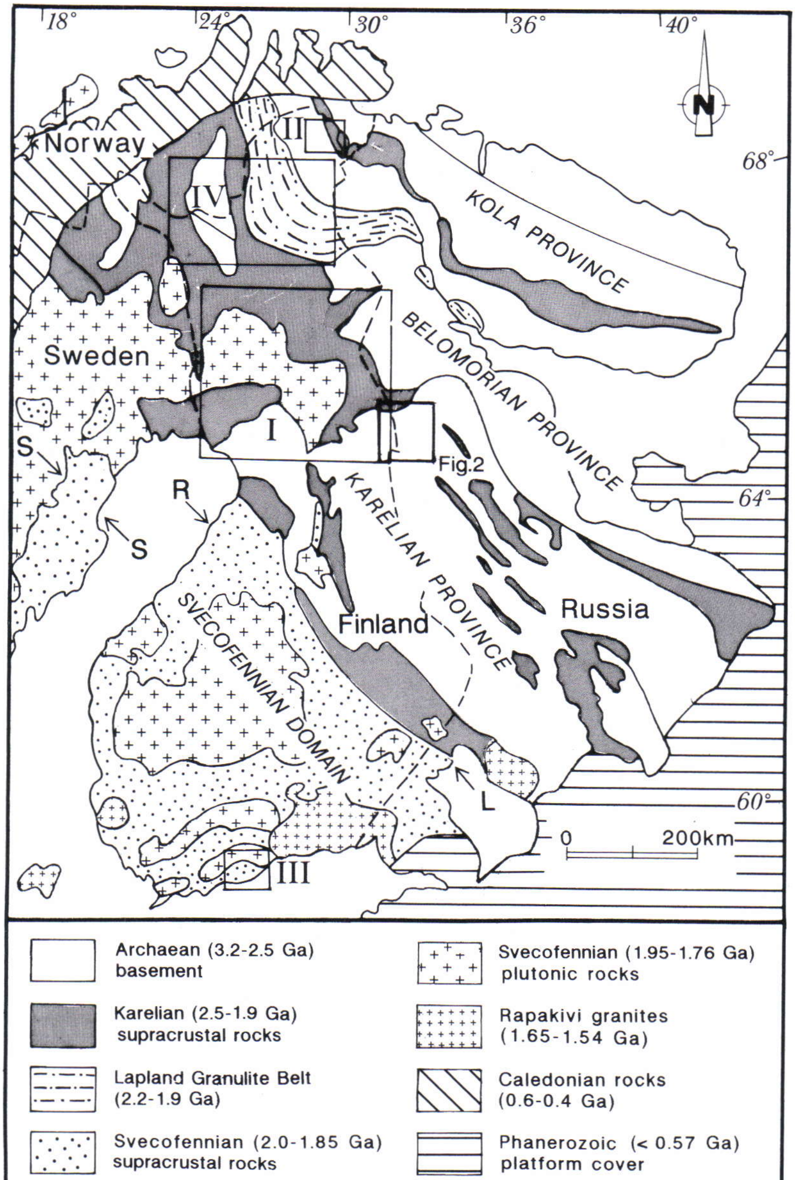

Fig. 1. Generalized geological map of the Fennoscandian Shield, modified mainly after Gaál and Gorbatschev, 1987; Gorbatschev and Bogdanova (1993); Korja et al. (1994) and Korja (1995). R - L = Raahe Ladoga Belt. S - S = Skellefte Belt. Areas of Fig. 2 and the original papers are outlined. For more detailed maps, see Fig. 1 in Paper I; Figs. 1 and 2 in Paper II; Fig. 1 in Paper III and Fig. 1 in Paper IV.

Paper II

Paper II deals with a 1.93 Ga old gabbro diorite intrusion surrounding a younger ultramafic intrusion in Tsuomasvarri, at the southern margin of the Pechenga-Polmak-Pasvik collision zone in the Kola Peninsula. In the gabbro diorite intrusion we isolated a primary remanence component (A') and a secondary component (D), which was also isolated in the ultramafic intrusion. The primary, well defined 1.93 Ga old component has not previously been isolated in the Fennoscandian Shield and so it is useful in constructing the APWP of Fennoscandia. The ultramafic intrusion is dominated by a reversely magnetized component (C),which is probably primary and is interpreted as being about 1.9 Ga in age. The ultramafic intrusion also carries a normal polarity component (A), suggested to be ca. 1.88 Ga in age and three other secondary components; D, B and F. The D component of the gabbro diorite and the ultramafic intrusions is similar to that isolated as a primary 2.44 Ga old component in the Koillismaa layered intrusions (Paper I). However, as will be shown later in this work, it probably represents a later Precambrian overprint in the Tsuomasvarri intrusion. The magnetization age of the B component is probably about 1.75 Ga, based on the K-Ar age data and on evidence of magmatic activity of this age in the area. The F component was interpreted as a Caledonian overprint. The results from the Tsuomasvarri intrusions illustrate the ability of ultramafic and mafic rocks to record several episodes of remagnetization in the northern Fennoscandian Shield, some of which are not readily detected by petrographic observations and radiometric determinations.

Paper III

The 1.63 Ga old Sipoo diabase and quartz porphyry dykes, which are related to the coeval Onas rapakivi batholith, carry a typical Subjotnian magnetization. The remanence is of reversed polarity in the diabases and of normal polarity in the quartz porphyry dykes. On the basis of palaeomagnetic and petrographic evidence, the diabase dykes have been partially remagnetized in the normal polarity direction as a consequence of hydrothermal fluids released from the quartz porphyry dykes and the rapakivi granite. The result verifies that the diabases are slightly older than the quartz porphyry dykes. Another component in the quartz porphyry dykes, showing higher coercivities than the primary Subjotnian direction, was also isolated. Similar results obtained elsewhere in southern Fennoscandia imply that this secondary remanence was acquired at about 1.35 Ga and is thus probably related to the plate tectonic activity that initiated the Sveconorwegian-Grenvillian orogeny.

Paper IV

The Laanila-Ristijärvi and Kautokeino diabase dykes represent young ca. 1.0 Ga mafic activity in the northern Fennoscandian Shield. They provide examples of formations not subjected to later remagnetizations but carrying only a primary remanence. The primary nature of the remanence was verified by baked contact tests. The Laanila, Ristijärvi and Kautokeino dykes, dated at 1042±50 Ma, 1013±32 and 1066±34 Ma (Sm-Nd method), respectively, have similar rock magnetic and petrological properties and thus may originate from the same magma source. However, the remanence directions of the Laanila-Ristijärvi and Kautokeino dykes differ significantly from each other. By comparison with other precisely dated Fennoscandian poles, the magnetization age of the Kautokeino dykes is ca. 900 Ma, which is about 100 Ma younger than that of the Laanila-Ristijärvi dykes. Local tectonic movements, delayed magnetization of the Kautokeino dykes or an erroneous Fennoscandian APWP were put forward as reasons for the discrepancy, but so far none of the explanations has been regarded as valid.

1. INTRODUCTION

In metamorphic terranes the natural remanent magnetization (NRM) of an igneous rock unit may consist of several distinct components. When a rock unit first cools below the blocking temperatures of its constituent magnetic minerals it acquires a primary thermoremanent magnetization (TRM). Later it may be subjected to metamorphic and tectonic processes, which produce secondary remanence components. These secondary components may be of thermal, chemical or viscous origin. Before paleomagnetic data can be correctly interpreted, the various components must be separated and correlated with the processes and geological events responsible for their formation. When a sufficient amount of reliable, well-dated palaeomagnetic data have been acquired, it may be possible to date and correlate the metamorphic events which have left their fingerprints on the remanences of rocks. Therefore, one of the main aims of this study as a whole is to show how palaeomagnetism can be used in tracing, identifying and dating the different geological processes which have modified rocks in the Fennoscandian Shield.

The Archaean and Palaeoproterozoic roks of the Fennoscandian Shield have often been partly or completely remagnetized during later geological events, resulting in the acquisition of multicomponent remanent magnetizations. The most significant event causing remagnetization was the Svecofennian orogeny at ca. 1.9 Ga (Papers I and II). But even younger Subjotnian dykes (ca. 1.6 Ga), which represent relatively late and fresh rocks in the Fennoscandian Shield, carry secondary magnetizations (Paper III). It is essential to understand remagnetization mechanisms because they may provide information on the processes taking place during the corresponding geological event. Emphasis is thus laid on the mechanisms which have produced the Svecofennian remagnetization and on the extent to which the pre-Svecofennian rocks have retained their original magnetization.

All the formations studied consist of Proterozoic intrusive rocks which acquired their primary remanences during cooling. Isolation and identification of the primary remanence are prime tasks in most palaeomagnetic studies (Paper IV), because a rock unit, and hence its primary remanence, can be precisely dated isotopically. Plate movement interpretations and plate reconstructions are mainly based on these isotopically dated magnetization events. In the Fennoscandian Shield there are still large age gaps with no or only a few well defined, precisely dated palaeomagnetic data. In particular, the period from the Archaean (ca. 3.4-2.5 Ga) to ca. 1.9 Ga lacks such key poles. Another purpose of this work has been to acquire reliable palaeomagnetic results from that time interval. A study in cooperation with the University of Oulu (Finland), the University of Toronto (Canada) and the Academy of Sciences in Petrozavodsk (Russia) has begun in order to obtain new geological, geochemical and palaeomagnetic data on Palaeoproterozoic formations. Here, new preliminary palaeomagnetic results are given for 2.4-2.0 Ga old gabbronorite and diabase dykes in Russian Karelia.

2. TYPES OF SECONDARY MAGNETIZATION

Several types of secondary magnetization can overprint primary thermoremanent magnetization (TRM): chemical or thermochemical remanent magnetization (CRM or TCRM), partial thermoremanent magnetization (pTRM) and viscous or viscous partial thermoremanent magnetization (VRM or VpTRM). Superimposed primary and secondary remanences may be carried by two coexisting magnetic minerals or by coexisting multidomain (MD) and single-domain (SD) states of the same mineral (McClelland-Brown, 1982). By comparing unblocking temperature (T_{UB}) spectra and coercivity spectra of a certain remanence component in different rock types and formations, it may be possible to resolve the nature of the magnetization (see section 4.5.2). Petrographic studies on the grade of alteration and metamorphism are also important for identification of different remanence types. It is particularly important to distinguish pTRM and CRM because the T_{UB}'s of pTRM establish the temperature of the geological processes that caused the overprinting, whereas the T_{UB}'s of CRM are unrelated to the temperature at which the chemical change occurred (e.g. Buchan et al., 1977; Dunlop, 1979; McClelland-Brown 1981, 1982).

As will be shown in section 4.5, the Svecofennian orogeny has overprinted most of the older, Archaean and Palaeoproterozoic rock formations. Before the overprinting mechanism can be studied, the nature of the remanence, whether pTRM, CRM or VRM, must be understood. These remanence types are therefore briefly outlined below.

2.1. Partial thermoremanent magnetization

Primary TRM is produced when a rock cools below the blocking temperature (T_B) of its magnetic minerals and is aligned with the ambient geomagnetic field (e.g. Stacey and Banerjee, 1974; O'Reilly, 1976). T_B, the temperature at which the relaxation time becomes long during cooling over a range of temperatures, depends on the effective grain sizes, composition and cooling rate. The T_{UB} (unblocking temperature) observed in thermal demagnetization experiments in the laboratory is the temperature at which the stable remanence of a grain size is lost. Relaxation time is the time any one particle needs for its magnetization to relax to its thermal equilibrium value (Stacey and Banerjee, 1974; Tarling, 1971). It is heavily dependent on the ratio v/T in which v is the grain volume and T the absolute temperature (Irving, 1964). When the relaxation time is long, the remanent magnetization is stable over geological time and is resistant to later changes in the geomagnetic field.

TRM is the form of remanence typically acquired by igneous rocks. The most typical Fe-Ti oxide in igneous rocks is titanomagnetite ($Fe_{3-x}Ti_xO_4$), a solid solution with compositions between magnetite (Fe_3O_4) and ulvöspinel (Fe_2TiO_4). The Curie temperature of titanomagnetite decreases nearly linearly with increasing percentage of titanium, being 580°C for pure magnetite (x=0) and -153°C (x=1) for ulvöspinel (e.g. Stacey and Banerjee, 1974). The Curie temperatures observed in most Precambrian basic igneous rocks are close to pure magnetite in composition (see e.g. Piper, 1987). The Curie temperature, which is thus specific to different minerals and dependent on their composition, is the temperature at which (ferro- and ferri)magnetic properties are lost on heating (Tarling, 1971).

Partial thermoremanent magnetization (pTRM), when occurring as secondary magnetization superimposed on the primary thermoremanence, is acquired over a specific temperature range as a rock cools after reheating. The primary TRM is replaced over the temperature range of the secondary pTRM. Mechanisms for producing pTRM include burial in regional metamorphism and advective heat flow in contact metamorphism. pTRM does not involve changes in mineralogy. However, increased temperatures typically cause chemical and mineralogical changes which may produce chemical remanence (see below). If the chemical changes are small, suggested, for instance, by the fresh appearance of minerals, it is possible to isolate the remanence direction at a specific temperature range and to date the secondary thermal remanence. On the basis of the highest T_{UB}, maximum temperature to which the rock has been subjected can be estimated. The effect of prolonged heating during burial on magnetization is taken into account by correcting the laboratory T_{UB} to correspond to the time of burial (Pullaiah et al., 1975).

2.2. Chemical remanent magnetization

CRM is acquired through the growth of new ferrimagnetic minerals or the alteration of pre-existing magnetic or silicate minerals. It does not require increased temperatures but does involve changes in grain volumes. If temperature is also increased, CRM will be accompanied by pTRM, and the remanence is then called partial thermochemical remanent magnetization (pTCRM).

The direction of CRM must be interpreted with caution, as many studies (see Paper III) have shown that the direction may be influenced by both the ambient geomagnetic field and the internal field produced by existing ferrimagnetic minerals (e.g. Heider and Dunlop, 1987). CRM may be very resistant to demagnetization, with T_{UB}'s and coercivities comparable to or even higher than those of TRM. The reason is related to the relaxation time. As noted above, CRM is acquired through increase of grain volume. For SD grains the increase in grain volume will result in a longer relaxation time and therefore an increase in T_B (e.g. Butler, 1992). However, relaxation time will decrease if grain growth passes into the MD size ranges. Stable CRM is acquired when the grain size remains in the single-domain/pseudo single-domain (SD/PSD) region. Stable CRM can also occur when magnetically unstable large MD grains are subdivided into smaller regions by exsolution (e.g. Stacey and Banerjee, 1974).

The CRM blocking temperature is primarily controlled by the volume of the magnetic grains. The temperature at which the chemical alteration takes place may control the extent of the reaction, but may have no direct control over the CRM blocking temperature (McClelland-Brown, 1982). For example, CRM with high T_B's may arise at very low temperatures through a mechanical transport mechanism such as the penetration of fluid through a rock unit. In such case there will be no significant contemporaneous thermal resetting of the remanence.

2.3. Viscous remanent magnetization

VRM or VpTRM is acquired at low temperatures over a long period. Magnetic minerals do not change, but the direction of magnetization is gradually aligned with the ambient magnetic field direction. Normally VRM has low coercivities and T_{UB}'s. Most rocks investigated in this study carry a VRM aligned in the present Earth's magnetic field. However, long-lasting deep burial accompanied by hydrothermal fluids may be an effective mechanism for acquisition of VpTRM in the early history of the rocks.

3. ANALYTICAL METHODS

3.1. Isolation of remanence components

The conventional demagnetization methods applied in the present study are described in Papers I, II, III and IV. In the preliminary study on Russian Karelian dykes, the remanent magnetization of most of the specimens was measured with a JR4 spinner magnetometer (Geofizika Brno) or one built in the laboratory (Geological Survey of Finland, GSF). Stepwise alternating field (AF) and thermal demagnetization were carried out using Schonstedt instruments. Detailed thermal demagnetization was conducted on 18 specimens, which were cleaned in 3°C steps between 560° and 600°C in order to isolate a small high temperature component. In the two-step cleaning method used for another 16 specimens, samples were first demagnetized with AF cleaning in 13 steps up to 100 mT, with thermal demagnetization in nine steps up to 560°C, and then in 3°C steps between 560° and 580°C or 600°C. In these detailed demagnetization experiments, remanent magnetization was measured with a SQUID magnetometer.

As described in Papers I, II, III and IV, standard multicomponent analysing methods comprising vector plots (Zijderveld, 1967; Torsvik, 1986; Leino, 1991) and principal component analysis (Kirschvink, 1980) were applied to the data in order to isolate individual remanence components. The results of the Russian Karelian dykes were analysed with a 'Tubefind' program (Leino, 1991). Statistical means were calculated according to Fisher (1953). The palaeopoles and drift maps were compiled and plotted with GMAP (Torsvik et al., 1990) and MapDraw (compiled by M. Leino, GSF) programs.

3.2. Identification of remanence carriers

Identification of the magnetic carriers of each component of remanent magnetization may help establish how well the primary magnetic minerals have survived hydrothermal processes and how well they have retained their initial magnetic direction e.g. during metamorphism and hydrothermal alteration. It may then be possible to unravel the successive events (oxidation, burial, metamorphism, etc.) which have affected the rocks since their emplacement (see e.g. Dinarès-Turell and McClelland, 1991).

In most cases it is difficult to determine which particular phase of a magnetic mineral is the carrier of an observed remanence component, because there are typically several modes of occurrence of the same mineral in altered and metamorphosed rocks (Papers II, III, and IV). Extraction of oriented mineral grains and their measurement to isolate the carrier of a specific component has been used in detailed studies of remanence carriers (Buchan, 1978; Geissman et al., 1988). However, this technique is too time consuming for routine work. In this study of dykes in Russian Karelia, the magnetic mineralogy of samples has so far been studied by thermal demagnetization and reflected light microscopy alone. The optical resolution (> 10 μm) is not high enough, however, to isolate the magnetic minerals which are the most probable carriers of hard and stable remanence. The diameter of PSD magnetite grains is 1-10μm and that of SD grains even smaller (e.g. Butler, 1992). In most cases discussed (Papers III and IV) the VRM component is carried by microscopically visible MD grains. Hence, the hard magnetization, whether primary or secondary, is most likely carried by optically invisible grains, and petrographic studies alone cannot reveal the remanence carrier. Rock magnetic studies using isothermal remanent magnetization (IRM) and determination of hysteresis properties are therefore needed, as pointed out in Papers II, III and IV. Such studies have not been conducted on the Russian Karelian dykes as yet, but will be included in further investigation.

Remanence intensity and susceptibility may decrease with increasing hydrothermal alteration or as a result of thermochemical overprinting (Lapointe et al., 1986.; Li and Beske-Diehl, 1993; Papers II, III and IV). For example, intensive alteration may cause dissolution and replacement of titanomagnetite by non-magnetic minerals, resulting in a decrease in magnetic intensity. Petrophysical properties of the rocks such as density and the Koenigsberger's Q-ratio are useful tools for studying alteration and metamorphism within a rock suite and between different areas (see Puranen, 1989; Henkel, 1994). Petrophysical properties of the investigated rocks are therefore described in Papers II, III and IV and will be studied for the Russian Karelian dykes in the future.

4. MULTICOMPONENT REMANENT MAGNETIZATIONS AS REFLECTORS OF THE GEOLOGICAL EVOLUTION OF THE FENNOSCANDIAN SHIELD - DISCUSSION

4.1 Outline of 2.4 - 2.0 Ga geological events in the northern Fennoscandian Shield

The northeastern Fennoscandian Shield (Fig. 1) is mainly composed of Archaean crust divided into the Karelian, Belomorian and Kola Provinces (Gaál and Gorbatschev, 1987). The Karelian Province consists of Late-Archaean (3.1 - 2.6 Ga) granitoids and greenstone belts, which are partly covered by Palaeoproterozoic (Karelian) formations. The Belomorian Province adjoins the Karelian Province in the northeast but differs from it in being characterized by repeated Late-Archaean and Palaeoproterozoic metamorphism and deformation, forming a high-grade mobile belt (Gorbatschev and Bogdanova, 1993; Bibikova et al., 1994). From isotopic, structural and metamorphic studies, it is inferred that the Karelian and Belomorian-Kola Belts collided in the Late Archaean, at ca. 2.5 Ga (Bibikova et al., 1994; Bogdanova, 1994; Bridgwater et al., 1994). The collision was followed by an extensional stage at 2.5-2.35 Ga, when regionally extensive igneous mafic to syenitic intrusions and mafic dykes were emplaced in an intra-plate rift environment accompanied by voluminous volcanism and sedimentation. In the Belomorian belt south of Moncha, at the bottom of the bay of the White Sea (see Fig. 1, Paper II), the Kolvitsa metagabbro-anorthosite has been dated at 2.45 Ga (Balagansky et al., 1994) and the bimodal Tolstik gabbro-granite intrusion in the range of 2435 - 2405 Ma (U-Pb zircon age) (Bogdanova and Bibikova, 1993). Both plutons are cut by coeval or slightly younger mafic dykes. All the sequences in the Belomorian belt were subjected to high grade metamorphism and shearing at 2.42-2.0 Ga (Balagansky et al., 1994; Bogdanova, 1994; Bridgwater et al., 1994).

Intracratonic rifting of the Archaean crust at about 2.4 Ga also took place in the Kola Province, where the layered gabbro intrusion of General'skaya Gora (Mt. Generalskaja in Fig. 1, Paper II) gives a Sm-Nd age of 2453±42 Ma (Bakushkin et al., 1990). The General'skaya Gora intrusion is separated from the Archaean host by a thrust fault, formed in the early stages of development of the Pechenga zone (Melezhik and Sturt, 1994). This zone forms part of the Polmak-Pasvik-Pechenga-

Imandra/Varzuga Greenstone Belt (PV Greenstone Belt) transecting the Kola Province (Pechenga-Varzuga zone in Fig. 1, Paper II). The zone along which the Archaean Sorvaranger belt in the north and the Inari Terrain in the south (see Fig. 2, Paper II) were amalgamated about 2.0 - 1.9 Ga ago is called the Kola suture belt (Berthelsen and Marker, 1986; Paper II). The Monchegorsk intrusion (Moncha in Fig. 1, Paper II) in the central part of the PV Greenstone Belt gives a U-Pb (Zr) age of 2493±7 Ma (Balashov et al., 1993). The Panskie Tundry layered gabbro intrusion (Panski-Fedorova, Fig. 1, Paper II) has a U-Pb age of 2471±4 Ma and the youngest Imandra lopolith close to Moncha (Fig. 1, Paper II), has a U-Pb age of 2396±7 Ma (Mitrofanov et al., 1991).

Rifting also took place in the Karelian Province, where the mafic-ultramafic layered intrusions of Peräpohja (western intrusions, Fig. 1, Paper I) and Koillismaa in Finland and the Olanga complex in Russia (Fig. 2) were emplaced coevally with or only slightly later than most intrusions in the Kola Province. The Koillismaa layered intrusions are dated at 2436±5 Ma (U-Pb age on zircon, Alapieti, 1982; Paper I). The layered intrusions of the Olanga Complex yield Sm-Nd ages of 2406-2439 Ma (Turchenko et al., 1991). Coeval Lapponian and Sumian (2500-2400 Ma) volcanism took place in central Lapland, Kuusamo and Russian Karelia simultaneously with the emplacement of the layered intrusions and boninitic dykes (Turchenko, 1992; Luukkonen, 1992; Vuollo, 1994).

In the Karelian Province, the emplacement of the 2.4 Ga layered intrusions and associated dykes and volcanites was followed by extensive sedimentation and volcanism in Jatulian time, that is about 2.2-2.0 Ma ago, in an extensional regime suggesting transition from cratonic to orogenic conditions (e.g. Gaál and Gorbatschev, 1987; Gaál, 1990). The Jatulian rifting was marked by low-Al tholeiitic (Karjalitic) sills and NW-SE trending dykes (Vuollo, 1994) at the western boundary of the Archaean craton, where a passive margin had developed. This rifting triggered the evolution of the Svecofennian domain (Gaál, 1990). At about 2.1 Ga, another swarm of NW-SE trending Fe-tholeiitic dykes was emplaced in eastern and northern parts of the shield (Vuollo, 1994).

4.2. Previous palaeomagnetic studies on 2.4 - 2.0 Ga formations

Palaeomagnetic studies of the Koillismaa layered intrusions (Mertanen et al., 1987; Paper I) were important in establishing a 2.4 Ga old palaeopole for Fennoscandia (Elming et al., 1993). Reliable data were already available on Archaean (Neuvonen et al., 1981) and Svecofennian rocks (see e.g. Pesonen and Neuvonen, 1981; Pesonen et al., 1991), but data on the 2.4 Ga and 2.2-2.0 Ga old formations were poorly constrained. Four remanence components isolated in the layered intrusions were interpreted as being 2.44, 2.1, 1.9 and 1.75 Ga old in the light of geological evidence (Paper I; Alapieti et al., 1990). However, only the 1.9 Ga old remanence direction could be convincingly linked with the Svecofennian orogeny on the basis of partially reset U-Pb ages. Interpretations of the origin of the other magnetizations were inconclusive. Since crystallization, the intrusions have undergone a complex history of deformation and metamorphism, at 2.2-2.1 Ga, 2.0-1.9 Ga and 1.9-1.8 Ga (Alapieti et al., 1990). However, it was shown that the eastern Koillismaa intrusions are less strongly metamorphosed than the western Kemi-Penikat intrusions, suggesting that a primary 2.44 Ga remanence might be isolated in the eastern intrusions. Unfortunately, no stable remanences were obtained from the western intrusions.

In the Koillismaa layered intrusions, the primary remanence direction (component D with an E-SE declination and moderate downward inclination; Paper I) was shown to have been acquired during tilting. However, its primary nature could not be adequately verified by baked contact tests. Later, the same remanence component was also isolated in the Palaeoproterozoic formations in Russia (Krasnova and Gus'kova, 1990; 1991). However, contact relations indicate that a similar remanence direction observed in the Tsuomasvarri ultramafic intrusion is younger than 1.93 Ga (Paper II). As a result, palaeomagnetic data on Tsuomasvarri led to uncertainty concerning the age of the D remanence in the layered intrusions. As will be discussed in section 4.4, the "D" remanence component in the Tsuomasvarri ultramafic intrusion is now thought to represent a Neoproterozoic (ca. 650 Ma) overprint in the area. The 2.44 Ga old remanence of the Fennoscandian Shield will be further discussed in the light of the new preliminary results presented in section 4.3.

Jatulian (about 2.2-2.0 Ga) remanence directions have also been a subject of controversy (Neuvonen, 1975; Pesonen, 1987). Several Jatulian dykes and intrusions, including the low-Al tholeiitic (Karjalitic) sills and dykes and Fe-tholeiitic dykes have been studied palaeomagnetically, but typically they do not carry stable remanences. One reason for the instability may be the hydrothermal alteration of the dykes. According to Vuollo (1994), the 2.2 Ga old low-Al tholeiitic dykes were crystallized under relatively high a_{H_2O} conditions. Furthermore, the primary mineralogy and texture of the 2.1 Ga old Fe-tholeiitic dykes have generally been destroyed (Vuollo et al., 1994).

In the 2.44 Ga layered intrusions, we isolated a shallow E-W directing remanence direction (component E) with both normal and reversed polarities which, from its pole position on the APWP of the Fennoscandian Shield, was interpreted as a Jatulian (2.2-2.0 Ga) overprint (Paper I). A similar direction, although imprecisely determined, has been obtained from ca. 2.0 Ga old Pialozero diabases and sedimentary rocks in the southeastern part of the Karelian Province (Khramov and Ignatieva, 1980). The remanence exhibits dual polarity, too, but the authors do not report positive field tests to indicate that the remanence is primary. A corresponding shallow westerly direction has also been isolated in ca. 2125 Ma andesites and pillow basalts in the northern Pechenga zone in the Kola Province (Torsvik and Meert, 1995). On the basis of a positive conglomerate test, the authors regard the remanence as primary.

A completely different remanence component, directed towards the NE with a moderate downward inclination, has, however, been isolated in the Jatulian diabase dykes of the Varpaisjärvi area in central Finland (Neuvonen et al., 1981; Neuvonen, 1992). One of the Varpaisjärvi dykes gave a Sm-Nd age of 2085±95 Ma (Toivola et al., 1991). On the basis of positive baked contact tests, the remanence is interpreted as primary (Neuvonen et al., 1981). Recently, Krasnova and Gus'kova (1991) have interpreted a similar down NE remanence, obtained from mafic dykes in Russian Karelia, as Jatulian (about 2.0 Ga) primary magnetization.

Thus, two different directions - a shallow up or down E-W direction and an intermediate down NE direction - have been reported for Jatulian (2.2-2.0 Ga) rocks. Both remanence directions were verified as primary by positive conglomerate and baked contact tests. However, the down NE direction is close to the B direction in the Koillismaa intrusions, which was regarded as a post-Svecofennian overprint (about 1750 Ma) on the basis of its low unblocking temperatures and its pole position in relation to the APWP (Paper I). The B direction was also isolated as an overprint in the Tsuomasvarri intrusion (Paper II) and in a small Archaean intrusion in the Kola Province in Finland (Isokivennokka granite; Mertanen and Pesonen, 1992). Elming (1994) reports a similar overprint direction in gabbro intrusions of Svecofennian age (1.84-1.80 Ga) in northern Sweden. He interprets this component as post-Svecofennian. Hence, one of the two directions of possible Jatulian age in Varpaisjärvi and Russian Karelia also occurs as a widespread post-Svecofennian (post 1.8 Ga) overprint in the northern Fennoscandian Shield.

To obtain reliable, dated palaeomagnetic results from the Palaeoproterozoic formations and eventually to resolve the origin of the remanence components isolated in the layered intrusions (Paper I), gabbronorite and diabase dykes in Russian Karelia have been studied. These dykes were chosen because they occur far from the collision zone of the Svecofennian orogen at the boundary between the Svecofennian Domain and the Karelian Province (Fig. 1) and are thus probably only little affected by orogenic events. In the following, the palaeomagnetic results from the Russian Karelian dykes are presented in detail to demonstrate the separation of remanence components. The main aim of the preliminary study was to isolate primary remanence components and to assess the effect of the Svecofennian orogeny far from the actual orogenic belt.

4.3. Preliminary palaeomagnetic results from the Palaeoproterozoic dykes in Russian Karelia

4.3.1 Geology and sampling of the Suoperä and Lake Pääjärvi areas

The Suoperä and Lake Pääjärvi areas in Russian Karelia are located in the northeastern part of the Karelian Province (Fig. 1) which is predominantly composed of Late-Archaean migmatites and granitoids (e.g.

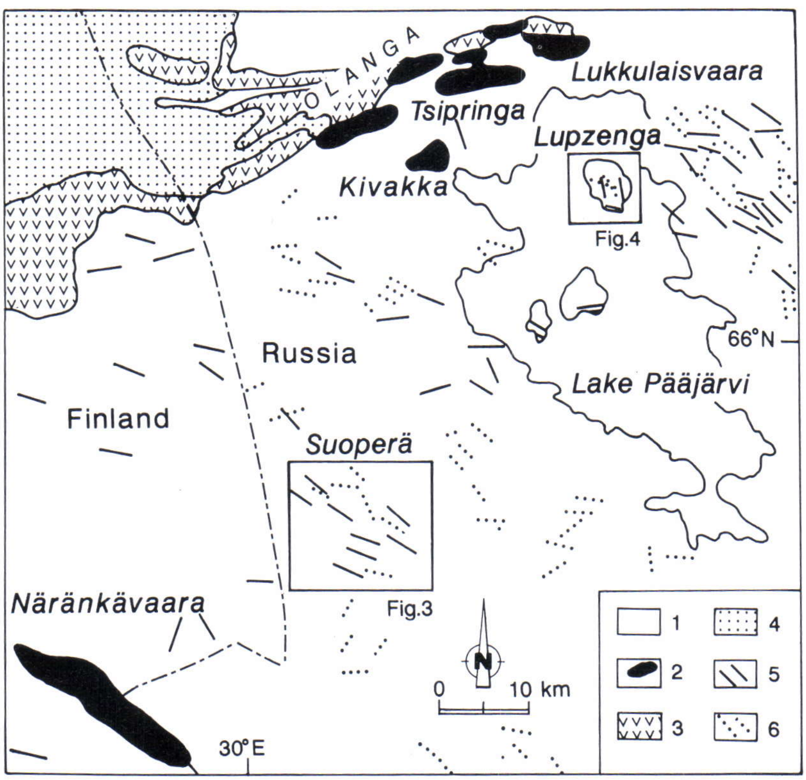

Fig. 2. Simplified geological map of the Näränkävaara layered intrusion and Olanga complex (Kivakka, Tsipringa and Lukkulaisvaara layered intrusions) and associated mafic dykes compiled from Turchenko (1992) and V. Stepanov and A. Slabunov (pers. comm.,1994). 1 = Archaean granite gneiss basement, 2 = Layered intrusions (2.45-2.35 Ga), 3 = Sumi-Sariolan (2.5-2.3) volcanic and sedimentary rocks, 4 = Jatulian (2.2-2.0 Ga) volcanic and sedimentary rocks, 5 = Mafic dykes of ca. 2.4 Ga age, 6 = Mafic dykes younger than ca. 2.4 Ga. Areas of Figs. 3 and 4 are outlined.

Luukkonen, 1992). These areas (Fig. 2) are part of the Pjaozero block, which comprises diorites, granodiorites, quartz diorites and gneissose granodiorites (Grishin, 1992). According to Zystra (1991), the dominant Archaean basement rock type in the Suoperä area is quartz diorite and in the Lake Pääjärvi area granodiorite. Several age determinations show that igneous activity took place at 2.4 Ga in the area. For the layered intrusion complex of Olanga (Fig. 2), Sm-Nd dating yielded ages of 2439±29 Ma for the Kivakka massif, 2406±128 Ma for the Lukkulaisvaara massif and 2336±236 Ma for the Tsipringa massif (Turchenko et al., 1991). The U-Pb zircon age of a granite porphyry at Tiroyarvi is 2449±3.7 Ma (Levchenkov et al., 1990). Sm-Nd dating of olivine gabbronorite dykes related to the Olanga intrusions yielded an age of 2457±88 Ma (Turchenko et al., 1991). The dykes investigated in this study are probably from the same swarm as the dated dykes. Using the U-Pb method on baddeleyite, J. Vuollo (pers. comm., 1995) obtained an age of 2446+5/-4 Ma for one of the gabbronorite dykes of this study.

To the author's knowledge, there is no direct evidence of penetrative tectonometamorphic events related to the Svecofennian orogeny in the area. However, an extensive and deep NE-SW trending fracture zone occurs just north of the study area and may have had some influence on the palaeomagnetic directions. The zone, which includes the several hundred kilometre long Kantalahti-Puolanka fracture zone (Elo, 1992; Ruotoistenmäki, 1992), extends ca. 300 km southwestwards as the Oulujärvi Shear Zone (Kärki et al., 1993). It cuts a number of different Archaean and Palaeoproterozoic lithological units, and is less than ca. 1800 Ma old as evidenced by the Oulu granite that is cut by the fracture (Vaasjoki, in Ruotoistenmäki, 1992).

Most dykes of the present study trend NW-SE, but there are also many trending E-W. It is, however, difficult to place the dykes in different age groups on the basis of their trends, because in some places the NW-SE dykes cut the E-W dykes and in others the E-W dykes cut the NW-SE dykes.

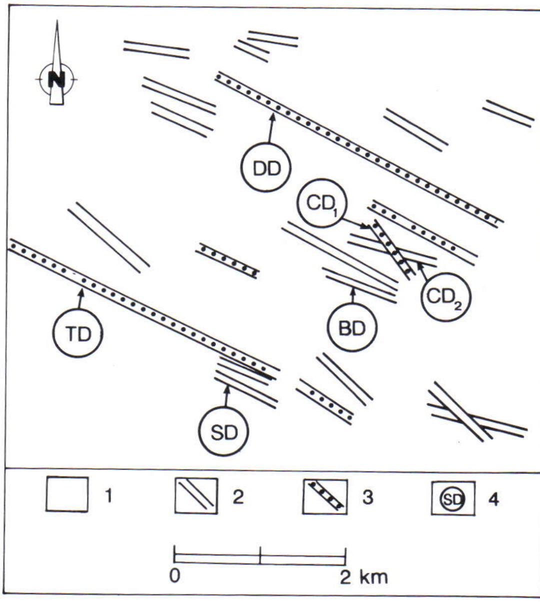

Fig. 3. Sampling localities in the Suoperä area (see Fig. 2) in Russian Karelia, based on geological map of Zystra (1991). 1 = Massive quartz diorite, 2 = Gabbronorite, 3 = Olivine gabbronorite, 4 = Dyke label.

In the Suoperä area (Fig. 3), preliminary palaeomagnetic results have been obtained from six subvertical dykes, all with a NW-SE trend. At one site (CD, Fig. 3), a NW-trending dyke (CD_1) cuts a dyke (CD_2) with a more prominent E-W trend. The minimum width of the dykes is 20 m. Five to twelve samples were taken from each dyke. The dykes are subophitic and consist of plagioclase and ortho- and clinopyroxene which have been variably replaced by amphibole and biotite, and minor apatite and zircon. Magnetite is the dominant iron-titanium oxide and olivine is present sporadically.

In the Lake Pääjärvi area, samples were taken from four crosscutting dykes at the island of Lupzenga (site XD, Fig. 4). The oldest dyke, which cuts an Archaean granite, is gabbronoritic and trends NW-SE. It is cut by thin

6-60 cm wide fine-grained diabase dykes, also trending NW-SE. According to J. Vuollo (pers. comm., 1994), the gabbronorites and thin diabase dykes have a similar, boninitic, composition. Both dyke types are cut by an Fe-tholeiitic (J. Vuollo; pers. comm., 1994) gabbroic diabase (gabbro diabase) dyke ("Older Jatulian dyke") with a more pronounced E-W trend, which in turn is cut by an EW-trending Fe-tholeiitic gabbro diabase dyke ("Younger Jatulian dyke"). Samples were taken from all four dyke types and the adjacent granite near the contact with the gabbronorite. The mineral composition of the Lupzenga gabbronorite dykes corresponds to that in the Suoperä area. Orientation, cross-cutting relations and mineral composition indicate that the dykes belong to the same dyke swarms as those studied by Krasnova and Gus'kova (1991). The dykes follow the same pattern, the oldest gabbronorites being cut by gabbro diabases.

4.3.2. Palaeomagnetic results

At least two stable magnetization components are recognized in dykes of the Suoperä

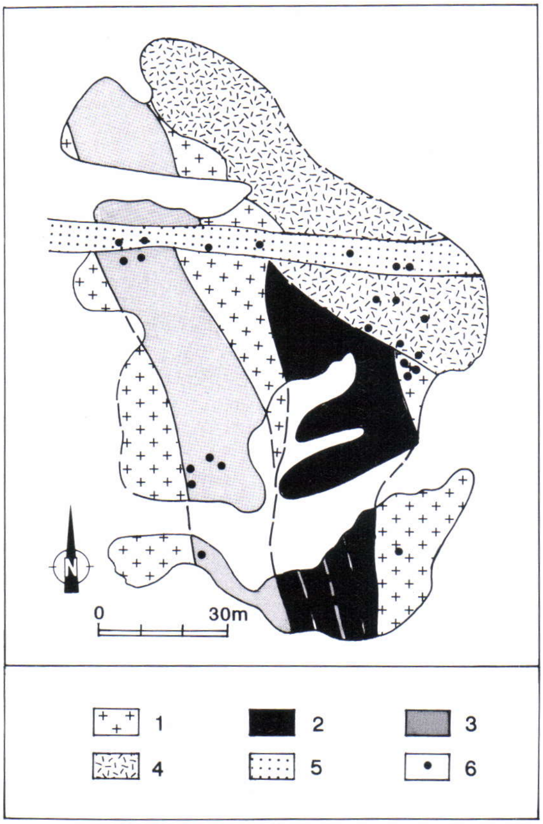

Fig. 4. Simplified geological map of the Lupzenga Island in Lake Pääjärvi (site XD, see Fig. 2) in Russian Karelia. Compiled after Stepanov (1994). 1 = Archaean granite, 2 = gabbronorite, 3 = diabase, 4 = Gabbro diabase "Older Jatulian dyke", 5 = Gabbro diabase "Younger Jatulian dyke", 6 = sample location.

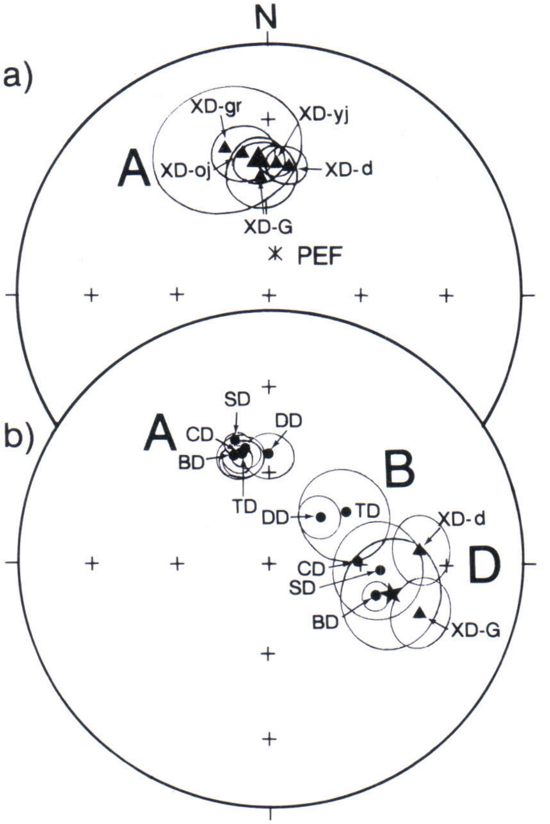

Fig. 5. a) Site mean directions for component A at the Lupzenga Island (site XD); XD-G = gabbronorite, XD-d = diabase, XD-gr = granite, XD-oj = "Older Jatulian dyke", XD-yj = "Younger Jatulian dyke", PEF = the Present Earth's Field direction, b) Circle = Site mean directions of components A, B and D in the Suoperä area, Triangle = Site mean directions of component D at Lupzenga Island, Star = mean directions of component D at BD, SD and XD-G.

and Lake Pääjärvi areas (Fig. 5). The D component has an intermediate down E-SE direction, whereas the A component is down to the N-NW. Some samples exhibit directions intermediate between A and D (component B) or swing between A and D.

4.3.2.1. Suoperä area

All dykes carry multicomponent remanent magnetizations (Fig. 5, Table 1). Two stable components were isolated in each thermally demagnetized sample. A steep low coercivity and low T_{UB} component, regarded as a viscous remanence acquired in the present Earth's magnetic field (PEF), was also isolated in each sample. On the basis of remanence directions and intensity decay curves, the Suoperä dykes (Fig. 3) form two different groups. Dykes BD, SD and CD_2 belong to Group I and dykes TD, DD and CD_1 to Group II. The Group I dykes carry both D and A components, whereas the Group II dykes carry mainly the A component and in some samples the B component.

AF demagnetization

Group I. All specimens from dyke BD show directional swings along great circles between A and D upon AF demagnetization (Fig. 6a). However, as shown by orthogonal plots (Fig.

Table 1. Mean palaeomagnetic data on the Suoperä gabbronorite dykes (Lat=65.7°N, Long=30.2°E)

Dyke	B/N/n	D (°)	I (°)	α95 (°)	k	Plat (°N)	Plong (°E)	dp (°)	dm (°)
Component A									
SD	1/3*/7	344.2	48.6	4.0	954	52.6	233.1	3.5	5.3
TD	1/11*/29	345.2	54.2	6.5	50	57.8	233.2	6.4	9.1
BD	1/8*/11	341.0	54.2	5.3	111	57.0	239.7	5.2	7.4
CD_1 and CD_2	2/7*/13	344.3	53.4	7.6	64	56.8	234.4	7.4	10.6
DD	1/6*/17	359.4	54.4	7.9	73	59.2	211.0	7.8	11.1
MEAN	5*/35/77	346.7	53.1	4.7	270	57.1	230.6	A95 = 6.1	
Component B									
TD	1/3*/5	57.0	60.8	14.9	70	-50.8	308.9	17.3	22.7
DD	1/4*/9	46.9	67.9	6.9	177	-62.2	309.3	9.7	11.6
Component D									
SD	1/4*/9	93.7	54.2	15.2	381	-29.9	281.22	15.0	21.4
BD	1/8*/11	106.7	53.5	4.6	147	-24.3	270.7	4.4	6.4
CD_2	1/1/1	89.3	61.6	-	-	-38.5	279.9	-	-
MEAN	3*/13	97.1	56.6	10.3	145.8	-31.0	277.1	A95 = 13.3	

B/N/n is the number of dykes/samples/specimens used in the statistical calculations; * denotes the statistical level used in mean calculations; D and I are the mean declination and inclination respectively; α95 is the radius of the circle of 95% confidence; k is Fisher's (1953) precision parameter; Plat and Plong are the palaeolatitude and paleolongitude for the Virtual Geomagnetic Poles of different dykes; dp and dm are the semiaxes of the oval of 95% confidence; A95 is the radius of the circle of 95% confidence of the mean poles. Component A has normal (N) polarity. Pole positions for components B and D have been inverted to reversed (R) polarity.

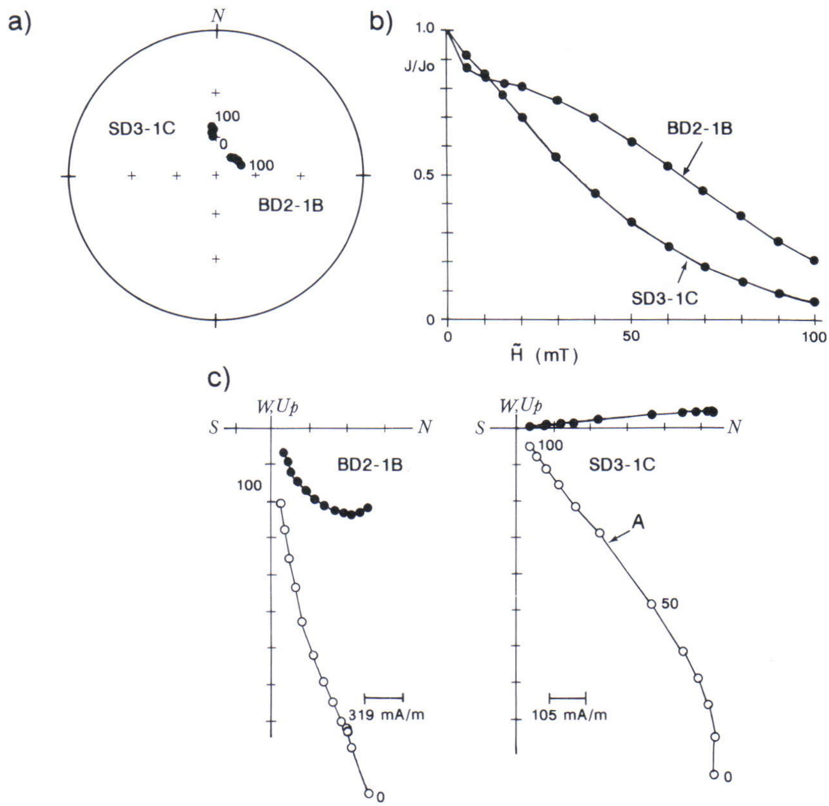

Fig. 6. Examples of AF demagnetization behaviour for Suoperä dykes BD and SD. a) Stereographic projection of directional data upon demagnetization, where PEF as in Fig. 5, b) relative NRM intensity decay curves, c) orthogonal demagnetization diagrams. Open (closed) symbols denote projections onto vertical (horizontal) planes. Numbers at demagnetization steps denote peak alternating field (mT). Dyke BD exhibits only an intermediate direction component. Note the high coercivities of both dykes.

6c), the Zijderveld trajectories are curved, indicating that at least two components are being simultaneously demagnetized. The high coercivities of the BD specimens are typical of fine-grained SD magnetite.

Specimens from dyke SD also have comparatively hard coercivities, although they are always softer than in dyke BD (Fig. 6). The vector plots decay straight to the origin. The samples are dominated by the A direction, which is maintained up to 100 mT (e.g. specimen SD3-1C, Fig. 6).

On the basis of geochemical analysis (J. Vuollo, pers. comm., 1995) the older dyke (CD_2) at site CD is boninitic and the younger one (CD_1) Fe-tholeiitic. Sample $CD_2$7-1A (Fig. 7) is from the interior of dyke CD_2 which behaves similarly to dyke BD upon AF cleaning. However, samples $CD_2$5 and $CD_2$6 (illustrated by specimen $CD_2$6-1B, Fig. 7) from the same dyke, taken near the contact with the younger crosscutting dyke, CD_1, are magnetically soft and carry only component A.

Group II. In AF demagnetization, the specimens from dykes DD, TD and CD_1 show consistent behaviour (Figs. 7 and 8). They are magnetically softer than Group I specimens and carry only component A, isolated at high

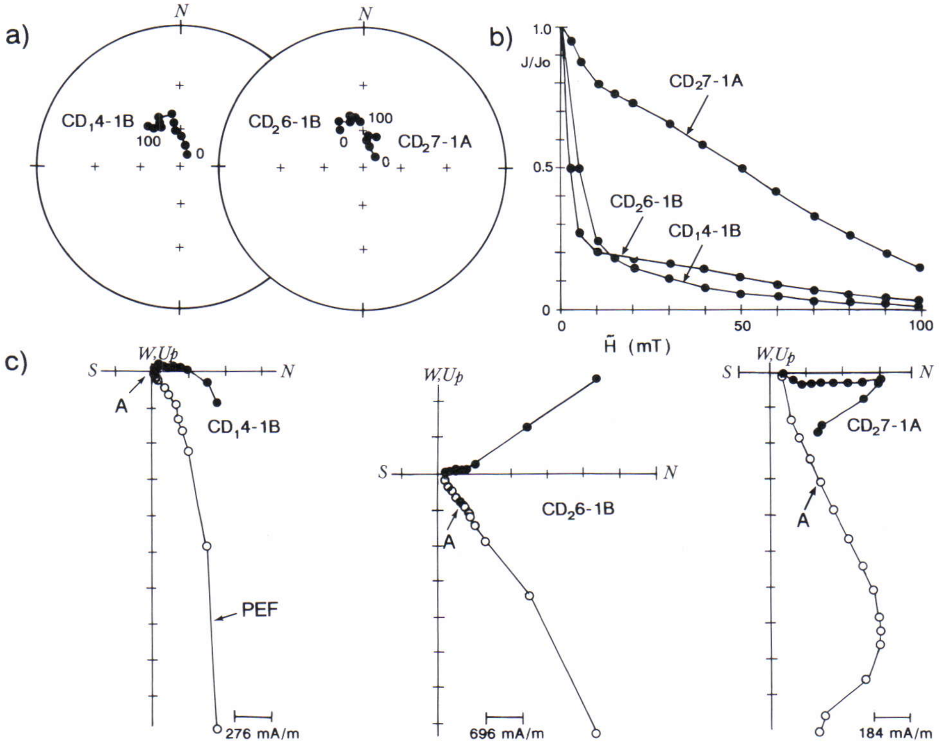

Fig. 7. Examples of AF demagnetization behaviour for Suoperä dykes CD_1 and CD_2. Specimens $CD_2$6-1B and $CD_2$7-1A are from the margin and interior of the same dyke, respectively. Note that specimens $CD_1$4-1B and $CD_2$6-1B, which exhibit only the A component (excluding a small VRM due PEF), are magnetically softer than sample $CD_2$7. Symbols as in Fig. 6.

coercivities up to 100 mT. Note that in all CD_1 samples, components A and PEF decay simultaneously, and result in a curved line in lower AF steps (Fig. 7C), whereas in sample $CD_2$6 from dyke CD_2 the A component is clearly separated above 10 mT. Like samples $CD_2$5-6 from the margin of dyke CD_2, the remanence intensity of dyke CD_1 decays exponentially with AF cleaning. According to Costanzo-Alvarez and Dunlop (1993) such behaviour indicates that magnetite must be of MD size and optically visible.

Thermal demagnetization

Group I. Thermal demagnetization of dykes BD, SD and CD_2 revealed two components: A and D. Examples of thermal demagnetization are shown in Figures 9 and 10. The intensity decay curves are square-shouldered. Component D is isolated in a narrow T_{UB} range of 570-600°C, indicative for low-titanium titanomagnetite. On the other hand, component A typically occupies a wide T_{UB} interval of 400-560°C. Because AF demagnetization proved ineffective in isolating different remanence components, especially in dykes BD and CD_2, some of the specimens were thermally demagnetized after AF treatment up to 100 mT (see section 3). Specimen BD2-1B is an example of a two-stage cleaning experiment (Figs. 6 and 9).

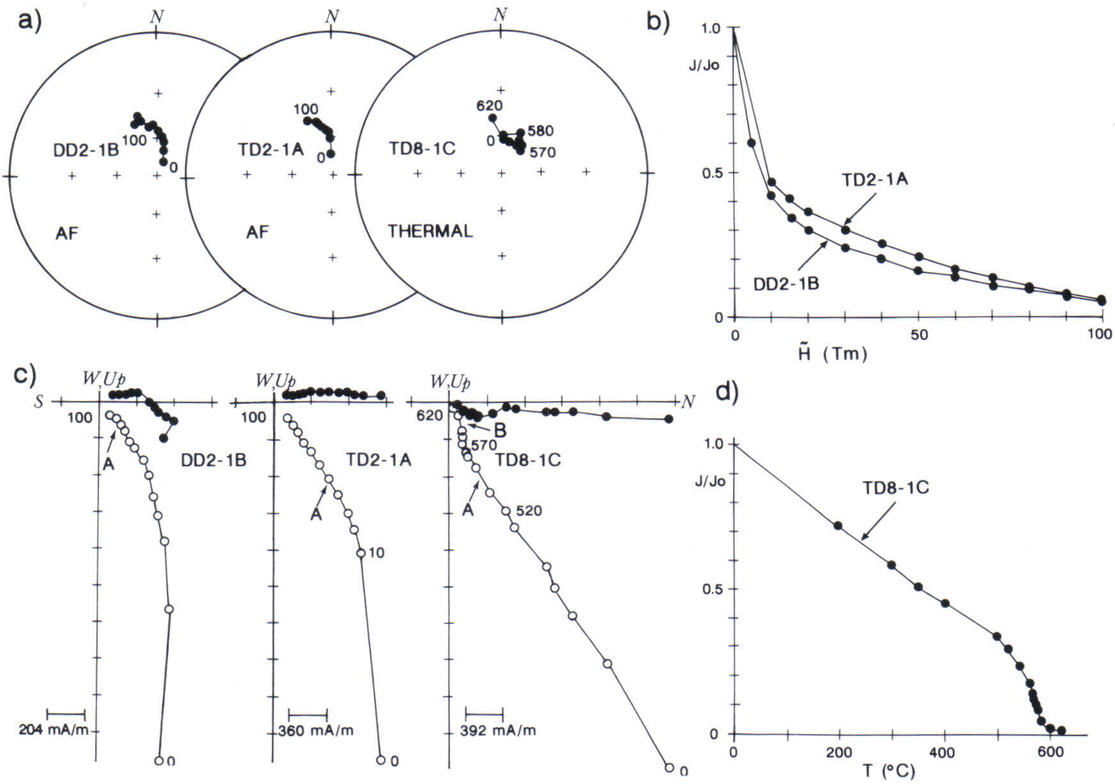

Fig. 8. Examples of AF demagnetization behaviour for Suoperä dykes DD and TD and typical thermal demagnetization behaviour for dyke TD. In a) and c) numbers by each demagnetization step for specimen TD8-1C denote temperature (°C), d) relative intensity decay upon heating. Other symbols as in Fig. 6.

Dyke BD has a larger D component than dykes SD and CD_2, in which the A component dominates (Figs. 9 and 10). Thermal demagnetization of samples $CD_2$5 and $CD_2$6 from dyke CD_2 near the contact with dyke CD_1 gives only component A, which is isolated in the whole temperature range of 200-580°C. In a few samples from dyke BD, component A was isolated as an intermediate temperature direction (400-560°C), while component D occupies the low (0-400°C) and high (560-600°C) T_{UB}'s. In orthogonal plots this is seen as a z-shaped decay towards origin (specimen BD6-1B, Fig. 9c). Such behaviour is common for superimposed remanence components if one is a chemical remanence (Schwarz and Buchan, 1989). In most samples the T_{UB}'s of components A and D partly overlap in the temperature range of 540-570°C. The overlap of two components further implies that the other one is probably of chemical origin, rather than a partial overprint. AF cleaning will not reveal two components if the coercivity of the superimposed remanence is identical. The remanent coercive force spectra of the constituent magnetizations overlap, but their blocking temperature spectra do not necessarily do so.

Because AF demagnetization isolated only component A in dyke SD, but thermal demagnetization revealed both components A and D, it is probable that the AF-cleaned A direction is contaminated by the underlying D direc-

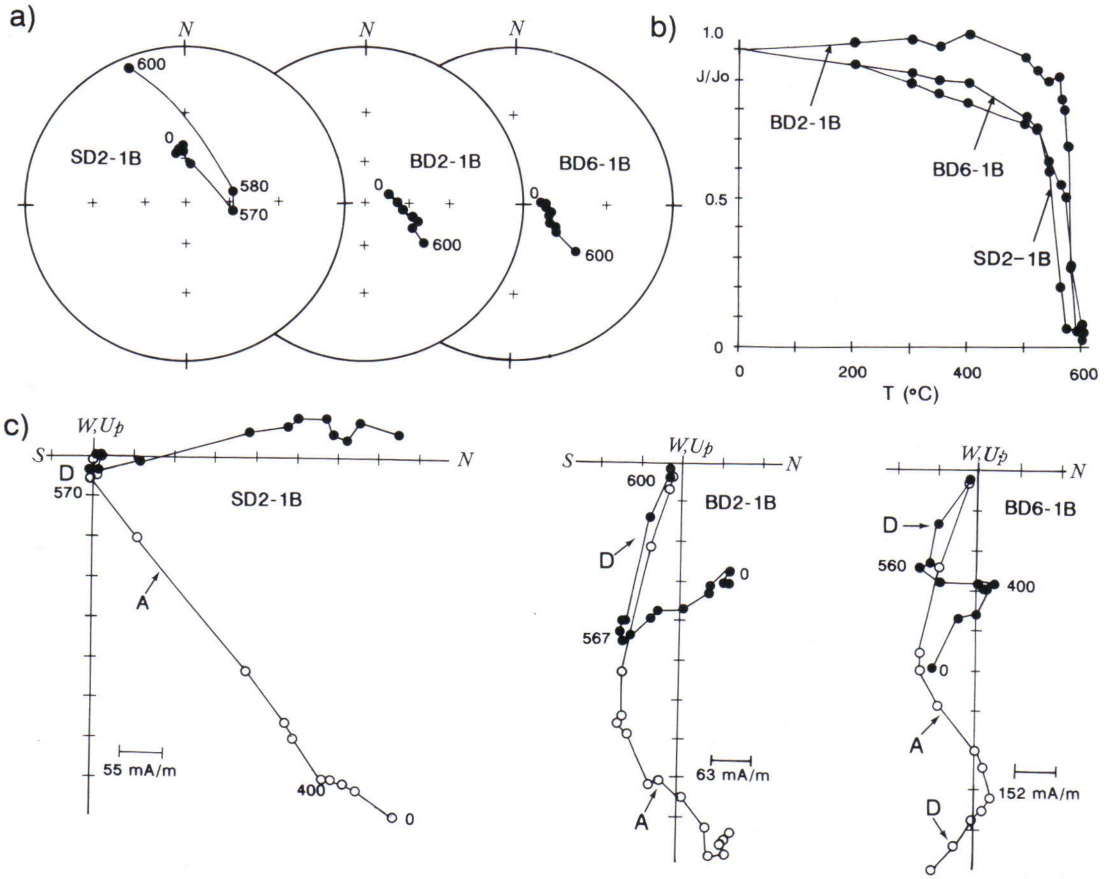

Fig. 9. Examples of thermal demagnetization behaviour typical of Suoperä dykes BD and SD. a) Stereographic projections, b) relative intensity decay upon heating, c) orthogonal demagnetization diagrams. Numbers by each demagnetization step denote temperature (°C). Other symbols as in Figs 6 and 9.

tion. Indeed, when the mean direction of component A was calculated separately for the AF and thermally demagnetized directions, the declination was about 10° higher for the AF cleaned directions. Therefore, only the thermally cleaned direction of component A was used for dyke SD (Table 1).

Group II. Dyke CD_1 was demagnetized with AF only. However, the similarity of the AF behaviour to that of samples $CD_2$5 and $CD_2$6 from dyke CD_2, suggests that the thermal behaviour would be similar, too.

Like dyke CD_1 and samples $CD_2$5 and $CD_2$6 from near the contact of dyke CD_2, most TD samples are totally overprinted by A. However, some TD samples reveal component B in the temperature range of 560-620°C, above the temperature range of 520-560°C of the A direction (Fig. 8). This direction has a comparatively steep downward inclination and NE declination. The direction is roughly the same as that regarded as an artefact in dyke BD. But, because component B is isolated as a high temperature component (possibly carried by low-titanium titanomagnetite) clearly distinct from the A component, it is probably real. The direction of B deviates clearly from that of D (Fig. 5) although both occupy a similar T_{UB} range. The intensity decay curve of dyke TD is clearly different from that of dykes BD and SD (Figs. 8 and 9). The mean direction of component A was calculated sep-

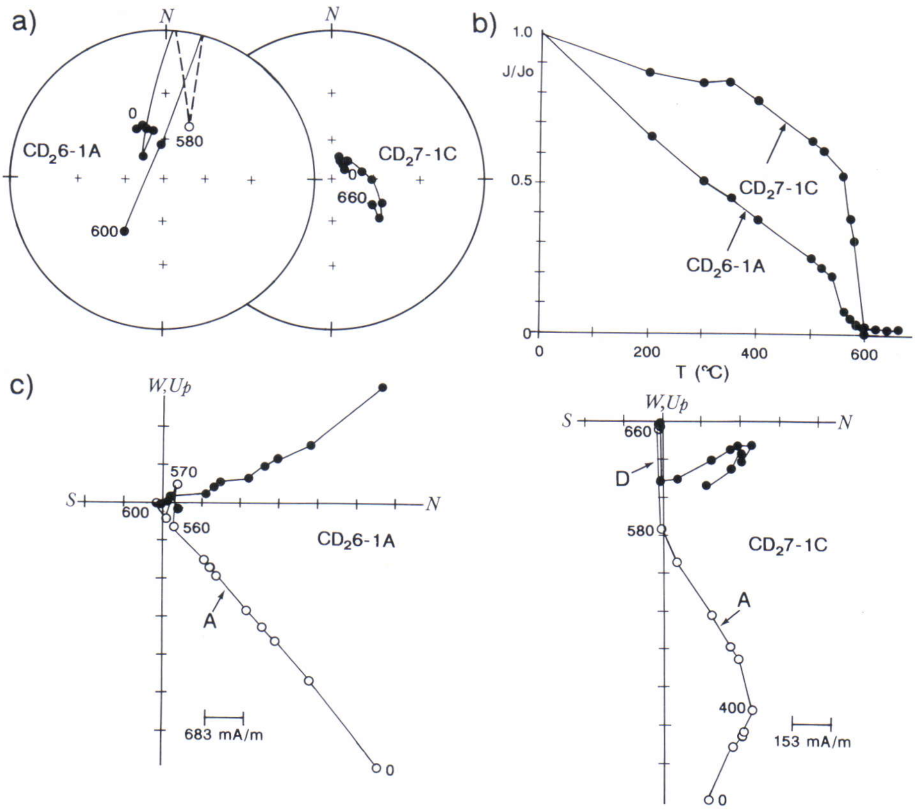

Fig. 10. Examples of thermal demagnetization behaviour for Suoperä dyke CD_2. Note the difference in the intensity decay curves for the margin (specimen CD_26-1A) and interior (specimen CD_27-1C) samples. Symbols as in Figs. 6 and 9.

arately for AF and thermally demagnetized samples, but the results do not differ from each other. Hence, all directions were used for the mean calculations (Table 1).

Samples from dyke DD, demagnetized with AF only, resemble those of dyke TD. Yet, as shown by the straight line on the Zijderveld plot in intermediate AF demagnetization steps (e.g. specimen DD2-1B, Fig. 8), dyke DD may also carry component B. Owing to the curved trajectory in most of the samples, and thus the uncertainty in separating the PEF, A and B directions these results should be regarded with caution.

4.3.2.2. Lake Pääjärvi area (Lupzenga Island)

The dykes at site XD on Lupzenga Island (Figs. 2 and 4) were cleaned in the same way as Suoperä dykes. The mean directions are shown in Figure 5 and Table 2. The remanence behaviour of the gabbronorite and the thin diabase dykes cutting the gabbronorite is almost identical to that of dykes BD, SD and CD_2 in Suoperä. Examples of AF demagnetization for the gabbronorite and diabase samples are shown in Figure 11. In the gabbronorite an intermediate direction similar to that in dyke BD in Suoperä is observed between A and D. However, the vector plot

Table 2. Mean palaeomagnetic data on the Lupzenga (XD) dykes and host granite (Lat=66.2°N, Long=30.9°E)

Dyke	B/N/n	D (°)	I (°)	α95 (°)	k	Plat (°N)	Plong (°E)	dp (°)	dm (°)
Component A									
granite	1/4*/9	343.5	39.5	22.6	18	44.9	232.7	16.2	27.1
gabbronorite	1/7*/11	357.3	51.6	11.1	30	56.0	215.0	10.3	15.2
diabase	1/6*/16	8.6	48.0	6.5	107	52.5	198.5	5.6	8.5
older Jatulian	1/6*/14	349.5	43.1	9.8	48	48.3	225.3	7.5	12.2
younger Jatulian	1/4*/7	3.5	46.8	5.4	288	51.8	206.0	4.5	7.0
MEAN	5*/27/57	356.0	46.2	8.1	90	51.4	216.2	A95 =9.4	
Component D									
gabbronorite	1/5*/9	108.3	36.8	10.5	54	-11.8	276.4	7.2	12.3
granite, cont.	1/3*/5	112.0	41.0	20.1	39	-13.1	271.7	14.8	24.4
diabase	1/5*/11	84.7	39.2	10.2	57	-22.3	296.3	7.3	12.2

B/N/n is the number of formations/samples/specimens. For other explanations, see Table 1.

decays straight to the origin and is not curved at higher coercivities as in dyke BD. AF demagnetization of the diabases reveals two components: a low coercivity component D and a high coercivity component A, which is the opposite of the thermal behaviour of dyke BD, for instance. The Zijderveld trajectory is, however, curved and may indicate simultaneous removal of two components. Therefore the low-coercivity directions were not used in the mean calculations.

Thermal demagnetization of the gabbronorites and diabases reveals two components both in the samples that were stepwise heated after AF demagnetization and in those that were only thermally cleaned (Fig. 12). The D component unblocks within the narrow temperature range of 560-580°C. Component A unblocks below 560°C. In some samples, illustrated by specimen XD34-2C (Fig. 12), the orthogonal plot shows a z-shaped trajectory similar to that seen in samples from Suoperä, although the A directions are more scattered in the samples from Lupzenga. The directions within samples are consistent, but some of the samples have declinations of up to 30°, which is unusual for the typically well-grouped A directions with a declination of around 350°. It is thus possible that there are still two unresolved populations of components. However, because they unblock in the same temperature range, they have been grouped together with the typical NW direction of A to give an overall mean.

All samples from the two crosscutting Fe-tholeiitic gabbro diabase dykes ("Older and Younger Jatulian dykes", Fig. 4) on Lupzenga Island carry the A component whether demagnetized with AF or thermally. Most of the thermally demagnetized samples reveal only the single A component up to temperatures of 580°C. In particular, the "Older Jatulian dyke" shows a very consistent A direction, resembling dyke TD in Suoperä.

In one thermally demagnetized sample of each of the "Old and Young Jatulian dykes" (Fig. 13), another component, close to the direction of component B, was erased above the tem-

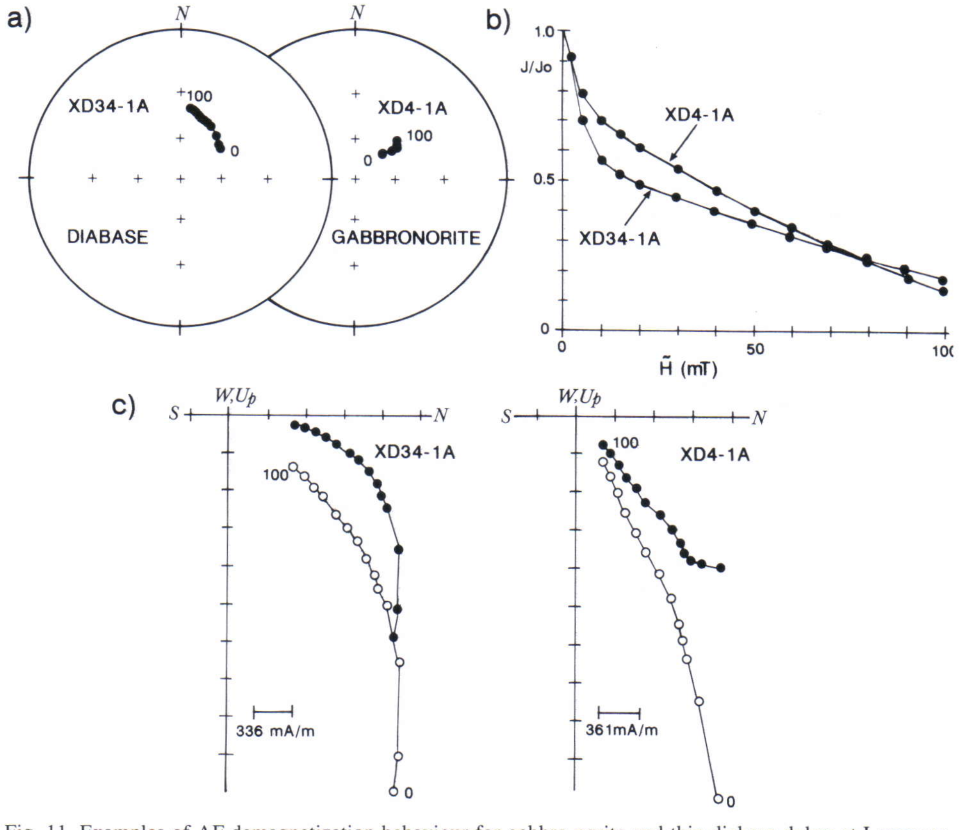

Fig. 11. Examples of AF demagnetization behaviour for gabbro norite and thin diabase dykes at Lupzenga Island. Note the similar intensity decay curves and the curved Zijderveld trajectory for the diabase. Symbols as in Fig. 6.

perature of the A component. In the older dyke, this direction, isolated between 560° and 567°C, is D=68° and I=63°. In the younger dyke, erased between 560° and 570°C, the direction is D=40° and I=55°. Because these directions were obtained in only one specimen from each dyke, they are not used further. Unlike the gabbronorites and the thin diabase dykes, component D was not found in these "Jatulian" dykes, even though most of the specimens were thermally demagnetized in 3° steps through the unblocking range typical of D. Gabbronorite samples taken from the contact with the older dyke reveal component A in most samples. One baked sample exhibits a higher T_{UB} component close to the direction revealed in the older dyke. But, conclusive interpretations about the primary nature of the high T_{UB} component of the older dyke cannot be drawn from only one sample. The same applies to the baked contact test of the younger dyke, as only one sample shows results indicative for a positive baked contact test.

To determine whether component D in the gabbronorite is primary, four samples were taken from the Archaean host granite within a few centimetres - one metre of the gabbronorite. They show stable behaviour, and component D is clearly isolated in the temperature range of 540-580°C (Fig. 14). The granite also carries the A overprint, although more scattered due to the weak intensities which typically prevent

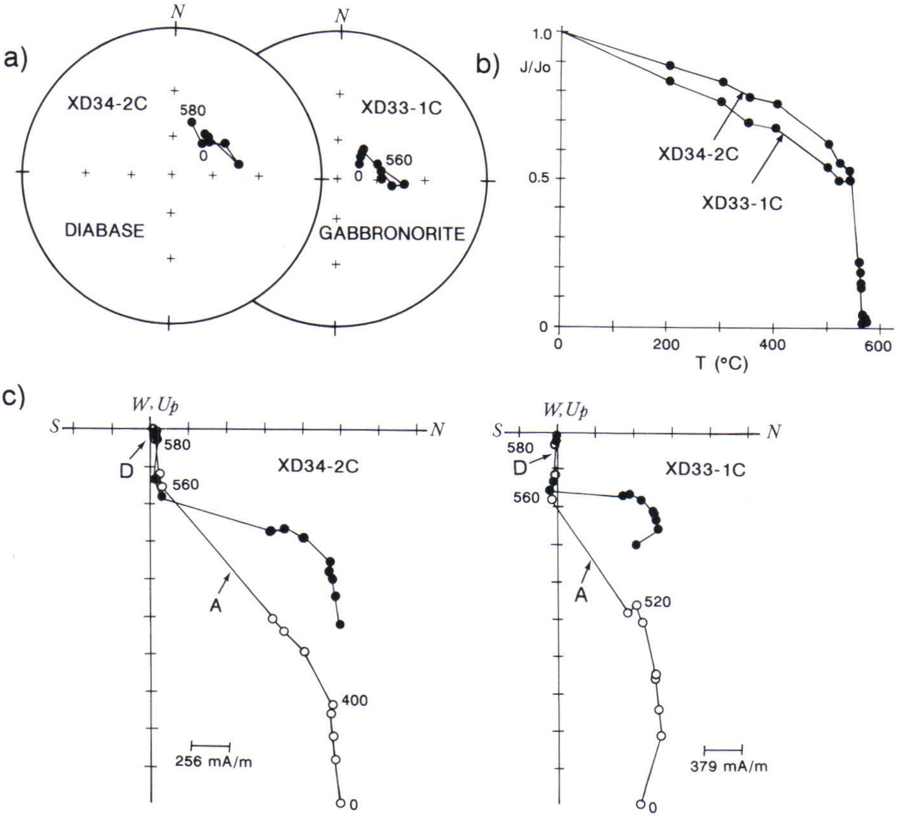

Fig. 12. Examples of thermal demagnetization behaviour for gabbronorite and thin diabase dykes at Lupzenga Island. Symbols as in Figs. 6 and 9.

granites from being good recorders of the past magnetic field. Samples were not collected from the unbaked granite, and so the results of the baked contact test are inconclusive.

Even so, the absence of component D from crosscutting "Jatulian" dykes supports the interpretation that component D is the primary thermoremanence in the gabbronorites, a concept that is further supported by the occurrence of D only in the samples with the highest coercivities and a narrow T_{UB} range between 560 and 580°C, which is typical of the thermoremanence of igneous rocks (e.g. Butler, 1992). As all the dykes and host rocks studied carry a similar A direction, A must be a regional overprint.

4.4. Comparison of 2.4 - 2.0 Ga palaeopoles in Fennoscandia

Palaeomagnetic data on the Suoperä gabbronorite dykes and the Lupzenga dykes are listed in Table 1 and Table 2, respectively. Mean palaeopoles obtained here and in Paper I are given in Table 3. Palaeomagnetic poles from the Russian Karelian dykes and other published works are plotted in Figure 15. The grouping of D poles, thought to be about 2.44 Ga in age, is discussed in this section. The A poles, which are associated with the Svecofen-

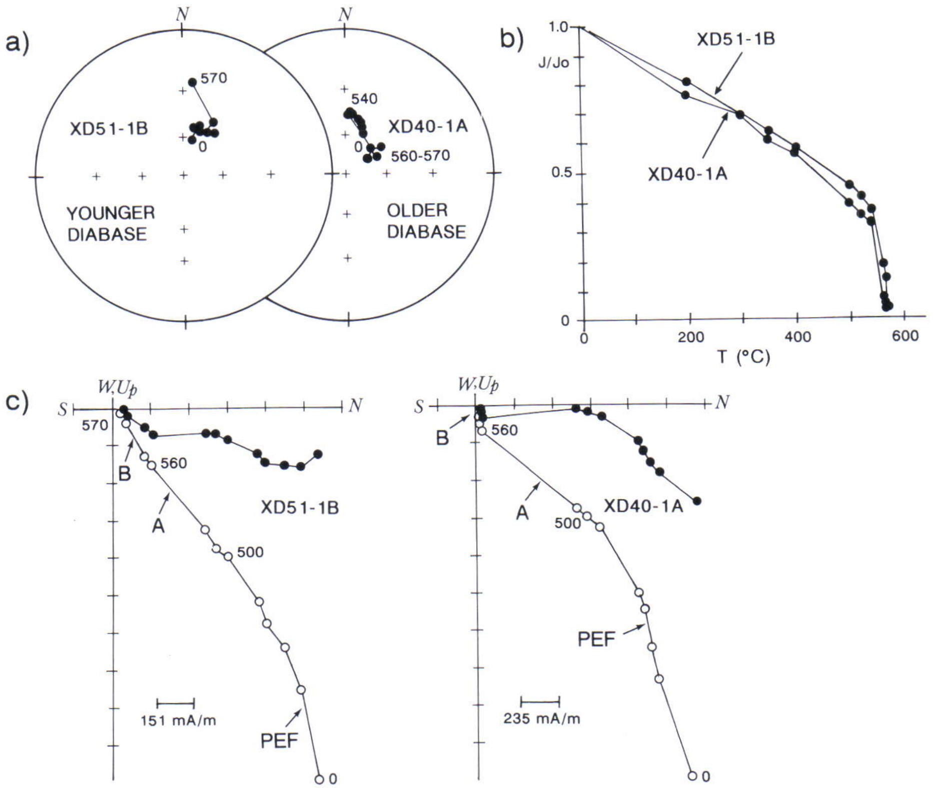

Fig. 13. Examples of thermal demagnetization behaviour for "Younger and Older Jatulian dykes". Note the more discrete unblocking temperatures than in gabbronorites and diabases in Fig. 12. Symbols as in Figs. 6 and 9.

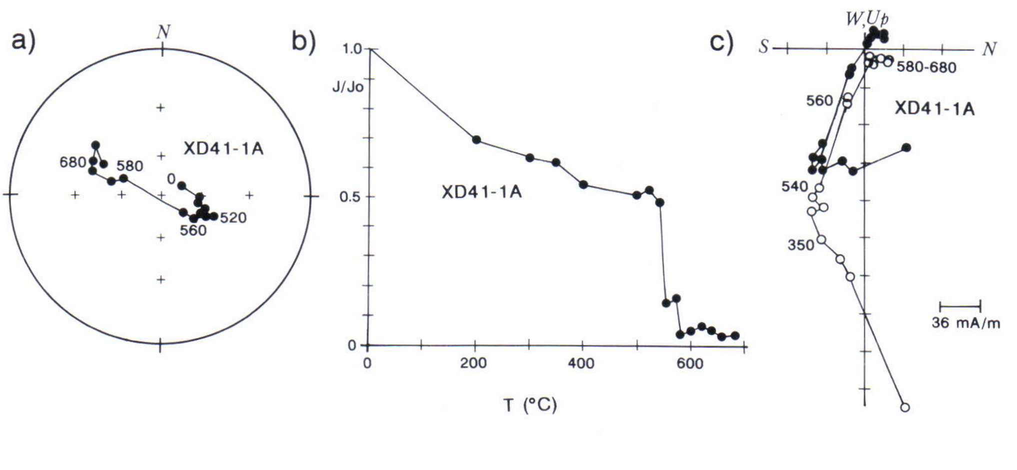

Fig. 14. Thermal demagnetization of a sample from baked Archaean granite, ca. 1 m from the contact with the gabbronorite at Lupzenga Island. Symbols as in Figs. 6 and 9.

Table 3. Mean palaeomagnetic poles and reference directions

Formation	Comp.	B/N/n	D'	I'	P_{lat}	P_{long}	A95	λref
Suoperä (SP)	A	5*/5/35	345.5	51.8	57.1	230.6	6.1	32
Lake Pääjärvi (XD)	A	5*/5/27	354.1	43.9	51.4	216.2	9.4	26
Mean XD/SP	A	10*/10/62	349.9	48.0	54.4	222.9	5.7	29
Koillismaa (L)	A	5*/32/105	347.5	36.7	45.5	224.5	5.3	20
Koitelainen (K)[1)]	A	1/3*/14	339.7	36.1	43.9	234.6	8.6	-20
Mean XD/SP	D	3*/3/17	100.3	46.3	-20.4	276.1	13.5	-28
XD - diabase	D	4/4/5*	81.1	36.9	-22.3	296.3	9.4	-21
Koillismaa (L)[1)]	D	4*/22/62	111.2	51.3	-20.0	265.0	22.0	-32

Formation at Suoperä (SP) and at Lake Pääjärvi (XD) denotes gabbronorite dyke (except XD - diabase), at Koillismaa and Koitelainen, layered intrusions (Paper I). Letters in parentheses refer to poles in Fig. 15. [1)] = Paper I, other poles from this study. Comp. = remanence component. Components A and D show normal and reversed polarities, respectively. B/N/n = number of dykes or intrusions/sites/samples. * denotes the statistical level used in mean calculations. D', I' = reference declination and inclination, calculated from the mean palaeomagnetic poles with respect to the reference location (Kajaani, 64.1°N, 27.7°E). P_{lat}, P_{long} = latitude (°E), longitude (°N) of the mean pole position, A95 = the radius of the circle of 95% confidence of the mean poles. λref = reference palaeolatitude.

nian orogeny, are considered in section 4.5.3.

The poles from dyke BD and SD of Suoperä and from the gabbronorite dyke (pole XD-G) at Lupzenga were used to calculate the mean pole, XD/SP, for component D (Fig. 15, Table 3). On the basis of similar magnetic behaviour, petrography and geochemistry (J. Vuollo, pers. comm., 1995) these dykes probably originated from the same magma source. Pole XD-d (Fig. 15, Table 3) from the thin diabases was not included in the mean D direction, because, although similar in geochemistry, the diabases are slightly younger than the gabbronorites, as suggested by the crosscutting relations. In addition, the pole from dyke CD_2 was not included in the mean D direction because it was obtained only from one sample (Table 1).

Dyke BD yields a U-Pb baddeleyite age of 2446+5/-4 Ma (J. Vuollo, pers. comm., 1995). Palaeopole XD/SP of component D from the dykes (Table 3) is close to the pole L (Fig. 15) from the Koillismaa layered intrusions (Paper I), which have a mean U-Pb zircon age of 2436±5 Ma (Alapieti, 1982). The high scatter of pole L is probably due to the small number of entries (5) in the mean calculation of individual intrusions. The scattered data reflect differences of direction among the intrusions, possibly indicating uncorrected local tectonic movement. On the basis of the similarity in the pole positions of XD/SP and L it is suggested that palaeopole D of the Karelian dykes dates to about 2.44 Ga.

Further support for a ca. 2.44 Ga age for the palaeopoles of the 2.44 Ga old layered intrusions and dykes is provided by palaeomagnetic studies conducted elsewhere in Russian Karelia. Krasnova and Gus'kova (1991) obtained similar poles from gabbronorites and mafic intrusions east of Lake Pääjärvi. They isolated component D (called C in their study) in the oldest dykes, but not in younger crosscutting dykes. They also obtained the same direction from a gabbro pluton and from the Archaean basement as an overprint. Pole Pj (Fig. 15) is the mean of these poles. The inclination of direction Pj (Krasnova and Gus'kova, 1991) is higher than that of the D component of this study, possibly reflecting incomplete removal of a young viscous component.

Most palaeomagnetic studies of Fennoscandia have claimed that there are still too few

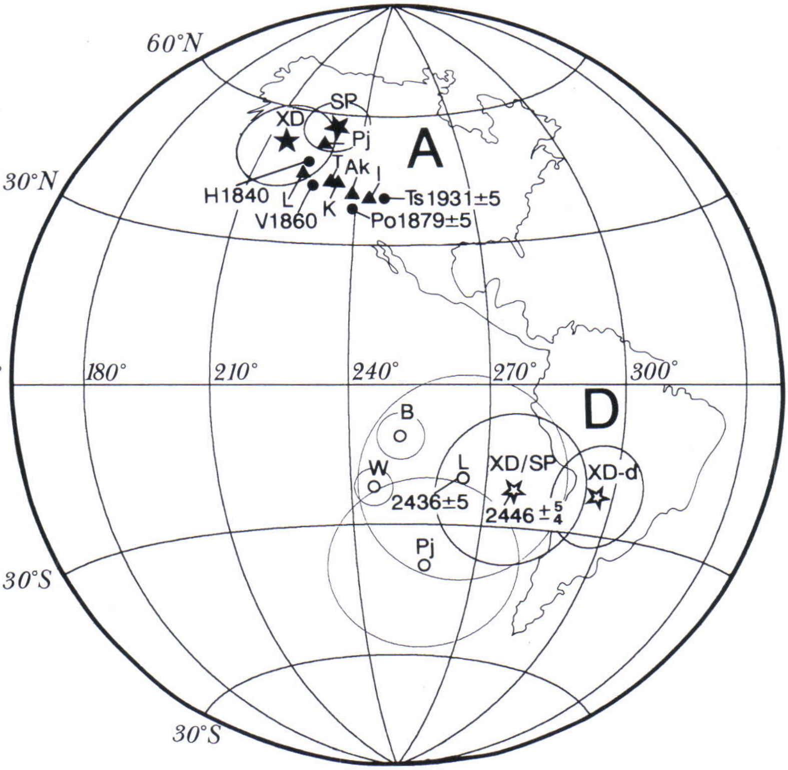

Fig. 15. Pole positions of units with U-Pb zircon or baddeleyite ages (Ma). Cones of circular standard deviation are shown only for poles of this study (stars) and for poles in Russia with a proposed age of 2.4 Ga. Letters A and D denote the remanence components. Primary poles obtained in other studies are shown as circles and secondary poles as triangles. Closed (open) symbols denote normal (reversed) polarity. XD/SP = mean of gabbronorite dykes at Suoperä and at Lake Pääjärvi, XD-d = diabase dykes at Lake Pääjärvi, L = Koillismaa layered intrusions (Paper I), B = Burakovskaya dyke, W = Vodlozero dyke (Krasnova and Gus'kova (1990), Pj = mean D and A components from east of Lake Pjaozero (Krasnova and Gus'kova (1991), Ts = Tsuomasvarri gabbro-diorite (Paper II), Po = Pohjanmaa gabbro-diorites (Pesonen and Stigzelius, 1972), V = Vittangi gabbro (Elming, 1985,1994), H = Haukivesi lamprophyre dykes (Neuvonen et al., 1981), I = Archaean Isokivennokka granite (S. Mertanen, unpublished data), T = Tsuomasvarri ultramafic intrusion (Paper II), Ak = Akanvaara gabbro (Mertanen and Pesonen, 1992), K = Koitelainen layered intrusion (Paper I), XD = mean Svecofennian overprint pole at Lake Pääjärvi, SP = mean Svecofennian overprint pole at Suoperä.

definitive palaeomagnetic data with precise isotopic ages to show relative movements within the Fennoscandian Shield in Archaean - Palaeoproterozoic time (Pesonen et al., 1989; Elming et al., 1993). Although this question still remains unresolved, there are indications favouring a single stable Karelian craton since 2.4 Ga. The concept is based on palaeomagnetic results obtained from the Burakovskaya and Koillismaa layered intrusions and associated dykes, representing two widely separated areas of the Karelian craton.

The Burakovskaya layered intrusion (2365±90 Ma, Sm-Nd mineral age, Amelin and Semenov, 1990) lies in the southeastern part of the Karelian Province (see Fig. 1, Paper II) and is coeval with the Koillismaa-Olanga layered complexes in the northwestern part of the province. The Burakovskaya intrusion consists of one large intrusion, small satellite intrusions and a swarm of mafic dykes. The complex is located in the Vodloezero block, east of Lake Onega in Russia. The Vodloezro block represents an Archaean high grade metamorphic terrane which, due to dry metamorphic conditions and low fluid activity, has preserved the primary remanence directions in most of its formations (Krasnova and Gus'kova, 1990). For instance, the Svecofennian palaeomagnetic overprint has not been found in this area and K-Ar dating of basement rocks gave only Archaean ages (Krasnova and Gus'kova,1990).

The palaeomagnetic directions of the main Burakovskaya intrusion itself are statistically too poorly defined to give satisfactory results for correlation. However, poles B (Burakovskaya dyke) and W (Vodloezero dyke) in Figure 15 were obtained from gabbro dolerite dykes genetically related to the Burakovskaya layered intrusion (Krasnova and Gus'kova,1990). The dykes have not been isotopically dated, but one of them carries a down SE remanence similar to that of the dated Burakovskaya intrusion and shows a positive baked contact test against an Archaean diabase dyke. From the similarity in the remanence directions of the dykes, Krasnova and Gus'kova (1990) conclude that the poles are 2.4 Ga in age.

The above palaeomagnetic results have important implications for tectonism within the Karelian Province. They demonstrate that no large-scale block rotations have taken place since 2.4 Ga. However, one cannot rule out horizontal lateral displacements without significant rotation as these would not be recorded in the remanence directions and pole positions. On the other hand, vertical movements, such as uplifts of adjacent blocks at different rates, are recorded in the palaeomagnetic directions (see Paper IV). The more westerly position of the poles from the Vodlozero block than those from the Pjaozero block may indicate either block movement or APW due to slight age differences between the intrusions in the two areas. Because the confidence cones (A95) of the poles intersect, the difference is within the limits of error. Hence, before definitive conclusions can be made about relative movements between different areas within the Karelian Province, more precisely-dated and well-defined palaeopoles are required.

Palaeomagnetic studies of relative tectonic movements between different provinces, e.g. Karelian, Kola and/or Belomorian Provinces, are still sparse. Some palaeomagnetic studies on the Belomorian belt have been carried out by Russian geophysicists, but so far the data are not precise enough for correlations to be made. An attempt to study intraplate movements between the Kola and Karelian Provinces was made in the study of the Tsuomasvarri area (Paper II), but due to the difference in age of the formations studied, no conclusions about the tectonism could be drawn. Torsvik and Meert (1995) have studied Palaeoproterozoic formations in the Pechenga zone, where the General'skaya Gora layered gabbro intrusion (Fig. 1, Paper II) of 2453±42 Ma (Sm-Nd, Bakushkin et al., 1990) is overlain by a full sequence of volcano-sedimentary rocks with ages between 2.3 and 1.8 Ga (Balashov et al., 1990; Hanski et al., 1990; Mitrofanov et al., 1991). Unfortunately, the General'skaya Gora intrusion is totally remagnetized, and so the 2.4 Ga magnetization could not be isolated (Torsvik and Meert, 1995). The remanence direction (component B1) of General'skaya Gora is close to typical Svecofennian directions, although Torsvik and Meert (1995) combine it with another secondary direction (component B2), which they relate to a Late Precambrian remagnetization event (see below).

In Paper II doubt was cast on the 2.4 Ga age of a pole that had previously (Paper I) been interpreted as primary in the Koillismaa intrusions, because it corresponded to the pole that was obtained in the Tsuomasvarri ultramafic intrusion, which is younger than 1.93 Ga. However, Torsvik and Meert (1995) suggest that in the Pechenga area, about 100 km southeast of Tsuomasvarri, a component which is close to the primary 2.44 Ga old D of this study represents a Late Precambrian (Vendian) remagnetization. It may be related to the emplacement of Vendian dykes with 546-580 Ma K/Ar and $^{40}Ar/^{39}Ar$ mineral ages in Varanger and Kola Peninsulas and which have similar palaeomagnetic directions (Torsvik et al., 1995). Since Paper II was written, new data has improved Late Precambrian - Palaeozoic APWP. Pole D from Tsuomasvarri now falls near the ca. 650 Ma segment of the path. Consequently, it now seems likely that the D component is a secondary overprint in the Tsuomasvarri intrusion but a primary one in the Koillismaa intrusions and Russian Karelian dykes.

Despite the results of the above new preliminary studies on the Russian Karelian dykes and those of previous studies in Fennoscandia, the Jatulian (2.2-2.0 Ga) remanent magnetization direction has still not been resolved. A Jatulian component was not clearly isolated in the Suoperä and Lake Pääjärvi dykes, although it may exist as component B (see Table 1). Krasnova and Gus'kova (1991) have isolated a similar component in both the youngest gabbro diabases and the oldest gabbronorites east of Lake Pääjärvi, as well as in Archaean basement rocks. They interpret it as a regional remagnetization at the time of emplacement of the gabbro diabases. However, there are no isotopic ages for the dykes or positive field tests to confirm that the remanence is primary. Evidently, more palaeomagnetic data are needed to determine the Jatulian remanent magnetization direction, and hence, position of the Fennoscandian Shield in Jatulian time.

4.5. Svecofennian orogeny

4.5.1 Geological outline

In recent years, several models have been proposed for the evolution of Svecofennian orogenic events (see e.g. Gaál and Gorbatschev, 1987; Gaál, 1990; Ekdahl, 1993; Korja et al., 1994; Lahtinen, 1994; Korja, 1995). In the Svecofennian orogeny new juvenile crust was formed in a collision between the Svecofennian island arc complex from the SW and the Archaean Karelian continent from the NE at about 1.9 -1.87 Ga ago (Huhma, 1986). The oceanic island arcs were produced by early subduction of oceanic crust at 1.93-1.9 Ga (Gaál, 1990). In the collision, Karelian rocks (Kalevian metaturbidites and 1.97 Ga ophiolites) were thrust eastwards onto the Archaean craton (Koistinen, 1981; Kohonen, 1995) with the results that in the collision zone, the thickened crust was composed of a mixture of Archaean, Karelian and Svecofennian material. The collision led to synkinematic deformation (1.89-1.88 Ga) which was accompanied by low-pressure - high temperature metamorphim, mostly at amphibolite facies and locally at granulite facies (Korsman et al., 1984; Korsman, 1988). The deformed rocks were intruded by synkinematic granitoids at 1.89-1.87 Ga. Subsequently, the northern parts of the island arc system were stabilized about 1.88-1.85 Ga ago (Vaasjoki and Sakko, 1988; Vaasjoki et al., 1994). The southern Svecofennian Domain experienced granitic magmatism and high grade metamorphism about 1.84 -1.81 Ga ago due to a heat flux which is attributed to magmatic underplating (Korja et al., 1993) as the collision was followed by an extensional stage causing thinned crust in its lower parts. The youngest, postkinematic granites were emplaced in southern Finland at 1.80-1.79 Ga (Vaasjoki et al., 1994). Later, in the Subjotnian time (1.65-1.54 Ga), the southern Svecofennian Domain was intruded by rapakivi granites and associated diabase and quartz porphyry dykes (Haapala and Rämö, 1992; see also Paper III).

The effect of the heat flux associated with the synkinematic and late kinematic Svecofennian orogenic events in the eastern and northern Karelian Province is not so well identified as it is in the Svecofennian Domain. According to Gaál and Gorbatschev (1987), the Svecofennian orogeny mainly affected areas close to the orogenic belt, and was elsewhere restricted to rifts and fractures and to mafic igneous activity. Recent studies have shown that the collision caused a rise in temperature in the Archaean craton, too (Korja et al., 1994). The collision is recorded in a variety of ways. The Archaean komatiites show disturbed REE compositions (Gruau et al., 1992), and the oxygen and hydrogen systematics of the gold mineralizations hosted by the greenstone belt were disturbed (Karhu et al., 1993). K-Ar biotite and hornblende ages were reset due to heating of the crust within the Late Archaean area of eastern Finland (Kontinen et al., 1992). Furthermore, as shown here, the primary remanent magnetization directions of the Archaean and Palaeoproterozoic pre-Svecofennian rocks were systematically overprinted due to the Svecofennian orogeny.

4.5.2. Effect of the Svecofennian orogeny on palaeomagnetic directions

The Suoperä and Lake Pääjärvi gabbronorite dykes in Russia are used as an example to illustrate the mechanism of Svecofennian remagnetization. The dykes are ideal for this purpose because, in addition to the primary component D, they show the dominance of the Svecofennian remanence (component A). This has been separated in all samples of different rock types throughout the whole area and thus probably represents a regional overprint. The direction of A corresponds to the Svecofennian direction isolated elsewhere in Fennoscandia (see later in this section).

Type of remagnetization

The fact that the unblocking temperatures of component A are lower than those of component D indicates that A is related to thermal activation of the crust. The Koillismaa layered intrusions behave in a similar manner to the gabbronorite dykes in Russia. Thermal demagnetization gives low to intermediate T_{UB}'s for component A and high T_{UB}'s for component D, with the latter component isolated in a narrow unblocking temperature spectrum (see e.g. Fig. 10, Paper I). Some of the Suoperä and Lake Pääjärvi dykes are completely overprinted by A, whilst others show only a partial overprint.

Component A could represent purely a partial thermoremanence acquired during uplift and cooling following the Svecofennian orogeny. The most typical T_{UB}'s for component A are between 400° and 560°C. Thus, if component A represents pTRM, the temperature of metamorphism was high enough to unblock the grains with T_{UB}' s below 560°C but insufficient to reset those above 560°C. Theoretical calculations for pure SD magnetite (Pullaiah et al., 1975) show that a T_{UB} of 560°C requires a peak reheating temperature of about 550°C if maintained for about 10 Ma. Calculations for samples containing a variety of grain sizes of pure magnetite (Middleton and Schmidt, 1982) suggest a maximum reheating temperature of about 450°C. The temperatures are reasonable in regard to the amphibolite facies metamorphic grade proposed for the Karelian Province (Tuisku and Laajoki, 1990). However, certain features of the demagnetization results indicate that A does not represent a pure partial thermoremanence, but that thermochemical activation is a more probable mechanism for the acquisition of component A.

In some thermally demagnetized dykes T_{UB}'s of component A occupy intermediate temperature ranges, whereas in the lower and higher temperature intervals, component D is erased (Figs. 9 and 12). The same feature was observed in the layered intrusions (see Fig. 13, Paper I). If the A overprint was pure pTRM, it would have reset all grains with T_{UB}'s below the peak metamorphic temperature (ca. 550°C, at least in Koillismaa and Russian Karelia). Therefore, the A component is likely a CRM or pTCRM. CRM may be acquired at temperatures as low as ca. 300°C (e.g. Butler, 1992), that is, far below the peak T_{UB}. The partial overlap of T_{UB} spectra of components A and D, seen in most samples, may not, however, be as strong an indication of CRM. As McClelland-Brown (1982) pointed out, the presence of two coexisting minerals or the coexistence of SD and MD states of the same mineral can cause overlap of pTRM spectra, too. In the Suoperä and Lake Pääjärvi samples, magnetite is the only remanence carrying Fe-oxide mineral. But, as evidenced by petrographic studies, the grain sizes are heterogeneous, composed of visible MD states and probably of SD/PSD states. Thus it is difficult to distinguish with certainty between pTRM and CRM. Hopefully, the hysteresis and SIRM studies, to be conducted in the future on the Karelian dykes, will shed more light on the magnetic states of the remanence carrying minerals.

Petrography

For the preliminary study of Russian dykes, at least two polished thin sections were studied from each dyke. The degree of alteration was determined mainly on silicate minerals. On the basis of thermal demagnetization behaviour, it was expected that the samples with a dominantly Svecofennian direction would be more altered than those exhibiting mainly primary remanence. Such a correlation is indeed seen, for instance, in dyke SD from the Suoperä area, where the alteration is clearly related to hydrothermal activity and retrogressive metamorphism, as evidenced by the complete replacement of pyroxenes by amphiboles (typically blue-green hornblende in addition to light green primary amphibole), chlorite and biotite, sericitization of plagioclase and formation of new magnetite. Specimen SD3-1C (Fig. 6) represents just such a highly altered sample and carries only component A. At site CD, the older dyke, CD_2, at the contact of the younger one, CD_1, is thoroughly altered and carries only component A. Far from the contact, however, the CD_2 dyke is fresh and carries the D component, as well as superimposed A. A positive correlation between hydrothermal alteration and remagnetization was also demonstrated in the Subjotnian diabase dykes of Sipoo (Paper III), where the most altered dyke was heavily overprinted (shown also as decreased susceptibility and remanence intensity values) and the fresh dykes carried the primary remanence.

There are some cases in which remagnetization does not appear to be correlated with the degree of alteration. Some samples exhibiting only the Svecofennian overprint have primary minerals which are only slightly altered, for instance in dyke TD, where plagioclase is fresh and the pyroxenes show only minor uralitization at the grain margins. Consequently, the occurrence of the Svecofennian overprint cannot always be directly related to circulation of hydrothermal fluids and alteration of the rock. Conversely, heavily-altered rocks may also be the carriers of the primary remanent magnetization, as is evident, for instance, in dyke SD, which exhibits a small primary D component in addition to a large A component (Fig. 9). One explanation for the preservation of the primary D component may be in the small disseminated magnetite inclusions in plagioclase which give this mineral a cloudy, brownish appearance (see e.g. Halls and Palmer, 1990). On the other hand, if the alteration was partly caused by deuteric reactions shortly after emplacement, the resulting remanence would have had essentially the same direction as the primary D remanence. However, because all samples carry superimposed A component, it is more likely that the main alteration is due to hydrothermal activity related to the Svecofennian orogeny.

In most samples the only indication of secondary growth of magnetite is the intergrowth of magnetite with biotite, which has replaced pyroxenes at the grain margins. Another indication of secondary, newly formed magnetite is the presence of small disseminated grains within amphiboles, which may be responsible for the high coercivities of the A component. The largest magnetite grains are most probably the carriers of the viscous PEF component (see Papers II, III and IV).

Mechanism of Svecofennian remagnetization

Observations made during this study and palaeomagnetic studies conducted elsewhere in Fennoscandia (see below) suggest that the dominant mechanism of Svecofennian remagnetization is thermochemical resetting introduced by warm hydrothermal fluids. The fluid flux is spatially restricted, and thus the impact of remagnetization may vary within areas and even within a single dyke. For instance, as pointed out by Halls (1991), the dyke margins (e.g. those of CD_1 and CD_2 in Suoperä and gabbronorite dyke and "Older Jatulian" dyke in Lupzenga) may be more vulnerable to remagnetization if they are fractured. A corresponding case is found in the Varpaisjärvi area, in the Raahe-Ladoga belt, where 2680 Ma old granulite-grade enderbites (Paavola, 1984) are cut by ca. 2.1 Ga old Jatulian dykes. At a distance from the dykes the enderbites have retained their primary Archaean remanence (Neuvonen et al., 1981) but near the dyke margins, the enderbites exhibit the Svecofennian remagnetization, which totally overprints the Jatulian direction (S. Mertanen, unpublished data). This type of remagnetization may be due to intense fluid flux which is localized in altered zones around metaintrusive bodies as a result of stress variations and strain incompatibility effects related to rheological contrasts between intrusions and host rocks (see Oliver et al., 1990). Hydrothermal activity and retrogressive mineral reactions typically take place at shear zones (e.g. Etheridge et al., 1983; McCaig, 1988). The Svecofennian overprint at and near the dyke margins in Russian Karelia and Varpaisjärvi area suggests that the dykes have provided channels for free movement of the fluids responsible for the remagnetization.

The sources of the fluid or fluid paths, which are a subject beyond the scope of this paper, comprise surface derived meteoric waters, deep-seated metamorphic or magmatic waters, and mixture of surface-derived and metamorphic waters (see e.g. McCaig, 1988). In recent studies, remagnetization was used to date fluid migration and orogeny (Oliver, 1986; Miller and Kent, 1988; McCabe and Elmore, 1989; Kotzer et al., 1992). Studies on young Palaeozoic orogenic belts, e.g. the Appalachians in eastern North America, have suggested that fluids released from the orogenic belt have moved laterally for several hundreds of kilometres (Oliver, 1986). The lateral migration of orogenic fluids may result from movements of thrust sheets at the collision zone which drive fluids toward the craton. Consequently, a low-temperature chemical remagnetization which is spatially restricted due to channelling of the fluids, may be acquired (op. cit). Such a mechanism could be a cause of the Svecofennian remagnetization, too. However, more work is still needed to prove this hypothesis.

Because the Karelian Province has experienced amphibolite facies metamorphism (Korsman et al., 1984; Tuisku and Laajoki, 1990) and the remagnetization is recorded over a large region, a further cause of remagnetization must be uplift and cooling of the crust following the Svecofennian orogeny. For instance, preservation of the Archaean remanence direction in the Archaean basement rocks of the Soilu area, close to the

Koillismaa layered intrusions, was demonstrated in Paper I, but in most samples the Archaean direction is partially overprinted by the Svecofennian direction (S. Mertanen, unpublished data). According to Neuvonen (1975; 1992), the Archaean basement rocks outside the Varpaisjärvi granulite grade metamorphic block have been totally remagnetized in the Svecofennian direction. Inside the Varpaisjärvi block, the rocks yield only Archaean K-Ar biotite and hornblende ages (Kontinen et al., 1992) implying that prevailing dry conditions were unable to unblock the biotite (blocking at ca. 300°C) and hornblende (blocking at ca. 500°C) systems. Kontinen et al. (1992) have suggested that the mechanism for the reset K-Ar ages of the Archaean basement rocks was deep burial as a consequence of thrust nappes of the orogen. However, far from the collision zone, amphibolite-facies metamorphism and remagnetization are not necessarily due to deep burial alone; hydrothermal fluids may have played an important role in metamorphism and remagnetization of the whole Archaean basement. Because the Svecofennian remagnetization most probably represents pTCRM, the T_{UB}'s are not directly related to the peak metamorphic temperatures attained during the Svecofennian orogeny.

4.5.3. Timing of the Svecofennian remagnetization

Due to collision of the pre-Svecofennian island arc with the Archaean craton, the Svecofennian orogeny has probably been the most conspicuous modifier of remanent magnetizations in the Fennoscandian Shield. However, considering the many tectonomagmatic events that took place simultaneously in the northeastern part of the Shield over a time interval of 1.9-1.8 Ga, the Svecofennian orogeny in the main collisional zone was probably not the only cause of remagnetizations. For example, collisional events at ca. 1.9 Ga in the Lapland granulite belt and in the Polmak-Pasvik-Pechenga belt of the Kola Peninsula probably affected the remanent magnetizations in the northern parts of the Karelian and Kola Provinces. High-grade metamorphism also took place in the Belomorian belt at about 1.9 Ma concomittantly with the high-grade metamorphism of the Lapland granulite belt (e.g. Balagansky et al., 1994; Bogdanova, 1994; Bridgwater et al., 1994).

The timing of the Svecofennian remagnetization in the central Karelian Province can be related to the resetting of K-Ar ages of the Archaean basement, where the average reset age of hornblende is 1850 Ma and of biotite 1795 Ma (Kontinen et al., 1992). Tectonometamorphic studies and isotopic age determinations have shown that the temperature rose in eastern Fennoscandia in the post-orogenic stage of the Svecofennian orogeny approximately 30 Ma after the collision (e.g. Korja et al., 1994). According to Pajunen (1993) and Korja et al. (1994), fluid-induced metasomatic alteration characterizes Proterozoic shear and fracture zones in the southern part of the Archaean basement complex of central Finland. The occurrence of cordierite-sillimanite-bearing assemblages in these fracture zones indicates that the peak metamorphic temperature was ca. 600°C. The metamorphic-metasomatic alteration has been dated at 1852±2 Ma (U-Pb) on xenotime (YPO_4), which crystallized contemporaneously with cordierite (Pajunen, 1993). U-Pb ages of monazite and titanite indicate that, in the Kainuu schist belt, peak amphibolite facies regional metamorphism took place at about 1870 Ma (Laajoki and Tuisku, 1990).

The palaeomagnetic method can only establish an absolute age for the Svecofennian remagnetization by comparing the overprint palaeopoles with palaeopoles of isotopically dated rocks exhibiting well defined remanence directions (see discussion in Paper IV). Most previous palaeomagnetic studies on the Svecofennian orogeny have been carried out on primary remanences of 1.9-1.8 Ga mafic intrusions. The majority of the results have been obtained from the Svecofennian Domain or at the transitional zone between the Svecofennian Domain and Karelian Province - the Raahe-Ladoga Belt (see Neuvonen, et al., 1981; Pesonen et al., 1991, Elming et al., 1995). Palaeomagnetic data have also been reported for units north of the Skellefte District (Piper, 1980; Elming, 1982, 1985, 1994).

Primary Svecofennian palaeopoles

Some well-defined palaeopoles ($\alpha95 < 7°$ and $k > 70$ in all except one) from mafic igneous intrusions and dykes from the Fennoscandian Shield are plotted in Figure 15 (component A). They record the primary TRM acquired during cooling of the intrusions. The formations have been dated by the U-Pb on zircon and the ages range from 1930 to 1840 Ma. The oldest dated palaeopole (Ts, Fig. 15) is from the 1.93 Ga old Tsuomasvarri layered intrusion in the Kola Province (Paper II); the other poles derive mainly from the Raahe-Ladoga and Skellefte Belts or from close to them. Pole Po (Fig. 15) derives from the Pohjanmaa gabbro-diorite intrusions (Pesonen and Stigzelius, 1972), one of the palaeomagnetically studied intrusions dated at 1879±5 Ma (Ekdahl, 1993). According to Pesonen and Stigzelius (1972), the Pohjanmaa gabbro-diorites represent the youngest phase of synkinematic magmatism within the Raahe-Ladoga Belt. Pole V (Fig. 15) derives from the 1860 Ma old synkinematic Vittangi gabbro intrusion north of the Skellefte Belt (Elming, 1985, 1994). Pole H (Fig. 15) is from the 1837-1840 Ma (Huhma, 1981) N-S trending Haukivesi lamprophyre dykes at the Raahe-Ladoga Belt (Neuvonen et al., 1981) representing the post-kinematic stage of the Svecofennian orogeny. These primary palaeopoles define the Svecofennian track of the APWP of the Fennoscandian Shield (Pesonen et al., 1989, Elming et al., 1993, 1995), which trends from SE to NW with decreasing age.

Secondary Svecofennian palaeopoles

Secondary Svecofennian palaeopoles, isolated as overprints (Fig. 15) on older (often primary) directions, have been obtained in many parts of the Kola and Karelian Provinces. Pole I in Figure 15 derives from the Archaean Isokivennokka granite (see Kesola, 1991; Paper II) in the Kola Province, from north of the Kola suture belt. Due to weak NRM intensities and instability of most of the samples, an Archaean remanence could not be isolated. However, the typical Svecofennian remanence was obtained. Pole I is close to the primary pole Ts (Fig. 15) in the Tsuomasvarri gabbronorite intrusion, which lies south of the Kola suture belt (Paper II). Secondary pole T (Fig. 15) in the Tsuomasvarri ultramafic intrusion was interpreted as representing thermochemical remanence acquired during later metamorphic events following emplacement of the intrusion. In the Tsuomasvarri-Isokivennokka area resetting of remanent magnetization may be related to the collision that took place at the Kola suture belt about 1.9-1.8 Ga ago (Berthelsen and Marker, 1986; Melezhik and Sturt, 1994).

Pole Ak (Fig. 15) is a Svecofennian overprint in the 2.37 Ga (U-Pb, Zr; H.Huhma, pers. comm., 1987) Akanvaara gabbro intrusion, in the northern part of the Karelian Province (Mertanen and Pesonen, 1992, see Paper II for location). It corresponds to pole K obtained in the c. 2.45 Ga Koitelainen layered intrusion (Paper I). Both intrusions are located ca. 30-50 km south of the Lapland granulite belt. Sm-Nd dating of the metasedimentary gneisses of the Lapland granulite belt indicates that metamorphism took place in two stages, at ca. 1950 Ma and 1870 Ma (Daly and Bogdanova, 1991). These metamorphic events are probably responsible for the two overprint components of the Koitelainen intrusion, A' and A, which have totally replaced the primary remanence of this body. Thus, all the secondary "A" poles from Isokivennokka, Tsuomasvarri, Akanvaara and Koitelainen are located among the dated primary poles and record remagnetization at ca.1.90-1.87 Ga.

Secondary pole L (Fig. 15, Table 3) derives from the Koillismaa layered intrusions. In Paper I, an overall mean was calculated for the Svecofennian overprint direction (component A, Tables 1 and 2), including the Koillismaa and Koitelainen intrusions. To examine the direction and age of overprint A particularly in Koillismaa, pole L was recalculated by excluding the A direction of the Koitelainen intrusion. Pole Pj (Fig. 15) represents an overall mean of the Svecofennian partial overprint on different rock types east of Lake Pääjärvi (Krasnova and Gus'kova, 1991). The new poles of the gabbro noritic and diabase dykes of Suoperä (SP) and Lake Pääjärvi (XD) (Fig. 15, Table 3) are consistent with pole Pj, confirming the occurrence of coeval regional remagnetization in the Russian Karelian area. From comparison with the dated poles, it can be inferred that remagnetization in the Koillismaa and Russian Karelian areas took place at or later than 1840 Ma.

The younger remagnetization ages in Koillismaa and Russian Karelia may be related to the metamorphism in the Belomorian Belt. The Belomorian Belt lies only about 50 km from the Russian Karelian area, whereas the Svecofennian front is about 500 km away. Belomorian metamorphism is, however, not necessarily the only cause of remagnetization in the Koillismaa - Russian Karelia area; another may be hydrothermal fluids, which are related to collision and metamorphism at the Svecofennian orogen (see discussion on page 33).

5. PALAEOLATITUDES AND ROTATION OF THE FENNOSCANDIAN SHIELD

The latitudinal position and orientation of the Fennoscandian Shield during the Proterozoic (Fig. 16) was recently calculated at the Third Scandinavian Palaeomagnetic Workshop in Norway, using seven key palaeomagnetic poles (see Pesonen 1995). Each key pole represents the mean of primary poles from three separate studies. U-Pb ages are available for each key pole (Mertanen and Pesonen, 1995). Figure 16 has been supplemented with data obtained from this study (see Table 3).

Previously, the mean D palaeopole from 2.4 Ga old formations (Elming et al., 1993) was not considered sufficiently well-dated to be a key pole. However, the new preliminary data on the gabbronorite dykes obtained here are in agreement with the previous poles and lend support to the primary origin of the D remanence. Nevertheless, it should be stressed that the primary nature of the D remanence has still not been verified by baked contact tests. The present data show that at ca. 2.44 Ma, Fennoscandia was located at a latitude of about 30° (Table 3).

Whether the 2.44 Ga D palaeopole represents a north or a south pole is not known. The typical duration of polarity interval of the Earth's magnetic field, either normal (down pointing magnetization) or reversed (up pointing magnetization), is about 5 Ma and during the last 5 Ma the average time has been about 0.25 Ma (e.g. Butler, 1992). Therefore, within the limits of dating error, either polarity is equally possible.

In this paper and in Elming et al. (1993), the polarity assigned to pole D is different from that used in Paper I. The new polarity definition places the Fennoscandian Shield in a southern rather than a northern latitude and reorients the Fennoscandian Shield by 180°. This choice of polarity was made to minimize the APWP between adjacent poles and to make the polarity match that of Laurentia (see Buchan and Halls, 1990), derived from 2.4

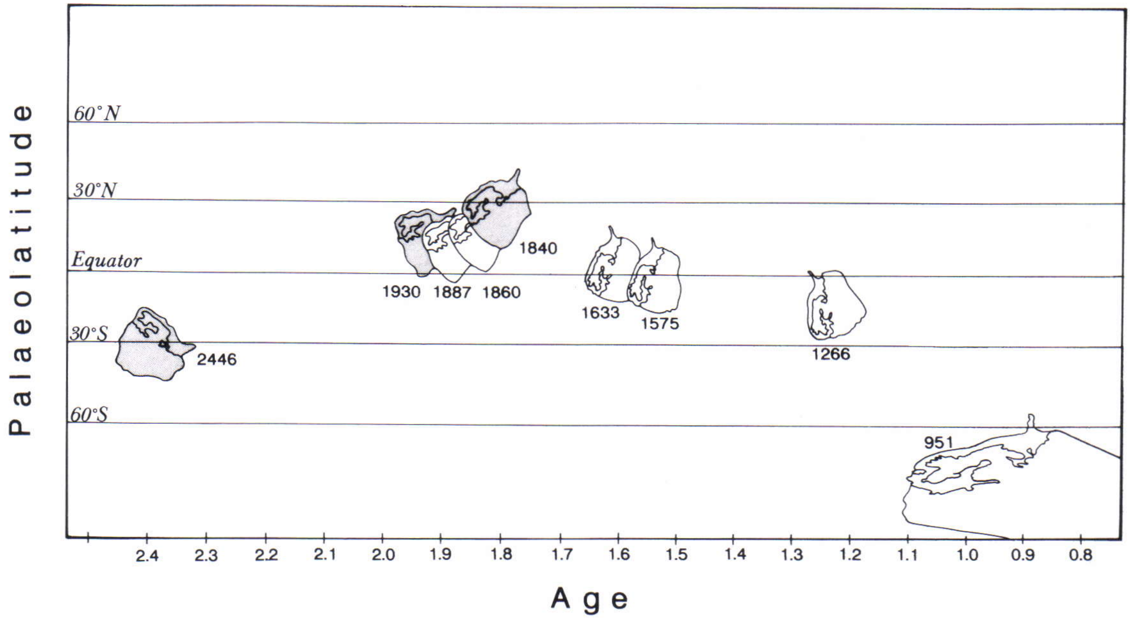

Fig. 16. Drift map of Fennoscandia based on U-Pb-dated palaeopoles. The shaded reconstructions are based on this study (2446 Ma and 1840 Ma) and on Paper II (1930 Ma). The map has been compiled by using GMAP (Torsvik et al., 1993) and MapDraw (compiled by M. Leino, GSF).

Ga old dual polarity Matachewan dykes (Bates and Halls,1990).

Palaeoclimatological evidence for the latitude of Fennoscandia at ca. 2.4 Ga is controversial. Sedimentological studies in Finland have confirmed a continental-wide glaciation at 2.5-2.3 Ma (Marmo and Ojakangas, 1984, Strand and Laajoki, 1993), indicative of high palaeolatitudes. However, recent studies by Sturt et al. (1994) indicate that 2.4 - 2.3 Ga regoliths are widespread, though sporadic, in both the Karelian and Kola Provinces in the eastern Fennoscandian Shield. On the basis of high carbonate content of the regoliths, Sturt et al. (1994) suggest that the regoliths were formed at low palaeolatitudes under arid/subarid palaeoenvironment and not in glacial environment at high palaeolatitudes.

As shown by the above studies, the Fennoscandian Shield experienced both glaciation at high latitudes and regolith formation at low latitudes between 2.5 and 2.3 Ga. Current palaeomagnetic data indicate that the Fennoscandian Shield drifted from high latitudes in the Late Archaean (Elming et al., 1993) to latitudes of 30° at ca. 2.4 Ga, a movement that is consistent with both sedimentological indications. At present we have no reliable evidence of the latitude of the Fennoscandian Shield at ca. 2.3 - 2.0 Ga. According to Kohonen and Marmo (1992) and Marmo (1993), the glaciation at 2.5-2.3 Ga was followed by a drastic change in climatic conditions at ca. 2.2. Ga, indicated by a palaeosol which was deposited under warm and humid conditions at low palaeolatitudes. The minimum age of the palaeosol is based on crosscutting mafic dykes dated at 2.2 Ga (Vuollo, 1991). In Canada, significant latitudinal movement during 2.3-2.0 Ga is evident from recent palaeomagnetic studies with precisely dated, well-defined palaeopoles at 2214, 2167 and 2114 Ma suggesting significant latitudinal drift and considerable rotation of the Canadian Shield

(Buchan et al., 1994). On the other hand, as noted by Marmo (1993) and Strand and Laajoki (1993), it is not easy to explain all climatological changes solely by latitudinal dependence. Referring to several successive glaciations in the Palaeoproterozoic Huronian Supergroup in North America, Strand and Laajoki (1993) conclude that continental-wide glaciation may also be related to climatic fluctuations and atmospheric changes. In order to correlate palaeomagnetic data accurately with climatological evidence we need precisely dated palaeopoles and precisely dated sedimentary sequences.

By the beginning of the Svecofennian orogeny, the Fennoscandian Shield had drifted to the Equator. The position of the Fennoscandian Shield at 1930 Ma has, however, been calculated from palaeopole A' obtained from the Tsuomasvarri gabbro-diorite intrusion in the Kola Peninsula (Paper II) and thus represents only a single area. It is therefore still uncertain whether the whole Fennoscandian Shield was located near the Equator at 1930 Ma. Magmatic activity was widespread, and has been precisely dated at that time in the Svecofennian Domain, but reliable palaeomagnetic data are lacking. At about 1890 Ma, Fennoscandia maintained its position near the Equator, but at 1860 Ma started to drift towards slightly higher latitudes. Latitudinal positions at 1887 and 1860 Ma (Fig. 16) have been calculated from mean palaeopoles obtained from mafic intrusions and dykes in the Raahe-Ladoga-Skellefte zone. The latitude and orientation of the Fennoscandian Shield during the late stages of the Svecofennian orogeny, calculated from the secondary palaeopoles of this study (Table 3), show further drift towards higher latitudes. As discussed earlier, the timing of the Svecofennian overprint is uncertain. An age of about 1840 Ma has been suggested only on the basis of the position of the overprint palaeopole relative to dated poles. $^{40}Ar/^{39}Ar$ age determinations on the Russian Karelian dykes are currently being made (by P. Layer) and may shed light on the age of this overprint.

The Svecofennian orogeny was followed by emplacement of rapakivi granites and associated diabase and quartz porphyry dykes in the central and southern Svecofennian Domain at about 1.67-1.54 Ga. At that time, the Fennoscandian Shield had once again drifted towards lower latitudes and was located at the equator at about 1.6 Ga (see also e.g. Pesonen et al., 1989, Elming et al., 1993). The position of the Fennoscandian Shield at 1632 and 1575 Ma has been calculated from the poles of Subjotnian dykes in southern Finland (see Paper III for discussion). Extensive dyke magmatism took place in southern Fennoscandia in Jotnian time at about 1270 Ma and marked the beginning of the Sveconorwegian orogeny. At 1270 Ma, the Fennoscandian Shield was at a palaeolatitude of about 20° as shown by mean palaeopoles obtained from Jotnian dykes in Finland.

The palaeopoles and position of the Fennoscandian Shield at 1.0-0.9 Ga are discussed in Paper IV. The position of Fennoscandia at 951 Ma (Fig. 16) is based on the mean of three poles ranging in age between ca. 980 and 945 Ma obtained from Sveconorwegian terranes in southern Norway. They all indicate a high southern palaeolatitude. The previously defined Sveconorwegian loop (Pesonen and Neuvonen, 1981; Pesonen et al., 1989; Elming et al., 1993) is not considered reliable because of the lack of U-Pb age determinations. However, the less reliable poles from the 1.0-0.85 Ga period show that a loop involving APW drift from low latitudes at ca. 1.0 Ga to high southern latitudes at ca. 0.95 Ga and back to low latitudes at ca. 0.85 Ga, may still exist (see Paper IV).

A discussion of the relative kinematic histories and reconstructions of the Fennoscandian Shield and the Laurentian Shield during the Proterozoic is beyond the scope of this study. It is necessary to note, however, that recent palaeomagnetic studies indicate that the

Shields probably moved together at 1.9 -1.8 Ga and at 1.27 Ga (Park, 1994; Buchan, 1995; Pesonen, 1995). This may have crucial implications for interpretations of the tectonothermal evolution of the Fennoscandian Shield.

6. CONCLUSIONS

The following conclusions are drawn from the present study:

1) Studies of multicomponent remanent magnetizations combined with petrophysical and petrographic investigations are a useful tool for interpreting secondary thermotectonic events in the Fennoscandian Shield. In particular, the thermal and thermochemical overprint associated with the Svecofennian orogeny is recorded in the remanence directions of pre-Svecofennian rocks.

2) Due to the Svecofennian orogeny at ca. 1.9-1.87 Ga, the whole Karelian Province and the parts of the Kola Province for which palaeomagnetic data are available are partially or completely remagnetized.

3) The overprint represents a partial thermochemical magnetization acquired during cooling and uplift following the Svecofennian orogeny. It is suggested that hydrothermal fluids played an essential role in remagnetization of the Archaean terrane. The Palaeoproterozoic dykes and intrusions cutting the Archaean basement also produced channels for hydrothermal fluids and thus were vulnerable to partial remagnetization.

4) Preservation of the primary remanence is not restricted to petrographically fresh rocks, because heavily altered rocks can also be carriers of a component of the primary remanence. Conversely, samples that show only minor alteration (e.g. uralitization along margins of pyroxenes) may be completely overprinted. More rock magnetic and petrographic studies on Fe-Ti-oxide minerals are still needed to resolve the origin and mechanism of the Svecofennian remagnetization.

5) The maximum age of Svecofennian remagnetization in the Koillismaa - Lake Pääjärvi region has been dated at ca. 1840 Ma on the basis of the pole positions relative to the isotopically dated primary poles. The Svecofennian overprints in the north of the Karelian Province, close to the Lapland Granulite belt, and in the Pechenga region are older, ca. 1.9-1.87 Ma, and are related to the tectonic events of these areas.

6) New preliminary palaeomagnetic results from gabbronorite dykes (2446+5/-4 Ma, U-Pb age on baddeleyite) in Russian Karelia indicate that the dykes acquired their primary remanent magnetization (component D) at the same time as the Koillismaa layered intrusions acquired theirs (2436±5 Ma, U-Pb age on zircon). Baked contact tests have not been carried out on the gabbronorite dykes, but the remanence is inferred to be primary from the absence of component D in the younger cross-cutting dykes.

7) Palaeomagnetic results from diabase dykes genetically related to the 2365±90 Ma old (Sm-Nd mineral age) Burakovskaya layered intrusion in the southeast of the Karelian Province (Krasnova and Gos'kova, 1990) are consistent with results from the north of the province, suggesting that no major tectonic movements have taken place between the regions since the emplacement of the formations. More reliable palaeomagnetic data integrated with isotopic studies are still needed to confirm this assumption.

8) Palaeomagnetic results in the Fennoscandian Shield in the Jatulian time (2.2-2.0 Ga) are still controversial and require more detailed studies on precisely dated formations.

9) According to this study, the Fennoscandian Shield was located at an intermediate southern latitude (ca. 30°S) at 2.44 Ga. Dur-

ing the early stages of the Svecofennian orogeny, the Fennoscandian Shield was close to the palaeoequator and shifted to slightly higher northern latitudes in the late stages of the orogeny, at ca. 1.84 Ga. In Subjotnian time, ca. 1.6 Ga ago, the Fennoscandian Shield was again located at the equator. During Jotnian time (ca. 1.27 Ga), the Fennoscandian Shield started to drift towards higher palaeolatitudes. At about 950 Ma, the Fennoscandian Shield was located at high southern latitudes.

ACKNOWLEDGEMENTS

First, I would like to thank the head of the palaeomagnetic laboratory of the Geological Survey of Finland, Lauri Pesonen, for giving me the opportunity to do research in palaeomagnetism and for providing unfailing support during the many stages of the work. I also thank Matti Leino for all his efforts in the palaeomagnetic laboratory, especially for making essential computer programs more accessible to users. I am further indebted to Hilkka Kumpunen and the summer assistants who made most of the palaeomagnetic measurements.

I am grateful to the whole field group in Russia for our successful cooperation, in particular to Jouni Vuollo who saw to all the arragements, Mauri Terho who was responsible for the hard drilling, and Henry Halls and Vesa Nykänen who helped to collect the hand samples. My three field trips in Russia were an unforgettable experience thanks to the congenial atmosphere created by our Russian colleagues and assistants: Vladimir Stepanov and Alexandr Slabunov, Ada, Lena and Alexandra Stepanov, and Tarmo Kulmala.

Discussions with Pentti Hölttä, Lauri Pesonen and Jouni Vuollo, who read the first version of the manuscript, are highly appreciated. The manuscript was officially reviewed by Ken Buchan from the Geological Survey of Canada and Raimo Lahtinen from the Geological Survey of Finland. Their suggestions and comments, which greatly helped to improve the manuscript, are acknowledged with gratitude. Special thanks go to Tapani Rämö for reading and correcting the final version of the manuscript.

Salme Nässling is thanked for the patience and care with which she drew the figures, Sini Autio for editing the manuscript and Gillian Häkli for revising the language.

Finally, my warmest thanks are addressed to my mother who has always encouraged me and provided irreplaceable help by looking after my son during the many evenings and weekends, that I have been working.

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