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A GEOLOGICAL INTERPRETATION OF THE UPPER CRUST ALONG FIRE 1

By
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The 500 km long FIRE 1 deep seismic reflection profile crosses over a major Palaeoproterozoic plate boundary in the Fennoscandian Shield – the Svecofennian – Karelian suture zone, where Archaean Karelian craton and its cover have been juxtaposed with the Palaeoproterozoic Svecofennian island arc rocks during the Svecofennian Orogeny, c. 1.9 Ga ago. The upper crustal reflection properties along FIRE 1 are described and the correlation with surface geology is discussed. In addition, a geological interpretation or a vertical lithological map of the uppermost 8 km of the crust is presented.

The upper crustal reflectors correlate well with surface geology in the scale of hundred meters. Major lithological units have different seismic patterns that change across the boundaries. The most prominent reflections seem to originate from lithological contrast associated with mafic intrusions, dykes or volcanic units. Shear zones are seen as white bands disrupting the seismic patterns.

The upper crust along FIRE 1 is composed of ten units belonging to three tectonic domains: Archaean nucleus, A-P boundary zone and Proterozoic Central Finland. Of these, the upper crustal structure of the Archaean Nucleus is dominated by Archaean structures, the A-P boundary zone by both Palaeoproterozoic extensional structures related to rifting and collisional structures related to Svecofennian orogeny and the Proterozoic Central Finland domain is characterized by both collisional and gravitational collapse structures of the Svecofennian orogen.

During the continental arc/continent collision the Western Kianta block, Kainuu Belt, Kajaani, Rautavaara, Iisalmi Complexes as well as Savo Belt belonging to the Archaean Proterozoic boundary zone were all thrust sequentially on the Eastern Kianta block belonging to the Archaean Nucleus. The scale of thrusting varies from less than 1 km to 10–20 km thick stacks where the deformation is concentrated mainly to the block boundaries. The Central Finland Granitoid Complex, which is further divided into Pihtipudas and Keuruu blocks, is a shallow (3–8 km) upper crustal unit, whose lower surface is a detachment zone. The detachment surface is associated with upper crustal graben and horst -structures.

Key words (GeoRef Thesaurus, AGI): crust, upper crust, bedrock, deep seismic sounding, reflection methods, FIRE, deep-seated structures, Proterozoic, Archean, Finland

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INTRODUCTION

The FIRE 1 deep seismic reflection profile crosses over a major Palaeoproterozoic plate boundary in the Fennoscandian Shield, the Svecofennian – Karelian suture zone (Bowes and Gaál 1981, Koistinen 1981), where Archaean Karelian craton and its cover have been juxtaposed with Palaeoproterozoic Svecofennian island arc rocks during the Svecofennian Orogeny, c. 1.9 Ga ago. Later more data from the Fennoscandian Shield such as the first well-documented Palaeoproterozoic ophiolite, Jormua Ophiolite (Kontinen 1987, Peltonen and Kontinen 2004), and a paleosubduction zone imaged by deep seismic reflection profiles (BABEL Working Group 1990) have favored the Palaeoproterozoic plate tectonics.

On the Karelian craton side, the Archean-Proterozoic boundary zone is approximately 150 km wide, NW-SE striking zone where the Archaean and its cover have been overprinted by 1.9–1.8 Ga metamorphism and deformation (Koistinen 1981, Huhma 1986, Kontinen et al. 1992, Pajunen and Poutiainen 1999, Peltonen & Mänttäri 2001). On the Proterozoic side, the Archaean-Proterozoic isotopic boundary is sharp and lies within 20 km of the exposed suture (Huhma 1986, Lahtinen & Huhma 1997). The Svecofennian and Archean-Proterozoic boundary zone are characterized by thick crust (55–65 km) with thick high velocity lower crust (v_p >7 km/s) at crustal depths greater than 35 km (Korja et al. 1993) as well as thick, high velocity lithosphere (> 200 km) (Sandoval et al. 2004). The boundary zone is characterized by Bouguer anomaly maxima (Ruotoistenmäki et al. 2001), SW dipping crustal conductors and abrupt change in crustal thickness from 58 km to 42 km within 20 km in distance (Korja et al. 1993). It also hosts a set of SE-NW striking magnetic and Bouguer anomaly lineaments interpreted as strike slip zones (Koistinen & Saltykova 1999).

As an explanation for the formation of the Archaean-Proterozoic boundary zone several plate tectonic scenarios have been proposed. The models can be classified into three categories: continental arc/continent collision zone (e.g. Gaál 1990, Lahtinen 1994), back-arc/retro-arc basin related to NE-directed subduction occurring further SW (e.g. Hietanen 1975, Gáal 1986) and a strike-slip model (e.g. Park 1985), where all Palaeoproterozoic parts are considered exotic terranes. Lahtinen et al. (2005) suggested that the Svecofennian Orogen have formed in four partly overlapping orogenies and they favour the continent-continent model for the Archean-Proterozoic boundary zone. The zone was thickened due to the E-W collision of a microcontinent (Keitele) and an attached island arc (Savo Belt) on the Archaean Karelian craton margin during the Lapland-Savo orogeny. Further thickening in the boundary zone occurred during N-S compression caused by the collision of an island arc and Bergslagen microcontinent on the southern margin of the Keitele microcontinent during the Fennian orogeny. Mafic underplating (Korja 1995, Korsman et al. 1999) have been suggested as the balancing process allowing for the preservation of the thick crust.

In order to understand how the currently exposed units of the crust have been formed it is pertinent to know their depth extent as well as the geometry of the structures associated with their emplacement. In 2001, a consortium consisting of the Geological Survey of Finland, and Universities of Helsinki and Oulu, with Russian Spetsgeofizika S.G.E. as a contractor, shot a deep seismic reflection survey called FIRE 1 (the Finnish Reflection Experiment 1) across the Archean-Proterozoic boundary zone to unravel the architecture and evolution of this major Palaeoproterozoic collisional structure (Kukkonen et al. 2006, *this volume*). Kukkonen et al. (2006, *this volume*) describe the acquisition and processing of the FIRE 1 data and outline the preliminary crustal features.

In this paper, we describe and discuss the correlation of the surface geology with the upper crustal reflection properties along FIRE 1. The upper crustal structures are better described with migrated sections where dip move out correction (DMO) has been applied. For this paper purpose DMO-sections to the depth 8 km have been calculated. These are used to identify those geological structures and formations that are exposed at surface and have a reflection response.

GEOLOGICAL AND GEOPHYSICAL OUTLINES

In the following, the geological and geophysical bedrock units have been described from the profile point of view. The profile has been divided into 10 main units by using both lithological and aeromagnetic maps. Established unit boundaries and unit nomenclature (Nironen et al. 2002) have been favored when possible. Some new boundaries have been suggested and a few new units have been differentiated.

The FIRE 1 line is a crooked line with a general trend that is perpendicular to the Archaean-Proterozoic (A-P) boundary. Along the FIRE 1 profile, the area east of the A-P boundary can further be divided into seven geological units: *Eastern Kianta block, Kuhmo Greenstone Belt, Western Kianta block, Kainuu Belt, Kajaani Complex, Rautavaara Complex and Iisalmi Complex*. The Eastern and Western Kianta blocks, and the Kuhmo Greenstone Belt are late Archaean (3.0–2.7 Ga) in age (Sorjonen-Ward & Luukkonen 2005). Palaeoproterozoic cover is only found on the western margin but Palaeoproterozoic mafic dykes (2.4–2.0 Ga) are abundant in these units. In this study, the Kainuu Belt comprises only of Palaeoproterozoic allochthonous (2.0–1.9 Ga) cover sequences, including turbidites and ophiolites. The occurrence of an Archaean sliver and associated autochthonous cover further divides the Kainuu Belt into eastern and western segments where ophiolites are found only from the latter (Kontinen & Meriläinen 1993). The heterogenous ***Kajaani Complex*** is composed of Archaean rocks with remnants of Palaeoproterozoic cover units (2.4–2.0 Ga), mafic dykes (2.4–2.0 Ga), alkalic gneisses (1.96 Ga) and granites (c. 1.8 Ga) (Havola 1997). A narrow sliver of the Rautavaara Complex (Paavola 2003) composed of Archaean rocks and Palaeoproterozoic cover sequences, is juxtaposed between Kajaani and Iisalmi Complexes. The Iisalmi Complex (c. 3.2 Ga) is older than the c. 2.8–2.7 Ga Rautavaara Complex (Mänttäri & Hölttä 2002). Together with the Pudasjärvi Granulite Belt (3.5–2.8 Ga, Mutanen & Huhma 2003) it may form an older Archaean terrane along the western edge of the Karelian craton. Palaeoproterozoic cover is found on the western margin and Palaeoproterozoic mafic dikes (2.2–2.0 Ga) as well as a bimodal association of granites and gabbros (1.88–1.86 Ga) as intruding the complex (Paavola 2001, Ruotoistenmäki et al. 2001).

The Palaeoproterozoic part, west of the A-P boundary, is part of the Svecofennian Domain and it can be further divided into three units: *Savo Belt* and the *Pihtipudas and Keuruu blocks* of the Central Finland Granitoid Complex (Fig 1.). The Savo Belt comprises 1.92 Ga volcanic and sedimentary rocks and gneissic tonalites, and it has been interpreted to represent rocks from a primitive island arc (Lahtinen 1994, Kousa et al. 1994). The Central Finland Granitoid Complex consists of granitoids and associated gabbros with some subvolcanic to volcanic rocks and few remnants of supracrustal sequences. The plutonic rocks can be divided into 1.89–1.88 Ga and 1.88–1.87 Ga syn- and post-kinematic intrusions (Nironen et al. 2000). The Central Finland Granitoid Complex also delineates approximately the projected boundaries for the hidden 2.1–2.0 Ga Keitele microcontinent (Lahtinen et al. 2005).

The bedrock in the Eastern Kianta block (EK) is composed of migmatic banded tonalites and trondhjemites, and more intrusive-type tonalites, granodiorites and granites, which locally show cutting relationships with the migmatitic granitoids. The remnants of Archaean supracrustal sequences, mainly banded amphibolites, are E-W directed and moderately dipping (30°–70°) to the north. There are also cataclastic shear zones and diabase dykes parallel to the banded amphibolites. The E-W trending structures are cross-cut by NW-SE directed shear zones, diabase dykes and some granitoids (Luukkonen 1993, 2001). There are also NE-SW trending fracture zones and diabase dikes but these are less abundant than the NW-SE directed ones. All the diabases are Palaeoproterozoic in age but the E-W directed ones are usually older and wider than the others (Luukkonen 2001). Just north of FIRE1 a Palaeoproterozoic rapakivi-type granite intrusion (2.4 Ga) has been found (Luukkonen 1993, Korsman et al. 1997, Rämö and Luukkonen 2001).

The Kuhmo Greenstone Belt (KGB) bisects the Kianta Terrane into western and eastern halves. The belt consists of ultramafic to felsic metavolcanic rocks as well as chemical and clastic metasedimentary rocks. The volcanic rocks started to erupt at 2.79 Ga. About 50 Ma later took place widespread granodioritic to tonalitic magmatism and deformation affecting both the migmatitic blocks and greenstone belt. During this tectonic event the contacts between the greenstone belts and the older migmatitic granitoids (TTG-series) were tectonically modified. At present, the Kuhmo Greenstone Belt is a rather narrow (less than 10 km in width and only 3 km at FIRE1 transect) synclinorial structure (Sorjonen-Ward and Luukkonen 2005 and references therein).

The Western Kianta block (WK) is characterized by migmatic tonalite-trondhjemites, and dominant intrusive-type tonalites, granodiorites and granites. Along FIRE line 1, E-W directed Archaean amphibolites are absent, which seems to be characteristic to

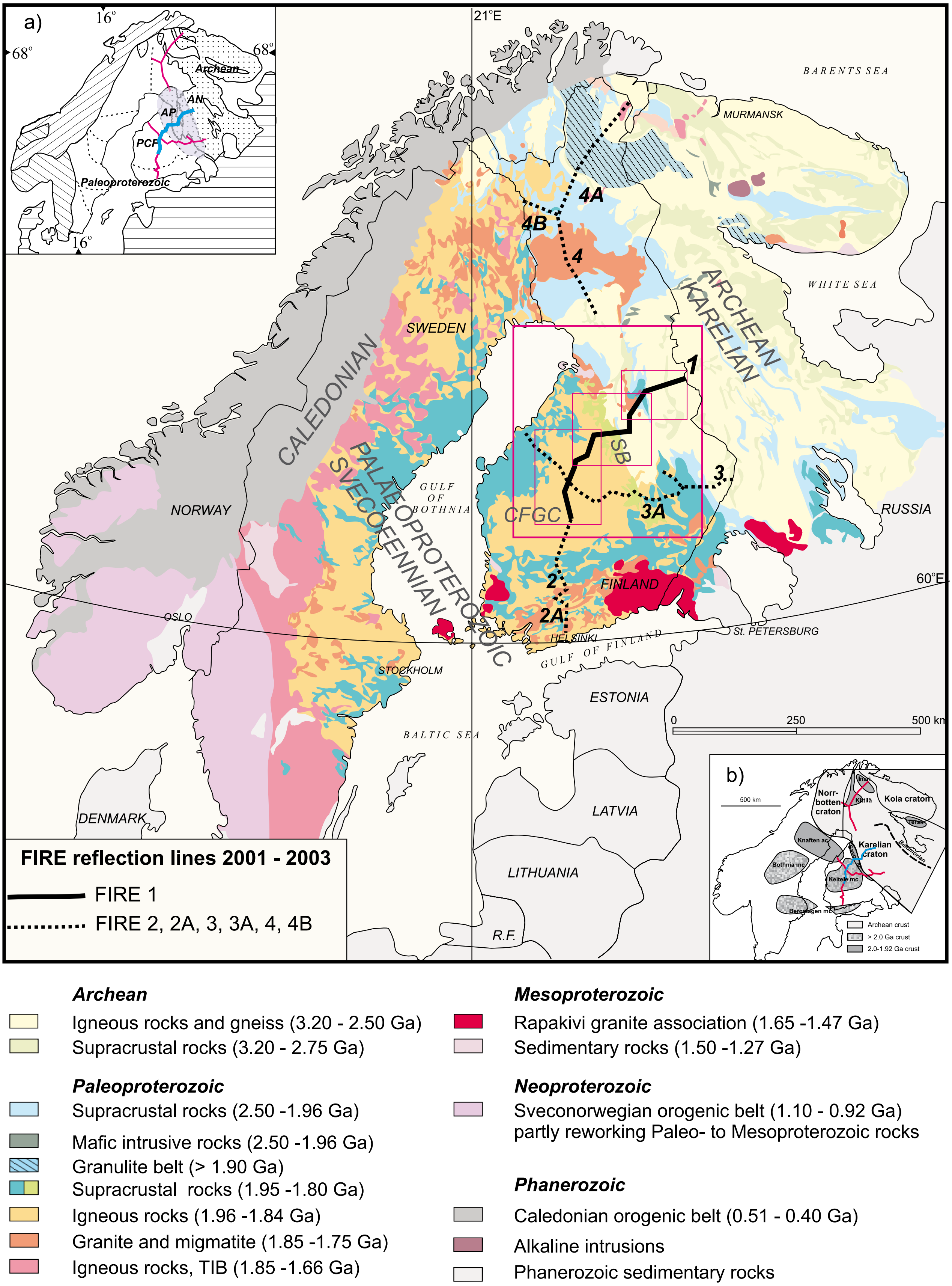


Figure 1. Geological map of the Fennoscandian Shield modified after Koistinen et al. (2001). FIRE 1 marked as a thick line. Figures 2, 5, 8 and 12 are outlined. a) Major tectonic domains of the Fennoscandian Shield. AN – Archaean Nucleus, A-P – Arhaean-Proterozoic boundary zone, PCF – Proterozoic Central Finland. b) Distribution of cratons, microcontinental nuclei, island arcs in the Fennoscandian Shield after Lahtinen et al. (2005).

the Western Kianta (Hyppönen 1973). The Western Kianta block is also transected by a wealth of Palaeoproterozoic diabases. They are steep to shallow dipping and striking dominantly in NW-SE to NWW-SEE direction (Hyppönen 1983) and only rarely in NE-SW direction. The scarce E-W structural trends, shear zones and diabases are mainly found close to KGB (Hyppönen 1973). Closer to the Kainuu Belt, remnants of pelitic paragneisses are more abundant (Kontinen & Meriläinen 1993). The autochthonous cover rocks at western margin include quartzite, dolomite, black schist, metadolerite and volcanic rocks. Magnetotelluric data indicates that the graphitic schists of the autochthonous sequence continue westwards under the Kajaani and Iisalmi Complexes (Korja & Koivukoski 1994, Korsman et al., 1999).

The Svecofennian orogeny caused a structural and thermal overprint on the Archaean terranes. The thermal overprint is well-recorded by the diabase dykes in the Kianta blocks. The diabases in the Western Kianta often show pervasive schistosity, differing from the relatively unstrained diabases of the Eastern Kianta (Kontinen et al. 1992, Kontinen 2002). The Svecofennian thermal peak is dated by U-Pb method on a xenotime to c. 1.85 Ga (Pajunen & Poutiainen 1999).

The Kainuu Belt (KB), a Palaeoproterozoic inlier between Archaean Kianta block and Kajaani and Iisalmi Complexes, is composed of allochthonous pelitic and turbiditic sequences (Kontinen & Meriläinen 1993) and has been further divided into eastern and western segments. The absence of volcanic or tuffitic intercalations is a characteristic feature of the allochthonous sedimentary sequences. Only turbidites are found in the eastern segment but ophiolites occur as tectonic slivers within allochthonous cover in the western segment. The 1.95 Ga old Jormua ophiolite (J; Fig. 2) is a fragment of Red Sea-type crust comprising of metabasalts derived from E-MORB to OIB-like sources (Kontinen 1987, Peltonen and Kontinen 2004).

The boundary between the Kainuu Belt and the Kajaani Complex (KC) is defined here as the thrust zone between the allochthonous units of the Kainuu Belt and the para-autochthonous units of the Kajaani Complex (Korsman et al. 1997). The heterogenous Kajaani Complex includes Archaean migmatitic granitoids with few Archaean amphibolite and mica gneiss units and fragments. Palaeoproterozoic para-autochthonous cover sequence (quartzite, dolomite, metadolerite, volcanic rocks) in the Kajaani Complex occurs as a continuous unit in the north, east and south, and as fragments and remnants in the northwest and west. Palaeoproterozoic mafic dikes and sills intrude the autochthonous cover and the well-preserved Archaean granitoids in the southern part. Heterogenous alkalic gneisses (1.96 Ga) are found in the near vicinity of the southern branch of the cover sequence. Younger granites (c. 1.8 Ga Kajaani granite; Havola 1997) are widespread, often pegmatitic and they migmatize all the other rocks. The northern part of the Kajaani Complex is characterized by N-S and NW-SE directed fault and fracture zones. Because they displace thrust zones and they lack associated diabases they are interpreted to be relatively young in age (Havola 1981, 1997).

The proposed boundary between the Kajaani Complex and the Rautavaara Complex (RC) is overprinted by the 1.8 Ga granites, and is inferred from the magnetic data. On FIRE 1, only a narrow sliver of the Rautavaara Complex is encountered and it comprises Archaean migmatites, a thin sheet of Palaeoproterozoic cover in the south and cross-cutting granite (1.8 Ga) (Paavola 2001). The southern boundary zone towards the Iisalmi Complex coincides with a swarm of strongly deformed diabases in the Hatulanmäki area (Paavola, 2003). The Rautavaara Complex, in general, is characterized by Archaean migmatites with micagneiss melanosome, and the occurrence of granitic dykes and related migmatization. Other characteristic features are the locally intensive Svecofennian deformation, seen as gentle lineation dipping towards SW or S, and the rareness of wide and continuous diabase dykes (Paavola 2001, 2003).

The Iisalmi Complex (IC) has paleosome ages of 3.2–3.1 Ga and it is older than the c. 2.8–2.7 Ga Rautavaara Complex (Mänttäri & Hölttä 2002). The 3.2 Ga ages are also older than the 3.0–2.7 Ga ages obtained from the Kianta block (cf. Sorjonen-Ward & Luukkonen 2005). As the metamorphic grade increases inwards within the Iisalmi Complex, it is further divided into Kukkopuro, Saavanmäki, Kulvenmäki (P. Hölttä 2005 pers. com.) and Vieremä blocks. Amphibolite banded tonalitic-trondhjemitic migmatite is the dominant Archaean lithology in the northern part of the Iisalmi Complex (Kukkopuro block), where amphibolites occur also as larger lenses. Granulite facies mineral parageneses are often found in areas south of Sukeva in the Saavanmäki granulite block and Kulvenmäki high pressure eclogitic granulite block (P. Hölttä 2005 pers. com.). Granulite-facies domains host also more homogenous tonalites and 2.7 Ga quartz diorites (enderbites). Leucodiorites of 2.7 Ga age intrude the migmatites. Along FIRE1 line the Archaean migmatites, amphibolites and granulites are shallow to moderately dipping (Paavola 1991, 2001, 2003).

The Palaeoproterozoic diabases are very common in the Iisalmi Complex. Their width varies from few meters to 150 m and they are up to 3–4 km in length.

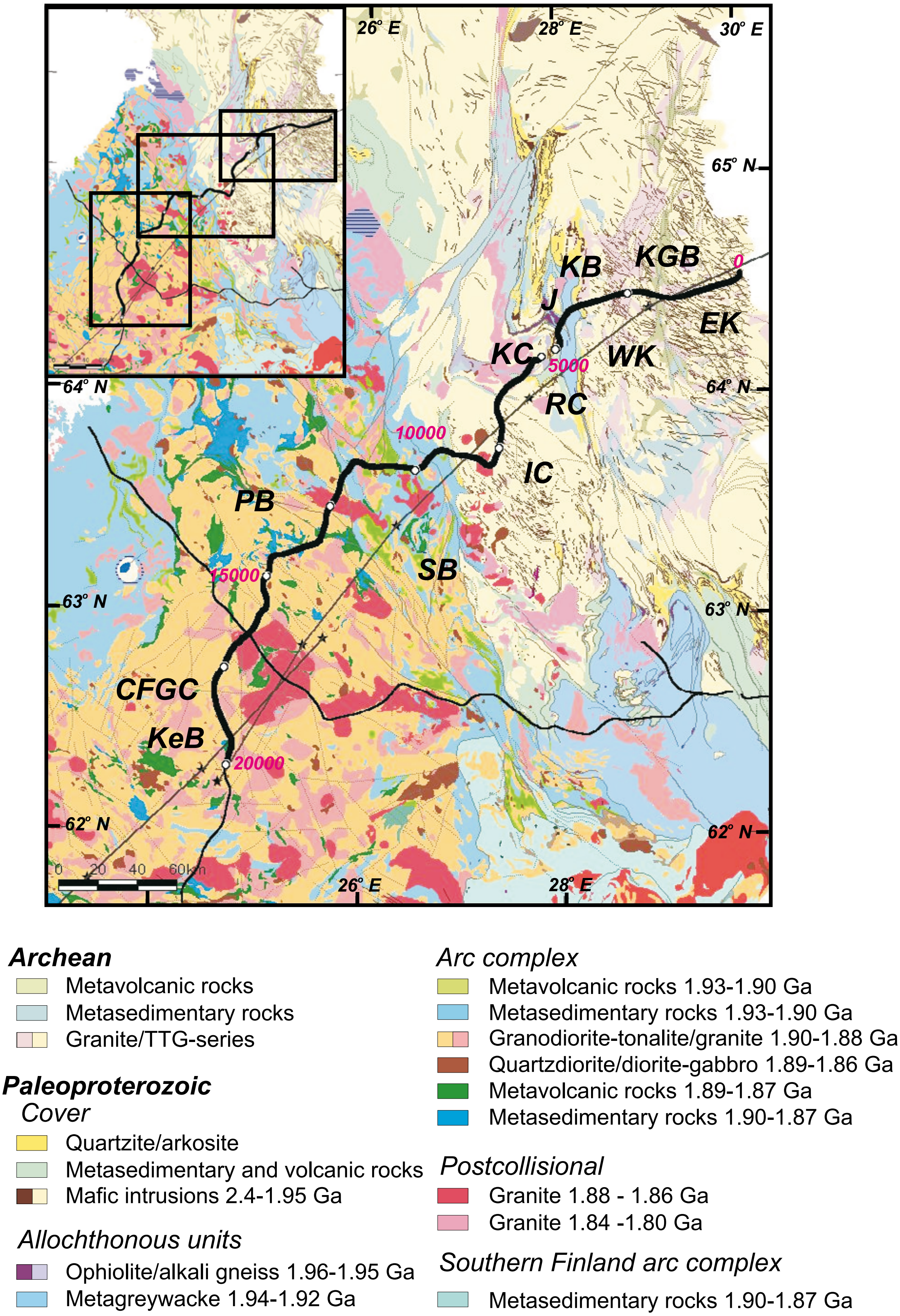


Figure 2. Geological map of the study area modified after Koistinen et al. (2001). FIRE 1 marked as a thick line. Shotpoints of deep seismic refraction lines SVEKA81 & 91 (Grad & Luosto 1987, Luosto et al. 1994) are shown with stars. Figures 5, 8 and 12 are outlined in the insert Figure 2a. EK – Eastern Kianta Block; KGB – Kuhmo Greenstone Belt; WK – Western Kianta Block; KB – Kainuu Belt; KC –Kainuu Complex; RC – Rautavaara Complex; IC – Iisalmi Complex; SB – Savo Belt; PB – Pihtipudas Block; KeB – Keuruu Block; CFGC – Central Finland Granitoid Complex.

The diabases are normally steeply dipping and they run mainly in the NW-SE direction, which is also the most common direction for faults and fracture zones. The diabases are variably affected by the Svecofennian deformation and metamorphism. While diabases in granulite domains have often at least partly preserved their primary mineralogy, diabases found in shear zones are fully recrystallized and strongly lineated (Paavola 1999, 2001, 2003).

Close to the suture, the Archaean-Proterozoic boundary zone (A-P) is characterized by the intrusion of a younger, 1.86–1.85 Ga old, bimodal suite of granites, diorites, hybrid gabbro-granite stocks, and related dyke rocks (Ruotoistenmäki et al. 2001, Paavola 2001, 2003). FIRE1 crosses one of these hybrid intrusions (Kauppilanmäki) (Paavola, 1990). Although the most common direction for faults and fracture zones is the NW-SE direction also N-S and NE-SW Proterozoic shearing and fracturing is observed between Sukeva and Kauppilanmäki. Further to the west, concentric shearing and lithological layering is found around the 1.86 Ga old Kaarakkala intrusion. Fracture zones are locally associated with sheared and mylonitic tectonometasomatic quartz-epidote rocks (Paavola 2001, 2003).

At the western edge of the Iisalmi Complex, highly strained mylonites and thrust zones separate the Vieremä block from the granulitic blocks. The Vieremä block consists of autochthonous/para-autochthonous Salahmi schist belt (Korkiakoski & Laajoki 1988), composed of mica gneisses and quartzites, and Archaean units. A characteristic structural feature in the whole border zone is a post-1.89 Ga lineation, which plunges gently to moderately SW and overprints axial plane foliation. Postcollisional 1.82 Ga reactivation has been suggested to have caused faulting, thermal resetting and intrusion of granite dykes in the Salahmi area (Pietikäinen ja Vaasjoki 1999). The Archaean units, Salahmi cover sequences and the Svecofennian units to the west have all been thrust NE on the western part of the Iisalmi Complex and at present they form an imbrication structure (Tuisku & Laajoki 1990).

The suture between Svecofennian Domain and Karelian Craton lies within the contact between the Salahmi schists and the schists of the Savo Belt (SB). The Savo Belt is few tens of kilometres wide, SE-NW trending belt consisting of metavolcanic and metaturbiditic rocks of the Pyhäsalmi formation that are interlayered with gneissic tonalites (1.92 Ga; Kousa et al. 1994, Lahtinen 1994). The volcano-sedimentary complex was later intruded by the “synkinematic” granodiorite, tonalite, quartz-diorite plutons, associated volcanic rocks and mafic dykes (1.89 – 1.88 Ga) as well as by post-kinematic pyroxene bearing granitoids starting at 1.885 Ga (e.g. Marttila 1981, Korsman et. al. 1999).

In the western parts of the Savo Belt, locally well-preserved, bimodal volcanics belonging to the Pyhäsalmi formation are found (Kousa et al., 1994). The type areas Pyhäsalmi and Mullikkoräme, hosting massive sulphide ores, occur as isolated blocks surrounded by younger plutonic rocks. The Pyhäsalmi area is transected by abundant mafic and intermediate dykes that are weakly foliated and younger (1.89–1.87 Ga) than the surrounding supracrustal rocks (Marttila 1993). The porphyry dykes are thin but the diabase dykes are up to 10 m wide. The dykes west of Pyhäsalmi are oriented in E-W direction and the dykes east of Pyhäsalmi in NW-SE direction. The mica gneisses west of Pyhäsalmi formation belong to the Savo Belt whereas the adjacent younger metavolcanics belong to the Central Finland Granitoid Complex (Kousa & Lundqvist, 2000). The Savo Belt does not have any observable suture against the Central Finland Granitoid Complex (CFGC).

The Savo Belt and the Archaean-Proterozoic boundary as whole developed to a major strike-slip shear zone (Koistinen & Saltykova 1999), seen in the abundant occurrence of ductile to brittle shear zones and faults in the Savo Belt along the FIRE1 line. These structural trends are displayed as negative lineaments transposing older structural patterns on magnetic and Bouguer anomaly maps (Ruotoistenmäki et al. 1997a,b). Apart from the prominent NW-SE directed structural trend, a NE-SW-trending Oulujärvi shear zone is transecting the Pyhäsalmi area (Kärki et al. 1993). One characteristic feature of the Savo Belt is the occurrence of granulite facies rocks and associated pyroxene-granitoids that show contact aureoles. The age of peak metamorphism is c. 1.885 Ga (e.g. Korsman et al. 1999) followed by cooling to about 550˚C at 1.85 Ga and 300˚C at 1.78 Ga (Haudenschild 1995). One of the granulite domains is exposed just south of the FIRE line 1 suggesting that similar domains might also be found in the deeper parts of FIRE 1.

The Central Finland Granitoid Complex (CGFC) is characterised by synkinematic granitoids with small remnants of supracrustal units of volcanic and sedimentary rocks. The Central Finland Granitoid Complex is further dived into *Pihtipudas and Keuruu blocks* based on their reflective properties and differences in lithologies. *The Keuruu block* is dominated by synkinematic granodiorites (Nironen 2003) intruded by quartz monzonite, porphyritic granodiorite, and small gabbro intrusions. But the postkinematic granitoid intrusions and younger volcanic rocks typical of the Pihtipudas block have not been identified in the Keuruu block.

The 1.89–1.87 Ga volcanic rocks to the west of the western mica gneisses in the Pyhäsalmi area are included in the the *Pihtipudas block*. These younger volcanic rocks show mature arc affinity (Kousa et al. 1994). Further SW, along FIRE1 line, a granodiorite that is locally strongly gneissic, is found. The granodiorite comprises mafic enclaves and 1–20 m diabase dykes pointing to bimodal magmatism and magma mingling. The granodiorite is cut by a WNW-ESE trending 7–8 km wide porphyritic Pyytsalo granite (1.87 Ga, Kousa et al. 1994), which is locally exceptionally rich in K-feldspar, contains abundant epidote locally, and has syenite in fault and shear zones (Marttila 1993). The supracrustal rocks in Pihtipudas have been correlated with the Pyhäsalmi younger volcanics and the 1.89–1.87 Ga Ylivieska supracrustal rocks (Kousa & Lundqvist 2000). Both granodiorites and Pyytsalo granite cuts the Pihtipudas supracrustal rocks. Within Pyytsalo area, the fault structures start to change from N-S and NW-SE trending to WNW-ESE trending and the Pihtipudas volcanic belt is crosscut by a strong shear zone in WNW-ESE direction.

The area SW from Pihtipudas is dominated with plutonic rocks varying from gabbro to granite. The most common rock type along FIRE1 line is a granodiorite with small remnants of supracrustal units of volcanic and sedimentary rocks. According to Nykänen (1963) some gabbro-diorites are co-magmatic with the mafic volcanic rocks, which are correlated with the Pihtipudas volcanic rocks (Kousa & Lundqvist 2000).

The characteristic feature of the western part of the Pihtipudas block is the abundant occurrence of 1.89–1.87 Ga postkinematic plutonic and subvolcanic rocks intruding the synkinematic granitoids. Nironen (2003) has divided the lineaments in the CFGC into three groups: 1) 20–40°, 2) 120–135° and 3) 0°. Group 2 faults deform Group 1 faults and cros-scut also the postkinematic rocks in the Pihtipudas block. The magnetic anomaly map (Fig. 12b) suggests that the group 3 faults have an old component, which is cross-cut by group 1 and 2 faults. A group 3 fault (17250) has a dextral horizontal component and it marks a zone separating the Pihtipudas and Keuruu blocks.

DATA

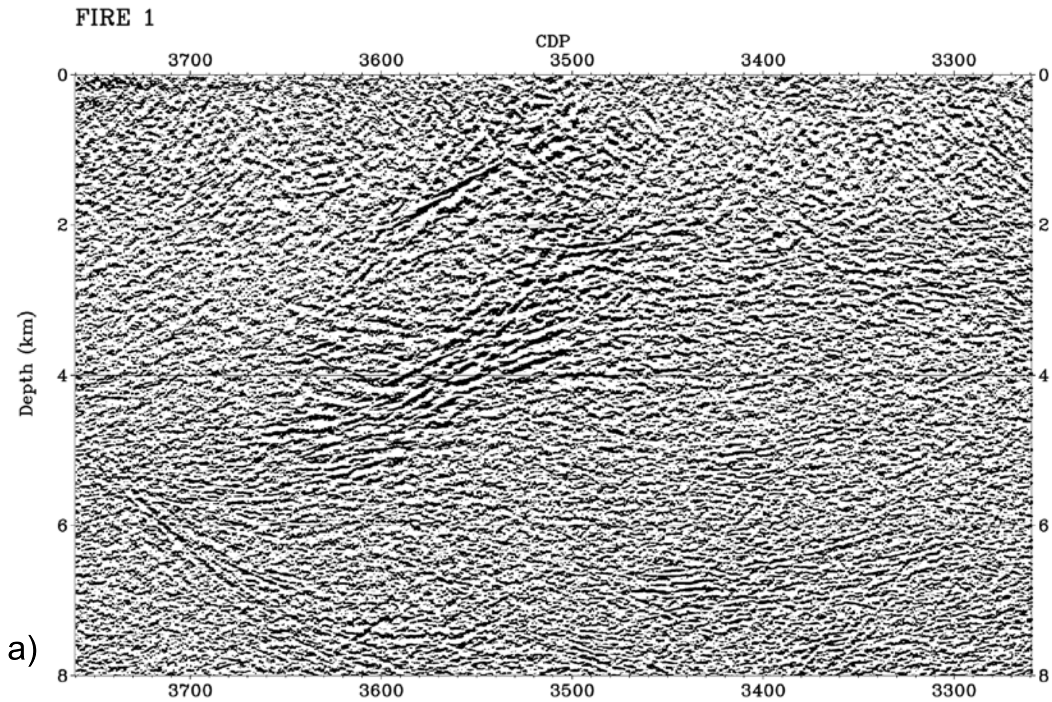
In this paper, the near vertical reflection data are the migrated DMO stacked sections from FIRE 1. Acquisition parameters and full processing sequence have been described by Kukkonen et al. (2006, *this volume*). In addition, DMO-stacked sections have been calculated to image better the steeply dipping reflections. The 1D-velocity function used in migration and depth conversion is the average function from the wide-angle SVEKA81&91 profiles (Fig.1, Grad & Luosto 1987, Luosto et al. 1994).

Both variable area wiggle-trace sections (Fig. 3a) and combination sections (Fig. 3b), where colour-coded smoothed envelope -section is overlain by biased variable area plot, are used as basis for the detailed interpretations. In the smoothed envelope sections, the data are smoothed over a 0.25 km x 0.06 km window in lateral and vertical directions, respectively (for details see Korja & Heikkinen 1995). Major structures have been cross-checked against the crustal-scale automatic line-drawings and grey-scale envelope sections described in Kukkonen et al. (2006, *this volume*). For printing purposes the data are presented as grey-scale variable intensity sections (Figures 6–7, 9–11, 13–15). All the sections are plotted without normalisation i.e. the amplitudes of the different areas in each section are comparable.

For a good reflection, a large impedance contrast (e.g. large density and velocity differences at contacts) is needed. For example, a strong reflection may arise from a contact between mafic rocks embedded in granite/quartzite environment. The amplitudes of the reflections coming from layered structures are further increased by constructive interference, if internal layer spacing and layer thickness is approximately one quarter of the dominant wavelength (λ = 100 m), in this case 25 m. However, a detectable reflection signal can originate from a structure that has a minimum thickness in the order of one eight of a wave length. For example a detectable reflection maybe acquired from either a thick layer (>10 m) of mafic volcanic rocks/dyke embedded in granite/quartzite environment. Transparent zones on the other hand, could develop from a large shear one or a set of closely spaced fracture zones.

Although seismic reflection method studies the material properties mainly in vertical dimensions the structures have continuity also in lateral dimensions. With reflection method two reflection points are observed as two separate objects when their distance is greater than Fresnel radius $r = (z\,\lambda/2)^{1/2}$. Because migration enhances lateral resolution, the calculated distances are always maxima estimates. For example, in FIRE 1 data, a set of reflections is observed as a continuous reflection surface when the points are less than 300 m apart at the depth of 1 km with increasing the distance to 900 m at the depth of 8 km.

The FIRE 1 profile is 500 km long and consists of 20000 CMP points. The data is shown in a sequence of



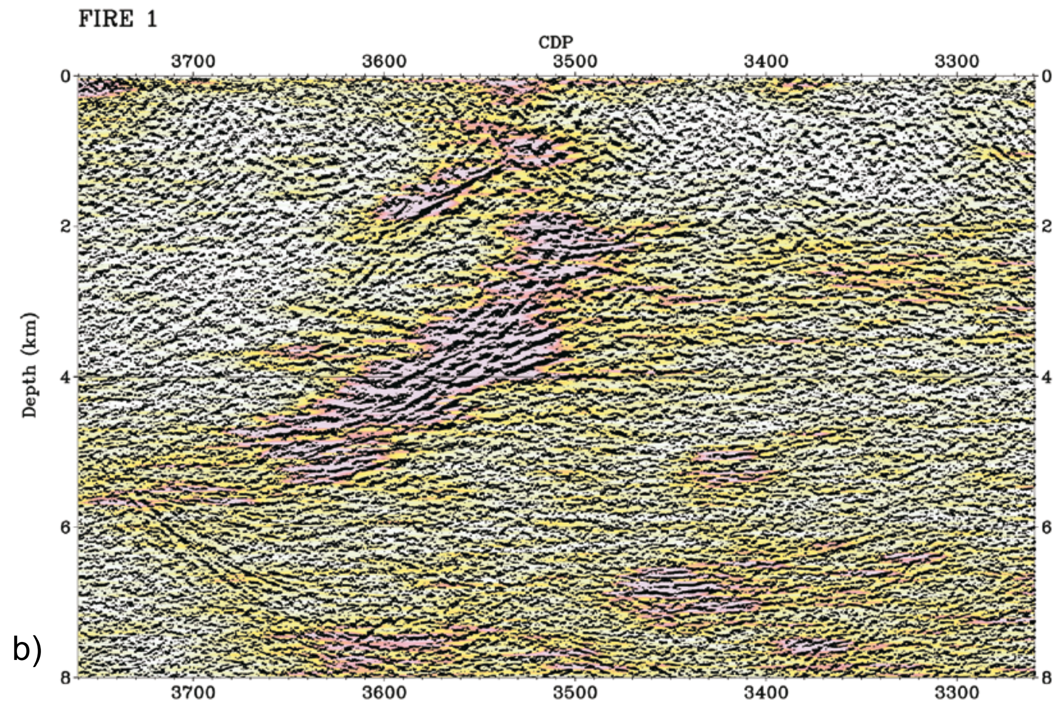
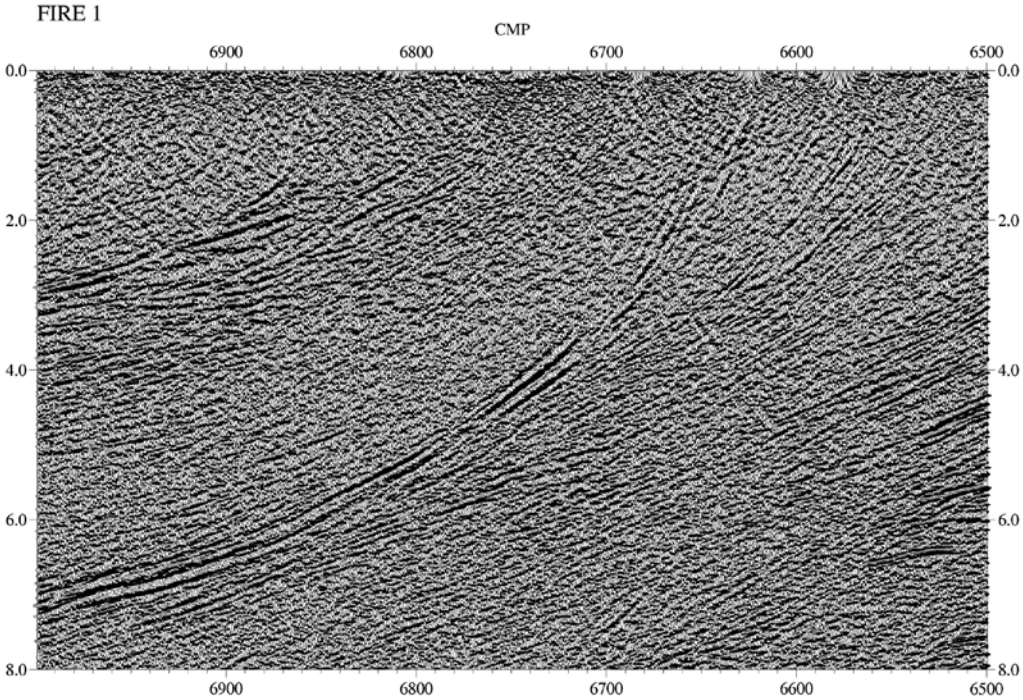


Figure 3 a and b. Examples of the quality of the migrated DMO stacked sections from FIRE 1. a) A variable area wiggle-trace section between CMP points 3250–3750. b) A combination section, where colour-coded smoothed envelope –section is overlain by biased variable area plot, between CMP points 3250–3750.

8 sections (middle panel in Figures 6–7, 9–11, 13–15) each displaying 2500 CMP points (62.5 km) and a depth interval between 0 and 8 km. In the sections, 100 CMP points correspond to 2.5 km in distance or 400 CMP points correspond to 10 km. The most prominent features of the reflection sections are shown in the line drawings below (lower panels in Figures 6–7, 9–11, 13–15). Above the sections correlative lithological maps (upper panels in Figures 6–7, 9–11, 13–15) along the section are shown.

In the following, the most prominent features and differences in reflectivity patterns are described by using grey-scale variable intensity sections (Figs. 6–7, 9–11, 13–15). These are followed intimately by correlation of the reflective properties with surface geology and geophysics. The interpretations are complemented



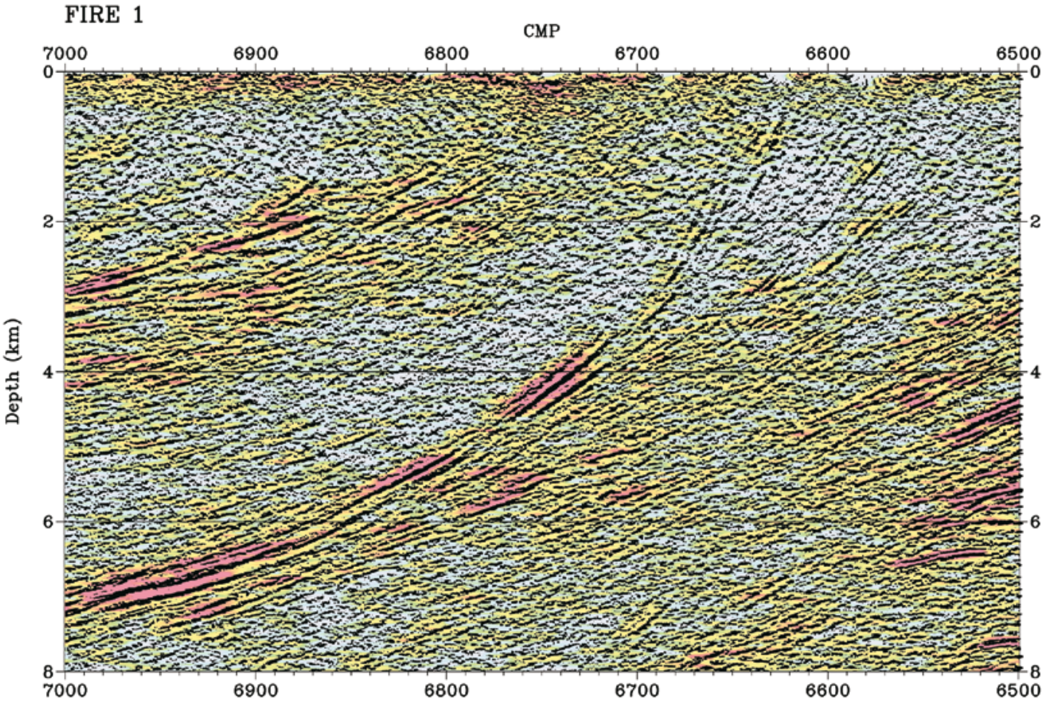


Figure 3 c and d. c) A variable area wiggle-trace section between CMP points 6500–7000.
d) A combination section, where colour-coded smoothed envelope –section is overlain by biased variable area plot, between CMP points 6500–7000.

by a geological interpretation in Appendix 1. For the geological interpretation, the line drawings have been collected together and are displayed one continuous profile, where the differences in the reflectivity patterns are emphasised by the different colours associated with lithological units. The profile is displayed as a straight section although it is a crocked line with several major turning points. The turning points are marked with red lines in the line drawings and the geological interpretation. The cross-point of FIRE profile 1 and 3 is marked with a blue line.

True strike direction can be determined whereas dip angle is always dependent on the interception angle (a) between the geological structure and the reflection line sin (Apparent dip, β) = sin(True Dip, α) * sin (a). The dip-directions of the reflections have been determined from the turning points or crossing lines at the junction between profiles 1 and 3. If the interception angle is smaller than 90° then the apparent dip of the reflection is less than the true angle. Because the dip angle is almost always apparent we have abandoned the term from the following descriptions if the interception angel is between 70° and 90° degrees. If it is less than 70° degrees the term apparent is used and both the apparent dip as well as the interception angle is given. For those reflections not coming to the surface the measured dip is a priory apparent. A certain ambiguity always retains in the interpretations where two structures are crossing the line at the same locality. The problem is visualized in Figure 4 with an example where a reflection line is passing three planes with variable dips and interception angles.

THE REFLECTIVE UNITS AND THEIR CORRELATION TO THE SURFACE

In the following the upper crustal units are described and correlated with geological and geophysical structures at the surface. Because FIRE 1 hosts large regional differences in both intensity and style of reflectivity it can be divided into 10 seismically different segments in the upper crust with some overlap in vertical dimension. The crustal segments are (Fig. 2): Eastern Kianta block (0 – 1900), Kuhmo Greenstone Belt (1900 – 2000), Western Kianta block (2000 – 3750), Kainuu Belt (3750 – 4500), Kajaani Complex (4500 – 6200), Rautavaara Complex (6200 – 6650), Iisalmi Complex (6650–9390), Savo Belt (9390 – 12100), Pihtipudas block (12100 – 17400) and Keuruu block (17400 – 20000).

The uppermost crust of the Archaean Karelian Domain east of the Kainuu Belt (0–3750; Fig. 5) can seismically be divided into three units: Eastern Kianta block, Kuhmo Greenstone Belt and Western Kianta. In the Eastern Kianta block the apparent reflections dip dominantly to the E-NE, in the Western Kianta the reflections dip to the W- SW. The Kuhmo Greenstone Belt is shallow, poorly reflective block between the eastern and Western Kianta blocks in the uppermost crust.

The **Eastern Kianta block** (0–1900; Fig. 6) has relatively high reflectivity and the prominent reflections have a gentle dip (15–35°) towards E-NE or are flat-lying. The dipping reflections tend to flatten out at depth (Example 1; CMP1500–1250). Near the surface, more steeply dipping reflections (60–70°) are spinning-off from the flat to gently dipping surfaces.

In the Eastern Kianta block, the FIRE1 line is running roughly in E-W direction, which is parallel to the general trend of the greenstone remnants running E-W directed and dipping moderately to steeply (30–70°) to the north. There are also shear zones and diabase dyke swarms running in this direction. The E-W trending structures, semi-parallel to the profile, are likely to give rise to the flat to shallow dipping reflections.

At surface the E-W trending greenstone remnants are cross-cut by E-W and NW-SE directed shear zones, diabase dykes, and aligned ellipsoidal granitoids. As these structures are at 45–90° to the reflection line they can give rise to steep to moderately dipping reflections. At surface, the steep reflections are associated with NW-SE-running mafic diabase dikes (CMP 900, 1020, 1300,) and shear zones (CMP 740) or their combination (CMP 430, 1400). The granitoid intrusions have weak reflectivity (CMP 1840–1420) but their contacts are sometimes associated with shear zones and/or diabase dyke swarms. Between CMP

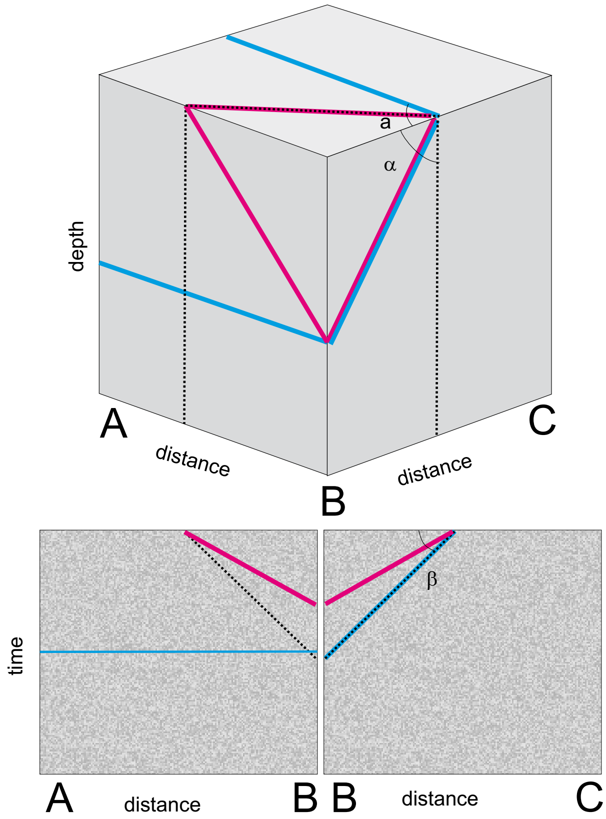


Figure 4. A schematic drawing on the effect of dip angle and angle of interception on crossing seismic sections AB and BC. The block diagram displays three planes (stippled, blue and red) with a common cross-point. The seismic sections AB and BC display their respective seismic images. Note that change in either dip-angle (α) or angle of interception (a) changes the apparent dip (β) of the reflection. Correlation with the surface structures is necessary for optimal interpretation of the cause of the reflections.

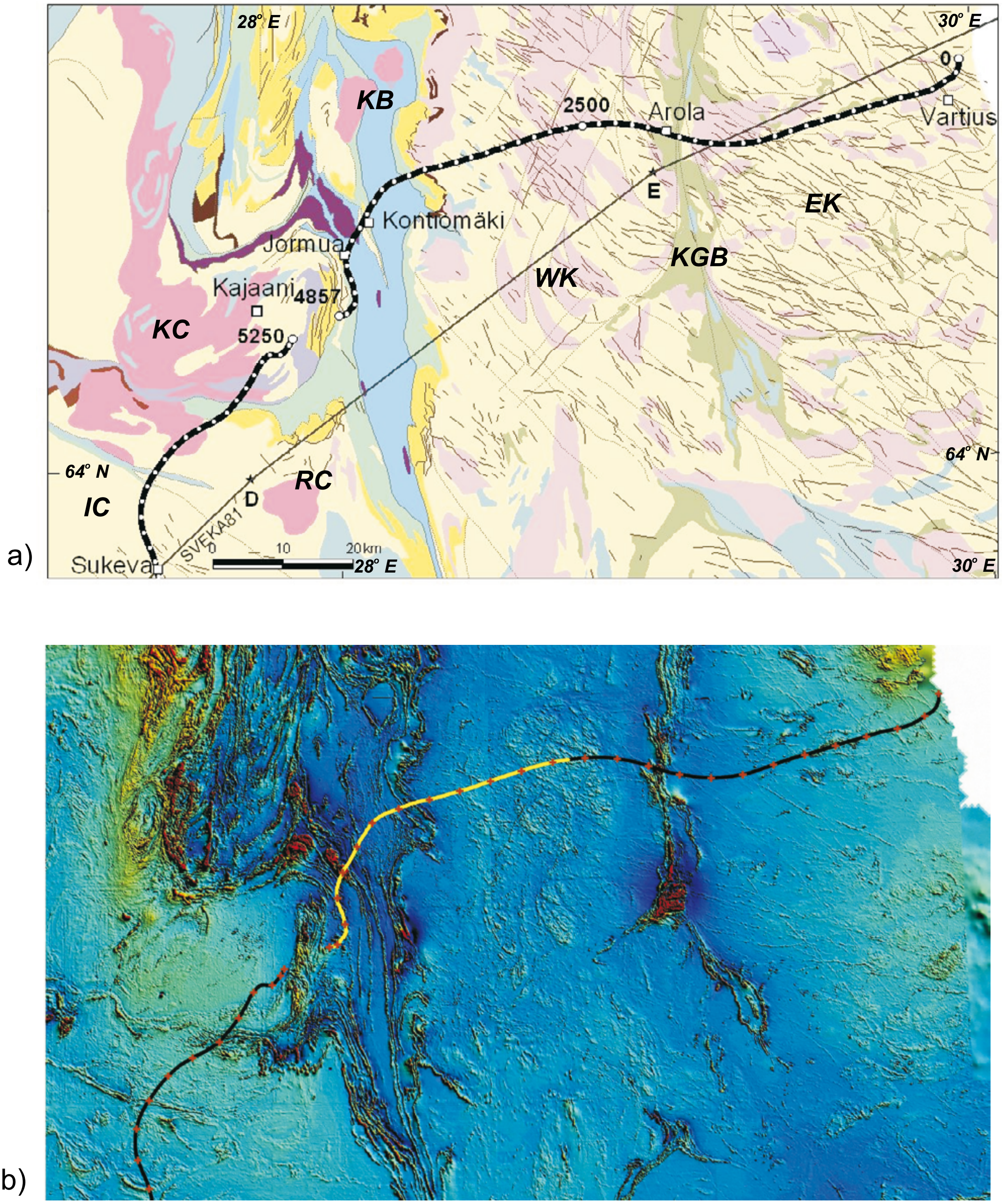


Figure 5. a) Lithological map of Eastern Finland between Vartius – Sukeva areas modified after Koistinen et al. (2001). FIRE1 marked as a thick line. EK – Eastern Kianta Block; KGB – Kuhmo Greenstone Belt; WK – Western Kianta Block; KB – Kainuu Belt; KC – Kainuu Complex; RC – Rautavaara Complex; IC – Iisalmi Complex;
b) A total intensity aeromagnetic anomaly map of Eastern Finland between Vartius – Sukeva areas (source GTK). Horizontal gradients are emphasized by vertical illumination of total intensity.

1750–1830, reflections apparently dipping 30° west are found. These are correlated with NW-SE diabases that are crossing the line at 30° angle. The dip is thus only apparent and the dykes are near vertical.

The Kuhmo Greenstone Belt (CMP 1900–2000) is a weakly reflective block with steep (60°) west-dipping contacts. The bottom of the belt is at 4 km depth where flat-lying reflections are found. The weak reflectivity is interpreted to image the small density-velocity contrasts among various mafic and ultramafic units

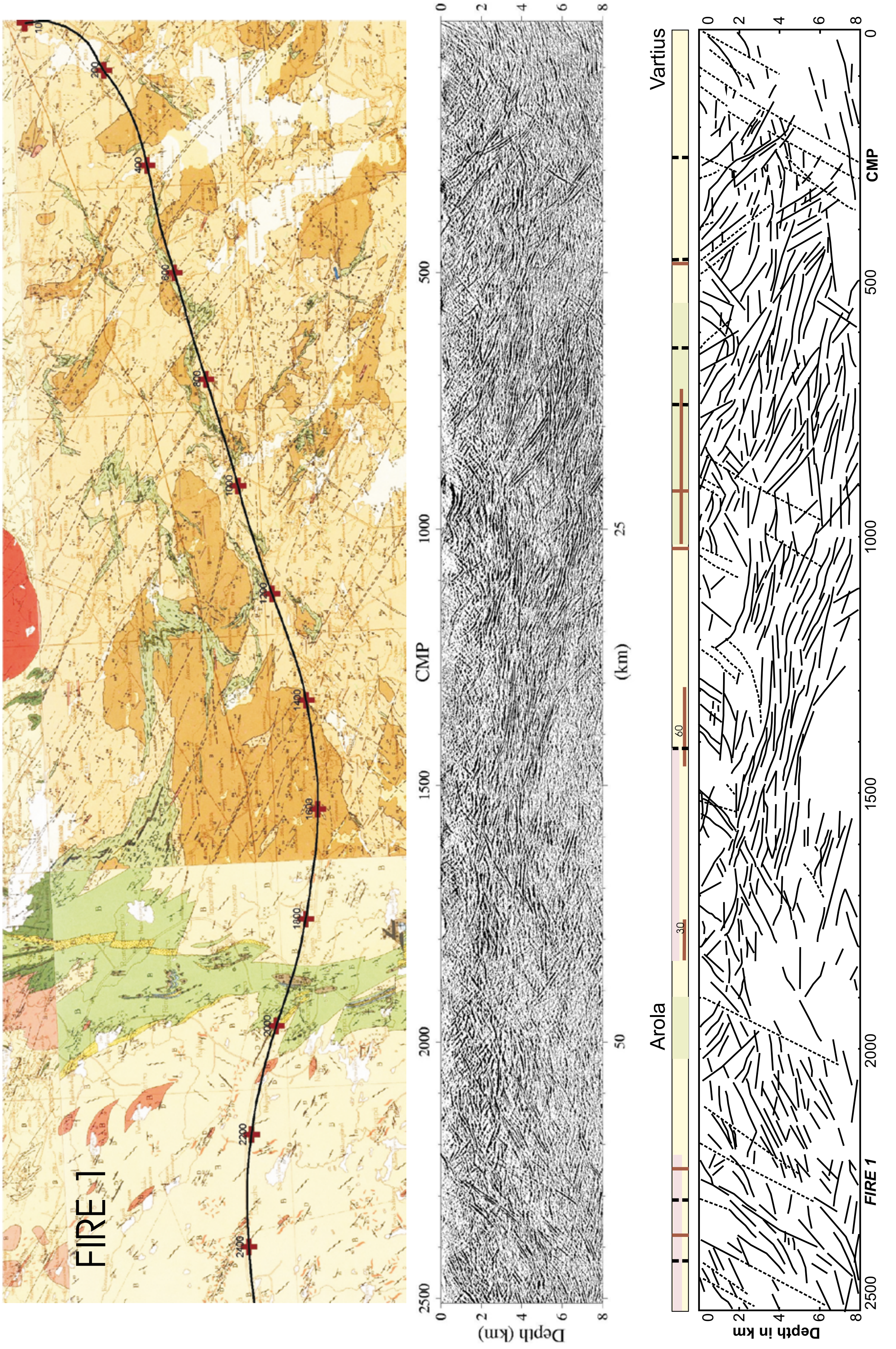


Figure 6. Uppermost 8 km of crust along FIRE 1 between CMP points 1–2500. Upper panel: Lithological map (Hyppönen 1973, Luukkonen 1993), middle panel; grey-scale variable intensity DMO section, lower panel; the lithology at surface and a line drawing. Major turning points are marked with a red line on the lower panel.

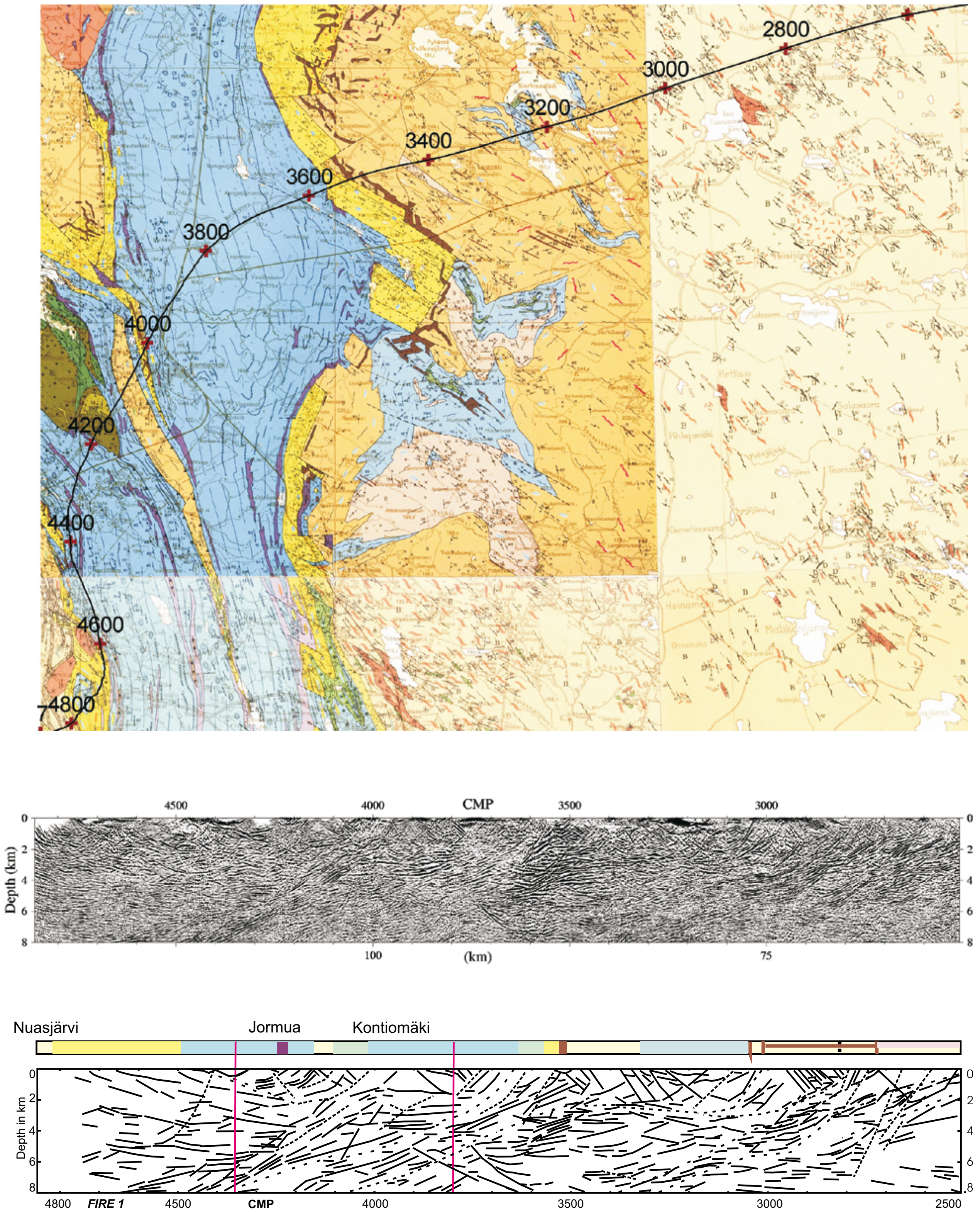


Figure 7. Uppermost 8 km of crust along FIRE 1 between CMP points 2500–5000. Upper panel: Lithological map (Hyppönen 1973, Havola 1981, Kontinen & Meriläinen 1993), middle panel; grey-scale variable intensity DMO section, lower panel; the lithology at surface and a line drawing. Major turning points are marked with a red line on the lower panel.

of the belt. The contacts are at surface sheared and tectonized (Hyppönen 1983).

The **Western Kianta block** (2000–3650; Figs. 6 and 7) is generally less-well reflective than the Eastern Kianta block. The more bright reflections form W-dipping zones where the dips vary between 45° and 20°. Some individual W-dipping reflections have steeper dips 60–70°. Between the westward dipping reflective zones there are apparently low to gentle (10–30°) eastward dipping individual reflections and zones between depths 3 km and 8 km. Close to the surface (0–2 km), more steeply (60–70°) eastward dipping reflections spin off from or displace the low angle reflections. The western edge of the Western Kianta block (Figure 7) is characterized by a prominent cluster of reflections that are apparently gently (35°) west-dipping to flat-lying. Where these reflections reach the surface (CMP 3500-3550), narrow layered intrusions (diabase sills) are found.

Both moderate to steep reflections correlate well with diabases (CMP 2230, 2350, 3030, 3030, 2700-3000) and shear zones (CMP 2290, 2400, 2700) at surface. The reflections associated with shear zones seem to penetrate deeper and displace reflections. The flat to shallow eastward dipping reflections at depth probably image Archaean structures, amphibolite remnants, deeper in the crust.

A set of steep shear zones at CMP 2170-2290 mark a major block boundary within the Western Kianta. The block to the west is associated with higher magnetic value levels and the occurrence of dominant granodiorites and tonalities. E-W diabase dykes or E-W Archaean structural trends are rare within this block.

The upper most crust in the west (0-4 km in depth, CMP 3150-3500) is characterized by low angle (15°), E-dipping layered reflectivity. The layered reflectivity is cross-cut by a swarm of steeply (60°) east-dipping reflections spinning off from the lowermost layers at the depths between 2 km and 3 km. At surface, the shallowly dipping layered structure correlates with Archaean migmatitic paragneisses, whereas the cross-cutting reflections correlate with Palaeoproterozoic diabases.

The Archaean-Proterozoic boundary zone (3650–12100; Figs. 5 and 8) is composed of both Archaean and Proterozoic units at surface. Kainuu Belt (3650 – 4500), Kajaani Complex (4500 – 6200), Rautavaara Complex (6200 – 6650), Iisalmi Complex (6650–9390) and Savo Belt (9390 – 12100) have been distinguished based on reflectivity. There are major turns of the profile at CMP 3770, 4350, 6620, 8200, 9500 that should be taken into account when interpreting the changes in dip directions along the profile.

The Kainuu Belt (3650–4500; Fig. 7) is characterized by 30–45° degrees westerly dipping bands of reflections, which are cut by steeper (< 60°) west-dipping poorly reflective zones outcropping between shotpoints CMP 3940–4020, 4100–4150.

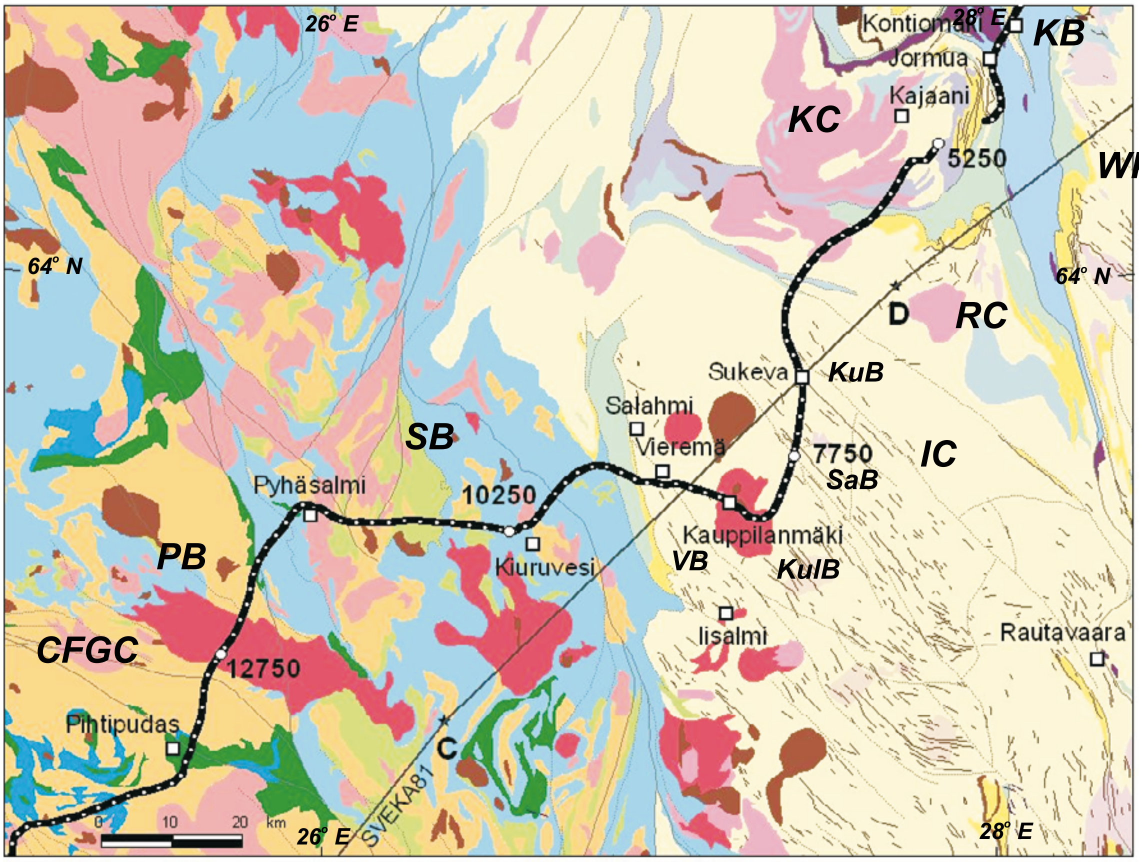
The gently dipping reflections correlate with the “allochtonous” graywacke-pelite association and steeper, poorly reflective structures correlate with shear zones and with an Archaean thrust sliver and associated autochthonous rocks at CMP 3940–4150. Another parallel set of bright reflections (30° and becoming steeper upwards) is found between CMPs 4300 and 4150, where they correlate with the Jormua ophiolite association (CMP 4150–4200).

At surface the **Kajaani Complex** (4500–6200) starts at CMP 4500 although at depth it continues northwards below the Kainuu Belt. In the eastern part is characterized by low to flat lying reflections cut by 20° degrees north dipping reflections (CMP 4160–4430, 4600–4760). At surface, the northerly dipping reflections correlate with western margin of the autochthonous units and other tectonic contacts found within the basement-quartzite-diabase association of the Kajaani Complex, where the schistosity is dominantly steep. After the southward turn of the profile at CMP 4350, the schistosity is semi-parallel to the profile causing the apparent shallow dip of the reflectors.

South of Kajaani (5300–6200; Figs. 8 and 9) the profile shows flat to shallow SW-dipping (0–35°, listric) layered reflectivity. Between CMPs 5300–5900 the layered reflectivity at surface correlates with the flat-lying Palaeoproterozoic alkali-gneisses (1.96 Ga) and younger granites (1.8 Ga) intruding the Kajaani Complex. At CMP 5900 a dipping reflection outcrops at surface where the northern contact of western finger of the Kainuu Belt is crossed. The dipping reflector flattens out when the line direction becomes parallel to the strike of the schist belt at CMP 5970. Underneath the continuous sets of reflections there are clusters of bright reflections with both E-NE and S-SW dips. At the shallowest these reflection clusters are at the depth of 2 km at the central part of the Kajaani Complex. It is interpreted that the clusters are related to the diabase-quartzite association found above at the surface.

The Rautavaara Complex (6200–6650) is characterized by two continuous sets of bright reflections dipping 35–40° SW. Between CMPs 6520 and 6650 bright reflections that at surface are near-vertical (80–85°) turn shallower and continue as a band of 30° degrees south dipping reflections at the depth (Fig. 3 c, d; Fig. 9). Where the listric reflections come to the surface (CMP 6600) they coincide with a sheared supracrustal unit and associated thrust surface (Havola, 1997; Paavola, 2001) marking the tectonic contact zone between the Rautavaara and Iisalmi Complexes. Two of the bright listric reflections coincide with vertical dykes at surface (CMP 6620, 6650).

The change of line direction from SW-NE to more S-N directed at 6620 has only a minimum effect on the reflection directions indicating that the true dip direction is towards the SW. Flat-lying to shallowly east-dipping reflectivity between CMPs 6000 and 6600 is associated with S-type granite intruding both Kajaani and Rautavaara Complexes.



A)

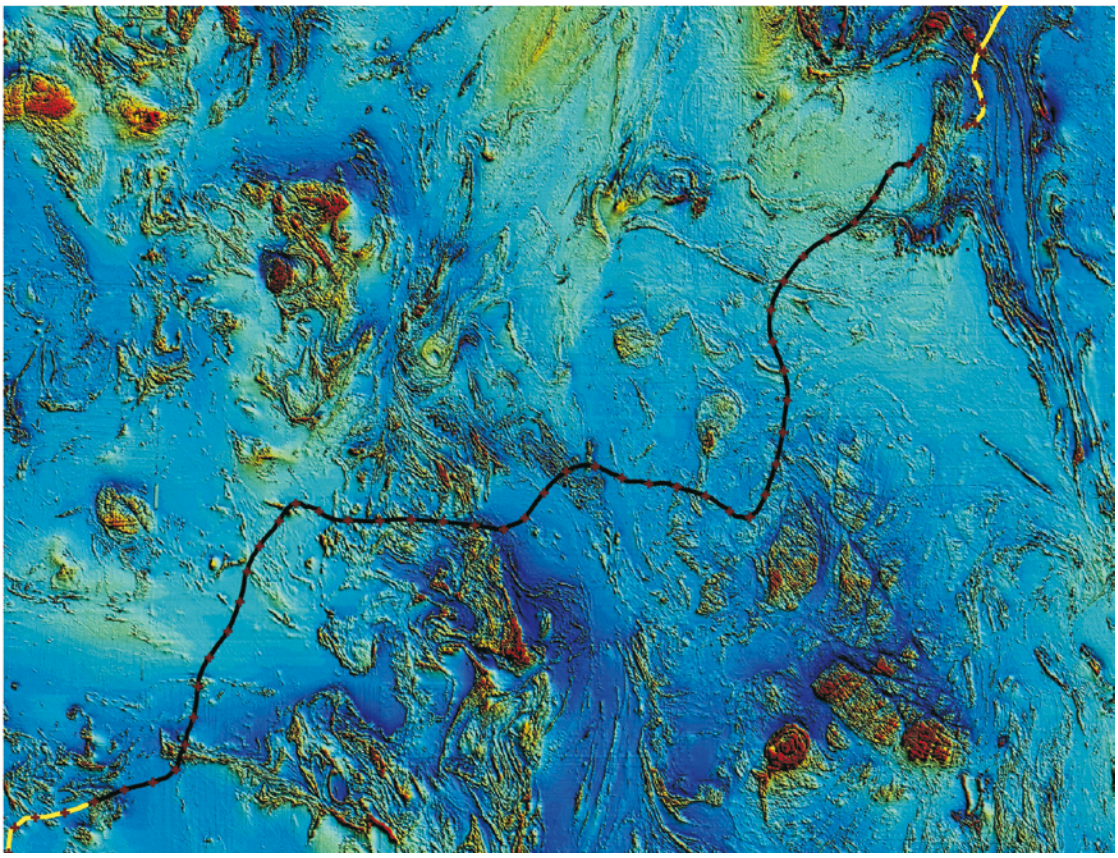

B)

Figure 8. a) Lithological map of Kainuu and Northern Savo between Kajaani – Pihtipudas areas modified after Koistinen et al. (2001). FIRE1 marked as a thick line. WK – Western Kianta Block; KB – Kainuu Belt; KC – Kainuu Complex; RC – Rautavaara Complex; IC – Iisalmi Complex; SB – Savo Belt; PB – Pihtipudas Block; KeB – Keuruu Block; CFGC – Central Finland Granitoid Complex.
b) A total intensity aeromagnetic anomaly map of Kainuu and Northern Savo between Kajaani – Pihtipudas areas (source GTK). Horizontal gradients are emphasized by vertical illumination of total intensity.

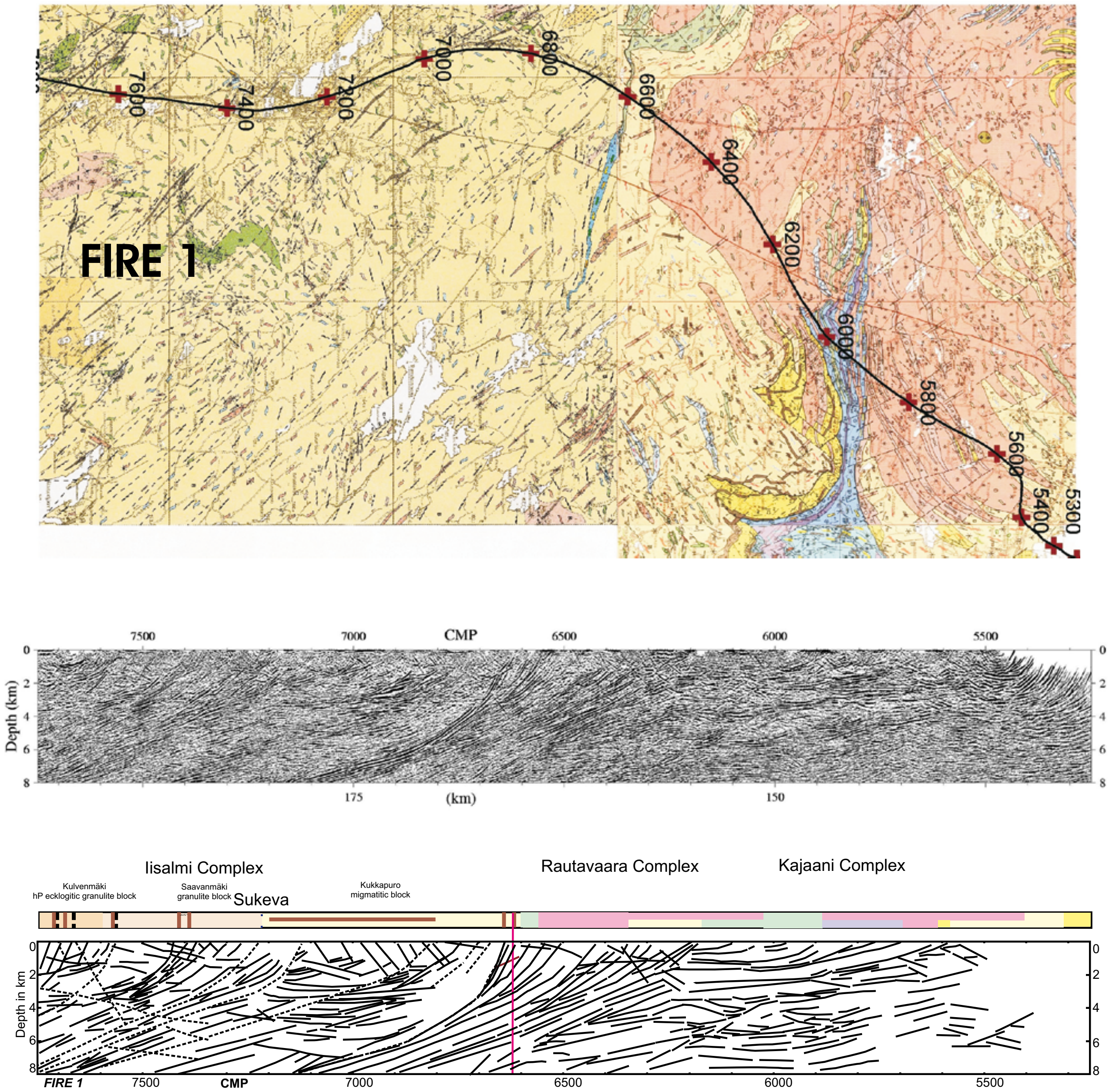


Figure 9. Uppermost 8 km of crust along FIRE 1 between CMP points 5250–7750. Upper panel: Lithological map (Havola 1981, Havola1997, Paavola 2003), middle panel; grey-scale variable intensity DMO section, lower panel; the lithology at surface and a line drawing. Major turning points are marked with a red line on the lower panel.

Based on the reflection properties the **Iisalmi Complex** (6650–9390; Fig.9) can further be divided into Kukkopuro, Kulvenmäki, Saavanmäki and Vieremä blocks. Between 6650 and 7760 there are three sets of reflections: shallow to moderately SW-dipping listric reflections, flat to shallow NE-dipping and moderately to steeply dipping reflections in the uppermost 2 km of the crust. The listric reflections are interpreted to arise from amphibolitic layers, mafic dykes and sills and shear zones. The NE-dipping reflections and the steep reflections are associated with mafic dykes.

Between CMPs 7200 and 7800, diabases and shear zones are striking NW-SE direction and they are at a 45° angle to FIRE line1. At CMP 7220 the line crosses a metamorphic block boundary between Kukkopuro amfibolite-facies block and granulite-facies Saavanmäki block. Within the Saavanmäki block (CMP 7220– 7600), the listric reflections dipping 55° at surface change gradually to reflections dipping 30° at the depth of 6 km. Where such reflections outcrop, diabases associated with shear zones are found between CMPs 7380–7420, 7560, 7640–7700. The bending reflections are cross-cut by weaker flat to shallow lying reflections at depths between 0 – 2.5 km. At the surface, the listric reflection at CMP 7760 is correlated with a metamorphic block boundary

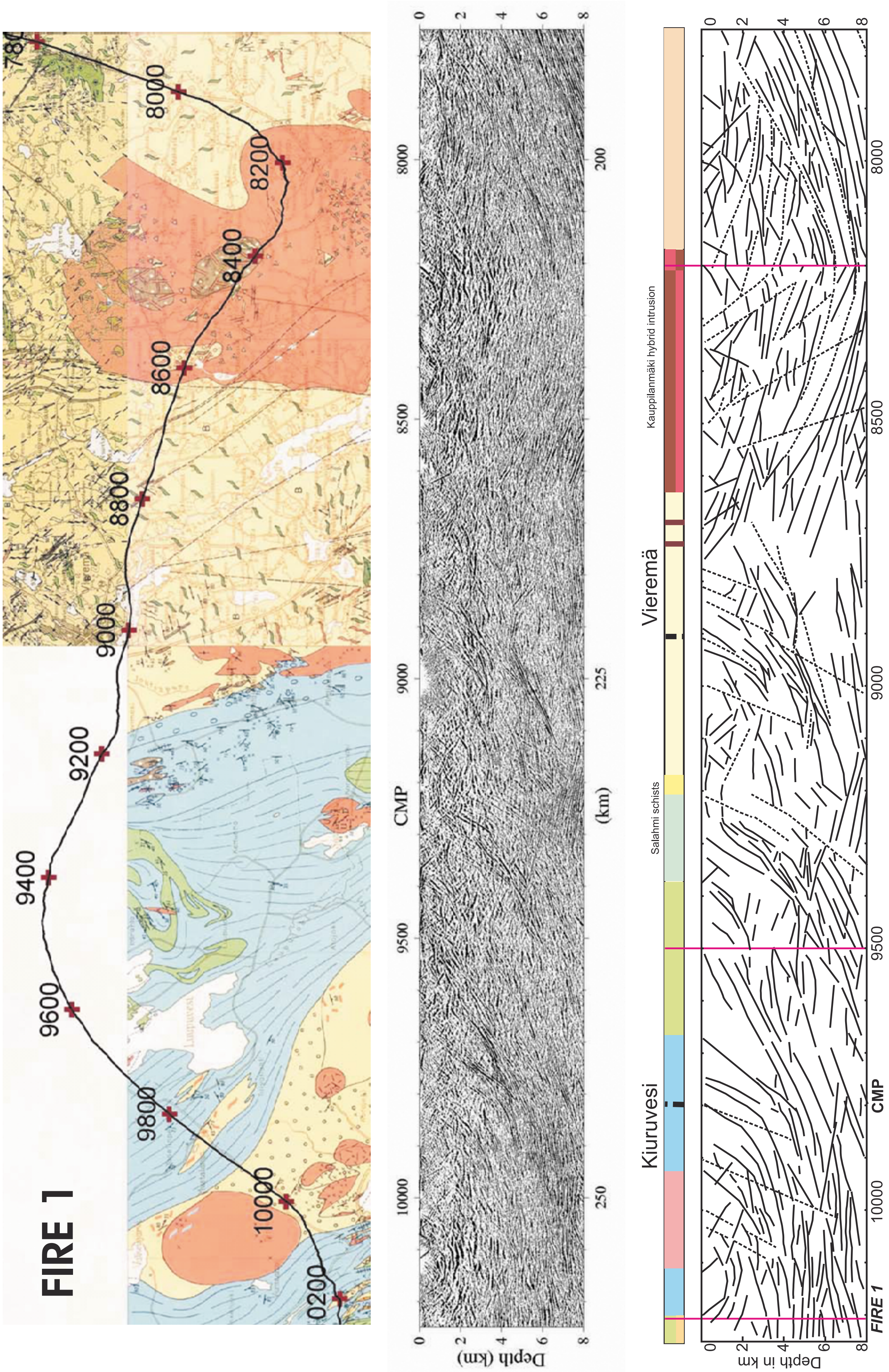


Figure 10. Uppermost 8 km of crust along FIRE 1 between CMP points 7750–10250. Upper panel: Lithological map (Marttila 1981, Paavola 1990, 2003), middle panel; grey shaded DMO section, lower panel; the lithology at surface and a line drawing. Major turning points are marked with a red line on the lower panel.

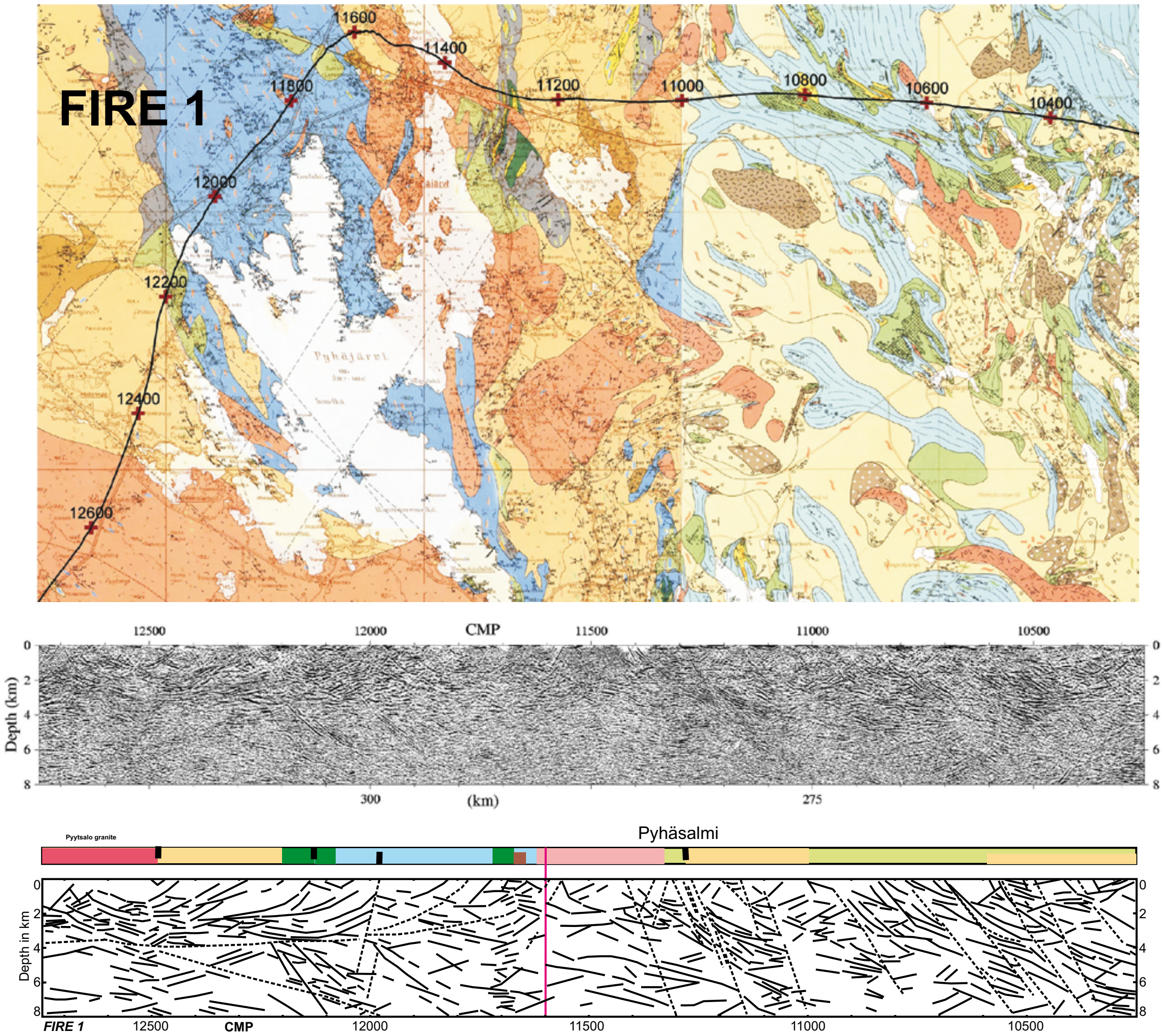


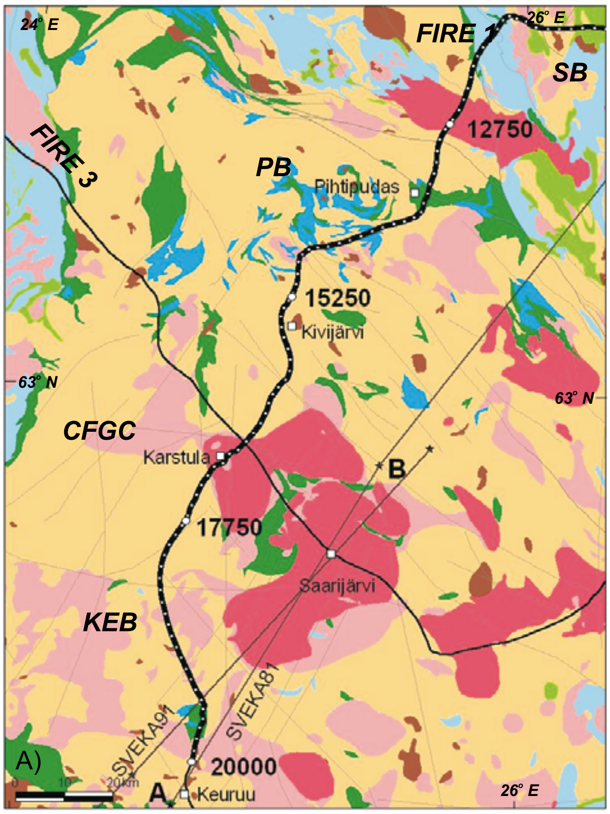
Figure 11. Uppermost 8 km of crust along FIRE 1 between CMP points 10250–12750. Upper panel: Lithological map (Marttila 1981, 1992), middle panel; grey-scale variable intensity DMO section, lower panel; the lithology at surface and a line drawing. Major turning points are marked with a red line on the lower panel.

between the Saavanmäki block and the Kulvemäki high pressure eclogitic granulites (P. Hölttä 2005, pers. comm.).

Within the Kulvemäki block (7760–8180; Figs. 9 and 10) the upper crustal reflectivity is weak and the reflections are flat to shallow dipping. At CMP 8200 the line takes a 100 degree turn to the west. In the uppermost crust, the dip of the reflections changes only little. Flat reflections change to shallow (15–20°) ones and it is interpreted that the shallow angle reflections are striking 350° and having a shallow dip towards the east. However, at the depths of 5 km to 8 km, the 30° dipping reflections change from S-dipping to E-dipping indicating that these structures are dipping SE. The shallow to moderately dipping reflections form a bowl structure whose contacts coincide loosely with the contact between high pressure granulites and amphibolite phases gneisses at CMP 7760 (P. Hölttä 2005, pers. com.).

Between CMPs 8150 and 8610 the shallow layered reflectivity within the area is correlated with the younger Proterozoic Kauppilanmäki granitoid intrusions. The western contact is sharp and associated with a steep (70°) east dipping reflector at CMP 8620. The eastern contact is more subtle but a change in general reflectivity pattern is observed. Close to surface between distances 8610 and 8730 reflections dipping 30° degrees east start of from the shallow dipping reflections. At surface these events can be correlated with mafic dykes at CMPs 8730 and 8680. Their magnetic anomalies can be followed to the 1.86 Ga old bimodal granite-gabbro Kaarakkala intrusion to the north (Fig. 8; Ruotoistenmäki at al. 2001).

Between distances CMPs 8780 and 9000 steeply



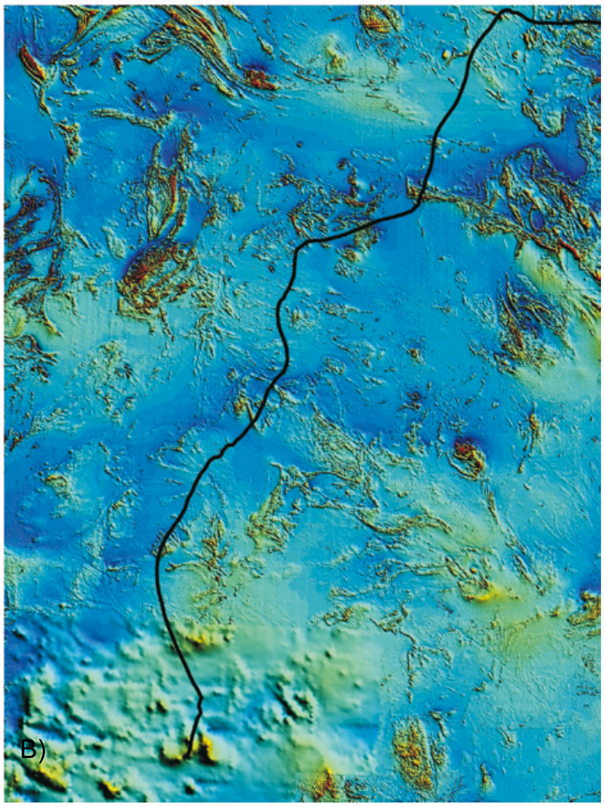


Figure 12. Lithological map of Central Finland between Pihtipudas – Keuruu areas modified after Koistinen et. al. (2001). FIRE1 marked as a thick line. SB – Savo Belt; PB – Pihtipudas Block; KeB – Keuruu Block; CFGC – Central Finland Granitoid Complex.
b) A total intensity aeromagnetic anomaly map of Central Finland between Pihtipudas – Keuruu areas (source GTK). Horizontal gradients are emphasized by vertical illumination of total intensity.

(70°) west dipping reflections are disrupting reflectivity that is dipping mainly to the west in the western parts and to the south in the eastern parts. A structural break takes places over a wider zone interpreted as a shear zone. At surface, the shear zone has been observed on magnetic map as well on outcrops (Figs. 8 and 10). The shear zone is the surface expression of the eastern edge of the Raahe-Ladoga shear zone and separates the strongly tectonized amphibolite-facies Vieremä block from the rest of the Iisalmi Complex. On a magnetic map the anomalies have been transposed in N-S direction.

Between CMP 9000 and 9250 the reflectivity is moderately west dipping or flat dipping in the uppermost 2 km. Across another steep poorly reflective structure at CMP 9220 the reflectivity changes to mainly steep (70°) and west dipping. At surface the change loosely correlates with Palaeoproterozoic autochthonous sequence outcropping between CMP points 9180 and 9390. At CMP 9390 the reflectivity changes abruptly to moderately west-dipping (25–30°) where at surface the crosses the thrust contact of the mica gneisses of the Svecofennian Savo Belt.

Across the **Savo Belt** (9390 – 12100; Figs. 10 and 11) the FIRE 1 profile changes direction from ESE-WNW to NE-SW at 9500 and to E-W at 10210. The changes in line direction are seen as changes in dip direction along the profile. Between depths 0–8 km and distances CMP 9390–10800 the reflections form a large open synformal structure, where the reflection dip towards the apparent west between CMP 9400–9920, sole out to flat-lying between CMP 9920–10200 and dip eastwards between CMP 10200–10800. Poor reflectivity is found where the reflections are flat-lying. In the west, the synformal structure is disrupted by steep reflections towards which the eastward dipping reflections are bending. At surface the weakly reflective area correlates with granitoid intrusions and the dipping bright reflections with the supracrustal sequences. The near vertical disruptive structures are interpreted as shear zones.

Another minor open synform in the upper most 4 km is found between distances CMP 10900 and 11300. At surface volcanic rocks are outcropping on the limbs of the synform, the centre part is characterized by tonalite intrusion. The synformal structure is disrupted by a set of steep reflections CMP 11030, 11050, 11200, 11260, 11280, dipping apparently 65°

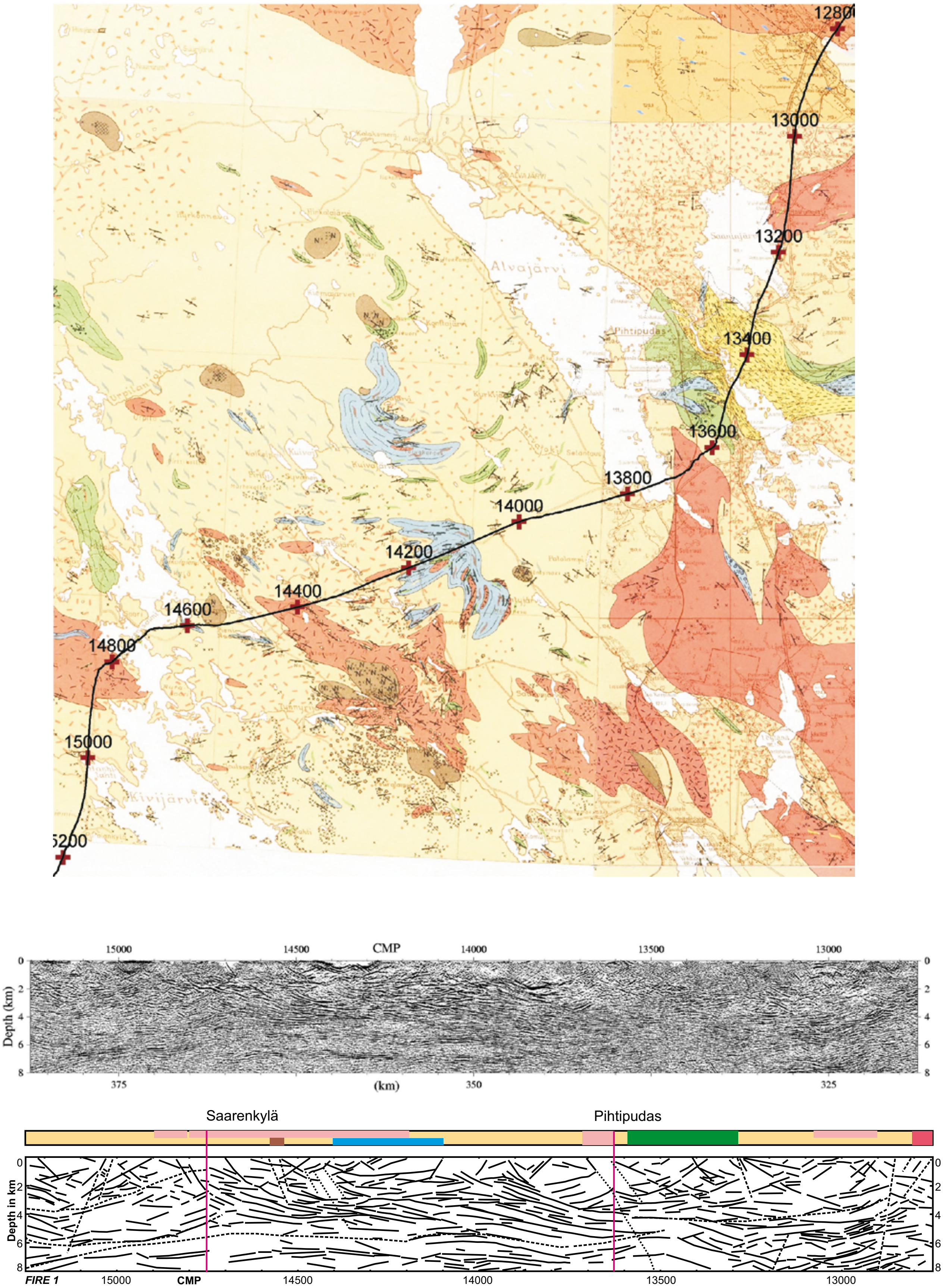


Figure 13. Uppermost 8 km of crust along FIRE 1 between CMP points 12750–15250. Upper panel: Lithological map (Nykänen 1963, Salli 1971, Marttila 1992), middle panel; grey-scale variable intensity DMO section, lower panel; the lithology at surface and a line drawing. Major turning points are marked with a red line on the lower panel.

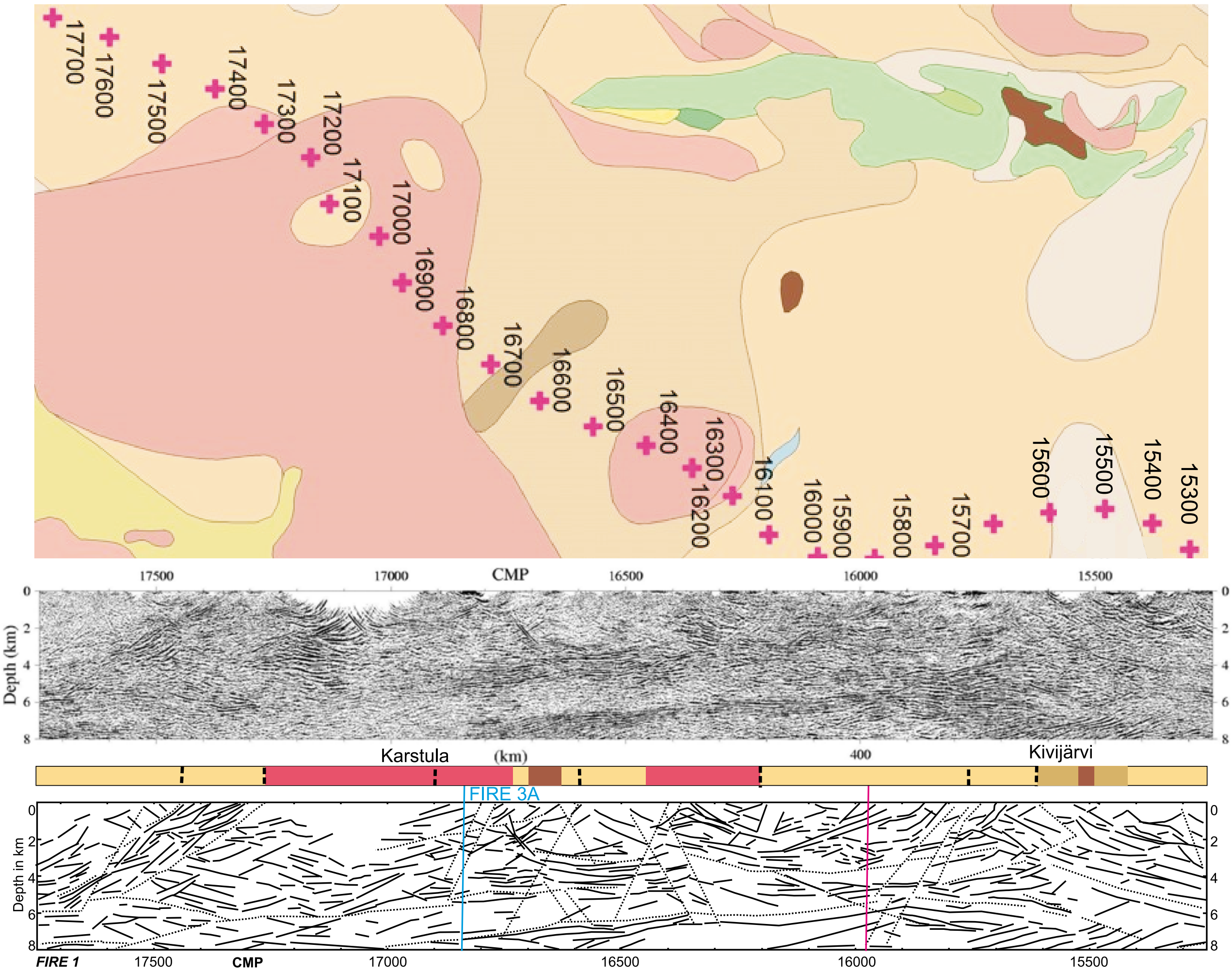


Figure 14. Uppermost 8 km of crust along FIRE 1 between CMP points 15250–17750. Upper panel: Lithological map (Nironen et al. 2002.), middle panel; grey-scale variable intensity DMO section, lower panel; the lithology at surface and a line drawing. Major turning points are marked with a red line on the lower panel.

to the east. At surface the area is transected by several sets of shear zones.

Between CMP 11300 and 11600 the reflections are shallow E-dipping (20–30°) or flat lying in the upper most crust (3 km). Further down the area is characterized by weak flat-lying reflectivity. At surface, this unit is associated with granitoid intrusion. At CMP 11600 the line takes an abrupt turn of 100° degrees towards the south-southwest. Between CMP 11600 and 11980 the upper crustal reflections dip shallowly (20°) towards the N and are cross-cut by listric S-dipping reflections down to the depth of 4 km, where they terminate at southward dipping, low angle reflections. The north-dipping reflections disappear across a near vertical reflection (80°) at CMP 11980. South of the vertical structure (11980–12410) the reflections are dipping southwards and soling out at the depth of three kilometres. At surface, the reflections between CMP 11300 and 12410 are associated with a migmatitic gneiss unit and a minor volcanic unit and intruding gabbro. Between CMPs 12410 and 12200 the S-dipping reflections are associated with a small volcanic unit.

It is interpreted that the N-dipping structures are associated with the Savo Belt and that the S-dipping structures are associated with the Central Finland Granitoid Complex. The reflection section suggests that the change from Savo Belt to Central Finland Granitoid Complex is gradual and that the structures related to the Central Finland Granitoid Complex are overprinting the structures related to the Savo Belt.

Based on the reflective properties the **Central Finland Granitoid Complex** (12100 – 20000; Fig. 12) is divided into Pihtipudas (12100 – 17400) and Keuruu blocks (17400–20000). **The Pihtipudas block** (12100 –17400; Figs. 13, 14 and 15) is characterized by shallow to flat lying reflections soling out at depths between 3 km and 6 km. The shallow reflections are disrupted by steep reflections (CMP 15600–15650, 15750–15780, 16370, 16750, 17250, 17500, 17800, 17920, 18700, 18770, 18910) dipping to S (SW) or N (NE) terminating at the soling surfaces. The reflec-

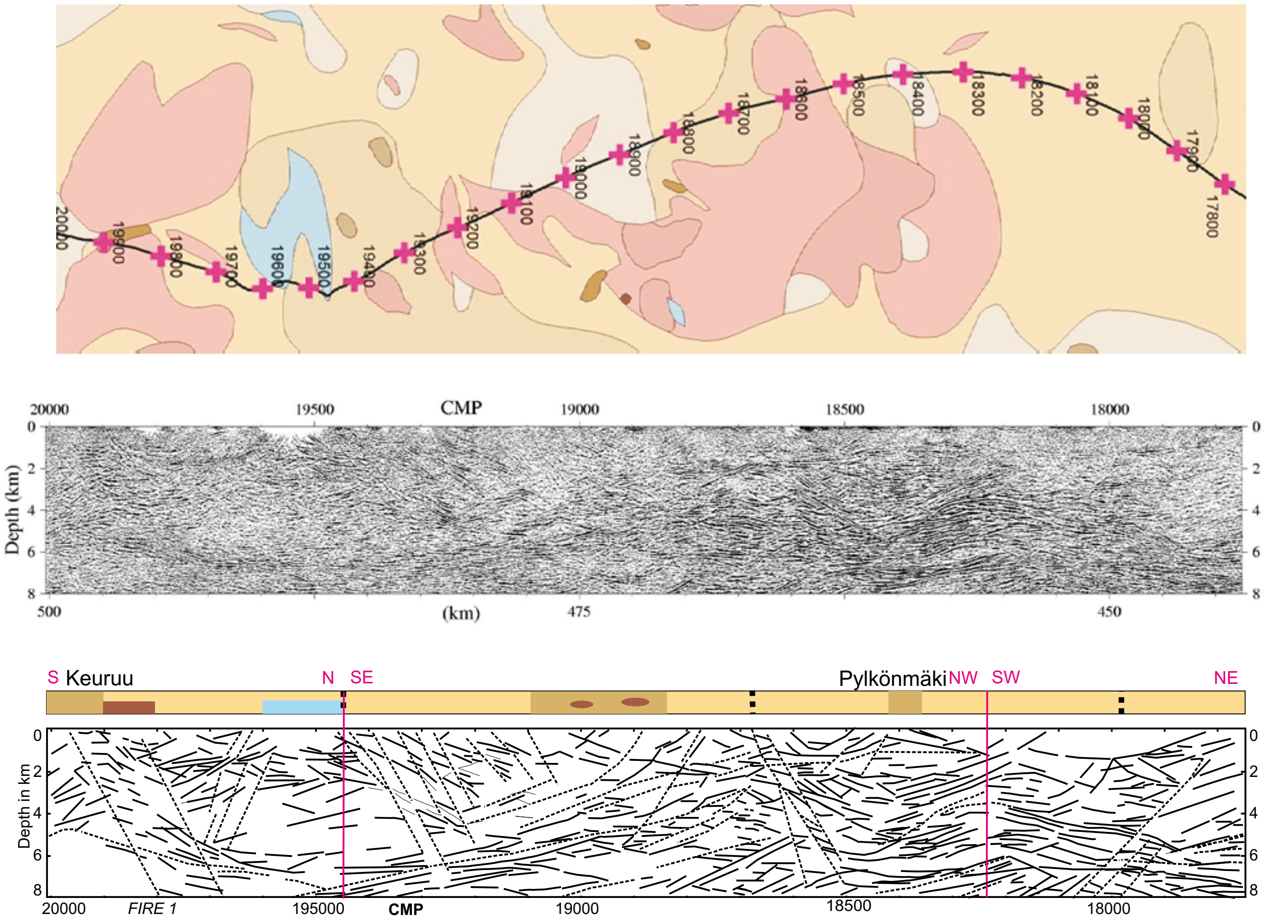


Figure 15. Uppermost 8 km of crust along FIRE 1 between CMP points 17750–20000. Upper panel: Lithological map (ref.), middle panel; grey-scale variable intensity DMO section, lower panel; the lithology at surface and a line drawing. Major turning points are marked with a red line on the lower panel.

tions are associated with linear magnetic anomalies, interpreted as fault zones, striking NW-SE. The faults delineate a graben and horst structure. Between CMPs 12800 and 13620 the reflections dip to the south and change to E-NE dipping where the line takes a turn to the W-SW. The geometry of the dips suggests that their true dip direction is to the E-SE and that they are striking NE-SW.

Some of the steep faults (CMP 12800, 16400, 16800, 17250) are at the approximate location of a post-kinematic granite contacts others are at or close to the contacts of gabbros (CMP 14550, 15600). Gabbroic rocks are associated with weak horizontal reflections that are placed above bright 20° degrees reflections (CMP 14530–14580, 15470–15450). The post-kinematic granites are associated with weak reflectivity. The reflective structures come closest to the surface where the Pihtipudas volcanic unit (13560–13280) and Saarenkylä volcanic and supracrustal units (CMPs 14100–14700) are exposed.

The Keuruu block (17400–20000; Fig. 15) is rather poorly reflective with minor shallow dipping (20°) reflections flattening out at the depths of 2 km, 6 km and 10 km. The ones flattening at 10 km are dipping S-SW and are cutting the other ones. The general flattening surface is deeper than in the Pihtipudas block. The shallow dipping reflections are cut by near vertical (72°) reflections that at surface correlate with shear zones (CMP) and/or point like gabbroic intrusions (CMP 18900, 19910). The FIRE 1 line takes a broad turn from S-SW to SE around 18250, which causes the reflections that flatten out at the depth of 6 km to change direction from apparently N-dipping to S-dipping indicating that the true dip direction is to the E and that the strike of the structure is N-S directed.

There are also steep reflections dipping north and continuing to the depths of 20 km at CMPs 19900 and 19450. These structures form block boundaries between tilted upper and middle crustal blocks.

A GEOLOGICAL INTERPRETATION

The upper crust along FIRE 1 (Fig. 16 and Appendix 1) is composed of ten units belonging to three domains: Archaean nucleus, A-P boundary zone and Proterozoic Central Finland. Of these, the upper crustal structure of the Archaean nucleus is dominated by Archaean structures, the A-P boundary zone by both Palaeoproterozoic extensional structures due to rifting and collisional structures related to Svecofennian orogeny and the Proterozoic Central Finland domain by both collisional and gravitational collapse structures of the Svecofennian orogen.

The nucleus is a collage of Archaean blocks and complexes. Each of the units has a different seismic character and the character changes at the block boundaries, which are normally tectonized. The E-W striking internal layering of the granitoid gneisses and greenstone remnants within the Eastern Kianta block are imaged as flat or shallowly dipping reflections. These are cross-cut by NW-SE directed steep reflections interpreted as shear zones and diabase dykes. Also NE-SW directed diabases are found. Both late Archaean granitoids as well as early Palaeoproterozoic diabases seem to have used the shear zones as pathways during emplacement. The mafic magmas have also occupied flat lying layer boundaries (layered intrusions). The Kuhmo Greenstone Belt is a weakly reflective block with steep west-dipping, tectonic contacts.

The Archaean structural trends are dominantly E-W in the Eastern Kianta block and close to the Kuhmo Greenstone Belt in the Western Kianta block. On the contrary the Kuhmo Greenstone Belt is oriented N-S. One possible explanation is that the Kuhmo Greenstone Belt is located in an Archaean N-S strike-slip zone overprinting and turning the E-W structures and thus, forming a crustal depression favourable for the preservation of the Kuhmo Greenstone Belt.

The majority of reflection structures in the Eastern Kianta are interpreted to originate from Archaean collisional and extensional processes (Belomorian). These structures have been reactivated and enhanced by early Palaeoproterozoic mafic magmatism and shearing related to extensional effects of numerous onsets of rifting. Although small-scale extensional features, e.g., faulting and jointing, are probable, it is proposed that only minor thinning of the Archaean crust occurred in the Eastern Kianta and Kuhmo Greenstone Belt during the Palaeoproterozoic rifting events. During the Svecofennian orogeny the Eastern Kianta block and Kuhmo Greenstone Belt suffered a thermal overprint caused by overriding crustal pieces. The relatively unstrained nature of the Palaeoproterozoic diabases (Kontinen 2002) suggests only limited deformation.

In the Western Kianta and especially in the block west of a set of steep shear zones at CMP 2170–2290 (see above) the diabase dykes change direction and they are locally highly strained (Kontinen 2002), and the overall background magnetization pattern changes to more rough. This abrupt change coincides with the non-reflective nature of upper crust. It is proposed that this is a major internal block boundary of the Western Kianta, which divides the Karelian Domain into stretched and non-stretched parts. The stretched part includes all the Archaean rocks west of this boundary. As an entity the Western Kianta block seems to comprise a rootless, allochthonous block that has been thrust on top of the Eastern Kianta and Kuhmo Greenstone Belt. The Western Kianta block is composed of small thrust sheets and sigmoidal thrust lenses bordered by westward dipping zones of high reflectivity.

The autochthonous units (marginal series) are deposited on the western edge of the Western Kianta. Together with the 2.4–2.0 Ga dykes they form a highly reflective unit. The contact between the autochthonous cover sequence and the eastern allochthonous unit of the Kainuu Belt is west dipping. Both the west-dipping allochthonous and the autochthonous units can be followed at least to the depth of 10 km based on the reflectivity pattern (Fig. 7). Magnetotelluric data (Korja & Koivukoski 1994) also suggests graphite-bearing metasedimentary units of the autochthonous series to continue to mid-crustal depths. It is interpreted that sheets carrying allochthonous units including metasedimentary rocks have been thrust on top of the autochthonous units.

The Archaean sliver and associated autochthonous cover, that divides the Kainuu Belt into eastern and western segments, tapers down to the south and coincides with ophiolite-related serpentinites of the western segment (Havola 1981). These serpentinites continue south to the Kaavi-Outokumpu area as squeezed fragments in a thrust/fault zone between two Archaean blocks (Korsman et al. 1997). On FIRE 1, the Archaean sliver within the Kainuu Belt is interpreted as a tectonic slice of the Kajaani Complex later transposed to vertical position during late-stage shearing (e.g., Kärki & Laajoki 1995). The ophiolitic remnants occur only in the western segment, which is a thin sheet (< 2 km) without continuation to a depth (Fig. 16). The Jormua ophiolite has been divided into blocks (Peltonen & Kontinen, 2004), which show a polarity in the occurrence of the EMORB dykes from western (none) via central (increasing to SE) to eastern block (voluminous with extrusive units).

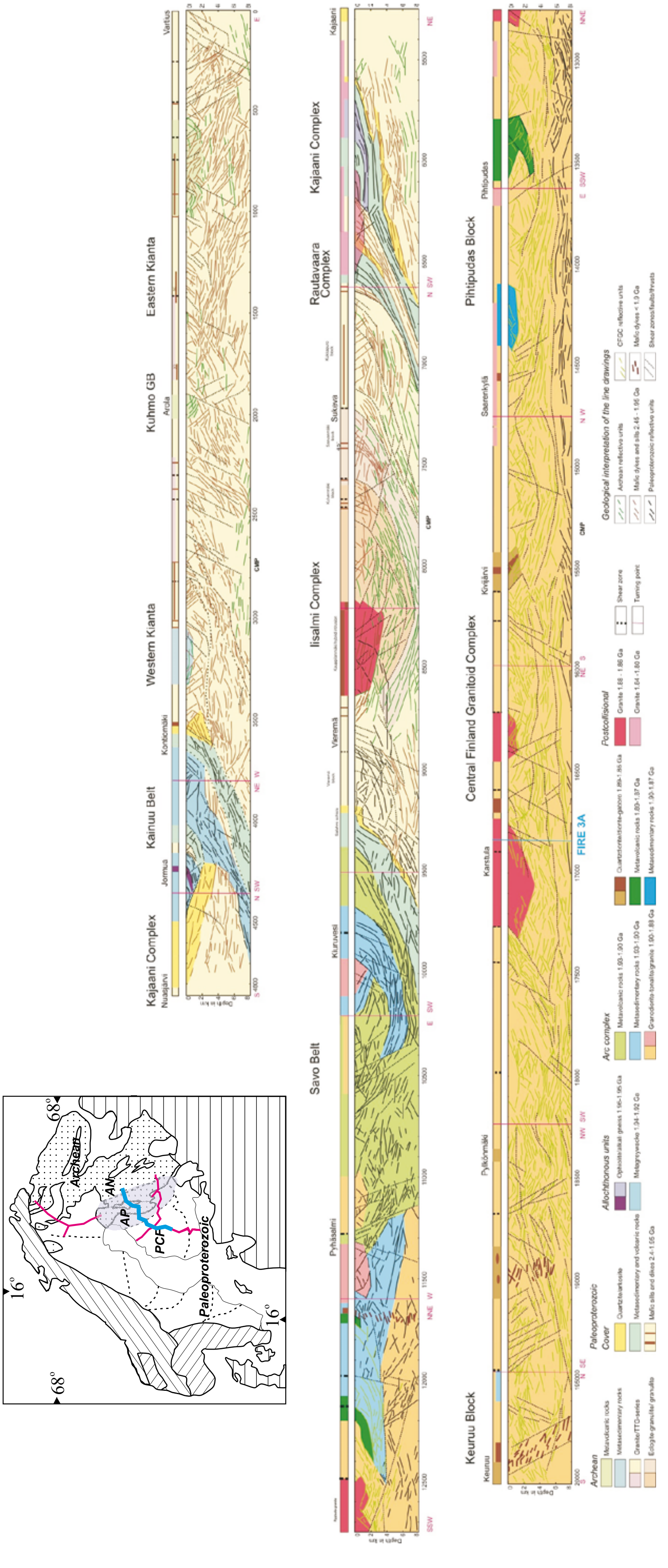


Figure 16. Geological interpretation of the upper structure of the upper crust along FIRE 1 (see Appendix for 1:250 000 scale)

The eastern block occurs as a tectonic sliver within turbidite sequences whereas the western block occurs within Archaean blocks and autochthonous rocks. It is proposed that the Jormua ophiolite, which is a remnant of a sub-crustal lithosphere (Peltonen & Kontinen 2004), back-thrust on the western margin of a Red Sea-type marginal basin during the onset of E-W collision. The following thin- and thick-skinned thrusting has been from the west, in the opposite direction, thus burying and preserving the ophiolite from erosion. Previously Peltonen & Kontinen (2004) have suggested a passive margin model where Jormua ophiolite has been transported as a nappe from the west onto an older rift basin.

The Kajaani Complex is composed of Archaean slivers interlayered with Proterozoic supracrustal units that are imaged as reflective thrust surfaces and at surface as ribbon like supracrustal remnants with strong schistosity and lineation (Havola 1997). Alkalic gneisses (1.96 Ga) that are related to rifting producing Jormua ophiolite (Peltonen & Kontinen 2004) are confined in a restricted area around the western branch of the autochthonous cover sequence and it remnants in the west. FIRE 1 profile follows semi-parallel the shallow SE dipping alkalic gneisses and thus the reflection image is a flat-lying bowl. The Kajaani Complex and the western segment of the Kainuu Belt have been thrust over the eastern segment of Kainuu Belt and western edge of the Western Kianta block. The Kajaani granite occurs as thin sheets intruding the stacked units within the Kajaani Complex.

A small sliver of the Rautavaara Complex is interpreted to occur between the Kajaani and Iisalmi Complexes along FIRE 1. It is seen as a strongly reflective unit arising from abundant diabases, which have been highly strained and flattened during Svecofennian thrusting. It seems that the Rautavaara Complex has been thrust on the Kajaani Complex. The western limb of the autochthonous cover of the Kajaani Complex can be followed with depth (Fig. 16 and Appendix) indicating a closed basin between the Rautavaara and Kajaani Complexes.

The stack continues to the west-south west where Iisalmi Complex has been thrust on top of the Rautavaara and Kajaani Complexes. The Iisalmi Complex is composed of Archaean slices whose contacts are flat to shallow dipping to the southwest. The internal metamorphic block boundaries at surface are imaged as listric reflections and interpreted as major Archaean thrust surfaces. The layered reflectivity pattern of the granulite slivers is interpreted to be caused by amfibolitic layers. The shallow dipping structures are crosscut by steep to vertical reflections that at surface are correlated with diabase dykes and shear zones. The diabase dykes seem to branch out from more shallow dipping structures at depth suggesting that some of the old surfaces may have been reused during the diabase emplacement. In the end, some of these surfaces have been reused as stacking surfaces during the Svecofennian collision. The generally little strained nature of diabase dykes indicate that the Iisalmi Complex has acted as a more intact unit relative to the highly strained Rautavaara Complex during crustal-scale stacking. One of these intact units is the granulitic Kulvenmäki and Saavanmäki blocks which occupy a bowl-shaped unit in the uppermost crust. A crustal scale shear zone separates the strongly tectonized amphibolite-facies Vieremä block from the rest of the Iisalmi Complex. The reflection data suggest that both thrusting and later strong shearing has affected this block. The Salahmi autochthonous cover sequence at the western margin of the Vieremä block may be few kilometres thick. The suture between the Savo Belt and the Iisalmi Complex lies on the thrust surface along which, the Savo Belt (Pietikäinen & Vaasjoki 1999) has been stacked on top of the Vieremä block.

The Savo Belt (1.92 Ga), along FIRE 1 line, is composed of alternating volcanic- and sediment-dominated units, where the more well preserved areas, e.g. Pyhäsalmi, are exposed in a narrow block with vertical shear zone contacts. The structure of the Savo Belt has been radically modified by steep to moderate dipping shear zones and the voluminous occurrence of 1.89–1.88 Ga granitoids, mafic rocks and diabases. The bending reflections indicate upward movement of the blocks and could be related to the metamorphic block structure observed in the Pielavesi-Kiuruvesi area (Haudenschild 1995, Korsman & Glebovitsky 1999). In the eastern part of the Savo Belt, a long wavelength synclinorium has been developed on the thrust unit. In the central part of the Savo Belt, the FIRE 1 line cross cuts the belt semi-parallel to the main strike of volcanic units, which is seen as shallow to flat-lying reflections.

The western part of the Savo Belt is composed of the Pyhäjärvi mica gneisses, which have been interpreted stratigraphically to occur between the 1.92 Ga Savo Belt and younger 1.89–1.88 Ga volcanic rocks (Kousa et al. 1994). Based on the FIRE1 data the mica gneiss unit might be divided into two parts where the western part would underlie the younger volcanic rocks. The east dipping part, starting under the granite, could possibly underlie the 1.92 Ga volcanic sequence. The strong reflections below the Pyhäjärvi mica gneisses in upper crust are interpreted as feeder dykes for the younger 1.89–1.88 Ga volcanics and gabbros.

The change in the dominant direction of 1.89–1.87 Ga dykes from NW-SE (east) to E-W directed (west) in the Pyhäsalmi area (Marttila 1993) indicates an important block boundary. Possible explanations are either a relatively late rotation of the eastern block to the NW-SE direction parallel to the Archean-Proterozoic boundary or the blocks have had different extensional regimes during dyke emplacement.

The Central Finland Granitoid Complex is only a few kilometres thick, highly reflective upper crustal feature. It is composed of graben and horst structures where the grabens are occupied by poorly reflective granitoids and horsts by well reflective volcano-sedimentary / supracrustal remnants. The formation of the granitoid complex is interpreted be related to upper crustal extension. When intruding the crust gabbros seem to have left horizontal steps behind indicating stepwise rise of the magmas. The near vertical shear zones seem to dip to the south in the northern part and to the north in the southern part indicating crustal scale tilting of the blocks. Nironen (2003) proposed that the dextral Group 2 (120–135º) faults have controlled the emplacement of many postkinematic granitoids, and that the style of faulting changed from transpressional to transtensional or extensional during 1.88–1.87 Ga.

DISCUSSION

In general, the reflectivity of the crystalline bedrock is weaker than that of sedimentary basins (Milkereit and Eaton 1998) and thus the signal to noise ratio is often poor. For a better ratio, a high fold of the data is required. In FIRE project a 90- fold of the data has been used (Kukkonen et al. 2006, *this volume*) and the reflection data is of good quality and the reflection patterns can be correlated with geological structures.

Theoretically the crystalline bedrock shows two end member classes of strong, prominent reflections: contacts between mafic and felsic units and high-density mineralization hosted in common silicate rocks. Of these particularly dikes and sills belonging to the first group have proven to be excellent markers as they show continuity and subhorizontal contacts that are easy to detect with the reflection method (Milkereit and Eaton 1998). In the Sudbury structure, a detailed correlation of the seismic reflection images with deep bore hole measurements of both density and seismic P-wave velocity have revealed that most of the reflections arise from lithological contacts and some from faults and deformation zones. Fractures in the uppermost 300 m may destroy the seismic response of lithological contacts. The most prominent reflections are associated with mafic intrusion boundaries and sulphide mineralizations (Milkereit et al. 1994, Wu et al. 1995). Because the crystalline crust is seldom layered it cannot be approximated as two dimensional. New seismic reflection techniques have been developed for the three-dimensional crystalline bedrock (e.g. Eaton et al. 2003).

In the upper crust of the FIRE 1 profile the reflections are interpreted to be caused by lithological and tectonic contacts, shear zones or their combination. A correlation with deep bore hole measurements is available in the easternmost part of FIRE 1 profile where seismic velocity and density measurements are available from two industrial boreholes about 20 km south of FIRE 1 (Anttila et al. 1999). Seismic velocity and density measurements also exist for a set of hand samples collected from the Archaean area by the Geological Survey of Finland. Based on these results it is interpreted that the layered reflectivity Archaean gneiss and granulite blocks could arise from alternating layers of granitoid gneissic rocks (5335 km/s, 2718 kg/m^3) remnants of amfibolite (6010 km/s, 3016 kg/m^3) layers (Silvennoinen 2004). One of these contrasts is observed at surface in the Iisalmi Complex where a well reflective structure coming to the surface is coinciding with a sheared amfibolite layer. Other well-reflective lithological contacts are formed by the mafic dykes and sills (layered intrusions) from which also high velocity and density (6484 km/s, 3051 kg/m^3) have been observed in deep boreholes (Anttila et al. 1999, Silvennoinen, 2004).

The mafic dykes seem to follow a horse tail pattern, where a thick and bright reflection packet branches out as a set of minor reflection curving upwards towards the surface. Such weak reflections are more difficult to identify both on the reflection section and from the bedrock. These sets of reflections are, however, broadly coinciding with swarms of diabases at the surface. The most prominent reflections outcropping at surface are associated with the layered sills. These sills are normally thicker than dykes and occur intruding the cover rocks implying a shallower intrusion depth relative to dykes. If the diabases are very steep they are devoid of reflectivity and they are observed indirectly as disrupted seismic signature. In many places diabases occur in shear zones and it is impossible to tell from the seismic section, which was first the shear zone or the diabase. But from tectonic point of view it is clear that diabases are most likely to intrude existing weaknesses of the crust. They are merely imaging the existing zones of weaknesses caused by extensional stress field. The horizontal reflections probably image the roofs of crustal magma chambers or those horizontal weakness zones where the magma pressures where equivalent to the overlying lithostatic pressures.

Shear zones at surface are often displayed as white, steeply dipping bands crossing the other structures. The pattern can be understood to develop as a consequence of shear deformation destroying the coherency of the bedrock units or their elastic properties. Thrust zones on the other hand are generally listric in shape and usually highly reflective suggesting that the thrust surface may have developed on existing competence differences. At places the listric reflections flatten out at depth and join a band of reflections interpreted as detachment zones or basal thrusts. It is suggested that the associated high reflectivity is caused by a set of low velocity faults occurring within the higher velocity material. If the shear zones were 25 m wide and spaced 25 m apart they would give high impedance contrast causing high reflectivity.

The outcropping igneous bodies are very shallow, only 1 to 4 km in depth. In many cases, the contacts with the surroundings are associated with dipping reflections interpreted as shear zones. The intrusions have flat to shallow dipping internal layering and a sharp lower contact. Some of the intrusions take the form of half graben. The intrusions have probably emplaced as nearly horizontal layered intrusions that grew with time and they were fed via shear zones (Petford et al. 2000).

Pre-existing weaknesses are the most likely places for failure to take place. Such reactivation of fault and shear zones is seen at places where mafic dykes have first either intruded a fault or created a fault and then later the dykes have been sheared. Structural inversion is suggested to have taken place within the Kainuu Belt, which first develops as an extensional basin and later inverts into a thrust stack. Similarly structural inversion has taken place within the Archaean units west of the Kuhmo Greenstone Belt. In the Iisalmi Complex, the Archaean thrust surfaces between blocks of different metamorphic grade were later intruded by Proterozoic diabase dykes. The crustal-scale and local thrusting is displayed as shear zones and highly strained dykes and block margins. The thickness of thrust sheets varies from a few km, in the Rautavaara and Kajaani Complexes, and the Vieremä block, to 10–20 km in the Western Kianta and Iisalmi Complexes.

The effects of three dimensions are well displayed in the Archaean Kianta block, were at surface steeply dipping diabase dykes are observed to have a wide range of apparent dips depending on the interception angle. Near vertical structures (80°) are observed when the interception angle is close to 90°, apparent dips around 30° are found when the interception angle is less than 30° and flat structures when the line is running parallel to the diabases or other studied structures. Other effects related to three dimensional effects are structures with changing dip such as curving reflections and apparent synform and antiform structures.

The high velocity body found in the upper parts of the SVEKA81 profile can loosely be correlated with the allochthonous Iisalmi Complex hosting high pressure mafic granulites. The low velocity layer beneath the Central Finland Granitoid Complex is situated just below the highly reflective unit, which in turn seems to be associated with a velocity increase. It is interpreted that the low velocity layer in the velocity model of Grad and Luosto (1987) is more related to the disappearance of unusual amounts of mafic material rather than an introduction of abnormally low velocity material.

The results of this study indicate that the Eastern Kianta together with the Kuhmo Greenstone Belt and adjacent area to the west form the least stretched part of the Karelian craton whereas the Archaean to the west is composed of more stretched units, which have been thinned considerably during Palaeoproterozoic rifting stages, leading finally to the break-up of the craton. Later a marginal basin also forms close to the craton edge. The complex structure of the thinned continental margin leads also into complex stacking pattern during collisional thickening. All these features favour the continental arc/continent collision model (e.g. Gaál 1990, Lahtinen 1994, Kohonen 1995, Lahtinen et al. 2005) for the Archean-Proterozoic boundary. Lahtinen et al. (2005) also suggested that the thickening of the boundary zone was due to two collisional stages; E-W (Lapland-Savo orogeny) and N-S (Fennian orogeny).

CONCLUSIONS

The upper crustal reflectors correlate well with surface geology in the scale of hundred meters. The integrated use of surface geology and aeromagnetic anomaly maps together with seismic reflection data enables a vertical continuation of the surface structures and lithologies and thus it is possible to create a vertical lithological map of the upper crust.

When interpreting the crustal sections great care has to be taken to track all the major changes in the profile geometry as changes in interception angle change the dip directions markedly. Interception angle may also change naturally along the geological structures. Cross-cutting relationships should be determined from wiggle-trace sections.

It is concluded that major lithological units have different seismic reflectivity patterns that change at the boundaries. Sometimes the boundaries have a positive reflection response or they may be transparent but sometimes the boundaries may be drawn where the seismic pattern changes. The most prominent reflections seem to originate from lithological contrast associated with mafic intrusions or volcanic units. Shear zones are seen as white bands disrupting the seismic patterns.

The upper crust along FIRE 1 is composed of ten units belonging to three tectonic domains: Archaean nucleus, A-P boundary zone and Proterozoic Central Finland. Of these, the upper crustal structure of the Archaean nucleus is dominated by Archaean structures, the A-P boundary zone by both Palaeoproterozoic extensional structures due to rifting and collisional structures related to Svecofennian orogeny and the Proterozoic Central Finland domain by both collisional and gravitational collapse structures of the Svecofennian orogen.

The Eastern Kianta together with Kuhmo Greenstone Belt and adjacent area to the west is considered to be the least stretched part of the Karelian craton. These areas show dominantly Archaean structures overprinted by Palaeoproterozoic diabases and minor shearing and faulting. The Western Kianta is composed of more stretched units, which have been thinned considerably during rifting stages leading finally to the break-up of the craton. The scale of thrusting varies from less than km to 10–20 km stacking where the effects of thrusting are mainly seen along the block boundaries.

The eastern segment of the Kainuu Belt has been squeezed between Western Kianta Block and Kajaani Complex. Kajaani Complex and the western segment of the Kainuu Belt, including the ophiolites, form an upper crustal allochthon, which has been thrust on the eastern segment of the Kainuu Belt. The reflection result supports the idea that the ophiolites have been part of a Red Sea- type marginal basin that has closed during collision. The Rautavaara and Iisalmi Complexes have been thrust on the Kajaani Complex. The continuations in depth of the western limb of the autochthonous cover of the Kajaani Complex show a separate basin between Rautavaara and Kajaani Complexes. As a whole it is interpreted that the Western Kianta block, Kainuu Belt, Kajaani, Rautavaara, Iisalmi Complexes as well as Savo Belt were all thrust sequentially on the Eastern Kianta block during the Svecofennian orogeny.

The Savo Belt is a synclinal exotic terrane with open internal folding. It has been squeezed between Iisalmi Complex in the east and an unexposed complex to the west. It has been thrust on the Iisalmi Complex. Later shearing has enabled vertical movements and the formation of the block structure.

The Central Finland Granitoid Complex is a shallow (3–8 km) unit, whose lower surface is a detachment zone. The detachment surface is associated with upper crustal graben and horst -structures. The complex is an upper crustal extensional structure that developed on top of the Keitele microcontinent and schist belts surrounding it.

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Appendix:

A geological interpretation of the upper crust along FIRE 1.

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