

DOCTORAL THESIS

Crustal Structure, Tectonics,
and Magmatic Evolution of the
Estonian Proterozoic Basement:
Geophysical and Geochemical
Insights

Juan David Solano Acosta

TALLINN UNIVERSITY OF TECHNOLOGY
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Declaration:

Hereby I declare that this doctoral thesis, my original investigation and achievement, submitted for the doctoral degree at Tallinn University of Technology, has not been submitted for any academic degree elsewhere.

Juan David Solano Acosta

signature

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TALLINNA TEHNIKAÜLIKOOL
DOKTORITÖÖ
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**Eesti proterosoilise maakoore ehitus,
tektoonika ja magmaliste protsesside
areng: geofüüsikaline ja geokeemiline
vaade**

JUAN DAVID SOLANO ACOSTA

Contents

List of Publications and Manuscripts	7
Author's Contributions to the Publications and Manuscripts	8
Abbreviations	9
1 Introduction	10
2 Objectives of the Dissertation	16
3 Geological setting and Literature Background	17
3.1 Tectonic Setting	17
3.2 Estonian Domains	18
3.3 Mesoproterozoic AMCG Rapakivi Magmatism	19
4 Methodology	21
4.1 Datasets	21
4.1.1 Geophysical Datasets	21
4.1.2 Geochemical Datasets	21
4.1.2.1 Alutaguse and Southern Svecofennian (SS) Domains	21
4.1.2.2 Rapakivi Dataset (Märjamaa–Kloostri)	29
4.2 Geophysical Modelling	32
4.2.1 Overview and Software Workflow	32
4.2.2 Geophysical Preprocessing and Filtering	32
4.2.3 Derivative Attributes and Structural Edges	32
4.2.4 3D Joint Inversion	33
4.3 Geochemical Analysis	33
4.3.1 Alutaguse and SS Data Integration and Classification Procedures ...	33
4.3.2 Processing Major-Element Data: Märjamaa–Kloostri Rapakivi Gran- ites	35
5 Results and Discussion	37
5.1 Estonian Basement Architecture	37
5.2 Relationships Between the Alutaguse Zone and the South Svecofennian (SS) Region	38
5.2.1 Geodynamic Implication	41
5.2.2 North Eastern Estonian Potential-Field Analysis and Exploration Implications	42
5.3 Geophysical and Petrogenetic Insights from Märjamaa and Kloostri Rapakivi	49
5.3.1 Geophysical Results	49
5.3.2 Petrogenetic Analysis	53
5.3.3 Geodynamic Features	57
6 Conclusions	63
Bibliography	84
Acknowledgements	85

Abstract.....	87
Kokkuvõte	89
Appendixes	91
Appendix I (Publication I)	91
Appendix II (Publication II)	121
Appendix III (Publication III)	161
Appendix IV (Manuscript I)	183
Appendix V (Manuscript II)	217
Appendix VI (Manuscript III)	299
Curriculum Vitae	314
Elulookirjeldus.....	319

List of Publications and Manuscripts

The present Ph.D. thesis is based on the following publications and Manuscripts, which are referred to in the text by Roman numerals.

- I J D Solano-Acosta, A Soesoo, and R Hints. New insights into the crustal structure across Estonia using satellite potential fields derived from WGM-2012 gravity data and EMAG2v3 magnetic data. *Tectonophysics*, 846:229656, 2023. doi:[10.1016/j.tecto.2022.229656](https://doi.org/10.1016/j.tecto.2022.229656)
- II J D Solano-Acosta, A Soesoo, and R Hints. Geochemistry, provenance, and tectonic setting of Paleoproterozoic metasedimentary and metavolcanic units of the Estonian Alutaguse region, eastern Fennoscandia. *Estonian Journal of Earth Sciences*, 74:61–82, 2025. doi:[10.3176/earth.2025.05](https://doi.org/10.3176/earth.2025.05)
- III J D Solano-Acosta, A Soesoo, and R Hints. Integrated Geophysical and Emplacement Modelling of the Märjamaa and Kloostri rapakivi Granitoids, Estonia: Insights into Intrusion Geometry and Tectonic Controls. *Precambrian Research*, 430:107938, 2025. doi:[10.1016/j.precamres.2025.107938](https://doi.org/10.1016/j.precamres.2025.107938)

Manuscripts

- I J D Solano-Acosta, A Soesoo, and R Hints. An updated review of major element geochemistry and petrogenesis of the oxidised Märjamaa and Kloostri Rapakivi Granites, Estonia. 2025. Manuscript Under review
- II J D Solano-Acosta, S Graul, A Soesoo, R Hints, T All, J Vind, L Maala, and S Liivamägi. Comparative geochemistry, provenance, and tectonic evolution of Paleoproterozoic metavolcanic and metasedimentary units in the Alutaguse Zone (North Estonia) and Southern Svecofennian domains (Russia–Finland): Implications for Eastern Fennoscandian crustal development. 2025. Manuscript Under review
- III J D Solano-Acosta, S Graul, A Soesoo, R Hints, T All, J Vind, and S Liivamägi. Exploration of critical elements in the concealed North Estonian basement and its Eastern Fennoscandian context. 2025. Manuscript Under review

Author's Contributions to the Publications and Manuscripts

- I In **Publications I–III & Manuscript I**, I was the lead author, responsible for conceptualising the study, processing and analysing the data, conducting the simulations, preparing the figures, and drafting the manuscripts. The co-authors contributed through supervision, guidance, and scientific discussion.
- II In **Manuscripts II–III**, I led the research, performed the data analysis, designed the figures, and wrote the manuscripts. The co-authors contributed in an advisory capacity and edition. These manuscripts involved participation from the Estonian Geological Survey (EGT) as part of a collaborative effort within the EU Horizon DEXPLORE project.

Abbreviations

AMCG	Anorthosite–Mangerite–Charnockite–Granite
AI	Hashimoto Alteration Index
Amph	Amphibolite
CCPI	Chemical Compositional Proximity Index
CIA	Chemical Index of Alteration
CIPW norm	Cross, Iddings, Pirrson, and Washington normative mineral calculation
CIW	Chemical Index of Weathering
CMB	Core–mantle boundary
CPD	Curie Point Depth
Crd	Cordierite
CRM	Critical Raw Materials
EEC	East European Craton
EGT	Estonian Geological Survey (Eesti Geoloogiateenistus)
EMAG2v3	Earth Magnetic Anomaly Grid (2-arc-minute Version 3)
Ga	<i>Giga-annum</i> , meaning billion years ago
Grt	Garnet
HF	Heat Flow
HFSE	High Field Strength Elements
ICV	Index of Compositional Variability
LOI	Loss on ignition
Moho	Mohorovičić Discontinuity
MTZ	Mantle transition zone
PAAS	Post-Archean Australian Shale
PIA	Plagioclase Index of Alteration
PPDZ	Paldiski–Pskov Deformation Zone
REE	Rare Earth Elements
RLSZ	Raahe–Ladoga Suture Zone
RTP	Reduced-to-pole (magnetic transformation)
SCLM	Subcontinental Lithospheric Mantle
SFSZ	South Finland Shear Zone
Sil	Sillimanite
SI	Sericitisation index
SO	Svecofennian Orogeny
SS	South Svecofennian
UC	Upper continuation
UCC	Upper Continental Crust
VMS	Volcanogenic Massive Sulphide
WIP	Parker weathering index
WGM-12	World Gravity Model 2012

1 Introduction

The present-day Baltica, also known as the East European Craton, occupied a central position within the Nuna supercontinent and comprises the Archean–Paleoproterozoic cratons of Fennoscandia, Volgo-Uralia, and Sarmatia in northeastern Europe. These cratonic blocks evolved independently prior to their amalgamation at ca. 1.8–1.7 Ga [7–12]. Fennoscandia, incorporates the central–southern crust representative of the Svecofennian Orogeny (SO; 1.92–1.80 Ga), composed of several NW–SE-trending, 100–300 km-wide tectonic megadomains bounded by major dextral shear zones that developed successively between 1.86 and 1.75 Ga during crustal accretion younging toward the southwest, producing igneous assemblages decreasing in age in the same direction [8, 13]. These structures, despite intervening microcontinents, record episodic rollback of a long-lived south-southwest-retreating subduction system, with Paleoproterozoic crustal growth continuing through several subduction collisional episodes [8, 14, 15]. A major tectonic reorganisation occurred at 1.82–1.80 Ga, when the oblique collision of Volgo-Sarmatia and Fennoscandia disrupted crustal growth [8, 10, 11, 16–18], halted northeastward accretion of the Bergslagen–Livonia microcontinent (1.87–1.75 Ga), and generated the transpressional architecture of southern Fennoscandia [13, 19]. In the southeast, this collision triggered a ~45° anticlockwise rotation of Sarmatia and crustal extension between 1.80 and 1.75 Ga [7, 8, 15], followed by widespread Nuna-related AMCG (Anorthosite–Mangerite–Charnokite–Granite) magmatism between, of which ~1.6–1.4 Ga rapakivi granites represent its major expression extending over central Fennoscandia [10, 11, 18, 20, 21]. The crystalline basement from Estonia to northern Poland forms the concealed southern continuation of the Fennoscandian crustal province [8, 17, 20–25].

Deciphering the structure and composition of the Estonian Precambrian basement enhances the understanding of Fennoscandian tectonic amalgamation and the subsequent emplacement of anorogenic magmatism. The Estonian crust is unusually thick and preserves Paleoproterozoic fabrics from the SO, linking northern Estonia to southern Finland and the Bergslagen microcontinent [8, 14, 26–30]. However, the extent to which these inherited structures influenced later tectono-magmatic events remains debated [8, 23, 31, 32], particularly in view of the widespread bimodal AMCG magmatism across Baltica [33–36]. Accordingly, this dissertation engages with the following overarching questions: (i) How do magnetic and gravity anomalies, together with crustal layers such as the Moho and Conrad discontinuities, constrain the physico-mechanical layering of the Estonian crust?; (ii) Does the NE portion of the Estonian basement (i.e. Tallinn–Alutaguse–Jõhvi corridor) represent a continuation of the Southern Svecofennian system or a distinct terrane? [8, 24, 28–30, 37–41]; and (iii) How did inherited SO fabrics control the emplacement of the Märjamaa–Kloostri rapakivi granitoids along the Åland–Paldiski–Pskov Deformation Zone (PPDZ) within Paleoproterozoic crustal structures, and what do their geochemical signatures reveal about Mesoproterozoic A-type magmatism? [11, 21, 33, 42–48]. To resolve these questions, potential-field modelling was integrated with tectonic and geochemical interpretations to constrain Svecofennian crustal architecture, evaluate the influence of inherited shear zones, and define the geodynamic conditions that governed different magmatic events from during the SO to the AMCG event related to the breakup of Nuna [10, 11, 47, 49–51]. Results also allow the identification of metalliferous signatures and target zones.

The Precambrian crystalline basement of Estonia, composed of Paleo- to Mesoproterozoic metamorphic and igneous rocks (1.9–1.5 Ga), is entirely buried beneath 100–900 m of Neoproterozoic–Devonian shallow-marine sandstones, mudstones, and carbonates that thicken southward [17, 25, 26, 39] (Figs. 1). With no basement outcrops exposed at the

surface, geological reconstructions rely on drill-core material together with geophysical and geochemical datasets [17, 23, 25, 38]. Quaternary glacial deposits further modify the topography, producing lake depressions and upland ridges [25].

Northern Estonia hosts Paleoproterozoic metavolcanic and metasedimentary rocks intruded by 1.92–1.88 Ga orogenic granites—correlating with units in southern Finland and Swedish Bergslagen [8, 28–30, 40, 41, 52, 53], whereas the southern Estonian–Latvian granulite belt comprises younger 1.84–1.77 Ga rocks [8, 17, 39], together forming two lithotectonic domains separated by the PPDZ [8, 24, 28, 38]. These domains, bounded by NW–SE Svecofennian shear zones (*Publications I & III*), are interpreted as late-Svecofennian accretionary belts formed during the 1.83–1.80 Ga collision between Fennoscandia and Volgo-Sarmatia (Fig. 2a–b) [8, 14, 16, 22]. Integration of geological and geophysical datasets further subdivides the basement into six structural–petrological zones (Fig. 2c–d): Tallinn, Alutaguse, Jõhvi, West-Estonian, Tapa, and South-Estonian—each characterised by distinct lithological and geochemical assemblages, metamorphic grades, and potential-field signatures [28, 41] (*Publication I*).

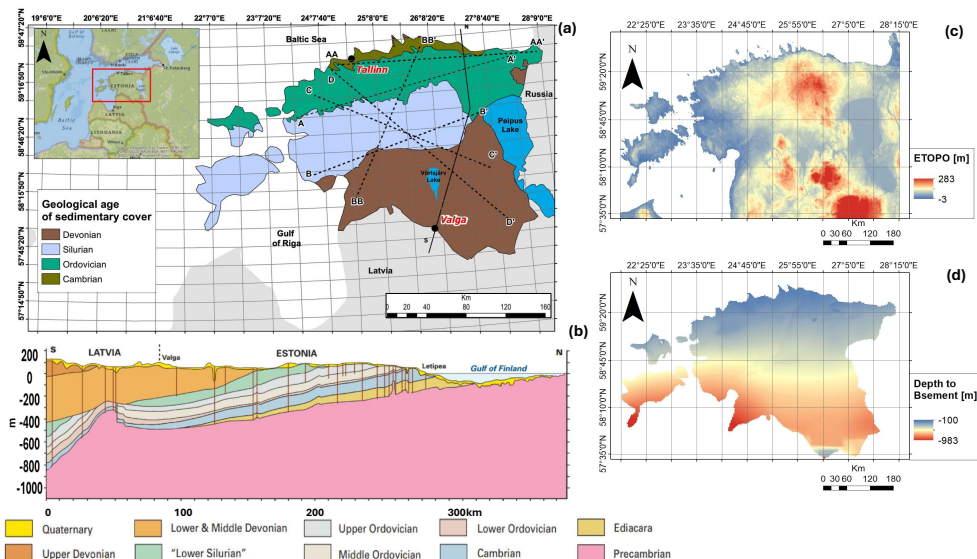


Figure 1: Geological framework of the Estonian Paleozoic units. (a) Generalised map of Estonia showing the sedimentary cover and outcrop areas of Lower and Middle Palaeozoic rocks (after Puura & Vaher, 1997 [54]). Dashed lines indicate WGM-12 and EMAG2v3 profiles used for regional potential field analysis (*Publication I*); the solid line marks the cross-section shown downwards.; (b) Geological cross-section through Estonia, modified from Bauert et al. (2015) [55], with vertical lines representing drill core locations.; (c) ETOPO1 global relief model integrating land topography and ocean bathymetry at 1 arc-minute resolution.; (d) Depth to the crystalline basement derived from drill core data, displayed as contour lines representing distance from the surface to the basement. Data sourced from the Estonian Geological Survey (EGT): <https://geoportaal.ee/eng/INSPIRE/INSPIRE-services/GE-Geology-p761.html>.

Positioned at the center of the Nuna (Columbia) supercontinent, Baltica experienced Paleoproterozoic crustal growth through the Sarmatia–Volgo Uralia collision around 2.20–2.05 Ga, followed by significant arc accretion in Fennoscandia during the SO [8, 14, 18, 56, 57]. This orogeny generated nearly 1×10^6 km² of new crust through episodic arc–continent amalgamation [51, 58]. The central and southern SO comprise NW–SE megadomains (100–

300 km), bounded by dextral shear zones that formed at 1.86–1.75 Ga during southwest younging accretion, with igneous ages decreasing toward this direction [8, 13]. These domains document episodic rollback of a long-lasting south–southwest-retreating subduction system and crustal growth via multiple subduction collisions [14, 37, 51, 53, 59, 60].

In Fennoscandia, the microcontinents Keitele and Bothnia, situated along the passive margins of the Archaean Karelia and Norrbotten cratons, have sparked new hypotheses about orocline tectonics in the Svecofennian terranes from around 1.87 Ga [16, 27, 52, 61]. Similar accretionary tectonism around 1.83–1.82 Ga has also been proposed for Bergslagen in south-central Sweden [8, 13]. The Bogdanova et al. (2015) [8] Trans-Baltic tectonic model (Fig. 2a), shows that the developments in the southern half of the SO at 1.87–1.75 Ga involved continuous and single-polarity north-eastward accretion of the microcontinent Bergslagen with Livonia, and presumably also large parts or even the whole of Amberland [8]. In southwestern Fennoscandia (Fig. 2b), three major arc systems developed: the Savo Belt (1.93–1.80 Ga), the Central–Western Finland Complex including the Central Finland Arc zone (CFAC) and Granitoid complexes (1.90–1.87 Ga), and the Southern Svecofennian (SS) domain (1.90–1.82 Ga), comprising the Häme and Uusimaa belts and the Saimaa and Ladoga zones [8, 10, 62]. The Häme and Uusimaa belts (1.90–1.88 Ga) extend to Saimaa and Ladoga interacting with the Archean crust [10, 51, 63], and their arcuate geometries reflect buckling along the Bothnian and Saimaa oroclines [10, 27, 51]. The Uusimaa belt continues beneath the Gulf of Finland, toward the Tallinn domain [8, 17, 37, 38, 53], adjacent to the northern border of the Bergslagen microcontinent (Fig. 2a) [8]. The Ladoga zone represents an eastern Svecofennian continuation along the Archean boundary [64–67].

Within this trans-Baltic framework, the ca. 1.92–1.88 Ga Tallinn, Alutaguse, and Jõhvi corridor host a variety of Fe–S–Si ore bodies [8, 28, 41, 68], constituting a portion of a Bergslagen-type metallogenic belt and the Uusimaa region of southern Finland [40, 69, 70]. This belt is associated with 1.91–1.89 Ga felsic volcanic and subvolcanic rocks, which are accompanied by semi-coeval turbiditic, carbonate, and skarn successions [8, 13, 40, 41]. Subsequent geological processes in northern Estonia included extensive migmatization and the intrusion of ~1.79 Ga granitoids [17, 23, 25, 39], linking this region to the Finnish late Svecofennian granite-migmatite (LSGM) [8, 52]. Within this magmatic–volcanic–sedimentary corridor, the Alutaguse domain, composed of aluminous and graphitic gneisses, meta-volcanic rocks, quartzites, and carbonates intruded by minor (post?) Svecofennian small granitoids and veinlets, is interpreted as a ca. 1.88 Ga rifted back-arc segment of the Uusimaa–Tallinn arc system (*Publication II & Manuscript II*). This back-arc basin was subsequently sealed and deformed during post-Svecofennian compression, promoting the development of polymetallic mineral systems [17, 68] (e.g., VMS-type Cu–Pb–Zn deposits) (*Manuscripts V & VI*), whereas the Jõhvi domain likely represents a high magnetic anomaly [71], interpreted as a part of the Bergslagen-type magmatic–volcanic–sedimentary skarnized member [8, 41], characterised by granulite-facies metamorphism and local base-metal sulphide mineralisation, interpreted as metamorphosed products of volcanosedimentary sequences and/or hydrothermal protoliths that underwent regional high-grade metamorphism [40, 41, 72].

The emplacement of Fennoscandian AMCG rapakivi intrusions spans 1.8–1.0 Ga, with a peak between 1.6 and 1.4 Ga during the breakup of Nuna (Columbia) [44, 57, 73–75] (Fig. 2a). These intrusions range in size from small plutons to batholiths covering up to 18,000 km² [11, 20, 21]. Seismic and potential-field data indicate ascent focused along horst–graben systems and listric faults [32, 43, 76–79], coincident with crustal thinning and mafic underplating during Subjotnian diabase emplacement (1.65–1.55 Ga) [35, 42, 43]. In the absence of subduction signatures [42, 80], proposed mechanisms include cauldron

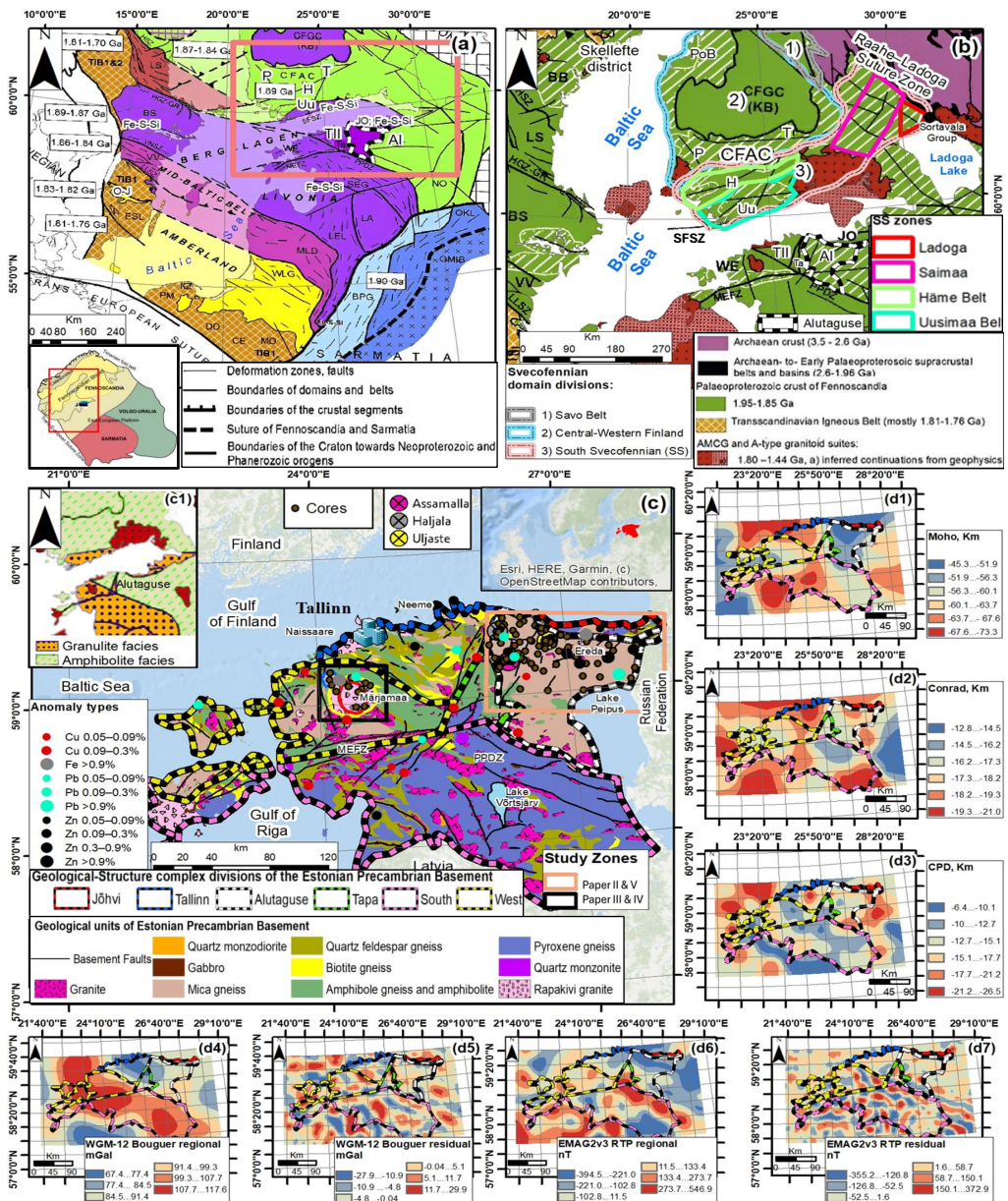


Figure 2: Geological setting of the study zone, highlighting key features: (a) Central and Southern

Svecofennian crustal structure crosses the Baltic Sea, encompassing Bergslagen, Livonia, Amberland, Keitele, and Bothnia zones, which may be considered microcontinents. The pinkish box enhances the study zone. Notice the box at the bottom left with Baltica, which enhances the position of the larger figure. The ages shown in the frames correspond to the main accretionary events within the specified zones. The dashed triangle lines represent the upper surfaces of dipping mantle reflectors [58]. The notation Fe-S-Si signals the existence of diverse iron and sulfide ore deposits.; **(b)** Major Palaeoproterozoic tectonic zones over the Fennoscandian area. The white diagonal ruling over the greenish area indicates Svecofennian sedimentary basins. This map depicts the South Svecofennian (SS) and Estonian Alutaguse zones. Figures a and b are modified from Bogdanova et al. 2015 [8], whose abbreviations are: AL – Alutaguse, BB – Bothnian, BS – Bergslagen, BPG – Belarus-Podlasie granulite belt, CE – Ciechanow, CFAC – Central Finland Arc zone; CFGC – Central Finland Granitoid zone (Keitele microcontinent). KB – Keitele microcontinent, JO – Jöhvi, LEL – Latvian-East Lithuanian, MLD – Mid-Lithuanian zone, NO – Novgorod, SEG – South Estonian granulite zone, Ta – Tapa, TIL – Tallinn, WE – West Estonian zone, WLG – West Lithuanian granulite zone. Deformation zones: HGZ-GR – Hagsta-Gävle-Rättvik Zone, HSZ – Hassela Shear Zone, MEFZ – Middle Estonian Fault Zone, PPDZ – Paldiski-Pskov Deformation Zone, SFSZ – South Finland Shear Zone, and for volcanic belts and sedimentary basins: Uu – Uusimaa, H – Häme, P – Pirkanmaa, T – Tampere, PoB – Pohjanmaa, O-J – Oskarshamn-Jönköping. Please refer to Bogdanova et al. 2015 [8] for further information and a comprehensive understanding.; **(c)** Geological map scheme of the Precambrian basement of Estonia, showing geochemical anomalies after Soesoo et al. (2020) [17], and (c1) an upper left-corner inset showing the distribution of granulite- and amphibolite facies metamorphic rocks. Reddish symbols correspond to Rapakivi lithologies. modified from Bogdanova et al. 2015 [8]. Assamallaa, Haljala, and Uljaste represent target exploration zones, and black dots indicate the investigated EGT cores. The geological data can be downloaded from the Geoportal of the Land Board of the Republic of Estonia (<https://geoportaal.ee/eng/INSPIRE/INSPIRE-services/GE-Geology-p761.html>).; **(d)** Geophysical maps of Estonian crust (Publication I) depicting: the (1) Moho discontinuity, (2) Conrad discontinuity, (3) Curie Point Depth (CPD), (4) WGM-12 regional Bouguer anomaly, (5) WGM-12 residual Bouguer anomaly, (6) EMAG2v3 regional magnetic anomaly, (7) EMAG2v3 residual magnetic anomaly.

subsidence [81–83], strike-slip faulting [84–86], and post-orogenic thermal relaxation [37].

In Baltica, rapakivi magmatism forms extensive AMCG suites generally linked to lithospheric extension, rifting, or mantle plumes [11, 44, 45, 50, 73, 87, 88]. A pure rift origin is questioned due to: coeval sedimentary basins are scarce [11, 89], magmatism persisted for 500 Myr [90], and thermal data do not require whole-lithosphere thinning [37]. Age-progressive hotspot tracks are also unsupported [11, 90]. Instead, paleomagnetic data place Baltica and Laurentia at low latitudes (ca. 1.65–1.45 Ga), favouring plume trapping, thermal insulation, and superswell-driven upwelling [11, 74, 89]. These conditions promoted mafic underplating, extension, and reactivation of Svecofennian structures, focusing AMCG emplacement [11, 42, 44, 45, 48, 50, 77, 80, 86] and generating hybrid SCLM (Subcontinental Lithospheric Mantle) melts [48, 50, 74].

Due to NE–SW Svecofennian compression established NW–SE to NNW–SSE shear corridors that were reactivated during Mesoproterozoic extension; structural deflections near rapakivi massifs and present shallow seismicity (Rg-wave propagation) point to (ongoing) oblique strike-slip [8, 11, 14, 18, 31, 32, 84, 91–93]. Within this oblique-extensional, corridor-focused regime, late Wiborg Suite intrusions (e.g., Märjamaa–Kloostri, Naissaare) formed sheet-like, tabular plutons along dilatational ramps (listric faults) imaged on Gulf of Bothnia and SW Finland profiles (Fig.2a-c) [32, 42, 43, 78–83, 91, 94].

In Estonia, rapakivi bodies are represented by the Riga Batholith and smaller intrusions—Märjamaa, Kloostri, Naissaare, Taebala, Neeme, and Ereda (Fig.2b-c) [33, 95]. The Märjamaa rapakivi granitoid and its Kloostri satellite, emplaced near the NW-trending Åland–PPDZ corridor, provide an ideal test for inherited structural control using integrated potential-field analysis and lineament mapping, and correlate the petrophysical responses with the geochemical signatures in the different phase-stages [33, 42, 43, 95, 96] (*Article III & Manuscript IV*).

This doctoral dissertation integrates geophysical modelling, structural and petrological analysis, and whole-rock geochemistry to refine interpretations of Precambrian basement architecture, tectonic evolution, and CRM potential in northern Estonia. It addresses four main themes: (1) lithospheric structure and crustal discontinuities of the Estonian basement constrained by gravity and magnetic datasets (*Publication I*); (2) assessment of sediment provenance and magma source characteristics in Alutaguse metasedimentary and metavolcanic rocks, constraining their relationships with the SS domains and the Bergslagen microcontinent and clarifying the geodynamic role of Alutaguse within the SO (*Publication II & Manuscript II*); (3) reconstruction of the petrogenetic evolution and emplacement of the Mesoproterozoic Märjamaa and Kloostri granitoids, linking pluton geometry to inherited Svecofennian fabrics and PPDZ reactivation under a transtensional regime (*Publication III & Manuscript I*); and (4) Identification of geochemical and geophysical anomalies that delineate CRM-bearing metalliferous exploration targets and prospective zones within the North Estonian basement (*Manuscript III*).

2 Objectives of the Dissertation

This dissertation investigates the crustal architecture, tectonic evolution, and magmatic processes of the Estonian Precambrian basement through an integrated geophysical, geochemical, and petrological approach. The primary research objectives are as follows:

- **Characterise lithospheric structure and deep crustal domains:** Model the Moho, Conrad, and Curie Point Depth (CPD) surfaces using global Bouguer (WGM-12) and magnetic (EMAG2v3) datasets to delineate geophysically coherent crustal blocks, identify aligned structural trends, and assess patterns related to distinct geodynamic evolutionary events (e.g., Svecofennian orogeny).
- **Reconstruct the geochemical and tectonic history of the Alutaguse zone:** Conduct major and trace element analyses of all metasedimentary and metavolcanic rock to assess lithological variability, source provenance, chemical maturity and tectonic setting, with an emphasis on the characteristics of the back-arc basin and correlations with the southern Svecofennian domains (Uusimaa, Häme, Saimaa, and Ladoga).
- **Establish a regional metallogenic and tectonic framework for the Alutaguse and Jõhvi domains:** Integrate and analyses the EGT-national Bouguer and magnetic datasets using the Python SimPEG-based inversion, along with geochemical and structural data, to identify crust-mantle interactions, compositional trends and metalliferous anomalies. The aim is to refine Svecofennian tectonic models and support further exploration of ore deposit targets.
- **Investigate the magmatic evolution and emplacement of rapakivi Märjamaa and Kloostri granitoids:** Integrate EGT-national Bouguer and magnetic datasets for derivative analysis and SimPEG-based inversion, combined with structural interpretation and petrological modelling, to constrain crystallisation conditions, ascent mechanisms, and chamber geometries within the AMCG Mesoproterozoic event and inherited PPDZ shear corridor arrangement.

3 Geological setting and Literature Background

3.1 Tectonic Setting

Following the Sarmatia–Volgo–Uralia amalgamation at ~ 2.1 Ga [57,97], Fennoscandia underwent protracted crustal accretion, arc magmatism, and basin closure between ~ 2.1 to 1.79 Ga, during the SO. Involving the Bergslagen and Livonia microcontinents, this evolution culminated in continent–continent collision during the Svecobaltic phase (1.84–1.79 Ga) [8,14,52,61,75,98–100] (Fig.2a).

Between 2.10–1.90 Ga, a pre-Svecofennian ocean separated proto-Fennoscandia from the Bergslagen block. By around 1.90 Ga, the Tallinn–Uusimaa arc system joined the Häme and Central Finland Arc Complex (CFAC), leading to crustal thickening and widespread magmatism in southern Finland, central Sweden, and northern Estonia. (Figs.2a,b) [8,10,24,37,39,58,61,101,102]. Crustal growth proceeded via alternating transpressional cycles along a southward-retreating active margin [8,31,102,103], consistent with tectonic switching [37,51,102,104,105]. The Skellefte, Tampere, Häme, and Uusimaa arcs have been interpreted as either discrete belts produced by subduction retreat or double-plunging zones [10,37,53,62], or as segments of a single curvilinear arc [51], all formed under a dominant N–NE stress regime [8,16,31,37,92,93,101].

Oblique accretion of southern terranes (Belarus–Baltic, Vitebsk, Osnitsk–Mikashevichi; 1.92–1.79 Ga) [8,106] led to amalgamation of Fennoscandia with Volgo–Sarmatia by ~ 1.77 Ga [8,14,75]. The Svecobaltic Orogeny (1.84–1.79 Ga) marked the final major crustal thickening before Gothian accretion (1.73–1.48 Ga) added westward-younging belts to Baltica's margin [7,8,10]. Paleogeographic reconstructions indicate assembly within Nuna (Columbia) (1.90–1.60 Ga) and subsequent long-distance drift, with rates up to 14 cm/yr during the SO, 6.5 cm/yr during Gothian, and 9 cm/yr during the Danopolonian orogeny [16,18,107–109].

The SS domain preserves the detailed record of accretionary and collisional events [10,27,51,58]. These domains later accreted against northern CFAC belts (i.e., Pirkanmaa ~ 1.90 Ga; Tampere ~ 1.92 Ga; Figs.2a,b) within a double-subduction system, culminating in closure of the Paleo–Svecofennian ocean at ~ 1.87 Ga [8,37,52,53]. The Häme belt (1.88–1.87 Ga) comprises amphibolite-facies metavolcanic – metasedimentary successions intruded by granites, migmatites, and pegmatites, with older arc-type units and younger rift-affinity basalts [37,53,62,101,102]. The Uusimaa belt (2.10–1.91 Ga) shows a dual rift–collision signature, hosting juvenile arc rocks at 1.85–1.87 Ga but retaining older isotopic signals from Paleoproterozoic crust [37,53,110–112]. Both belts contain 1.90–1.88 Ga metavolcanics and calcareous metasediments, especially in western Uusimaa [62,102]; whether they form distinct terranes or parts of a shared arc remains debated [8,37,53,102]. Eastward, metapelites, metagreywackes, and metavolcanics range from andalusite-bearing schists to diatexites. The South Finland Shear Zone (SFSZ; Fig.2b) is a major boundary marked by sillimanite-andalusite-grade metamorphism. Orijärvi records intense pre-metamorphic alteration, while West Uusimaa preserves granulite-facies assemblages formed at $\sim 800^\circ\text{C}$ and ~ 6 kbar, with three metamorphic events between 1.91–1.80 Ga [10,113].

Along the Raahe–Ladoga Suture Zone (RLSZ; Figs.2a–b), the Saimaa zone comprises mica schists, black shales, and diatexitic granulites, with zircon ages indicating a 2.4–2.2 Ga depositional gap [51]. A 1.92 Ga collision was followed by at least five deformation events between 1.9–1.8 Ga, and 1.91 Ga marine signals (metabasalt–shale) correlate with Bothnian oroclines and conductivity anomalies [63,114].

In the NW Ladoga region (SE Finland–Russian Karelia), the Archean–Proterozoic con-

tact follows the RLSZ (Fig.2b). Archean granite–gneiss domes are juxtaposed with Paleoproterozoic Kalevian sediments within a 50–150 km-wide zone. Ladoga granitoids show low ϵNd values indicating Archean input [64, 66, 67]. The complex includes metamorphosed Sortavala Group volcanics (2.06–1.98 Ga) and Kalevian siliciclastics (1.97–1.89 Ga), interpreted in a subduction setting. A NW-trending shear zone partitions two blocks; Sm–Nd data indicate 2.7–2.5 Ga sources, and detrital zircons span Proterozoic (2.0–1.9 Ga) and Archean (2.74–2.54, 3.01–2.90 Ga) populations [65]. South of the RLSZ, NE–SW shears and SE–NW folds accompany granitoids akin to the Central Finland complex; meta-sediments reflect erosion from Archean gneisses and Sortavala volcanics with Svecofennian assimilation [64, 67]. Western Ladoga volcanogenic units align with southern Savo and Tampere belts [67].

3.2 Estonian Domains

Building on the regional framework, the Estonian cover succession dips gently southward ($\sim 0.10\text{--}0.20^\circ$; 2–3.5 m/km) and thickens accordingly (Fig.1b,d). The basement structure is constrained primarily by drill cores, potential-field anomalies, and correlations with better-exposed Finnish and Swedish terranes [8, 24, 25]. Six structural–petrological zones (i.e., Tallinn, Alutaguse, Jõhvi, West-Estonian, Tapa, and South-Estonian) are distinguished by lithological, geochemical, magnetic, and gravity contrasts (Fig.2c) [8, 23, 28]. The PPDZ, a NW–SE dextral shear zone ~ 30 km wide, separates northern–eastern blocks from western–southern domains, while the E–W Middle Estonian Fault Zone (MEFZ) was reactivated during Mesoproterozoic extension, juxtaposing granulite- and amphibolite-facies blocks [24, 39].

NW-trending gravity and magnetic lineaments across the Baltic Sea suggest continuity of oblique Svecofennian crustal blocks into Estonia [10, 92, 115] (*Publication I*). The NW–SE lithological grain reflects inherited shear zones—PPDZ and its Finnish continuation (Sotunga–Jurmo Shear Zone, SJSZ)—linking further to the Hagsta–Gävle–Rättvik Zone (HGZ–GRZ) in Swedish Bergslagen area [8, 11, 14, 16, 19, 31, 86] (Fig.2a). These structures localize present-day seismicity, dominated by strike-slip mechanisms, along reactivated basement faults [31, 93].

In northeastern Estonia, the Tallinn and Alutaguse domains record amphibolite-facies metamorphism (ca. 3–5 kbar) and show strong lithological, geochemical, and structural correlations with the Uusimaa belt in southern Finland and the Swedish Bergslagen region [8, 28]. The Tallinn domain is interpreted as the northeastern continuation of the Uusimaa volcanic arc, whereas Alutaguse represents its genetically related back-arc basin [23, 37, 39] (*Publication II & Manuscript II*). Within the northeastern portion of the Estonian basement, the Tallinn Zone comprises a heterogeneous assemblage ranging from mafic amphibolite-facies metavolcanic rocks to metasediments, including amphibole gneisses, biotite–plagioclase gneisses, quartz–feldspar gneisses, mica gneisses, and subordinate sulphide–graphite gneisses and magnetite quartzites. This zone is interpreted as an arc-related volcanic–sedimentary domain developed along the Svecofennian active margin [25, 28, 68, 116]. The Alutaguse Zone is dominated by alumina-rich gneisses bearing biotite, cordierite, garnet, and sillimanite, together with biotite–plagioclase gneisses, sulphide–graphite gneisses (black schists), quartzites, carbonaceous and skarn-bearing rocks, and subordinate amphibole gneisses, amphibolites, and pyroxene gneisses. It is interpreted as a rifted back-arc domain characterized by extension, hydrothermal activity, and immature sedimentation adjacent to the Uusimaa–Tallinn arc system [17, 24]. The Jõhvi zone consists mainly of pyroxene gneisses, quartz–feldspar gneisses, biotite–plagioclase gneisses, amphibole gneisses, and garnet–cordierite gneisses, together with magnetite

quartzites responsible for the pronounced Jõhvi magnetic anomaly. Widespread migmatization and the development of charnockitic and enderbitic leucosomes indicate a high-grade marginal domain representing a deeply buried and thermally overprinted segment of the Svecofennian system [28, 40, 41, 117].

The Tapa Zone is characterized by alternating Fe- and Si-rich garnet–pyroxene quartzites, high-Al garnet–cordierite–sillimanite gneisses, and Ca-rich to Ca-poor pyroxene–amphibole and biotite-bearing gneisses, recording amphibolite- to granulite-facies metamorphism consistent with an inner-orogenic or lower-crustal setting during peak Svecofennian convergence [28]. The West Estonian Zone, bounded by NW-trending and E–W-trending tectonic zones, is dominated by amphibolite- to granulite-facies metasedimentary rocks comparable to those of southern Bergslagen, including amphibolites, biotite–plagioclase gneisses, quartz–feldspar gneisses, and minor pyroxene gneisses, and is interpreted as part of a high-grade Bergslagen-facing margin within the SO [8, 25, 28, 118]. In contrast, the South Estonian Zone consists predominantly of meta-igneous rocks with minor metasedimentary components and is marked by strong, lineated gravity and magnetic anomalies. Its assemblage of mafic and felsic granulites, including amphibole–pyroxene and biotite–hypersthene gneisses as well as quartz–feldspar gneisses, reflects a deeply exhumed lower-crustal domain associated with late Svecofennian collision and subsequent tectonothermal reworking [8, 25, 28, 118].

U–Pb zircon data divide the Estonian basement into three principal age groups. The oldest (1.92–1.80 Ga) includes North Estonian metavolcanics (1.92 ± 10 Ga) [8, 38, 39, 119], followed by southern granulites (1.83 ± 22 and 1.82 ± 7 Ga) [39, 119], Tapa tonalites (1.82 ± 26 Ga) [117], and Jõhvi belt magnetite gneisses yielding 1.87 ± 18 , 1.83 ± 10 , and 1.79 ± 19 Ga [8, 28, 40]. The second corresponds to tonalites in southern Estonia that date to 1.79 ± 16 Ga, charnockites and garnet gneisses to 1.76 ± 11 and 1.78 ± 2 Ga [39, 117] and a gabbro–norite dyke to 1.77 ± 20 Ga [117]. The third group (1.64–1.58 Ga) comprises A-type granitoids (Abja 1.64 ± 7 Ga [120]; Virtsu 1.61 ± 17 Ga [34]) and rapakivi intrusions at Märjamaa (1.63 ± 7 Ga), Naissaare (1.62 ± 7 Ga), and Riga (1.58 ± 2 Ga) [36].

3.3 Mesoproterozoic AMCG Rapakivi Magmatism

As Baltica underwent successive accretionary and crustal-growth processes during the Gothian Orogeny (1.73–1.48 Ga), the stabilized interior of the craton experienced a contrasting magmatic regime characterized by extensive bimodal igneous activity [10, 16, 18]. This intracratonic phase is recorded by the emplacement of rapakivi granites, gabbro–anorthosites, and associated dyke swarms. The 1.60–1.40 Ga Fennoscandian AMCG suite was emplaced within the Svecofennian Province and the Transscandinavian Igneous Belt (Fig. 2a) [7, 8, 15, 52]. These intrusions exhibit pronounced bimodal magmatism with gradational textures from gabbro–anorthosite to granite, accompanied by diabase swarms [44, 45, 121], and cluster into four principal age suites: 1.67–1.62, 1.59–1.56, 1.55–1.53, and 1.53–1.44 Ga, represented by the Wiborg, Åland, Salmi, and Ragunda suites. Together, they extend from Central Sweden to southern Finland, beneath the Gulf of Finland, and into Estonia, Lithuania, and Poland, reaching as far as the Danish island of Bornholm [11, 20, 21, 36, 42, 45].

In the Baltic countries, rapakivi intrusions are largely buried beneath Phanerozoic cover; their distribution and architecture are inferred from seismic and potential-field data [8, 20, 21]. In Estonia, the rapakivi suite includes the large Riga Batholith, as part of the Åland suite, and smaller–older Wiborg suite intrusions such as Märjamaa, Kloostri, Naissaare, Taebala, Neeme, and Ereda, with A-type stocks (Fig. 2c) [33, 36, 122]. Emplacement occurred through crustal-scale structures such as the PPDZ and MEFZ, reactivated under transten-

sional regimes and plume-related rifting during AMCG magmatism [8, 25, 42, 43, 74, 80]. Although fully buried, geometry and internal zonation are constrained by Bouguer and magnetic anomalies [17] (*Publications I & III*).

The Märajamaa–Kloostri plutons typify this magmatism. Märajamaa is a 40 × 25 km composite intrusion with three internal phases: xenolith-rich melanocratic granodiorites (Phase I); biotite–hornblende granites (Phase II); and a leucocratic granite with trachytic textures (Phase III; Kloostri satellite) [95, 122, 123]. Petrographically, two generations of anhedral quartz (20–30%) occur—one enclosed in K-feldspar/microcline, the other in the groundmass. Microcline (20–40%) forms perthitic phenocrysts (2–3 cm) with quartz, biotite, and rare titanite inclusions; plagioclase (20–40%) is euhedral oligoclase–andesine. Biotite (2–10%) is anhedral; hornblende occurs in Phases I–II and muscovite in Phase III. Accessory phases include magnetite, ilmenite (3–5%), apatite, fluorite, zircon, titanite, and epidote [25, 122].

In summary, Estonia records Paleoproterozoic arc–back-arc construction during the SO, followed by post-orogenic reactivation and Mesoproterozoic AMCG magmatism. The Tallinn–Alutaguse domains capture early Svecofennian crustal assembly, whereas the Märajamaa–Kloostri intrusions reflect later transtensional emplacement along the PPDZ during Nuna–Columbia evolution.

4 Methodology

4.1 Datasets

4.1.1 Geophysical Datasets

Bouguer (gravity) and magnetic potential datasets were used to model the Estonian crustal architecture, basement depth, and major lithospheric boundaries (*Publications I & III*). The WGM-12 Bouguer gravity grid (Fig.2d4, d5), based on satellite and terrestrial data [124, 125], and the EMAG2v3 magnetic grid (Fig.2d6, d7), compiled from airborne, marine, and satellite sources [126, 127], were used to detect large-scale density and magnetic contrasts *Publication I*. Elevation data from the ETOPO1 model (Fig.1c) .

At a local scale, the Alutaguse and Jõhvi domains were analysed using EGT Bouguer and (aero)magnetic datasets (Fig.3). Also, EGT potential models were used over a 45×40 km² grid defined over the Märjamaa–Kloostri rapakivi granites (Figs.4a,b; *Publication III*). The raw potential-field datasets are accessible via the EGT web platform [EGT web platform](#) ([hyperlink](#)).

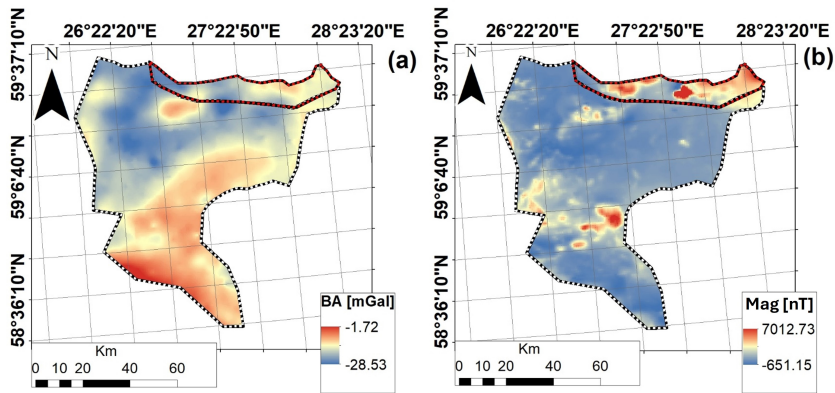


Figure 3: Alutaguse and Jõhvi EGT geophysical datasets including the (a) complete Bouguer anomaly (BA) and (b) magnetic anomaly, both provided by the Estonian Geological Survey (EGT).

4.1.2 Geochemical Datasets

4.1.2.1 Alutaguse and Southern Svecofennian (SS) Domains

This study integrates three complementary geochemical datasets from the Alutaguse zone to evaluate the compositional variability, provenance, and tectonic setting of Paleoproterozoic metasedimentary and metavolcanic rocks.

The first dataset consists of 229 whole-rock major element analyses compiled by Kivisilla et al. (1999) [116], originally obtained using classical wet-chemistry techniques from deep drill cores across the Alutaguse region. These analyses report major oxides in weight percent (wt.%), including SiO₂, TiO₂, Al₂O₃, Fe₂O_{3tot}, MnO, MgO, CaO, Na₂O, K₂O, P₂O₅, SO₃, and loss on ignition (LOI). Lithological classification was based on core descriptions by the Estonian Geological Survey (EGT), accessible via the public EGT dashboard [EGT dashboard](#). The full dataset is archived in the [SARV Geological Information System](#).

The second dataset comprises sixteen newly analysed samples—thirteen metasedimentary and three metavolcanic—collected for this research (*Publication II*). These samples were selected based on EGT drill-core logs and analysed for major elements using

X-ray fluorescence (XRF) and inductively coupled plasma–optical emission spectrometry (ICP-OES), and for trace and rare earth elements using inductively coupled plasma–mass spectrometry (ICP-MS). Sample preparation involved fine grinding and fusion with lithium tetraborate. These new analyses provide higher-resolution constraints on the geochemical variability, source characteristics, and depositional settings of the Alutaguse units.

The third dataset comprises recently released trace element analyses from EGT drill cores in the Alutaguse region, primarily from the Uljaste metalliferous target zone (Fig. 2c), available via the [eMaapõu geological portal](#). These samples were analysed predominantly for trace elements, with REE data limited to La and Ce. Major element oxides were not reported in this subset. Lithological classification follows both the core descriptions of Kivisilla et al. (1999) [116] and EGT drill-core documentation (Fig. 5), accessible through the [EGT “Puuraukude rakendus” web platform](#).

Combined, the three Alutaguse datasets comprise a total of 149 samples, with varying coverage of major, trace, and REE elements. Figures 2c and 5 illustrate the spatial distribution of drill cores and sampling intervals included in the Alutaguse compilation. Geochemical classification and petrogenetic modelling were applied to assess lithological variability, provenance, and depositional environments (*Publication II & Manuscript II*). Major, trace, and REE patterns were used to define the geochemical signatures of metavolcanic and metasedimentary rocks within the Alutaguse zone. The organised Alutaguse dataset used here is available at <http://dx.doi.org/10.3176/earth.2025.S05>

For comparison, 187 metasedimentary and metavolcanic samples from the Finnish SS zone— including the Uusimaa, Häme, and Saimaa domains (Fig. 2b) — were extracted from the national GTK geochemical database, as compiled and described by Rasilainen et al. (2007) [128] (*Manuscript II*). These samples, classified as “Schist, gneiss, amphibolite, and skarn rock” were filtered from the broader “sedimentary” and “volcanic” categories within the GTK dataset. In addition, 26 metasedimentary samples from the Ladoga domain were compiled from Kotova et al. (2009) [64], and 8 metavolcanic samples from Kotova and Podkovyrov (2014) [67] (*Manuscript II*).

Table 1 presents the minimum, maximum, and average major-element concentrations (wt.%), while Table 2 summarises the corresponding trace-element and REE concentrations (ppm) for both domains (*Manuscript II*).

		Metasedimentary Units																													
Group	Classification samples	# Samples	SiO ₂	TiO ₂	Al ₂ O ₃	Fe ₂ O ₃	MnO	MgO	CaO	Na ₂ O	K ₂ O	P ₂ O ₅	SO ₃	LOI	K ₂ O/Na ₂ O	Al ₂ O ₃ /TiO ₂	SiO ₂ /Al ₂ O ₃	Na ₂ O/K ₂ O	CaO/Al ₂ O ₃	CIA	CN	ICV	PIA	CIW	CIW/CIA						
Alutaguse High-SiO ₂ metasedimentary samples	Biotite gneisses	6	72.26, 64.87-75.99	0.46, 0.35-0.69	13.02, 10.35-16.07	4.54, 3.37-6.46	0.06, 0.02-0.19	18.0, 1.53-6.46	2.41, 1.50-4.78	2.30, 1.50-4.78	2.82, 0.06-0.19	0.06, 0.01-0.14	0.26, 0.03-0.68	1.03, 0.48-1.45	1.31, 0.55-2.47	1.31, 15.09-40.38	29.87, 15.09-40.38	5.72, 4.04-7.20	1.02, 0.41-1.82	0.18, 0.12-0.25	0.05, 0.05-0.09	0.07, 0.05-0.11	0.12, 0.09-0.16	57.45, 50.87-61.70	64.26, 60.82-68.40	1.16, 1.06-1.25					
	Biotite gneisses ± Grt ± Crd ± Sil	12	70.03, 63.56-76.09	0.52, 0.36-0.83	13.02, 10.36-15.97	6.29, 3.54-14.42	0.13, 0.01-0.13	6.29, 1.10-4.00	1.81, 1.07-4.19	2.03, 0.76-3.84	3.47, 1.56-5.86	0.09, 0.01-0.19	0.14, 0.05-1.97	0.40, 0.19-2.95	1.09, 0.41-6.87	1.31, 12.61-35.50	26.85, 10.00-7.32	5.50, 4.00-7.32	0.73, 0.15-2.46	0.17, 0.10-0.27	0.12, 0.02-0.10	0.07, 0.02-0.10	1.25, 1.01-2.04	68.00, 51.52-71.30	1.20, 55.05-80.72	1.14, 1.07-1.44					
	Garnet bearing mica gneisses ± Crd ± Sil	23	69.71, 64.87-76.89	0.56, 0.21-1.38	14.26, 10.21-18.59	5.86, 3.56-9.36	0.06, 0.01-0.12	1.80, 0.98-3.74	2.41, 0.85-4.59	2.30, 0.61-3.38	2.82, 1.01-4.40	0.06, 0.01-0.19	0.26, 0.02-0.17	1.03, 0.03-0.93	1.31, 0.70-2.04	1.31, 13.54-35.2	26.85, 9.90-68.39	5.50, 3.49-7.53	0.73, 0.28-2.85	0.17, 0.06-0.43	0.12, 0.10-0.27	0.07, 0.02-0.11	1.25, 1.01-2.04	68.00, 51.52-71.30	1.20, 55.05-80.72	1.14, 1.07-1.44					
	Cordierite bearing mica gneisses ± Grt ± Sil	16	69.37, 63.25-80.20	0.51, 0.21-0.92	14.34, 10.28-16.52	5.89, 2.23-10.13	0.06, 0.02-0.12	2.00, 0.74-5.10	1.90, 1.17-2.73	2.02, 0.80-2.73	3.53, 1.60-5.37	0.07, 0.01-0.22	0.31, 0.13-0.67	1.78, 0.66-2.84	1.84, 0.89-3.75	1.78, 13.96-60.29	1.84, 3.49-7.53	4.93, 3.83-7.80	0.60, 0.10-0.43	0.13, 0.08-0.17	0.07, 0.03-0.09	0.11, 0.07-0.14	63.33, 53.09-79.41	70.27, 63.81-82.93	1.18, 1.09-1.33						
	Garnet bearing mica gneisses ± Sil	5	68.92, 63.71-74.09	0.50, 0.07-0.81	13.70, 10.28-16.52	6.83, 4.73-9.92	0.12, 0.02-0.32	1.82, 1.00-3.54	1.92, 1.17-2.73	2.66, 1.69-3.05	1.45, 1.44-4.62	0.13, 0.05-0.21	0.27, 0.10-0.43	1.31, 0.52-2.29	1.31, 1.05-2.06	1.31, 13.96-60.29	44.15, 15.05-189.71	5.13, 3.49-7.53	0.95, 0.27-0.94	0.75, 0.41-1.33	0.19, 0.09-0.33	0.08, 0.05-0.19	66.71, 59.77-80.88	66.71, 57.77-73.68	1.17, 1.05-1.23						
	Graphite bearing mica gneisses ± Grt ± Crd ± Sil	14	68.96, 63.96-73.47	0.57, 0.05-0.86	14.36, 11.52-17.57	6.09, 4.02-10.98	0.09, 0.01-0.47	2.69, 1.10-1.86	1.88, 1.00-1.86	1.57, 0.20-3.74	3.77, 1.49-8.19	0.19, 0.02-1.20	0.45, 0.17-3.18	1.71, 0.65-5.86	1.38, 0.48-3.88	1.71, 17.41-270.40	5.05, 3.44-6.38	4.83, 3.64-6.38	0.57, 0.10-1.26	0.14, 0.01-0.25	0.17, 0.02-0.25	0.05, 0.02-0.08	68.48, 51.36-86.38	74.97, 63.41-91.42	1.21, 1.09-1.44						
	Biotite-plagioclase gneisses	5	67.07, 66.14-69.47	0.52, 0.44-0.77	15.13, 13.54-15.59	4.88, 4.59-5.37	0.07, 0.06-0.08	1.39, 1.10-1.86	4.23, 1.67-5.46	2.75, 1.80-3.29	3.49, 2.53-4.93	0.19, 0.17-0.22	0.27, 0.17-0.22	1.18, 0.72-1.74	1.39, 0.77-2.74	30.57, 17.51-35.07	4.45, 4.26-5.13	0.86, 0.36-1.30	0.28, 0.12-0.35	0.86, 0.36-1.30	0.28, 0.12-0.35	0.86, 0.36-1.30	54.13, 53.11-55.79	55.83, 53.11-55.79	62.91, 62.91-69.56	1.16, 1.10-1.27					
	Graphite bearing mica gneisses	3	66.85, 63.68-69.22	0.61, 0.16-1.00	13.71, 11.87-17.19	4.77, 5.37-9.90	0.04, 0.03-0.05	1.80, 0.85-2.48	2.09, 1.97-2.35	1.73, 1.43-2.40	3.54, 3.24-4.02	0.11, 0.10-0.12	2.04, 0.45-4.61	2.04, 1.25-2.81	2.04, 1.46-2.82	2.04, 12.02-73.88	3.76, 3.70-5.83	5.05, 3.35-6.69	0.50, 0.35-0.69	0.16, 0.10-0.20	0.19, 0.07-0.07	0.06, 0.05-0.07	63.50, 61.59-64.80	71.10, 70.04-71.67	1.21, 1.15-1.26						
	Biotite-plagioclase gneisses ± Amph	6	65.97, 63.59-72.43	0.53, 0.21-0.64	14.20, 14.20-15.24	5.82, 2.82-7.15	0.04-0.09	1.19-3.80	3.87-5.46	2.17-2.95	1.51-3.15	0.12-0.21	0.23-0.58	0.72-1.77	0.51-1.30	23.23-67.05	4.23-5.10	0.77-1.95	0.26-0.37	0.55-61.69	0.07-0.10	0.91-1.39	56.37-63.91	59.36-67.09	1.07-1.15						
	Biotite-Amphibole gneisses	1	65.61, 69.05, 63.25-80.20	0.67, 0.53, 0.05-1.38	14.09, 14.07, 10.21-18.59	5.62, 4.00, 1.02-14.42	0.03, 0.01-0.47	3.44, 0.61-5.10	4.98, 2.50, 0.20-5.76	1.78, 2.36, 0.55-3.84	2.45, 3.06, 1.01-8.19	0.20, 0.10-0.120	0.23, 0.03-0.41	1.53, 0.52-3.58	1.38, 0.35-9.81	21.00, 34.66, 9.90-270.40	4.66, 5.03, 3.49-7.80	0.73, 0.80, 0.10-2.85	0.35, 0.18, 0.01-0.43	0.32, 0.18, 0.05-0.41	0.35, 0.18, 0.05-0.41	0.32, 0.18, 0.05-0.41	62.36, 59.01, 49.57-75.85	63.26, 59.01, 49.57-75.85	70.65, 68.70, 55.05-91.42	1.13, 1.17, 1.05-1.44					
Biotite-plagioclase gneisses ± Amph	1	60.80, 60.80	0.63, 0.63	16.32, 16.32	7.02, 7.02	0.07, 0.07	3.17, 3.17	2.98, 2.98	2.24, 2.24	6.10, 6.10	0.42, 0.42	0.25, 0.25	1.66, 1.66	2.73, 2.73	25.89, 25.89	3.73, 3.73	0.37, 0.37	0.18, 0.18	53.89, 53.89	0.07, 0.07	1.36, 1.36	56.90, 56.90	68.92, 68.92	1.28, 1.28							
Alutaguse Low-SiO ₂ metasedimentary samples	Garnet bearing mica gneisses	5	59.97, 57.78-62.01	0.70, 0.50-0.94	15.03, 13.79-16.91	11.12, 9.48-14.01	0.28, 0.07-0.56	3.58, 2.54-5.36	2.74, 0.94-5.12	3.78, 0.36-3.21	0.78, 1.72-5.66	0.13, 0.05-0.28	1.32, 0.23-3.88	1.56, 0.43-2.77	5.37, 1.02-13.75	23.00, 16.11-29.54	4.01, 3.47-4.19	0.49, 0.07-0.98	0.18, 0.07-0.34	0.49, 0.07-0.34	0.18, 0.07-0.34	64.95, 55.23-77.80	0.04, 0.01-0.07	1.56, 1.41-1.77	73.63, 57.76-90.84	78.95, 65.63-92.58	1.22, 1.09-1.39				
	Cordierite bearing mica gneisses ± Grt ± Sil	6	59.64, 54.67-62.74	0.92, 0.70-1.78	17.50, 15.97-20.29	9.13, 7.61-22.39	0.08, 0.04-0.14	2.35, 1.10-3.09	2.35, 1.00-3.33	2.09, 0.61-3.43	3.18, 0.82-6.16	0.07, 0.05-0.11	0.30, 0.05-0.68	1.91, 0.87-4.20	1.91, 0.87-4.20	1.91, 9.54-29.04	2.19, 2.88-3.91	3.36, 0.24-1.16	0.69, 0.05-0.19	0.12, 0.05-0.19	0.07, 0.02-0.10	67.62, 55.28-85.47	0.06, 0.02-0.11	1.24, 1.03-1.65	76.37, 56.99-88.91	1.14, 61.36-89.43	1.10-1.27				
	Biotite gneisses	2	58.85, 56.02-61.69	1.13, 0.76-1.51	17.02, 16.91-17.13	9.33, 8.25-10.40	0.09, 0.08-0.09	3.65, 3.02-4.27	3.65, 2.68-6.18	1.29, 0.84-1.73	2.64, 2.52-3.12	0.30, 0.28-0.32	1.82, 0.53-1.50	1.82, 1.17-3.86	1.82, 11.38-22.32	2.51, 3.27-3.65	16.85, 3.22-3.65	0.56, 0.26-0.85	0.32, 0.28-0.36	0.70, 0.68-7.31	0.32, 0.03-0.06	1.38, 1.21-1.55	77.60, 72.15-83.05	80.45, 74.85-86.05	1.14, 1.10-1.18						
	Garnet bearing mica gneisses ± Crd ± Sil	14	58.08, 40.30-62.13	1.02, 0.70-2.88	17.80, 12.56-21.43	10.43, 5.56-19.60	0.13, 0.02-0.25	4.03, 2.53-8.56	4.03, 1.06-5.98	2.74, 0.51-3.04	3.78, 1.61-6.60	0.13, 0.02-0.43	1.32, 0.47-2.83	1.32, 0.80-5.66	1.32, 6.73-23.71	1.32, 2.08-4.81	19.89, 6.73-23.71	3.36, 2.08-4.81	0.58, 0.18-1.26	0.17, 0.05-0.39	0.58, 0.05-0.94	0.17, 0.02-0.09	66.23, 93.20-5	72.11, 53.22-88.56	76.74, 65.99-90.06	1.16, 1.08-1.29					
	Biotite-Amphibole gneisses	4	54.43-62.58	0.76-1.76	13.43-16.35	6.88-12.32	0.06-0.24	3.22-6.74	4.68-6.18	1.73-5.76	2.02-3.22	0.28-0.32	0.53-1.50	1.50-2.14	1.17-3.86	11.38-22.32	3.27-3.65	0.26-0.85	0.28-0.36	0.68-74.31	0.03-0.06	1.21-1.55	72.15-83.05	74.85-86.05	1.10-1.18						
	Biotite gneisses ± Grt ± Crd ± Sil	3	56.65, 48.26-61.57	1.43, 0.70-2.72	15.65, 14.05-18.08	14.54, 5.73-19.15	0.05, 0.01-0.11	3.43, 1.67-7.61	3.43, 1.10-2.50	3.43, 0.25-2.55	3.43, 0.86-7.14	0.07, 0.02-0.09	0.48, 0.22-0.98	1.20, 0.70-2.01	1.20, 0.75-26.16	1.20, 3.40-4.05	16.03, 3.40-4.05	3.63, 0.04-1.33	0.57, 0.07-0.17	0.10, 0.01-0.06	0.10, 0.01-0.06	65.48, 56.15-75.99	0.04, 0.01-0.06	1.80, 1.07-2.65	76.75, 61.83-89.67	82.74, 73.88-94.53	1.28, 1.05-1.47				
	Garnet bearing mica gneisses ± Amph	4	54.02, 52.78-56.09	1.62, 1.35-2.36	16.36, 16.01-16.70	12.35, 11.55-12.83	0.17, 0.15-0.19	3.67, 3.10-4.39	6.60, 6.17-7.02	0.75, 0.60-1.00	3.10, 2.40-4.06	0.49, 0.22-1.64	0.86, 0.22-1.64	1.32, 0.60-2.10	1.32, 2.63-6.80	1.32, 6.78-12.35	4.49, 3.23-3.37	10.70, 3.23-3.37	0.30, 0.15-0.38	0.26, 0.37-0.44	0.40, 0.37-0.44	73.76, 71.64-76.16	0.02-0.03	1.72, 1.57-1.95	84.19, 79.89-88.88	86.97, 82.94-89.41	1.18, 1.13-1.24				
	Graphite bearing mica gneisses ± Grt ± Crd ± Sil	26	52.99, 43.19-62.49	0.97, 0.47-2.68	15.77, 10.24-28.07	13.36, 8.46-19.04	0.14, 0.04-0.58	3.47, 2.29-7.65	3.47, 0.48-7.41	2.57, 0.48-7.41	3.53, 0.48-7.41	0.14, 0.02-0.85	4.81, 0.02-0.85	5.29, 1.18-16.60	5.29, 11.51-12.55	5.29, 11.51-12.55	18.53, 5.96-31.13	3.53, 1.54-5.10	0.41, 0.01-0.85	0.17, 0.03-0.46	0.17, 0.03-0.46	65.71, 54.74-88.38	0.04, 0.0006-0.09	1.75, 0.93-2.99	73.63, 57.40-99.73	78.69, 65.61-99.76	1.20, 1.11-1.40				
	Graphite bearing mica gneisses	25	56.06-59.59	0.40-1.56	10.73-20.37	8.46-19.81	0.02-0.10	1.67-1.88	0.42-4.37	0.21-2.05	1.68-7.06	0.04-0.24	0.05-13.73	2.24-10.22	0.82-27.80	1.24-30.39	1.96-5.16	0.04-1.22	0.03-0.36	52.97-79.17	0.01-0.07	1.31-2.75	56.09-94.88	66.18-95.97	1.09-1.41						
	Total	90	55.15, 36.06-74.04	0.94, 0.40-2.88	15.64, 10.24-28.07	12.62, 5.56-22.3.29																									

Metavolcanic Units																			
Group	Classification samples	# Samples	SiO ₂	TiO ₂	Al ₂ O ₃	Fe ₂ O ₃	MnO	MgO	CaO	Na ₂ O	K ₂ O	P ₂ O ₅	SO ₃	LOI	K ₂ O/Na ₂ O	Al ₂ O ₃ /TiO ₂	SiO ₂ /Al ₂ O ₃	Na ₂ O/K ₂ O	CaO/Al ₂ O ₃
Alutaguse Metavolcanic samples	Pyroxene gneisses + Amph	8	55.27, 49.40-59.39	1.04, 0.64-1.63	14.77, 12.56-17.30	11.06, 8.85-13.46	0.16, 0.12-0.26	6.18, 4.38-8.91	8.43, 6.83-11.26	1.11, 0.35-2.19	1.08, 0.38-1.77	0.23, 0.15-0.36	0.67, 0.08-1.66	1.59, 1.17-2.18	1.65, 0.19-3.55	15.56, 8.70-27.14	3.78, 3.18-4.55	1.50, 0.28-5.26	0.58, 0.42-0.75
	2-Pyroxene gneisses + Amph	20	52.66, 45.83-62.67	1.26, 0.57-2.90	13.44, 11.23-17.54	12.98, 7.14-18.65	0.19, 0.09-0.57	6.39, 3.60-12.06	9.67, 6.20-13.16	1.53, 0.35-2.63	0.83, 0.18-1.80	0.18, 0.06-0.29	0.88, 0.22-3.67	1.58, 0.36-2.81	0.74, 0.08-4.00	12.67, 3.88-26.68	3.97, 2.87-4.70	3.41, 0.25-12.52	0.74, 0.41-1.12
	2-Pyroxene gneisses	5	52.40, 50.07-54.69	1.49, 1.03-2.31	12.88, 11.29-13.71	13.72, 9.95-17.57	0.22, 0.18-0.25	7.69, 6.60-10.36	8.70, 2.83-11.68	0.53, 0.24-0.94	1.37, 0.10-4.49	0.11, 0.08-0.16	0.90, 0.11-2.05	2.50, 0.22-8.49	2.05, 0.25-4.79	9.33, 5.92-12.32	4.09, 3.65-4.54	1.28, 0.21-4.00	0.68, 0.21-0.92
	Amphibolite	4	52.23, 48.94-57.13	0.72, 0.58-0.96	10.53, 8.99-12.77	12.46, 10.81-13.52	0.19, 0.15-0.21	9.31, 5.07-13.13	8.89, 4.76-12.37	0.69, 0.49-0.96	1.71, 0.93-2.60	0.20, 0.17-0.21	3.08, 0.23-6.78	2.15, 1.78-2.88	2.50, 1.79-3.59	14.86, 13.30-17.18	5.06, 3.83-5.89	0.43, 0.28-0.56	0.86, 0.46-1.25
	Pyroxene gneisses	4	48.43, 38.96-52.21	1.45, 0.79-2.73	14.56, 12.19-16.14	13.18, 11.31-16.36	0.35, 0.18-0.65	8.76, 7.01-11.22	7.98, 2.91-11.18	0.68, 0.47-0.99	2.06, 0.44-6.77	0.26, 0.09-0.61	2.29, 0.23-7.15	2.03, 0.38-5.61	4.14, 0.45-14.47	12.67, 4.47-19.13	3.32, 3.20-3.51	1.18, 0.07-2.22	0.53, 0.24-0.74
	Total	41	52.68, 38.96-62.67	1.21, 0.57-2.90	13.46, 8.99-17.54	12.66, 7.14-18.65	0.20, 0.09-0.65	7.02, 3.60-13.13	9.07, 2.83-13.16	1.16, 0.24-2.63	1.15, 0.10-6.77	0.19, 0.06-0.61	1.19, 0.08-7.15	1.80, 0.22-8.49	1.58, 0.08-14.47	13.04, 3.88-27.14	3.99, 2.87-5.89	2.27, 0.07-12.52	0.69, 0.21-1.25
SS Zones Metavolcanic Samples	Uusimaa	31	52.45, 47.49-61.87	1.26, 0.24-2.84	16.23, 12.73-22.26	9.99, 2.69-18.62	0.17, 0.04-0.27	5.70, 1.57-11.07	9.81, 4.06-17.13	2.90, 0.75-5.05	1.06, 0.07-3.16	0.24, 0.02-0.97	0.19, 0.02-0.59		0.38, 0.07-1.40	18.10, 4.48-76.17	3.29, 2.15-4.67	4.58, 0.72-13.91	0.62, 0.29-1.30
	Häme	22	55.51, 47.76-62.38	1.18, 0.65-2.31	17.46, 15.05-20.33	17.46, 6.18-12.66	0.15, 0.08-0.23	4.29, 2.33-7.86	5.78, 3.72-14.54	3.09, 0.67-5.11	1.37, 0.28-3.37	0.32, 0.05-1.16	0.13, 0.02-0.47		0.59, 0.11-3.61	16.40, 6.50-28.89	3.20, 2.35-4.03	3.60, 0.28-9.27	0.43, 0.24-0.77
	Saimaa	5	51.36, 48.20-54.52	2.21, 0.72-3.60	14.55, 13.06-16.33	14.55, 8.36-18.83	0.23, 0.16-0.30	5.24, 2.43-8.08	7.69, 5.92-9.20	3.08, 2.36-3.67	1.08, 0.21-2.03	0.30, 0.12-0.46	0.11, 0.02-0.32		0.38, 0.06-0.74	9.34, 3.81-22.76	3.55, 3.26-4.17	5.64, 1.35-16.03	0.53, 0.45-0.56
	Ladoga	8	53.79, 46.65-59.22	1.37, 0.83-2.14	15.78, 12.00-18.33	10.38, 8.38-12.39	0.15, 0.09-0.27	6.45, 3.72-11.31	8.32, 6.03-12.86	1.92, 0.49-3.20	1.32, 0.48-2.06	0.51, 0.18-1.78			1.00, 0.31-2.47	12.58, 7.06-20.12	3.45, 3.03-4.17	1.67, 0.41-3.23	0.54, 0.36-0.84
	Total	66	53.55, 46.65-62.38	1.32, 0.24-3.60	16.46, 12.00-22.26	10.00, 2.69-18.83	0.16, 0.04-0.30	5.29, 1.57-11.31	8.72, 3.72-17.13	2.86, 0.49-5.11	1.20, 0.07-3.37	0.30, 0.02-1.78	0.14, 0.02-0.59		0.52, 0.06-3.61	16.20, 3.81-76.17	3.30, 2.15-4.67	3.98, 0.28-16.03	0.54, 0.24-1.30
PM			45.00	0.20	4.45	8.95	0.14	37.80	3.55	0.36	0.03	0.02			0.08	22.25	10.11	12.41	0.40
N-MORB			49.62	1.26	14.99	12.27	0.19	8.21	11.65	2.25	0.09	0.17			0.04	11.94	3.31	24.24	0.95
E-MORB			50.36	1.43	15.40	10.34	0.17	8.12	11.56	2.57	0.32	0.18			0.12	10.78	3.27	8.03	1.12

Group	Classification samples	# Samples	CIA	CN	ICV	PIA	CIW	CIW/CIA	WIP	AI	CCPI	SI	ACN/K	ANK	Mg#
Alutaguse Metavolcanic samples	Pyroxene gneisses + Amph	8	76.53, 64.03-86.73	0.04, 0.01-0.07	1.98, 1.53-2.47	80.47, 66.52-92.20	81.60, 69.25-92.85	1.07, 1.02-1.10	39.00, 28.60-52.31	43.23, 30.74-53.51	87.87, 75.74-93.76	0.53, 0.16-0.78	0.82, 0.62-1.07	5.74, 2.94-9.23	51.84, 40.16-61.27
	2-Pyroxene gneisses + Amph	20	71.39, 57.43-88.93	0.05, 0.01-0.08	2.50, 1.43-3.49	73.38, 57.60-90.50	74.66, 58.06-90.74	1.05, 1.01-1.13	42.23, 22.46-54.25	39.00, 24.14-58.57	87.27, 71.19-97.19	0.34, 0.07-0.80	0.66, 0.43-0.91	5.63, 2.43-14.33	48.88, 34.87-66.96
	2-Pyroxene gneisses	5	81.70, 63.22-90.34	0.02, 0.01-0.03	2.62, 2.39-3.13	86.31, 74.04-94.20	88.25, 81.58-94.46	1.09, 1.01-1.29	39.01, 23.64-77.49	49.40, 40.14-79.76	91.74, 79.40-97.85	0.57, 0.20-0.83	0.75, 0.53-1.18	9.59, 2.13-14.96	52.67, 42.74-63.64
	Amphibolite	4	72.19, 63.54-79.08	0.02, 0.02-0.03	3.28, 2.59-4.17	79.18, 70.64-84.66	82.44, 76.73-86.00	1.15, 1.09-1.21	48.12, 27.57-65.76	52.61, 31.82-71.60	89.69, 85.81-92.36	0.70, 0.64-0.78	0.58, 0.41-0.80	3.86, 2.37-5.46	57.74, 42.62-68.40
	Pyroxene gneisses	4	77.32, 57.90-86.86	0.02, 0.02-0.03	2.40, 1.84-3.32	83.43, 75.97-89.05	86.95, 83.21-89.36	1.16, 1.02-1.53	49.48, 31.45-93.78	54.63, 44.17-84.21	89.89, 78.20-95.20	0.54, 0.31-0.94	0.83, 0.68-0.98	7.07, 1.51-10.90	56.66, 53.12-61.13
	Total	41	74.31, 57.43-90.34	0.04, 0.01-0.08	2.48, 1.43-4.17	77.89, 57.60-94.20	79.63, 58.06-94.46	1.08, 1.01-1.23	42.49, 22.46-93.78	43.94, 24.14-84.21	88.42, 71.19-97.85	0.46, 0.07-0.94	0.71, 0.41-1.18	6.10, 1.51-14.96	51.54, 34.87-68.40
SS Zones Metavolcanic Samples	Uusimaa	31	60.90, 51.45-87.70	0.09, 0.02-0.16	1.95, 0.89-3.07	61.69, 51.92-87.97	63.57, 54.56-88.01	1.05, 1.00-1.17	58.12, 35.96-73.30	34.63, 18.65-49.61	77.64, 43.83-94.72	0.25, 0.07-0.58	0.70, 0.38-0.97	3.24, 1.70-13.86	52.20, 31.35-77.93
	Häme	22	60.59, 47.99-78.53	0.10, 0.02-0.15	1.52, 1.26-2.06	64.00, 47.76-87.21	64.00, 50.56-88.90	1.06, 1.01-1.15	58.95, 41.66-77.75	34.81, 26.07-50.00	73.17, 58.86-84.06	0.30, 0.10-0.78	0.88, 0.62-1.09	2.84, 1.54-4.74	48.00, 33.63-71.60
	Saimaa	5	56.36, 51.65-60.32	0.10, 0.08-0.12	2.32, 1.85-2.72	57.01, 52.00-62.21	59.13, 54.87-65.39	1.05, 1.01-1.10	59.07, 51.37-68.69	36.00, 25.99-49.02	81.03, 76.57-84.97	0.25, 0.06-0.42	0.72, 0.70-0.75	2.36, 1.81-2.71	41.76, 23.38-64.06
	Ladoga	8	68.29, 56.80-83.98	0.06, 0.02-0.10	1.95, 1.40-3.14	70.95, 57.57-89.68	72.93, 59.84-90.48	1.07, 1.02-1.11	51.13, 32.00-63.77	41.83, 28.98-56.56	82.11, 73.45-92.92	0.44, 0.24-0.71	0.82, 0.53-0.99	4.10, 2.35-7.24	52.93, 43.47-64.39
	Total	66	61.35, 47.99-87.70	0.09, 0.02-0.16	1.83, 0.89-3.14	62.54, 47.76-89.68	64.51, 50.56-90.48	1.05, 1.00-1.17	57.62, 32.00-77.75	35.66, 18.65-56.56	76.95, 43.83-94.72	0.29, 0.06-0.78	0.78, 0.38-1.09	3.15, 1.54-13.86	50.10, 23.38-77.93

Table 1: Average (Top) and minimum-maximum ranges (down) values of the major bulk-rock elemental data of the different metasedimentary and metavolcanic lithologies of the Alutaguse zone and the South Svecofennian (SS) zones. Major oxide values correspond to 100% normalised data. LOI corresponds to original values.

Metasedimentary																						
Units																						
Group	Classification samples	# Samples	V	Cr	Co	Ni	Cu	Zn	Sc	Mo	Rb	Ba	Th	U	Pb	Sr	Nb	Ta	Zr	Hf	Y	
Alutaguse High-SiO ₂ metasedimentary samples	Garnet bearing mica gneisses ± Crd ± Sil	16	107.81, 8.00-247.00	87.25, 16.00-203.00	15.36, 1.60-35.90	43.10, 2.50-92.40	55.01, 6.20-146.50	125.81, 21.00-371.00	12.95, 3.70-29.70	2.72, 0.73-5.79	109.60, 66.60-202.00	877.50, 460.00-1570.00	15.96, 1.61-71.00	2.31, 1.10-5.50	27.83, 14.50-65.50	193.28, 114.50-434.00	12.52, 1.40-45.30	0.64, 0.14-1.34	150.06, 67.20-217.00	4.42, 2.00-6.20	17.13, 7.60-31.20	
	Garnet bearing mica gneisses	4	144.12, 56.00-246.00	73.18, 57.00-91.00	34.20, 4.90-105.11	47.70, 20.40-104.00	45.66, 5.80-140.00	171.14, 26.00-308.00	13.71, 7.46-23.30	4.16, 0.91-11.90	80.11, 32.00-832.70	558.17, 10.90-23.50	17.35, 0.67-4.40	2.59, 14.20-35.20	24.04, 35.50-386.00	17.60, 8.30-214.00	13.58, 0.60, 0.31-0.89	0.60, 120.50-222.00	4.68, 3.60-6.30	17.76, 8.32-29.60		
	Cordierite bearing mica gneisses ± Grt ± Sil	5	159.80, 76.00-277.00	79.40, 53.00-97.00	30.24, 11.20-64.70	81.26, 32.60-195.00	184.16, 34.70-600.00	129.40, 84.00-202.00	19.84, 13.50-36.00	4.16, 1.13-11.75	122.82, 76.60-155.00	762.00, 190.00-1050.00	19.10, 7.95-42.60	2.86, 2.30-3.40	16.20, 7.50-22.00	27.62, 97.60-462.00	11.96, 8.80-17.40	0.73, 0.57-0.87	173.70, 137.00-204.00	5.00, 4.10-6.10	22.38, 15.90-36.90	
	Graphite bearing mica gneisses ± Grt ± Crd ± Sil	1	122.37, 122.37	129.65, 129.65	76.90, 76.90	30.37, 30.37	27.51, 27.51	116.38, 116.38	7.93, 7.93	0.33, 0.33	185.82, 185.82	844.42, 844.42	16.32, 16.32	2.26, 2.26	12.83, 12.83	112.52, 112.52	15.62, 15.62	1.13, 1.13	227.50, 227.50	6.53, 6.53	19.61, 19.61	
	Total	26	122.33, 8.00-277.00	85.13, 16.00-203.00	23.24, 1.60-105.11	50.41, 2.50-195.00	75.41, 5.80-600.00	125.81, 21.00-371.00	12.95, 3.70-36.00	3.06, 0.33-11.90	110.59, 59.40-202.00	795.45, 190.00-1570.00	16.56, 1.61-71.00	2.46, 0.67-5.50	24.16, 7.50-65.50	209.39, 35.50-462.00	12.67, 1.40-45.30	0.68, 0.14-1.34	160.87, 67.20-227.50	4.70, 2.00-6.53	18.47, 7.60-36.90	
Alutaguse Low-SiO ₂ metasedimentary samples	Graphite bearing mica gneisses ± Grt ± Crd ± Sil	104	208.63, 12.00-529.00	101.18, 14.00-432.00	31.38, 2.30-126.00	122.21, 8.30-385.00	190.05, 5.20-1060.00	270.00-410.00	19.05, 0.60-55.40	56.40, 0.47-28.70	15.40, 4.60-214.83	511.15, 30.00-254.00	13.21, 0.11-166.00	3.94, 0.30-17.20	104.07, 2.78-1430.00	133.43, 24.00-2230.00	9.03, 0.50-38.50	0.57, 0.05-2.24	134.01, 11.00-266.42	3.69, 0.40-7.78	20.96, 1.90-123.00	
	Graphite bearing mica gneisses	2	224.31, 214.97-233.65	105.62, 87.86-123.38	63.30, 52.74-73.87	55.53, 36.97-74.09	310.13, 41.48-578.78	270.00-410.00	16.64, 16.25-17.02	10.72, 8.40-13.04	159.72, 144.32-175.12	475.40, 283.49-551.97	16.96, 11.41-22.51	4.82, 2.87-6.16	7.06, 5.02-9.10	68.57, 25.99-111.14	16.32, 10.81-21.82	1.15, 0.98-1.32	112.81-159.98	4.04, 3.38-4.77	18.07, 15.67-20.48	
	Garnet bearing mica gneisses	1	115.22, 115.22	117.90, 117.90	67.64, 67.64	33.79, 33.79	8.36, 8.36	115.84, 115.84	35.14, 35.14	0.84, 0.84	156.58, 156.58	1707.24, 1707.24	5.44, 5.44	0.70, 0.70	32.88, 32.88	196.02, 196.02	13.08, 13.08	0.60, 0.60	300.37, 300.37	8.58, 8.58	82.47, 82.47	
	Total	107	207.51, 12.00-529.00	101.50, 14.00-432.00	32.31, 2.30-126.00	119.53, 8.30-385.00	190.60, 5.20-1060.00	270.00-410.00	15.61, 0.60-55.40	9.70, 0.47-28.70	111.08, 4.60-214.83	521.78, 30.00-254.00	13.22, 0.11-166.00	3.92, 0.30-17.20	100.84, 2.78-1430.00	133.42, 24.00-2230.00	9.29, 0.50-38.50	0.59, 0.05-2.24	136.30, 11.00-266.42	3.75, 0.40-7.78	21.53, 1.90-123.00	
	Uusimaa	33	63.87, 2.00-205.00	72.98, 19.00-136.00	10.01, 2.00-21.80	29.72, 14.00-58.80	23.36, 17.00-105.00	104.00, 33.00-216.00	14.93, 4.36-31.40	19.05, 32.50-935.00	56.40, 173.00-1220.00	511.15, 13.21, 3.94, 104.07, 133.43, 9.03, 0.57, 134.01, 3.69, 20.96, 1.90-123.00	11.86, 3.00, 35.63, 183.48, 12.38, 0.92, 226.11, 6.34, 35.97, 17.76, 8.32-29.60	11.86, 3.00, 35.63, 183.48, 12.38, 0.92, 226.11, 6.34, 35.97, 17.76, 8.32-29.60	11.86, 3.00, 35.63, 183.48, 12.38, 0.92, 226.11, 6.34, 35.97, 17.76, 8.32-29.60	11.86, 3.00, 35.63, 183.48, 12.38, 0.92, 226.11, 6.34, 35.97, 17.76, 8.32-29.60	11.86, 3.00, 35.63, 183.48, 12.38, 0.92, 226.11, 6.34, 35.97, 17.76, 8.32-29.60	11.86, 3.00, 35.63, 183.48, 12.38, 0.92, 226.11, 6.34, 35.97, 17.76, 8.32-29.60	11.86, 3.00, 35.63, 183.48, 12.38, 0.92, 226.11, 6.34, 35.97, 17.76, 8.32-29.60	11.86, 3.00, 35.63, 183.48, 12.38, 0.92, 226.11, 6.34, 35.97, 17.76, 8.32-29.60		
SS Zones High-SiO ₂ Metasedimentary samples	Häme	22	90.90, 54.40-131.00	125.77, 79.00-279.00	15.04, 8.97-35.90	46.30, 31.20-88.50	45.68, 17.00-164.00	111.90, 65.40-167.00	16.43, 11.00-22.30	195.05, 157.00-755.00	431.73, 2.58-17.40	12.05, 1.59-5.10	3.38, 16.00-39.80	30.63, 93.50-562.00	30.63, 93.50-562.00	12.05, 1.59-5.10	3.38, 16.00-39.80	30.63, 93.50-562.00	12.05, 1.59-5.10	3.38, 16.00-39.80	30.63, 93.50-562.00	
	Saimaa	40	95.35, 14.30-181.00	119.05, 31.00-190.00	12.59, 4.30-30.40	47.28, 14.00-85.00	31.82, 17.00-187.00	122.65, 27.70-220.00	16.16, 6.90-28.30	152.12, 10.60-234.00	567.08, 203.00-1150.00	12.18, 1.76-17.50	2.47, 1.60-40.90	32.02, 17.40, 31.91, 87.70-548.00	32.02, 17.40, 31.91, 87.70-548.00	17.48, 3.73-10.50	11.91, 0.48-1.23	183.71, 43.30-274.00	4.91, 1.16-70.31	26.36, 4.05-44.30		
	Ladoga	14	79.57, 37.00-171.00	94.93, 34.00-213.00	13.46, 4.80-35.00	37.11, 12.50-74.00	25.82, 15.00-197.00	63.93, 4.80-12.70	9.47, 4.80-12.70	83.79, 30.00-135.00	446.94, 146.00-848.00	7.52, 4.10-11.60	2.05, 7.70-31.00	16.36, 39.00-304.00	16.36, 39.00-304.00	8.54, 4.50-12.50	0.61, 53.00-211.00	100.86, 2.60-6.00	3.87, 9.20-21.77	26.18, 9.20-21.77		
	Total	109	82.89, 2.00-205.00	103.36, 19.00-279.00	12.41, 2.00-35.90	40.46, 13.30-88.50	45.68, 12.50-187.00	107.29, 4.36-31.40	14.98, 33.00-216.00	153.83, 146.00-848.00	534.92, 146.00-848.00	11.46, 4.10-11.60	2.76, 7.70-31.00	30.63, 93.50-562.00	30.63, 93.50-562.00	12.05, 1.59-5.10	3.38, 16.00-39.80	30.63, 93.50-562.00	12.05, 1.59-5.10	3.38, 16.00-39.80	30.63, 93.50-562.00	
	Uusimaa	13	74.20-157.00	50.00-361.00	10.40-31.10	19.00-68.20	17.00-126.00	82.70-176.00	14.50-35.70	66.80-338.00	170.00-126.00	334.22-70.00	0.69-4.36	20.00-55.40	127.00-501.00	5.73-28.20	0.43-1.55	97.60-264.00	2.51-8.07	17.00-67.60	4.71, 26.41, 26.41, 26.41	
SS Zones Low-SiO ₂ Metasedimentary samples	Häme	12	147.42, 81.00-308.00	135.41, 43.30-178.00	22.31, 14.60-34.20	66.57, 23.10-86.50	83.03, 17.00-407.00	174.00, 126.00-258.00	5.49-35.30	178.00-413.00	195.00-840.00	1.18-20.60	0.56-4.13	23.00-57.10	97.90-354.00	4.84-33.40	0.38-2.29	86.60-216.00	2.28-7.37	7.91-38.60		
	Saimaa	29	140.35, 93.40-342.00	154.14, 108.00-206.00	19.97, 12.70-36.50	71.01, 50.60-99.90	51.46, 17.00-217.00	159.70, 93.20-213.00	23.07, 9.47-32.70	186.21, 91.00-329.00	687.24, 264.00-1310.00	15.12, 8.82-25.40	2.89, 1.48-6.92	35.33, 21.00-47.30	159.60, 86.30-406.00	35.33, 11.70-23.00	15.17, 0.58-2.12	182.38, 116.00-278.00	4.92, 3.25-6.73	30.31, 9.24-47.90		
	Ladoga	12	159.17, 103.00-262.00	171.92, 101.00-275.00	68.75, 15.30-34.00	88.44, 39.00-92.00	263.09, 1.56-100.00	337.20, 35.00-351.00	26.80, 6.00-26.00	37.93, 50.00-246.00	323.10, 124.00-1691.00	7.82, 4.40-14.40	2.26, 2.00-4.20	238.24, 11.90-22.00	623.45, 56.00-445.00	10.23, 10.90-16.50	0.55, 0.05-0.20	90.75, 50.00-174.00	2.66, 2.70-5.10	22.52, 5.60-26.00		
	Total	66	140.33, 74.20-342.00	149.24, 43.30-361.00	21.49, 10.40-36.50	66.46, 19.00-99.90	88.44, 1.56-100.00	263.09, 35.00-351.00	26.80, 6.00-26.00	37.93, 50.00-413.00	323.10, 124.00-1691.00	7.82, 4.40-14.40	2.26, 2.00-4.20	238.24, 11.90-22.00	623.45, 56.00-445.00	10.23, 10.90-16.50	0.55, 0.05-0.20	90.75, 50.00-174.00	2.66, 2.70-5.10	22.52, 5.60-26.00		
	PAAS	107	83	17	44	25	71	112	13.6	350	22	140	12	1.5	550	30	64	71	26	4.5		
UCC			107	83	17	44	25	71	112	13.6	350	22	140	12	1.5	550	30	64	71	26	4.5	
Chondrite			56	2650	500	10500	120	310	2.3	5.92	7.25	1.57	3.82	0.24	0.9	2.41	0.237	0.613	0.093	0.457	0.148	
Metavolcanic																						
Units																						
Group	Classification samples	# Samples	V	Cr	Co	Ni	Cu	Zn	Sc	Mo	Rb	Ba	Th	U	Pb	Sr	Nb	Ta	Zr	Hf	Y	
Alutaguse Metavolcanic samples	2-Pyroxene gneisses + Amph	9	195.67, 70.00-382.00	207.00, 72.00-546.00	44.48, 34.20-59.40	132.48, 79.30-210.00	596.62, 55.20-2230.00	714.11, 118.00-2650.00	24.31, 7.30-44.80	2.81, 0.23-15.95	3.13, 0.60-9.40	75.56, 60.00-130.00	8.03, 0.05-21.60	4.07, 0.10-9.50	715.81, 1.40-4030.00	57.71, 27.40-260.00	5.36, 2.00-10.80	0.32, 0.13-0.64	83.26, 31.70-186.00	2.30, 1.20-5.00	17.00, 47.00-44.30	
	2-Pyroxene gneisses	11	215.50, 128.00-436.00	195.76, 47.00-938.45	195.76, 18.50-72.97	44.13, 30.56-134.52	140.26, 25.94-401.00	140.26, 104.00-689.20	37.21, 15.20-48.90	2.42, 0.29-11.40	43.89, 2.00-224.00	468.13, 50.00-3030.00	3.10, 0.04-10.90	1.54, 0.10-5.40	93.77							
	Pyroxene gneisses + Amph	2	131.00, 52.00-210.00	89.50, 74.00-105.00	14.40, 4.90-22.90	58.05, 16.80-99.30	101.25, 9.00-193.50	101.25, 69.00-193.50	14.10, 13.50-14.70	4.49, 4.49	79.87, 52.70-107.00	480.23, 350.00-900.00	11.28, 6.21-15.85	2.25, 0.90-3.60	11.28, 1380.00-1720.00	13.50, 24.60-313.00	10.15, 2.20-29.24	10.15, 0.48, 181.50	0.48, 0.20-2.08	167.00-169.00	1.40-4.30	10.90-28.00
	Pyroxene gneisses	10	204.23, 59.00-471.00	136.66, 53.66	47.64, 19.60	65.51, 16.80-92.00	130.40, 16.50-220.00	155.68, 66.00-290.00	20.12, 4.00-52.80	6.28, 1.02	54.32, 33.00-203.00	354.96, 150.00-710.00	12.14, 0.04-76.50	11.89, 0.20-2.70	1189, 117.50-5330.00	1739.15, 12.60-62.20	1739.15, 0.50-1.70	1739.15, 0.20-0.50	918, 53.00-202.00	2.70, 0.10-5.90	22.56, 6.90-34.00	
	Total	32	235.49, 29.00-471.00	149.75, 13.00-938.45	43.47, 4.90-72.97	88.44, 16.80-210.00	263.09, 9.00-2230.00	337.20, 66.00-265														

Metasedimentary Units														
Group	Classification samples	La	Ce	Pr	Nd	Sm	Eu	Gd	Dy	Ho	Er	Tm	Yb	Lu
Alutaguse High-SiO ₂ metasedimentary samples	Garnet bearing mica gneisses ± Crd ± Sil	48.01, 19.10-187.50	98.66, 32.40-436.00											
	Garnet bearing mica gneisses	50.57, 32.00-79.87	102.38, 63.30-145.73	15.90	52.80	6.39	1.30	5.11	1.69	0.32	0.92	0.13	0.80	0.12
	Cordierite bearing mica gneisses ± Grt ± Sil	53.94, 26.00-104.50	110.38, 55.70-223.00											
	Graphite bearing mica gneisses ± Grt ± Crd ± Sil	42.66	79.75	9.85	35.11	6.06	1.26	5.16	3.58	0.73	2.09	0.31	2.13	0.32
	Total	48.62, 19.10-187.50	99.40, 32.40-436.00	10.95, 7.10-15.90	37.97, 26.00-52.80	5.65, 4.50-6.39	1.15, 0.88-1.30	4.69, 3.80-5.16	2.92, 1.69-3.58	0.62, 0.32-0.80	1.77, 0.92-2.30	0.26, 0.13-0.33	1.71, 0.80-2.20	0.25, 0.12-0.32
Alutaguse Low-SiO ₂ metasedimentary samples	Graphite bearing mica gneisses ± Grt ± Crd ± Sil	39.01, 3.20-361.00	77.41, 10.15-500.00	8.68, 2.14-11.68	32.53, 10.51-43.52	6.15, 2.46-8.12	1.57, 0.97-2.02	5.95, 2.12-7.69	5.39, 0.75-9.17	1.24, 0.14-2.33	3.46, 0.37-6.83	0.52, 0.06-1.01	3.55, 0.39-7.12	0.56, 0.07-1.11
	Graphite bearing mica gneisses	41.06, 24.91-57.21	79.29, 51.65-106.92	9.47, 6.13-12.80	34.05, 22.28-45.82	5.99, 4.54-7.44	0.97, 0.66-1.27	5.39, 4.38-6.40	3.60, 3.35-3.86	0.72, 0.60-0.83	1.90, 1.49-2.31	0.27, 0.18-0.36	1.82, 1.08-2.56	0.30, 0.17-0.42
	Garnet bearing mica gneisses	27.63	46.20	4.97	17.07	3.28	1.74	5.34	10.66	2.94	9.59	1.61	11.85	1.89
	Total	38.93, 3.20-361.00	77.17, 10.15-500.00	8.51, 2.14-12.80	31.61, 10.51-45.82	5.83, 2.46-8.12	1.44, 0.66-2.02	5.70, 2.12-7.69	5.47, 0.75-10.66	1.27, 0.14-2.94	3.66, 0.37-9.59	0.56, 0.06-1.61	3.90, 0.39-11.85	0.61, 0.07-1.89
	Uusimaa	38.43, 16.00-54.30	77.61, 30.10-121.00	9.28, 3.18-16.00	35.20, 11.20-72.10	6.87, 1.48-18.60	1.27, 0.45-2.63	6.67, 1.07-20.70	5.80, 0.55-25.00	1.19, 0.12-5.50	3.44, 0.32-15.90	0.51, 0.04-2.40	3.38, 0.28-16.40	0.50, 0.05-2.48
SS Zones High-SiO ₂ Metasedimentary samples	Häme	35.57, 14.10-50.70	71.97, 27.30-99.00	8.48, 3.52-11.80	31.31, 14.10-43.90	5.93, 2.99-7.90	1.11, 0.73-1.51	5.46, 2.83-7.47	4.39, 2.03-5.88	2.42, 0.96-1.12	2.35, 0.97-1.46	2.35, 1.03-1.54	2.35, 1.19-1.48	0.34, 0.19-0.48
	Saimaa	41.22, 10.40-58.90	81.76, 17.00-114.00	9.65, 1.71-12.80	35.83, 5.48-48.60	6.44, 0.95-8.88	1.17, 0.33-1.75	5.69, 0.95-8.36	4.45, 0.68-7.75	0.87, 0.37-4.08	2.50, 0.05-0.60	0.36, 0.37-4.16	2.41, 0.06-0.55	0.36, 0.06-0.55
	Ladoga	23.39, 8.70-51.00	49.34, 17.80-106.00	5.39, 1.90-11.90	20.31, 7.50-45.00	3.65, 1.35-7.80	0.78, 0.48-1.30	3.19, 1.42-6.40	2.61, 1.51-4.20	0.56, 0.40-0.79	1.53, 1.05-2.80	0.25, 0.15-0.36	1.54, 1.10-2.30	0.22, 0.17-0.32
	Total	36.94, 8.70-58.90	74.36, 17.00-121.00	8.75, 1.71-16.00	32.73, 5.48-72.10	6.11, 0.95-18.60	1.14, 0.33-2.63	5.62, 0.95-20.70	4.61, 0.55-25.00	0.92, 0.12-5.50	2.64, 0.32-15.90	0.39, 0.04-2.40	2.58, 0.28-16.40	0.38, 0.05-2.48
	Uusimaa	40.29, 16.20-68.80	79.67, 32.90-139.00	9.41, 3.99-16.40	35.25, 15.00-62.60	6.55, 3.46-10.90	1.22, 0.78-1.84	6.07, 3.41-9.63	5.07, 2.74-9.31	1.04, 0.58-2.30	3.14, 0.92-2.58	0.47, 0.24-1.33	3.19, 1.60-9.17	0.48, 0.24-1.47
SS Zones Low-SiO ₂ Metasedimentary samples	Häme	41.68, 7.19-63.10	82.89, 17.50-124.00	9.84, 2.40-14.30	36.62, 11.30-53.10	6.67, 3.56-9.69	1.21, 0.99-1.44	6.03, 4.21-8.53	4.48, 1.84-6.45	0.72, 0.27-1.33	2.58, 0.67-3.84	0.38, 0.09-0.61	2.47, 0.56-4.22	0.36, 0.08-0.63
	Saimaa	48.56, 30.70-78.60	96.26, 62.80-158.00	11.36, 7.43-18.20	41.88, 27.50-66.80	7.55, 5.54-11.90	1.30, 0.76-1.57	6.68, 4.55-10.70	5.24, 2.34-7.84	0.52, 0.33-1.64	2.92, 0.69-4.79	0.43, 0.10-0.68	2.77, 0.70-4.21	0.41, 0.11-0.61
	Ladoga	22.79, 13.40-41.00	50.73, 27.70-81.00	5.48, 3.20-10.00	20.46, 11.90-37.00	3.92, 2.30-7.10	0.83, 0.41-1.52	3.40, 2.00-5.80	2.68, 1.25-4.40	0.57, 0.23-0.99	1.62, 0.79-2.80	0.25, 0.10-0.46	1.68, 0.60-2.80	0.24, 0.10-0.44
	Total	40.99, 7.19-78.60	82.28, 17.50-158.00	9.63, 2.40-18.20	35.72, 11.30-66.80	6.53, 2.30-11.90	1.18, 0.41-1.84	5.84, 2.00-10.70	4.61, 1.25-9.31	0.92, 0.23-2.30	2.67, 0.67-7.99	0.39, 0.09-1.33	2.60, 0.56-9.17	0.38, 0.08-1.47
PAAS		38.20	79.60	8.83	33.90	5.55	1.08	4.66	4.68	0.99	2.85	0.41	2.82	0.44
UCC		0.88	3.8	0.64	3.5	0.8	2.3	0.33	0.32	5.8	1	17	10.5	2.7
Chondrite		0.237	0.613	0.093	0.457	0.148	0.0563	0.199	0.246	0.0546	0.16	0.0247	0.161	0.0246
Metavolcanic Units														
Group	Classification samples	La	Ce	Pr	Nd	Sm	Eu	Gd	Dy	Ho	Er	Tm	Yb	Lu
Alutaguse Metavolcanic samples	2-Pyroxene gneisses + Amph	24.28, 6.90-56.60	55.75, 19.15-125.00											
	2-Pyroxene gneisses	13.98, 3.10-42.20	30.75, 4.85-79.60	1.92, 1.55-2.29	8.18, 6.55-9.81	2.25, 1.69-2.81	1.01, 0.92-1.10	2.65, 1.99-3.31	3.01, 2.10-3.92	0.69, 0.48-0.90	1.94, 1.34-2.54	0.30, 0.22-0.39	1.99, 1.50-2.49	0.31, 0.24-0.38
	Pyroxene gneisses + Amph	39.85, 32.60-47.10	76.90, 61.60-92.20											
	Pyroxene gneisses	67.74, 8.80-188.50	125.85, 23.88-400.00	3.30	14.60	3.72	1.39	4.26	4.73	1.08	2.98	0.44	2.97	0.42
	Total	35.29, 3.10-188.50	70.38, 4.85-400.00	2.38, 1.55-3.30	10.32, 6.55-14.60	2.74, 1.69-3.72	1.14, 0.92-1.39	3.19, 1.99-4.26	3.58, 2.10-4.73	0.82, 0.48-1.08	2.29, 1.34-2.98	0.35, 0.22-0.44	2.32, 1.50-2.97	0.35, 0.24-0.42
SS Zones Metavolcanic Samples	Uusimaa	14.48, 0.66-43.30	32.17, 1.73-95.20	4.31, 0.74-12.20	18.17, 1.60-50.40	4.17, 0.64-9.27	1.23, 0.33-2.50	4.48, 0.83-9.52	4.00, 0.81-1.84	2.24, 0.17-1.44	2.34, 0.07-0.75	0.33, 0.07-0.75	2.11, 0.42-5.17	0.31, 0.07-0.74
	Häme	20.11, 1.47-101.00	44.07, 4.52-210.00	5.74, 0.74-25.40	23.51, 3.72-91.90	5.09, 1.29-14.00	1.40, 0.31-3.59	5.17, 1.96-10.70	4.39, 1.81-8.15	0.89, 0.38-1.55	2.51, 0.04-4.39	0.36, 0.04-0.61	2.37, 1.05-4.19	0.35, 0.15-0.59
	Saimaa	16.19, 8.47-27.50	38.34, 18.60-58.90	5.45, 2.35-7.45	24.54, 9.78-38.10	6.24, 2.41-12.20	1.85, 0.74-3.74	6.84, 2.87-13.80	6.42, 3.12-12.70	1.26, 0.70-2.34	3.61, 2.12-6.17	0.52, 0.31-0.84	3.31, 2.03-5.35	0.48, 0.31-0.76
	Ladoga	26.26, 9.90-53.70	54.50, 23.20-112.00	6.66, 3.00-13.00	27.18, 12.90-49.90	5.02, 2.57-8.01	1.45, 0.78-2.12	4.86, 2.39-7.47	3.86, 1.83-5.18	0.75, 0.38-0.92	1.95, 1.10-2.33	0.29, 0.15-0.34	1.76, 1.06-2.11	0.27, 0.15-0.35
	Total	17.92, 0.66-101.00	39.31, 1.73-210.00	5.16, 0.74-25.40	21.53, 1.60-91.90	4.74, 0.64-14.00	1.36, 0.31-3.74	4.94, 0.83-13.80	4.29, 0.93-12.70	0.86, 0.17-2.34	2.40, 0.44-6.17	0.35, 0.07-0.84	2.25, 0.42-5.35	0.33, 0.07-0.76
PM		0.65	1.67	0.25	1.25	0.41	0.15	0.54	0.68	0.15	0.438	0.07	0.49	0.06

Table 2 Continue next page

Metasedimentary Units																						
Group	Classification samples	Th/U	Th/Ta	Zr/Sc	Th/Zr	Nb/Zr	Nb/Ta	Nb/Y	Zr/Y	Ba/Rb	La/Nb	Th/La	Sm/Yb	Eu ⁺	(Gd/Yb) _{cn}	(La/Sm) _{cn}	(La/Yb) _{cn}	ΣREE	LREE	HREE	HREE/LREE	(100) × TiO ₂ /Zr
Alutaguse High-SiO ₂ metasedimentary samples	Garnet bearing mica gneisses ± Crd ± Sil	8.20, 1.07-37.37	23.68, 11.50-52.99	14.43, 4.66-27.83	0.12, 0.02-0.69	0.09, 0.02-0.44	18.33, 10.00-33.81	0.79, 0.15-1.80	10.21, 3.71-19.41	8.26, 3.79-17.49	4.80, 2.44-15.79	0.31, 0.07-0.45										
	Garnet bearing mica gneisses	13.70, 2.48-35.27	30.57, 17.30-41.27	14.26, 7.63-20.30	0.11, 0.07-0.16	0.08, 0.07-0.10	24.38, 13.17-37.42	0.87, 0.57-1.57	10.68, 7.50-18.19	7.05, 4.75-9.39	3.91, 2.51-6.13	0.35, 0.29-0.43	8.00	0.27	4.81	4.8	36.09	311.58	307.1	4.47	0.01	0.32
	Cordierite bearing mica gneisses ± Grt ± Sil	6.49, 2.48-12.53	25.34, 10.19-48.97	10.05, 5.67-13.09	0.12, 0.04-0.31	0.07, 0.05-0.13	16.27, 14.10-20.00	0.57, 0.30-0.85	8.27, 5.53-10.35	6.05, 2.48-8.67	4.34, 2.36-6.01	0.34, 0.31-0.41										
	Graphite bearing mica gneisses ± Grt ± Crd ± Sil	7.23, 14.42	28.69	0.07	0.07	0.07	13.81	0.80	11.6	4.54	2.73	0.38	2.8	0.76	4.35	5.92	40.39	189.73	179.85	9.88	0.05	0.38
	Total	8.50, 1.07-37.37	24.18, 10.19-52.99	14.11, 4.66-28.69	0.11, 0.02-0.69	0.08, 0.02-0.44	18.45, 10.00-37.42	0.75, 0.15-1.80	9.91, 3.71-19.41	7.41, 2.48-17.49	4.42, 2.36-15.79	0.33, 0.07-0.45	4.30, 2.05-8.00	0.52, 0.27-0.76	4.58, 4.35-4.81	5.36, 4.80-5.92	38.24, 36.09-40.39	250.65, 189.73-311.58	243.48, 179.85-307.10	7.18, 4.47-9.88	0.03, 0.01-0.05	0.35, 0.32-0.38
Alutaguse Low-SiO ₂ metasedimentary samples	Graphite bearing mica gneisses ± Grt ± Crd ± Sil	4.29, 0.12-29.56	52.43, 0.14-221.25	17.25, 1.22-423.33	0.15, 0.00-5.28	0.08, 0.00-1.27	14.59, 0.03-6.03	0.59, 0.03-6.03	10.56, 0.53-133.68	5.05, 1.55-23.52	13.31, 0.35-68.83	0.33, 0.02-0.47	2.50, 0.79-6.27	0.45, 0.22-0.77	1.69, 0.36-5.18	3.60, 1.07-7.81	12.66, 0.91-67.97	160.25, 56.42-228.00	160.25, 39.93-212.67	15.64, 2.00-28.92	0.12, 0.02-0.41	0.77, 0.38-1.85
	Graphite bearing mica gneisses	4.76, 1.69-7.84	14.21-23.53	8.24, 1.22-13.54	0.12, 0.04-0.11	0.12, 0.10-0.14	13.77, 11.04-16.50	0.96, 0.53-1.39	7.86, 5.51-10.21	2.56, 1.96-3.15	2.46, 2.30-2.62	0.43, 0.39-0.46	0.37, 0.35-0.39	1.62, 0.87-2.36	3.55, 3.55-3.56	4.25-12.20	125.60-245.50	114.55-237.86	7.63-11.04	0.03-0.10	0.48-0.53	
	Garnet bearing mica gneisses	7.76, 9.13	8.55	0.02	0.04	0.04	21.96	0.16	3.64	10.9	2.11	0.2	0.28	0.35	2.21	3.58	11.33	146.09	106.22	39.87	0.38	0.23
	Total	4.33, 0.12-29.56	50.94, 0.14-221.25	16.97, 1.22-423.33	0.14, 0.00-5.28	0.08, 0.00-1.27	14.64, 0.03-6.03	0.59, 0.03-6.03	10.42, 0.53-133.68	5.05, 1.55-23.52	12.90, 0.35-68.83	0.33, 0.02-0.47	2.58, 0.28-6.91	0.43, 0.22-0.77	1.73, 0.36-5.18	3.59, 1.07-7.81	3.59, 0.91-67.97	174.94, 56.42-245.50	158.24, 39.93-237.86	16.70, 2.00-39.87	0.14, 0.02-0.41	0.67, 0.32-1.85
SS Zones High-SiO ₂ Metasedimentary samples	Uusimaa	4.56, 1.10-9.71	14.33, 4.79-33.67	17.14, 6.05-61.53	0.06, 0.02-0.09	0.06, 0.03-0.14	14.41, 8.08-32.91	0.42, 0.16-1.37	7.45, 4.01-22.80	5.93, 1.61-37.54	3.23, 1.72-4.30	0.31, 0.25-1.10	2.45, 1.03-5.30	0.77, 0.25-1.10	2.32, 0.93-6.61	3.02, 1.09-4.51	11.74, 0.86-40.84	191.15, 64.97-375.79	175.32, 63.48-304.43	15.83, 1.49-71.36	0.09, 0.02-0.23	0.29, 0.07-0.54
	Häme	3.82, 1.62-8.36	12.52, 6.72-22.31	11.07, 5.99-16.15	0.07, 0.02-0.12	0.07, 0.03-0.14	12.94, 7.83-18.72	0.48, 0.33-0.82	7.02, 3.88-11.10	2.49, 1.69-4.08	3.4, 1.80-5.2	0.34, 0.25-0.31	2.56, 1.02-3.11	0.71, 0.44-1.42	2.96, 1.12-4.39	7.52, 1.99-19.27	171.34, 70.75-236.86	159.82, 65.68-222.05	11.51, 5.08-15.74	0.07, 0.06-0.09	0.37, 0.25-0.58	
	Saimaa	5.25, 1.07-12.41	15.72, 3.70-34.38	12.45, 6.28-34.37	0.07, 0.04-0.11	0.07, 0.04-0.11	15.25, 7.84-20.63	0.42, 0.47-0.92	7.36, 3.79-12.01	4.26, 1.47-20.09	3.47, 2.61-5.61	0.29, 0.14-0.35	2.74, 1.49-3.83	0.65, 0.34-0.96	1.84, 1.30-2.70	3.94, 2.41-4.88	10.98, 6.34-17.55	181.76, 38.59-263.62	178.02, 36.82-250.45	11.78, 1.78-19.33	0.07, 0.05-0.10	0.37, 0.12-0.56
	Ladoga	3.86, 2.50-7.92	11.96, 7.00-15.71	12.05, 5.08-43.96	0.09, 0.02-0.12	0.10, 0.03-0.14	13.11, 11.14-14.52	0.61, 0.36-1.36	7.57, 3.33-20.29	5.83, 2.24-17.77	2.83, 1.54-5.00	0.34, 0.23-0.56	2.34, 1.43-3.55	0.64, 0.45-1.06	2.22, 1.14-5.95	4.07, 2.19-6.84	14.26, 2.83-44.54	113.23, 43.95-240.89	106.05, 39.17-229.40	7.18, 4.49-71.49	0.07, 0.05-0.12	0.74, 0.15-1.36
	Total	4.57, 1.07-12.41	14.17, 3.70-34.38	13.54, 5.08-61.53	0.07, 0.02-0.12	0.07, 0.02-0.19	14.26, 7.83-32.91	0.48, 0.16-1.37	7.35, 3.33-22.80	4.61, 1.54-5.81	3.21, 0.31-3.26	0.31, 0.14-0.56	0.33, 0.25-1.10	2.56, 0.63-6.61	2.03, 1.09-6.84	3.48, 0.86-44.54	10.93, 38.59-375.79	178.02, 36.82-304.43	165.64, 149-71.36	12.36, 0.02-0.23	0.07, 0.07-1.36	
SS Zones Low-SiO ₂ Metasedimentary samples	Uusimaa	5.00, 2.68-8.79	13.60, 5.01-24.25	7.59, 3.19-13.10	0.08, 0.03-0.14	0.08, 0.05-0.11	14.10, 10.00-21.36	0.40, 0.24-0.63	5.11, 3.66-7.43	3.16, 0.79-6.02	1.25, 0.86-4.62	0.30, 0.21-0.38	2.05, 0.84-2.96	0.70, 0.46-1.57	2.78, 1.17-8.16	3.36, 1.61-4.30	16.00, 4.50-66.11	172.76, 87.41-326.31	174.87, 76.44-308.85	14.29, 7.46-32.82	0.09, 0.06-0.14	0.51, 0.33-0.93
	Häme	5.20, 1.46-8.28	12.02, 3.14-19.28	9.69, 2.45-38.80	0.08, 0.01-0.12	0.10, 0.06-0.16	14.31, 11.52-16.19	0.82, 0.19-4.22	7.29, 3.38-26.93	2.66, 0.87-2.94	0.30, 0.16-0.37	0.32, 0.16-0.42	0.72, 0.37-1.18	0.46, 0.25-0.71	1.78, 1.08-2.41	3.16, 1.76-4.45	8.09, 2.08-12.58	196.98, 59.21-291.74	184.95, 47.42-274.09	12.03, 4.12-17.65	0.08, 0.02-0.25	0.64, 0.39-1.60
	Saimaa	6.00, 1.92-10.61	15.73, 6.27-25.81	8.31, 4.94-16.90	0.08, 0.04-0.14	0.08, 0.04-0.12	13.96, 10.26-26.67	1.05, 0.33-1.69	8.64, 4.23-17.32	4.01, 0.80-8.60	1.76, 1.96-4.13	0.36, 0.25-0.36	2.51, 2.14-9.79	0.59, 0.36-1.10	2.02, 0.97-2.82	3.85, 0.59-4.73	11.54, 1.47-17.51	115.14, 148.96-366.68	107.60, 139.92-345.77	7.54, 4.88-20.99	0.07, 0.02-0.08	1.14, 0.28-0.65
	Ladoga	2.76, 1.57-3.76	8.40, 4.73-11.45	7.74, 2.69-25.17	0.13, 0.03-0.17	0.10, 0.08-0.22	13.96, 12.02-16.47	1.05, 0.57-2.30	8.66, 3.05-26.96	4.01, 1.91-9.66	1.76, 0.98-2.63	0.36, 0.27-0.65	2.51, 1.62-3.83	0.59, 0.36-0.84	2.02, 1.14-2.68	3.85, 2.40-4.22	11.54, 4.29-17.76	115.14, 64.46-196.10	107.60, 61.01-183.42	7.54, 3.45-12.68	0.04-0.11	1.14, 0.66-2.00
	Total	5.07, 1.46-10.61	13.30, 3.14-25.81	8.32, 2.45-38.80	0.09, 0.01-0.17	0.10, 0.04-0.22	14.67, 10.00-26.67	0.66, 0.19-4.22	6.73, 3.05-26.96	3.54, 0.79-9.66	2.86, 0.98-4.62	0.32, 0.16-0.65	2.85, 0.84-14.12	0.67, 0.36-1.18	2.06, 1.33	3.57, 4.29	11.51, 9.2	194.61, 184.783	182.19, 171.82	12.42, 12.96	0.07, 0.08	0.63, 0.28-0.47
PAAS		4.71	11.4	13.12	0.05	0.09	14.84	0.7	7.77	4.91	2.5	0.35	2.05	0.64	1.33	4.16	9.2	184.783	171.82	12.96	0.08	0.47
UCC		3.89	10.5	13.97	0.06	0.06	12	0.55	8.64	4.91	2.5	0.35	2.05	0.64	1.39	4.16	9.26	146.37	136.28	10.09	0.07	0.33
Metavolcanic units																						
Group	Classification samples	Th/U	Th/Ta	Zr/Sc	Th/Zr	Nb/Zr	Nb/Ta	Nb/Y	Zr/Y	Ba/Rb	La/Nb	Th/La	Sm/Yb	Eu*	(Gd/Yb) _{cn}	(La/Sm) _{cn}	(La/Yb) _{cn}	ΣREE	LREE	HREE	HREE/LREE	δNb
Alutaguse Metavolcanic samples	2-Pyroxene gneisses + Amph	1.53, 0.50-3.92	24.91, 0.08-81.74	5.09, 0.71-13.25	0.07, 0.00-0.19	0.08, 0.03-0.29	16.30, 13.50-22.17	0.19, 0.07-0.36	3.07, 1.23-5.58	36.95, 13.83-100.00	4.64, 0.74-10.35	0.26, 0.01-0.41										0.12, -0.44-1.13
	2-Pyroxene gneisses	1.36, 0.40-3.83	5.06, 0.07-25.95	3.29, 0.03-0.10	0.03, 0.00-0.14	0.16, 0.06-0.34	14.82, 8.53-19.60	0.72, 0.13-1.05	4.91, 2.81-388.46	45.64, 1.38-3.38	0.18, 0.01-0.52	1.13	0.50, 0.45-0.55	0.98, 0.87-1.08	2.98, 1.63-4.33	3.50, 2.00-5.01	44.69, 35.29-54.09	35.96, 29.07-42.85	8.73, 6.22-11.24	0.24, 0.21-0.26	0.34, -0.15-1.13	
	Pyroxene gneisses + Amph	5.72, 4.54-6.90	22.73, 21.41-24.04	12.86, 12.52-13.20	0.06, 0.03-0.10	0.06, 0.04-0.07	23.03, 17.79-28.28	0.66, 0.35-0.98	10.96, 8.19-13.74	6.17, 5.70-6.64	3.93, 3.89-3.98	0.27, 0.19-0.35										-0.46, -0.47-0.45
	Pyroxene gneisses	5.82, 0.08-34.77	12.96, 0.08-45.00	8.26, 1.61-25.56	0.07, 0.00-0.42	0.13, 0.04-0.34	22.11, 4.08-40.00	1.14, 0.01-0.91	5.84, 2.03-121.21	34.47, 85.55-317.50	66.09, 0.00-0.41	0.14, 0.00-0.41	1.25	0.69	1.96	4.39	13.63	75.26	61.84	13.43	0.22	0.41, -0.63-1.44
	Total	3.07, 0.08-34.77	14.21, 0.08-45.00	5.95, 1.61-25.56	0.06, 0.00-0.42	0.09, 0.03-0.34	18.02, 4.08-40.00	0.70, 0.01-0.91	5.06, 2.03-121.21	37.24, 85.55-317.50	66.09, 0.00-0.41	0.14, 0.00-0.41	1.17	0.56	1.90	4.45	6.88, 35.29-75.26	54.88, 29.07-61.84	44.58, 6.22-13.43	10.30, 0.21-0.26	0.23, -0.63-1.44	
SS Metavolcanic Samples	Uusimaa	2.31, 0.89-3.20	5.61, 0.77-11.82	3.97, 0.13-14.49	0.02, 0.00-0.05	0.02, 0.01-0.12	15.00, 2.67-22.21	0.09, 0.03-0.75	4.53, 1.15-11.41	2.39, 0.80-56.52	2.39, 0.99-4.24	0.23, 0.03-0.36	2.39, 0.97-6.99	0.69, 0.48-1.99	1.24, 0.94-6.47	2.69, 1.04-5.34	3.67, 1.85-44.02	13.77, 9.31-230.49	13.77, 6.53-220.64	10.48, 2.27-24.27	0.08, 0.04-0.43	-0.07, -0.59-0.27
	Häme	1.97, 0.71-3.05	5.53-18.66	7.40, 0.77-43.95	0.02, 0.00-0.05	0.06, 0.03-0.09	15.06, 8.15-22.24	0.36, 0.06-1.62	5.71, 1.67-19.88	7.39, 1.40-16.55	2.04, 1.41-3.76	0.18, 0.02-0.40	0.68-8.33	0.41-1.01	0.83-2.64	0.93-4.50	1.07-15.77	30.34-468.15	13.95-456.59	4.94-20.46		

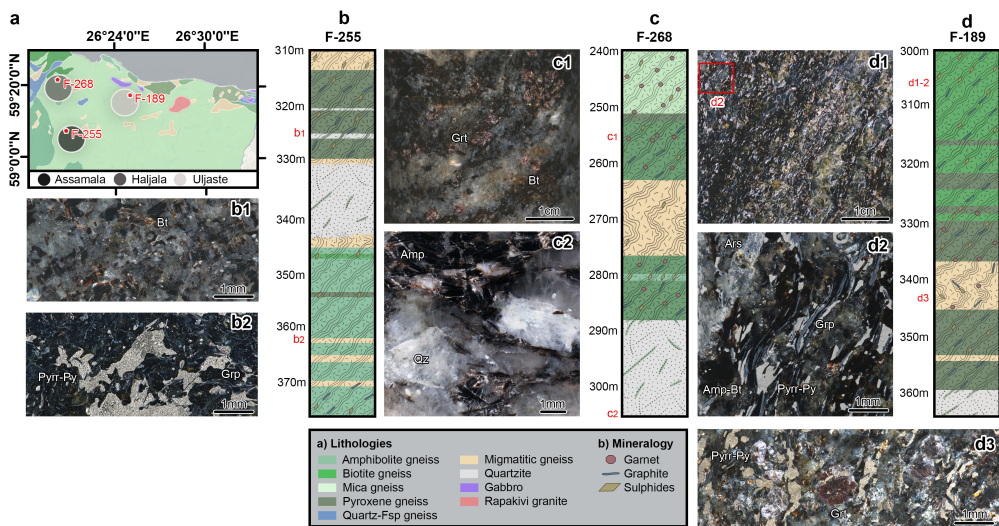


Figure 5: (a) Zoom within Alutagus domain, CRM target zones and representative Drill cores. A representative PPX-10x image of the analysed samples is also provided per drillcore. The drill core localities are (b)Haljala, (c) Assamalla, (d) Uljaste (Fig.1c). The Id-Number of each drill core is indicated within the reference map. Mineral abbreviations: Amp—amphibolite, Ars—arsenopyrite, Bt—biotite, Qz—quartz, Pyr—pyrrhotite, Py—pyrite, Grp—graphite, Grt—garnet

4.1.2.2 Rapakivi Dataset (Märjamaa-Kloostri)

Whole-Rock major elemental geochemical data for the Märjamaa pluton and Kloostri satellite were compiled from Kivisilla et al. (1999) [116], comprising 45 samples organised by drill-core positions, and per corresponding Phases (Fig.4c), based on Estonian Geological Survey drill core records. Elemental concentrations were normalised to 100 wt.%, and summary values are presented in Table 3.

Id	Kivisilla et al. (1999) [116] Id	Long	Lat	Core Id	Depth (m)	Phase	SiO ₂	TiO ₂	Al ₂ O ₃	Fe ₂ O ₃ _{tot}	MnO	MgO	CaO	Na ₂ O	K ₂ O	P ₂ O ₅	SO ₃	LOI	Fe ₂ O ₃	FeO	FeO*
3023300	374	24.43	58.980	302	330	I	74.30	0.10	12.56	1.18	0.01	0.40	1.77	3.19	5.97	0.01	0.07	0.44	0.54	0.58	1.07
3023320	375	24.43	58.980	302	332	I	63.45	1.08	13.66	7.08	0.04	1.83	3.78	2.93	4.58	0.46	0.22	0.90	2.82	3.83	6.37
3023344	377	24.43	58.980	302	334.4	I	67.26	0.75	13.06	5.12	0.11	1.14	3.27	2.90	4.91	0.34	0.40	0.73	2.11	2.71	4.61
30233442	376	24.43	58.980	302	334.42	I	67.14	0.80	13.14	5.18	0.11	1.13	3.42	2.90	4.73	0.37	0.40	0.69	2.11	2.76	4.66
3024031	380	24.43	58.980	302	403.1	I	73.00	0.11	13.34	2.05	0.01	0.43	1.35	2.80	6.31	0.03	0.08	0.49	0.94	1.00	1.85
3024051	381	24.43	58.980	302	405.1	I	65.95	0.78	13.79	5.25	0.01	1.40	3.04	2.84	5.59	0.34	0.22	0.80	2.24	2.70	4.72
3024384	385	24.43	58.980	302	438.4	I	65.56	0.76	13.75	5.88	0.01	1.41	3.03	2.99	5.47	0.32	0.20	0.63	2.51	3.03	5.29
3024600	388	24.43	58.980	302	460	I	64.85	0.82	13.34	5.73	0.10	1.04	4.86	2.48	5.23	0.35	0.22	0.96	2.33	3.06	5.15
3024785	390	24.43	58.980	302	478.5	I	66.01	0.93	13.50	5.74	0.02	1.60	2.93	2.88	5.05	0.39	0.20	0.77	2.37	3.03	5.16
3024805	391	24.43	58.980	302	480.5	I	65.85	0.90	14.21	6.17	0.14	1.23	2.86	2.75	4.77	0.33	0.25	0.54	2.48	3.31	5.55
3024850	393	24.43	58.980	302	485	I	66.32	0.80	13.78	5.48	0.14	1.39	2.94	3.00	5.10	0.30	0.22	0.54	2.29	2.86	4.93
3024874	394	24.43	58.980	302	487.4	I	65.60	0.88	13.09	6.10	0.11	1.65	3.25	2.76	4.89	0.43	0.50	0.73	2.47	3.26	5.48
9v3250	125	24.53	58.980	9v	325	I	64.53	0.98	13.02	6.94	0.14	1.59	3.65	2.47	4.43	0.46	0.55	1.22	2.67	3.84	6.24
9v3270	126	24.53	58.980	9v	327	I	63.13	0.88	13.78	7.13	0.28	1.75	3.35	2.46	5.07	0.47	0.44	1.27	2.84	3.86	6.42
F3052780	1374	24.51	59.100	F305	278	I	69.62	0.37	13.24	4.30	0.09	0.67	2.18	2.57	6.26	0.12	0.22	0.35	1.91	2.15	3.87
F3063064	1376	24.47	59.080	F306	306.4	I	67.40	1.11	13.56	4.88	0.10	1.38	3.11	2.57	4.80	0.32	0.22	0.55	1.96	2.63	4.39
F3063270	1377	24.47	59.080	F306	327	I	63.71	1.08	14.07	6.78	0.13	1.36	3.70	3.12	4.73	0.49	0.22	0.60	2.77	3.61	6.10
F3063430	1378	24.47	59.080	F306	343	I	67.48	0.65	13.59	4.79	0.09	1.26	3.12	2.89	5.17	0.30	0.22	0.43	2.01	2.50	4.31
F3152675	1420	24.35	59.120	F315	267.5	I	67.91	0.88	13.29	5.06	0.10	1.34	3.17	2.60	4.42	0.27	0.30	0.66	1.99	2.76	4.55
F3152813	1421	24.35	59.120	F315	281.3	I	66.53	0.90	13.99	5.64	0.12	1.55	3.29	2.30	4.32	0.34	0.27	0.76	2.14	3.14	5.07
F3153087	1422	24.35	59.120	F315	308.7	I	64.91	0.94	13.96	5.48	0.13	1.51	3.46	3.00	5.33	0.32	0.30	0.68	2.31	2.84	4.93
F3032833	1368	24.61	59.120	F303	283.3	II	69.32	0.53	13.29	4.47	0.06	0.67	2.07	2.44	6.28	0.14	0.22	0.52	1.97	2.25	4.02
F3033190	1369	24.61	59.120	F303	319	II	67.54	0.66	13.80	5.46	0.08	0.84	2.65	2.50	5.27	0.19	0.22	0.78	2.25	2.89	4.92
F3033338	1370	24.61	59.120	F303	333.8	II	69.72	0.17	15.78	1.49	0.03	0.38	1.62	2.67	7.44	0.03	0.23	0.44	0.71	0.70	1.34
F3042790	1371	24.56	59.140	F304	279	II	69.91	0.26	13.14	3.64	0.06	0.50	1.62	2.65	7.40	0.04	0.22	0.56	1.74	1.71	3.28
F3043035	1372	24.56	59.140	F304	303.5	II	71.03	0.19	13.62	3.23	0.07	0.54	1.67	2.58	6.57	0.05	0.22	0.22	1.47	1.59	2.91
F3142442	1411	24.39	59.180	F314	244.2	II	72.50	0.34	11.81	3.31	0.05	0.50	1.50	2.37	6.47	0.03	0.22	0.90	1.49	1.64	2.98
F3142558	1412	24.39	59.180	F314	255.8	II	73.66	0.17	13.09	2.24	0.03	0.42	1.01	2.53	5.99	0.03	0.22	0.61	0.99	1.12	2.01
F3142676	1413	24.39	59.180	F314	267.6	II	72.09	0.34	12.10	3.73	0.06	0.46	1.50	2.38	6.48	0.02	0.22	0.60	1.67	1.85	3.36
F3142906	1415	24.39	59.180	F314	290.6	II	73.24	0.22	12.92	2.16	0.03	0.46	1.01	2.56	6.31	0.03	0.22	0.84	0.97	1.07	1.94
F3192560	1438	24.294	59.153	F319	256	II	66.02	0.60	14.36	4.47	0.07	1.26	2.31	2.76	6.37	0.13	0.22	1.43	1.99	2.23	4.03
F3193090	1439	24.294	59.153	F319	309	II	70.61	0.43	12.65	3.15	0.07	0.63	2.13	2.57	7.03	0.05	0.22	0.46	1.47	1.51	2.83
F3242760	1456	24.2	59.090	F324	276	II	66.63	0.73	13.22	6.55	0.13	1.11	3.24	2.61	4.62	0.37	0.25	0.53	2.60	3.56	5.90
F3243450	1457	24.2	59.090	F324	330	II	66.06	0.52	13.74	6.03	0.11	0.54	2.65	2.80	6.25	0.19	0.22	0.89	2.68	3.02	5.42
F3243510	1458	24.2	59.090	F324	331.2	II	68.33	0.56	12.75	6.10	0.11	0.89	3.11	2.51	4.76	0.30	0.22	0.36	2.44	3.29	5.49
F3222408	1449	24.14	59.220	F322	240.8	III	70.12	0.63	12.19	4.58	0.13	0.80	2.63	3.00	4.93	0.08	0.23	0.68	1.93	2.39	4.13
F3222707	1450	24.14	59.220	F322	270.7	III	70.16	0.37	13.89	2.60	0.07	0.79	1.99	3.23	6.19	0.04	0.22	0.46	1.20	1.26	2.34
F3222760	1451	24.14	59.220	F322	276	III	75.34	0.08	12.89	1.05	0.01	0.22	1.34	4.67	3.88	0.01	0.22	0.28	0.47	0.52	0.95
F3232892	1452	24.13	59.170	F323	289.2	III	72.92	0.14	13.04	2.47	0.03	0.45	1.72	3.66	4.62	0.02	0.23	0.69	1.07	1.26	2.23
F3233042	1453	24.13	59.170	F323	304.2	III	72.49	0.25	13.27	2.11	0.04	0.53	1.71	2.80	6.00	0.03	0.22	0.54	0.94	1.05	1.90
F3233310	1454	24.13	59.170	F323	331	III	72.45	0.25	13.83	2.28	0.07	0.36	1.48	3.04	5.66	0.04	0.23	0.31	1.02	1.14	2.06
F3233330	1455	24.13	59.170	F323	333	III	71.54	0.33	13.24	2.36	0.07	0.62	1.89	3.19	5.98	0.03	0.22	0.53	1.08	1.16	2.13
F3282580	1467	24.17	59.130	F328	258	III	70.33	0.40	13.28	2.96	0.06	0.62	2.06	3.47	6.21	0.05	0.22	0.34	1.38	1.42	2.66
F3282810	1468	24.17	59.130	F328	281	III	69.84	0.44	13.33	3.06	0.07	0.58	2.31	3.39	6.25	0.05	0.22	0.46	1.43	1.47	2.76
F3283100	1469	24.17	59.130	F328	310	III	70.74	0.42	12.64	3.34	0.10	1.07	2.26	2.94	5.79	0.04	0.23	0.43	1.48	1.67	3.00

Table 3 Continue next page

Id	Phase	Geochemical Relations												CIPW after Buckle et al. (2023) [129]							
		$Mg\#$	$Na_2O + K_2O$	K_2O/Na_2O	FeO/MgO	$FeO^*/(FeO^*+MgO)$	MALI	ASI	A/NK	A/CNK	DF1	DF2	$T_{SiO_2}[^\circ C]$	$T_{MgO}[^\circ C]$	Qtz	An	Ab	Or	Ilm	Mag	Ap
3023300	I	25.16	9.16	1.88	2.68	0.73	7.39	0.84	1.07	0.84	-3.16	-1.82	670.98	556.79	29.89	2.36	27.11	35.50	0.19	0.80	0.02
3023320	I	20.52	7.50	1.56	3.49	0.78	3.73	0.85	1.40	0.82	-1.25	-1.44	824.50	645.53	19.44	10.78	25.16	27.46	2.08	4.16	1.08
3023344	I	18.24	7.81	1.70	4.03	0.80	4.54	0.84	1.30	0.82	-1.73	-0.68	770.57	713.70	24.49	8.25	24.87	29.45	1.45	3.11	0.80
30233442	I	17.86	7.62	1.63	4.14	0.81	4.20	0.84	1.33	0.82	-1.60	-0.67	772.35	713.90	24.82	9.03	24.85	28.32	1.55	3.10	0.87
3024031	I	17.34	9.11	2.25	4.29	0.81	7.76	0.96	1.17	0.96	-3.17	-2.85	689.33	557.12	29.16	5.21	23.89	37.54	0.21	1.37	0.07
3024051	I	21.02	8.43	1.96	3.38	0.77	5.39	0.87	1.29	0.85	-1.90	-2.85	789.20	556.90	20.80	8.49	24.38	33.45	1.50	3.29	0.80
3024384	I	19.37	8.46	1.83	3.75	0.79	5.44	0.86	1.27	0.84	-1.83	-2.84	794.68	556.79	19.83	8.03	25.56	32.74	1.46	3.69	0.75
3024600	I	15.40	7.71	2.11	4.94	0.83	2.85	0.73	1.37	0.72	-1.82	-0.74	804.72	706.33	21.17	9.97	21.33	31.40	1.59	3.43	0.82
3024785	I	21.80	7.92	1.75	3.23	0.76	5.00	0.90	1.32	0.87	-1.62	-2.04	788.34	602.77	22.37	9.13	24.68	30.21	1.79	3.49	0.92
3024805	I	16.67	7.53	1.73	4.50	0.82	4.66	0.98	1.47	0.95	-1.65	-0.41	790.59	730.35	23.50	12.21	23.58	28.54	1.72	3.65	0.77
3024850	I	20.29	8.10	1.70	3.54	0.78	5.16	0.89	1.32	0.87	-1.83	-0.43	783.93	730.47	21.78	9.17	25.65	30.47	1.53	3.36	0.70
3024874	I	21.35	7.65	1.77	3.32	0.77	4.40	0.86	1.33	0.83	-1.61	-0.71	794.04	713.71	22.72	9.05	23.71	29.38	1.70	3.65	1.02
9v3250	I	18.66	6.91	1.79	3.92	0.80	3.25	0.87	1.47	0.84	-1.34	-0.41	809.23	730.83	24.03	11.59	21.40	26.80	1.91	3.95	1.09
9v3270	I	19.71	7.53	2.06	3.67	0.79	4.18	0.92	1.45	0.88	-1.73	-0.16	829.06	781.35	20.10	11.84	21.27	30.60	1.70	4.21	1.12
F3052780	I	13.45	8.84	2.43	5.79	0.85	6.65	0.89	1.20	0.88	-2.82	-0.88	737.24	699.32	25.53	6.14	21.95	37.32	0.71	2.79	0.28
F3063064	I	22.01	7.37	1.86	3.19	0.76	4.26	0.92	1.44	0.90	-1.66	-0.31	768.68	706.68	26.25	11.39	22.02	28.67	2.13	2.87	0.75
F3063270	I	16.73	7.85	1.51	4.48	0.82	4.14	0.86	1.37	0.83	-1.44	-0.60	820.82	725.50	18.86	10.53	26.76	28.27	2.08	4.06	1.15
F3063430	I	20.79	8.06	1.79	3.43	0.77	4.94	0.87	1.31	0.85	-1.88	-0.98	767.45	699.43	23.48	8.94	24.66	30.86	1.24	2.94	0.70
F3152675	I	20.94	7.02	1.70	3.40	0.77	3.85	0.92	1.47	0.90	-1.47	-0.35	761.35	706.81	27.74	11.71	22.27	26.44	1.69	2.92	0.63
F3152813	I	21.56	6.62	1.88	3.27	0.77	3.33	1.00	1.65	0.97	-1.36	-0.44	780.93	719.64	27.38	14.29	19.72	25.87	1.73	3.15	0.80
F3153087	I	21.60	8.32	1.78	3.27	0.77	4.87	0.84	1.31	0.82	-1.87	-0.40	803.90	725.36	19.00	9.02	25.70	31.89	1.81	3.40	0.75
F3032833	II	13.04	8.71	2.57	6.00	0.86	6.65	0.92	1.23	0.91	-2.78	-0.96	741.50	671.97	26.10	6.85	20.84	37.46	1.02	2.88	0.33
F3033190	II	13.31	7.77	2.11	5.86	0.85	5.12	0.96	1.41	0.94	-2.11	-0.72	766.60	691.33	25.87	11.03	21.41	31.59	1.27	3.31	0.45
F3033338	II	20.32	10.11	2.79	3.53	0.78	8.48	1.03	1.27	1.03	-3.68	-1.72	735.79	627.65	21.72	7.92	22.74	44.29	0.33	1.04	0.07
F3042790	II	12.05	10.05	2.79	6.57	0.87	8.44	0.86	1.06	0.86	-3.60	-0.77	733.10	671.83	22.87	2.12	22.67	44.16	0.50	2.54	0.09
F3043035	II	14.32	9.15	2.54	5.38	0.84	7.48	0.95	1.20	0.95	-3.21	-1.12	717.25	682.40	26.49	6.23	21.99	39.05	0.36	2.14	0.12
F3142442	II	13.09	8.84	2.72	5.97	0.86	7.34	0.87	1.08	0.87	-3.29	-0.08	696.44	659.86	30.90	2.51	20.36	38.71	0.65	2.18	0.07
F3142558	II	15.79	8.52	2.37	4.80	0.83	7.51	1.05	1.23	1.05	-3.17	-1.33	680.01	627.43	33.16	4.86	21.60	35.77	0.33	1.45	0.07
F3142676	II	10.98	8.87	2.72	7.30	0.88	7.37	0.89	1.11	0.89	-3.36	0.35	702.14	672.06	30.07	3.20	20.37	38.72	0.65	2.45	0.05
F3142906	II	17.56	8.87	2.47	4.23	0.81	7.86	1.01	1.17	1.00	-3.30	-1.06	685.90	627.40	31.44	4.87	21.90	37.74	0.42	1.43	0.07
F3192560	II	21.95	9.13	2.31	3.20	0.76	6.82	0.93	1.26	0.92	-2.65	-0.77	788.09	682.24	19.37	8.15	23.77	38.38	1.16	2.95	0.31
F3193090	II	16.67	9.60	2.74	4.50	0.82	7.47	0.81	1.07	0.81	-3.33	-0.14	723.12	682.27	24.90	2.23	21.92	41.92	0.82	2.14	0.12
F3242760	II	14.53	7.23	1.77	5.29	0.84	3.98	0.90	1.42	0.87	-1.57	-0.65	779.47	725.67	25.61	10.84	22.34	27.61	1.41	3.81	0.87
F3243450	II	8.21	9.05	2.23	10.06	0.91	6.41	0.86	1.21	0.85	-2.65	-0.86	787.62	713.40	20.14	6.56	24.08	37.47	1.00	3.94	0.45
F3243510	II	12.73	7.27	1.90	6.17	0.86	4.16	0.88	1.37	0.85	-1.76	-0.86	755.45	713.50	28.00	9.58	21.43	28.39	1.07	3.57	0.70
F3222408	III	14.93	7.92	1.64	5.13	0.84	5.29	0.81	1.19	0.81	-2.10	0.61	730.17	725.80	28.22	5.31	25.65	29.46	1.22	2.83	0.19
F3222707	III	23.32	9.42	1.92	2.96	0.75	7.43	0.89	1.16	0.89	-2.91	-0.18	729.55	681.61	23.14	5.17	27.51	36.90	0.70	1.75	0.09
F3222760	III	17.32	8.55	0.83	4.29	0.81	7.21	0.90	1.08	0.90	-2.17	-1.90	656.18	557.06	30.94	2.77	39.75	23.07	0.15	0.68	0.02
F3232892	III	15.46	8.28	1.26	4.92	0.83	6.56	0.92	1.18	0.92	-2.32	-1.10	690.41	627.81	30.02	5.55	31.29	27.62	0.27	1.57	0.05
F3233042	III	20.08	8.80	2.14	3.58	0.78	7.09	0.94	1.20	0.93	-2.99	-0.68	696.60	645.62	29.25	5.98	23.91	35.79	0.48	1.38	0.07
F3233310	III	13.64	8.70	1.86	5.70	0.85	7.21	1.00	1.24	1.00	-2.87	-0.53	697.15	682.45	29.28	7.14	25.85	33.67	0.48	1.48	0.09
F3233330	III	20.74	9.18	1.88	3.44	0.77	7.29	0.88	1.13	0.87	-2.93	0.05	710.02	682.16	26.14	4.15	27.25	35.68	0.63	1.58	0.07
F3282580	III	17.22	9.69	1.79	4.32	0.81	7.63	0.82	1.07	0.82	-2.90	-0.38	727.12	671.53	22.53	2.31	29.62	36.99	0.76	2.02	0.12
F3282810	III	15.89	9.64	1.84	4.76	0.83	7.32	0.81	1.08	0.81	-2.89	-0.15	734.12	682.18	22.11	2.72	28.95	37.22	0.84	2.09	0.12
F3283100	III	24.32	8.73	1.97	2.80	0.74	6.46	0.83	1.14	0.83	-2.68	0.44	721.26	706.96	26.24	4.26	25.05	34.51	0.81	2.16	0.09

Table 3 Continue next page

Table 3: The major elemental data [wt.%] sourced from the Kivisilla et al. (1999) [116] dataset are organised by location, drill core, and the tripartite phases, as illustrated in Figure 4c. The data is normalised to 100%. The values for Fe_2O_3 and FeO were derived from $\text{Fe}_2\text{O}_{3\text{tot}}$ following the method outlined by Le Maitre (1976) [130]. The calculation used for plutonic rocks is as follows: $\text{FeO}^* = \text{Fe}_2\text{O}_3(0.8998) + \text{FeO}$. The analysis of major elemental relationships in Tripartite rapakivi samples, including temperature, pressure and CIPW. $Mg\#_{[\text{wt.\%}]} = 100 \cdot \frac{MgO}{MgO+FeO^*}$; $MALI_{[\text{wt.\%}]} = (\text{Na}_2\text{O} + \text{K}_2\text{O}) - \text{CaO}$; $ASI_{[\text{mol}]} = \frac{\text{Al}_2\text{O}_3}{\text{CaO}+\text{Na}_2\text{O}+\text{K}_2\text{O}}$; $A/NK_{[\text{mol}]} = \frac{\text{Al}_2\text{O}_3}{\text{Na}_2\text{O}+\text{K}_2\text{O}}$. Multi-dimensional discriminant functions DF1 and DF2, based on tectonic analysis, after Verma et al. (2013) [131]. $DF1 = 0.051 \ln \frac{\text{TiO}_2}{\text{SiO}_2} + 0.226 \ln \frac{\text{Al}_2\text{O}_3}{\text{SiO}_2} - 1.77 \ln \frac{\text{Fe}_2\text{O}_3}{\text{SiO}_2} + 1.83 \ln \frac{\text{FeO}}{\text{SiO}_2} - 0.065 \ln \frac{\text{MnO}}{\text{SiO}_2} + 0.134 \ln \frac{\text{MgO}}{\text{SiO}_2} + 0.225 \ln \frac{\text{CaO}}{\text{SiO}_2} + 0.742 \ln \frac{\text{Na}_2\text{O}}{\text{SiO}_2} - 1.78 \ln \frac{\text{K}_2\text{O}}{\text{SiO}_2} + 0.146 \ln \frac{\text{P}_2\text{O}_5}{\text{SiO}_2} - 2.12$. & $DF2 = 1.09 \ln \frac{\text{TiO}_2}{\text{SiO}_2} - 1.65 \ln \frac{\text{Al}_2\text{O}_3}{\text{SiO}_2} - 1.19 \ln \frac{\text{Fe}_2\text{O}_3}{\text{SiO}_2} + 1.03 \ln \frac{\text{FeO}}{\text{SiO}_2} + 0.82 \ln \frac{\text{MnO}}{\text{SiO}_2} + 0.026 \ln \frac{\text{MgO}}{\text{SiO}_2} + 0.023 \ln \frac{\text{CaO}}{\text{SiO}_2} + 0.212 \ln \frac{\text{Na}_2\text{O}}{\text{SiO}_2} + 0.085 \ln \frac{\text{K}_2\text{O}}{\text{SiO}_2} - 0.85 \ln \frac{\text{P}_2\text{O}_5}{\text{SiO}_2} + 2.54$. CIPW Abbreviations: Qtz=Quartz, An=Anorthite, Ab=Albite, Or=Orthoclase, Ilm=Ilmenite, Mag=Magnetite, Ap=Apatite.

4.2 Geophysical Modelling

4.2.1 Overview and Software Workflow

This study employs a multidisciplinary approach—integrating geophysical modelling, geochemical analysis, and tectonic interpretation—to investigate the structure and evolution of the Estonian Precambrian basement. It draws on global satellite potential fields, national EGT Bouguer and magnetic surveys (*Publications I & III*), and new geochemical data from Alutaguse drill cores, alongside legacy datasets for Alutaguse and SS (*Publication II & Manuscript II*) and the Märjamaa–Kloostri rapakivi bodies (*Manuscript I*). Data processing used Oasis Montaj for potential field filtering (*Publications I & III*), ArcMap v10.8 for spatial analysis and figures, and SimPEG for geophysical inversion (*Publication III*). Geochemical statistics and plotting were performed using Excel and Python modules (*Publication II & Manuscripts IV–VI*).

4.2.2 Geophysical Preprocessing and Filtering

All datasets were projected to UTM Zone 34N and processed using Oasis Montaj (Geosoft) and ArcMap v10.8. Magnetic data were reduced to the pole (RTP) to correct for inclination and declination effects and improve source–anomaly alignment [132, 133]. Spectral wavelength filtering, including Fourier-based phase-match high- and low-pass filters, was applied to the Complete Bouguer and RTP grids to separate regional and residual components, emphasizing short-wavelength anomalies related to upper-crustal sources [134].

4.2.3 Derivative Attributes and Structural Edges

To enhance structural interpretation, derivative filters were applied to the residual Bouguer and RTP datasets [135, 136]. Specifically, Tilt Derivative Ratio (TDR), Tilt Angle Derivative (TDX), the Horizontal Derivative of TDR (HDTDR), and Analytic Signal Derivative (AS) were calculated over the residual part as follows:

$$\text{TDR} = \tan^{-1} \left(\frac{\partial f / \partial z}{\sqrt{(\partial f / \partial x)^2 + (\partial f / \partial y)^2}} \right),$$

$$\text{TDX} = \tan^{-1} \left(\frac{\sqrt{(\partial f / \partial x)^2 + (\partial f / \partial y)^2}}{\partial f / \partial z} \right),$$

$$\text{HDTDR} = \sqrt{\left(\frac{\partial \text{TDR}}{\partial x}\right)^2 + \left(\frac{\partial \text{TDR}}{\partial y}\right)^2},$$

$$\text{AS} = \sqrt{\left(\frac{\partial f}{\partial x}\right)^2 + \left(\frac{\partial f}{\partial y}\right)^2 + \left(\frac{\partial f}{\partial z}\right)^2}.$$

where $\partial f/\partial x$, $\partial f/\partial y$, and $\partial f/\partial z$ represent the first derivatives of the potential field in the x , y , and z directions, respectively.

Lineament mapping identifies structural features from the intersection of sub-planar surfaces [137–139]. Geophysical lineaments were extracted from residual derivative grids using hillshade techniques to highlight trends [135, 137] (*Publications I & III*). Eight hillshaded raster images per grid were created in ArcMap 10.8, illuminated from various azimuths ($0^\circ, 45^\circ, 90^\circ, 135^\circ, 180^\circ, 225^\circ, 270^\circ, 315^\circ$) at a 45° solar elevation. These were combined into two composite grids: Combination 1 ($0^\circ, 45^\circ, 90^\circ, 135^\circ$) and Combination 2 ($180^\circ, 225^\circ, 270^\circ, 315^\circ$) [135, 137]. Lineaments were extracted using the *LINE* module in PCI-Geomatica 2012 and analysed in ArcGIS 10.8 using the *COGO* tool.

4.2.4 3D Joint Inversion

To constrain the 3D geometry of subsurface bodies, cross-gradient joint inversion of gravity and magnetic anomalies was performed using SimPEG ([Cross-gradient Joint Inversion tutorial](#)). This approach integrated anomaly distributions with petrophysical contrasts to model Rapakivi intrusion geometries (*Publication III*) and metalliferous domains in the Alutaguse region (*Manuscript III*).

4.3 Geochemical Analysis

4.3.1 Alutaguse and SS Data Integration and Classification Procedures

For the major-element dataset of Alutaguse and SS metasedimentary and metavolcanic samples, oxide totals were normalised to 100 *wt. %* with LOI excluded, as GTK samples do not report LOI [128] (*Manuscript II*). However, unnormalised LOI values were incorporated for those samples in which they were available, when relevant.

Metasedimentary samples were categorised by SiO_2 content following Verma and Armstrong-Altrin (2013) [140] classification, where High- SiO_2 defined as $> 63\text{wt. %}$ and Low- SiO_2 as $\leq 63\text{wt. %}$. Although a few lithologies straddle both groups, most show consistent silica-based distributions (Table 1; Fig. 6a1). Lithologies predominantly assigned to the High- SiO_2 group include: biotite gneisses ($\pm \text{Grt} \pm \text{Crld} \pm \text{Sil}$), garnet-bearing mica gneisses ($\pm \text{Crld} \pm \text{Sil}$), cordierite-bearing mica gneisses ($\pm \text{Grt} \pm \text{Sil}$), graphite-bearing mica gneisses ($\pm \text{Grt} \pm \text{Crld} \pm \text{Sil}$), biotite-plagioclase gneisses ($\pm \text{Amph}$), and biotite-amphibole gneisses. The Low- SiO_2 group includes similar lithologies but with lower average silica contents, majoritally conformed by graphite-bearing mica gneisses ($\pm \text{Grt} \pm \text{Crld} \pm \text{Sil}$), garnet-bearing mica gneisses ($\pm \text{Amph}$), and biotite-amphibole gneisses.

As the third dataset (from Uljaste; Fig. 2c) includes only trace elements and lacks major oxides, samples were assigned to hypothetical High- or Low- SiO_2 groups based on the average $\bar{\chi}_{\text{SiO}_2}$ values of each lithology. This classification uses silica distribution patterns established from the major elemental silica of the samples (Fig. 6a1; Table 2).

The SS dataset was processed using the same procedure to enable direct comparison (Table 1; Fig. 6a3). SS metasedimentary samples were grouped by both High/Low SiO_2 content and domain (Uusimaa, Häme, Saimaa, and Ladoga). Lithologies include micaceous gneisses ($\pm \text{Grt} \pm \text{Crld} \pm \text{Sil}$) [128], and quartz-micaceous gneisses from Ladoga [64]. All SS samples are of Svecofennian age [64, 67, 128], and none reported graphite.

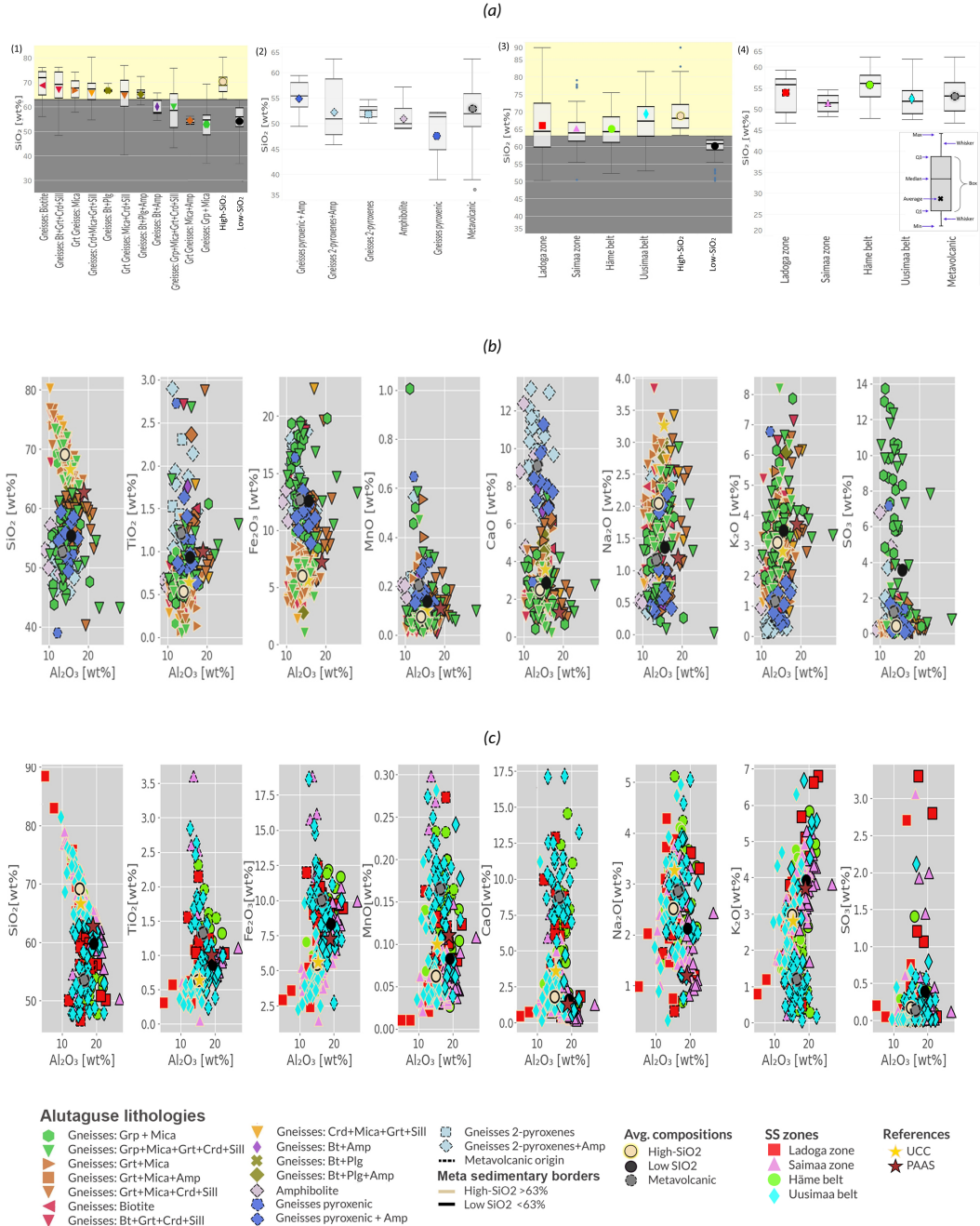


Figure 6: Boxplot and Harker diagram summary of metasedimentary and metavolcanic samples from the Alutaguse and South Svecofennian (SS) domains. (a) Boxplot of SiO_2 concentrations by lithological group for Alutaguse: (1) metasedimentary and (2) metavolcanic rocks; and for SS: (1) metasedimentary and (2) metavolcanic units.; (b) Bivariate Harker diagrams of Alutaguse samples and (c) SS samples, plotting major oxides vs. Al_2O_3 . Groups are distinguished as follows: High- SiO_2 (light brown), Low- SiO_2 (black), and metavolcanics (dashed grey).

No further subdivisions were applied to metavolcanic samples (Figs. 6a2,a4). Alutaguse metavolcanics include amphibolites, pyroxene gneisses ($\pm$ Amph), and two-pyroxene gneisses ($\pm$ Amph). In the SS dataset, metavolcanic samples correspond to amphibolites, hornblende gneisses, and mafic volcanites [128], as well as biotite-clinopyroxene and amphibole schists from the Ladoga domain [67].

4.3.2 Processing Major-Element Data: Märjamaa–Kloostri Rapakivi Granites

Figure 7 shows the major element distribution of the samples per emplacement phase. The granitic samples were analysed using the updated CIPW normative calculation implemented in the Python-based *webNORM* application [129], available at <https://webnorm.streamlit.app/> (*Manuscript I*). This tool builds on the algorithm of Verma et al. (2003) [141] and applies the Fe-correction method for plutonic rocks recommended by Le Maitre (1976) [130], which is widely used for igneous rock classification. CIPW norms, however, represent an anhydrous, idealised mineral assemblage and do not account for hydrous phases (e.g., biotite, hornblende) that are present in the Märjamaa–Kloostri rapakivi granites [95, 96]. Consequently, while normative mineral proportions (e.g., quartz, feldspar, Fe–Ti oxides) provide a useful framework for comparing bulk compositional trends and classification, CIPW phases such as pyroxenes and corundum are excluded, as they are not part of the typical rapakivi mineral assemblage [142, 143]. They cannot substitute for measured mineral chemistry when interpreting crystallisation conditions or magma evolution. For this reason, the normative data presented here are used exclusively to describe bulk chemical trends and feldspar-quartz-oxide proportions, without attempting to derive thermobarometric parameters (P – T – H_2O) or detailed fractionation pathways, which require mineralogical analyses [95, 142, 144].

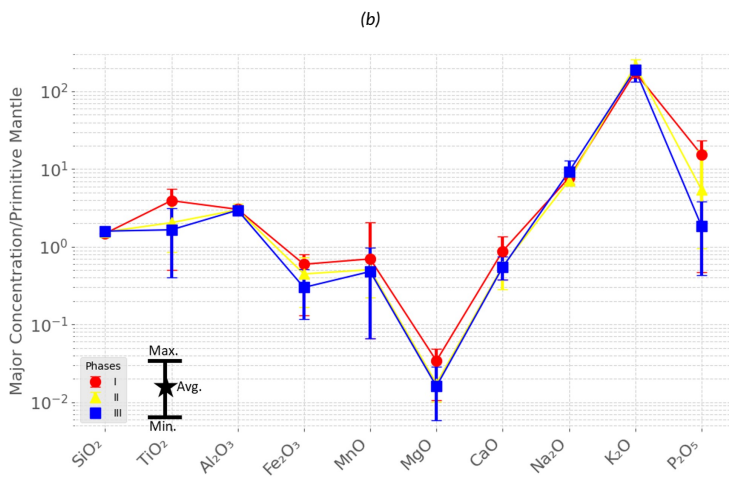
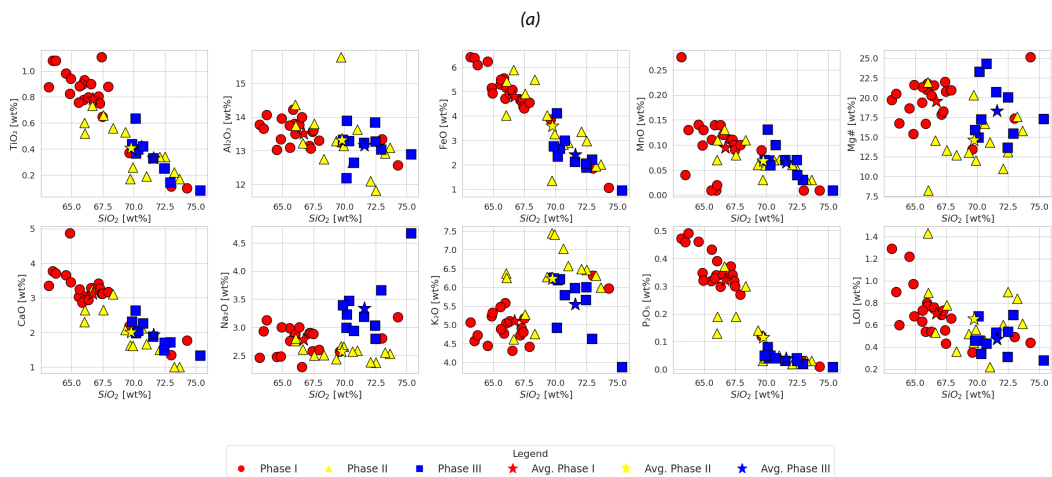


Figure 7: Major element distributions of samples from Kivisilla et al. (1999) [116], grouped into the three phases of the analysed rapakivi granites. The data are shown as (a) Harker bi-plots comparing SiO_2 with other major elements and (b) spider plot normalized to primitive mantle [145].

5 Results and Discussion

5.1 Estonian Basement Architecture

Results from the analysis of WGM-12 Bouguer gravity and EMAG2v3 RTP magnetic (*Publication I*) show Moho depths of 45–73 km (mean 60 km; Fig.2d1) and a Conrad at 12–21 km (mean 18 km; Fig.2d2) [124, 125]. Magnetic spectral analysis gives Curie Point Depths (CPD) of 6–26 km (mean 15 km; Fig.2d3) from EMAG2v3 [126, 127]. The shallowest CPDs and highest heat flow occur along the PPDZ and near the Märjamaa–Kloostri intrusions, indicating localized lithospheric thinning. Residual Bouguer and RTP highs align with known Cu–Pb–Zn and Fe occurrences in the Tallinn–Alutaguse–Jõhvi divisions, coinciding with strong gravity and magnetic highs (Fig.2d5,d7).

Residual derivatives from WGM-12 and EMAG2v3 consistently pick out NW–SE lineaments that match mapped basement faults and the direction of shear-wave fast polarisation, confirming a tectonic origin [93]. Seismicity (M 1.2–4.5) is upper crustal and mainly dextral strike-slip on NW–SE to NNW–SSE subvertical faults [31], reflecting post-Svecofennian reactivation comparable to 1.85–1.84 Ga shear corridors in southern Finland (Åland–Turku) [8, 14, 24, 146]. Rose diagrams show a dominant NW–SE fabric (Fig.8c), supporting SO block amalgamation followed by local extension and lithospheric weakening [8, 10]. High heat-flow zones (High tones) cluster where lineaments are dense, especially near Märjamaa and the PPDZ (Figs.8d,e). Southern Estonia couples higher heat flow with a shallower Moho (Fig.2d1), whereas the Tallinn zone shows lower heat flow and deeper Moho values. Alutaguse and Jõhvi Moho values are the lowest, maybe related to the back-arc structure of this part of the basement.

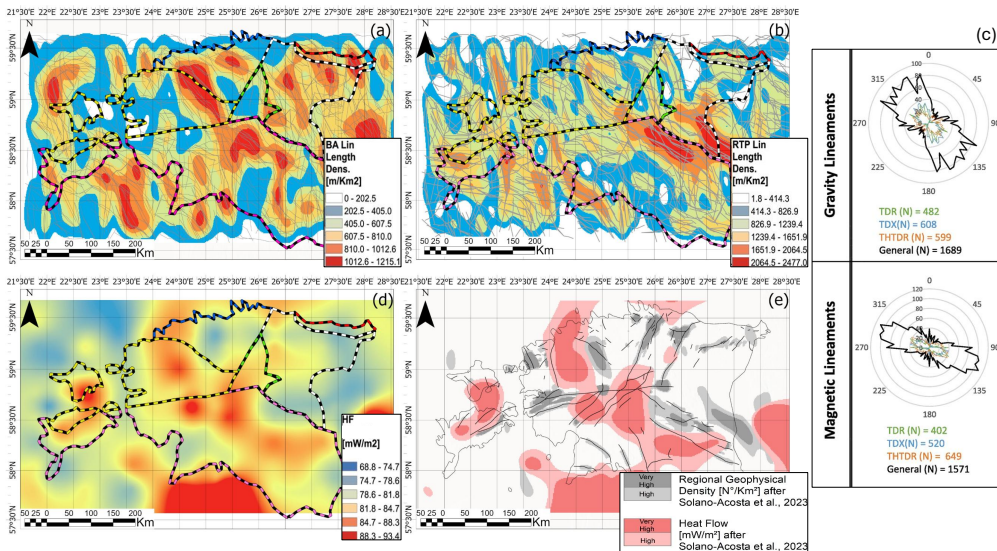


Figure 8: Density maps of the lineaments obtained from the residual part of the potential models. (a) Length density map of the gravimetric lineaments.; (b) Length density map of the magnetic lineaments.; (c) Rose diagrams from potential residual derivatives; (d) Heat flow (HF).; (e) A comparison between the zones displaying the highest HF values and those exhibiting the highest density of tectonic features (i.e., Lineaments and basement faults).

5.2 Relationships Between the Alutaguse Zone and the South Svecofennian (SS) Region

Publication II and Manuscript II integrate new and compiled geochemical datasets to characterise the lithological diversity and tectonic evolution of the Alutaguse zone within the context of the SO. Al_2O_3 -based Harker diagrams reveal distinct compositional trends for Alutaguse (Fig. 6b) and the SS domains (Fig. 6c). High- SiO_2 metasediments from both regions cluster near the UCC reference, whereas Low- SiO_2 samples align more closely with PAAS. Major elemental data analysed per domain enhance their regional trends (Fig. 9a).

Several weathering indices were calculated for the metasedimentary and metavolcanic samples. Among these, the Alutaguse and Saimaa samples exhibit the highest mean degrees of weathering (Fig. 9b; Table 1), whereas the Ladoga, Häme, and Uusimaa samples appear to be largely chemically fresh. Alutaguse and SS metasediments yield average values indicative of moderate weathering ($\text{CIA} < 80$) and chemical immaturity ($\text{ICV} > 1$, $\text{SiO}_2 < \text{Al}_2\text{O}_3$; Table 1), making them appropriate for provenance and tectonic interpretations. $\bar{\text{LOI}}$ values in Alutaguse metavolcanics are generally low, suggesting limited alteration. LOI data are not available for SS metavolcanics [67, 128], although still exhibiting lower metavolcanic index values than Alutaguse (e.g., Al-CCPI-SI , Table 1). Together, these trends suggest a minor open-system overprint, implying that the measured geochemical signatures predominantly preserve the original magmatic compositions [147–149].

Across the High- SiO_2 group, Alutaguse forms the most mafic-rich and metal-enriched pole, with higher Fe_2O_3 , MgO , SO_3 and K_2O , lower Al_2O_3 and Na_2O (Table 1), stronger transition-metal enrichment (V, Cr, Ni, Cu, Zn), higher Pb (Fig. 9c), Ba and Sr, and the highest ΣREE average with the lowest HREE/LREE (Table 2). CaO and MnO are significantly high in Alutaguse and Uusimaa (Fig. 9a), which is relatively felsic, with higher SiO_2 averages, less Fe–Mg–S enrichment, and stronger HREE and classical Eu anomalies, defining the felsic arc member. Häme and Saimaa High- SiO_2 samples display higher Al_2O_3 , Na_2O , and TiO_2 , lower Fe–Mg– SO_3 , higher HREE and stronger Eu anomalies, and less extreme LREE dominance. Ladoga shows the widest SiO_2 range, lowest Al_2O_3 and ΣREE , quartz-rich compositions, and lower metals and Ba–Sr than Alutaguse, forming the most quartz-rich and REE-poor end-member.

In the Low- SiO_2 group, the contrasts intensify. Alutaguse has the lowest SiO_2 and Al_2O_3 , the highest Fe_2O_3 , MgO , SO_3 , and the broadest ranges, with high CaO and MnO (shared only with Uusimaa) (Fig. 9a), very low Na_2O , extreme maxima in transition metals and Pb (Fig. 9c), high ΣREE and strong LREE dominance with low HREE/LREE. Uusimaa and Saimaa Low- SiO_2 samples are more siliceous, have higher Al_2O_3 , higher Zr–Hf–Y, and more balanced REE patterns with stronger Eu anomalies (Fig. 9d). Low- SiO_2 Häme, Saimaa, and Ladoga samples do not reach the Alutaguse Fe–Mg–S–metal enrichments, instead remaining more felsic, less metal-rich, and more compositionally balanced.

Metasedimentary tectonic discrimination diagrams based on discriminant functions effectively separate arc, rift, and collision settings [140, 150]. High- SiO_2 samples from Alutaguse and the SS mainly plot in arc domains (Figs. 10a1,a2), whereas Low- SiO_2 samples group in rift settings (Figs. 10b1,b2). Stable trace elements (Th, Sc, La, Zr) help identify source rocks and tectonic regimes [151–153]. Triangular plots (Th–La–Sc, Sc–Th–Zr/10) place most samples in the continental island-arc field (Figs. 10c1,c2), supporting derivation from active continental margin environments [154].

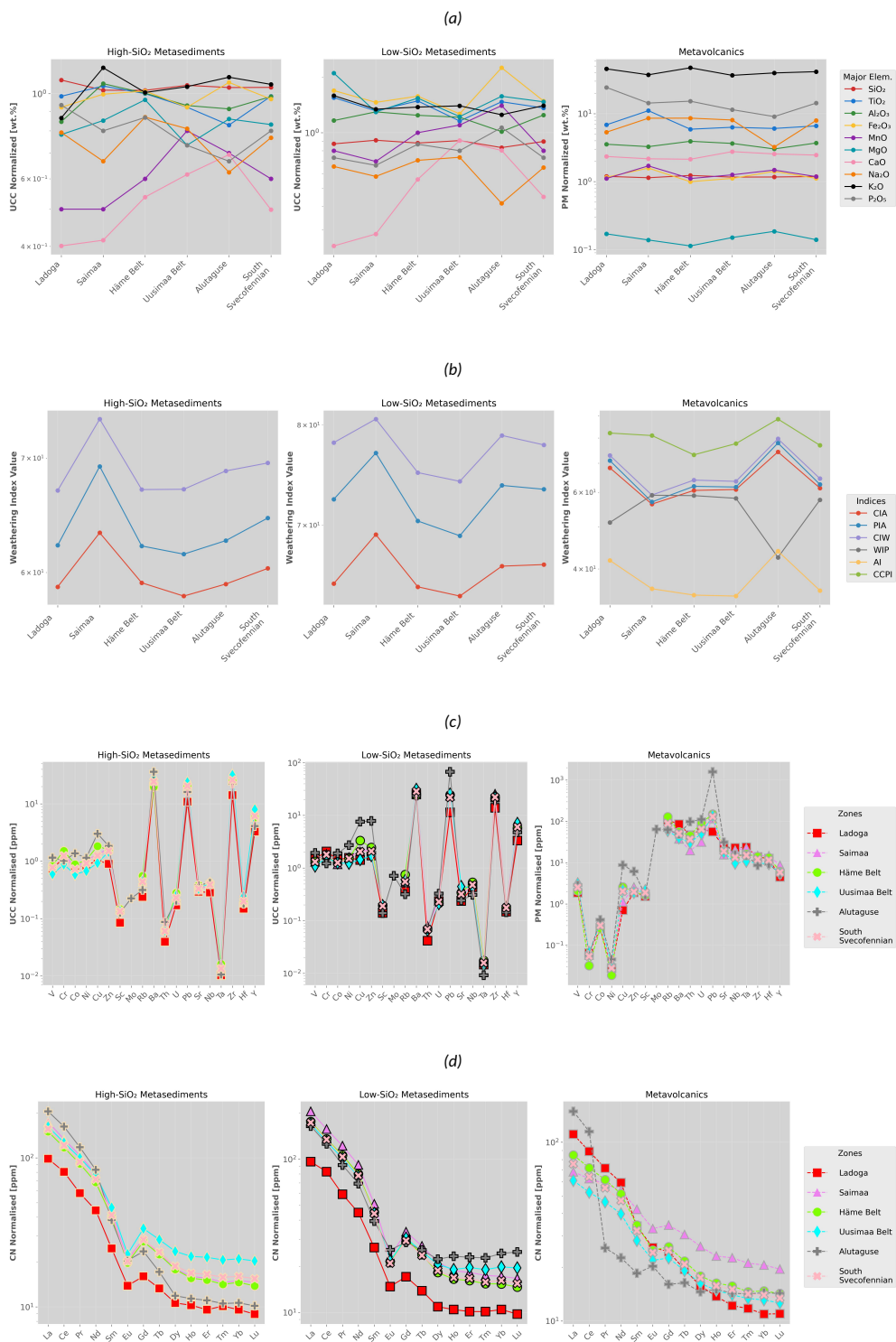


Figure 9: Bivariant average geochemical trends of the Alutaguse and South Svecofennian (SS) zones: (a) Major elemental data and (b) Weathering indices, both as shown in Table 1, (c) Trace elements, and (d) Rare Earth Elements (REE) as detailed in Table 2.

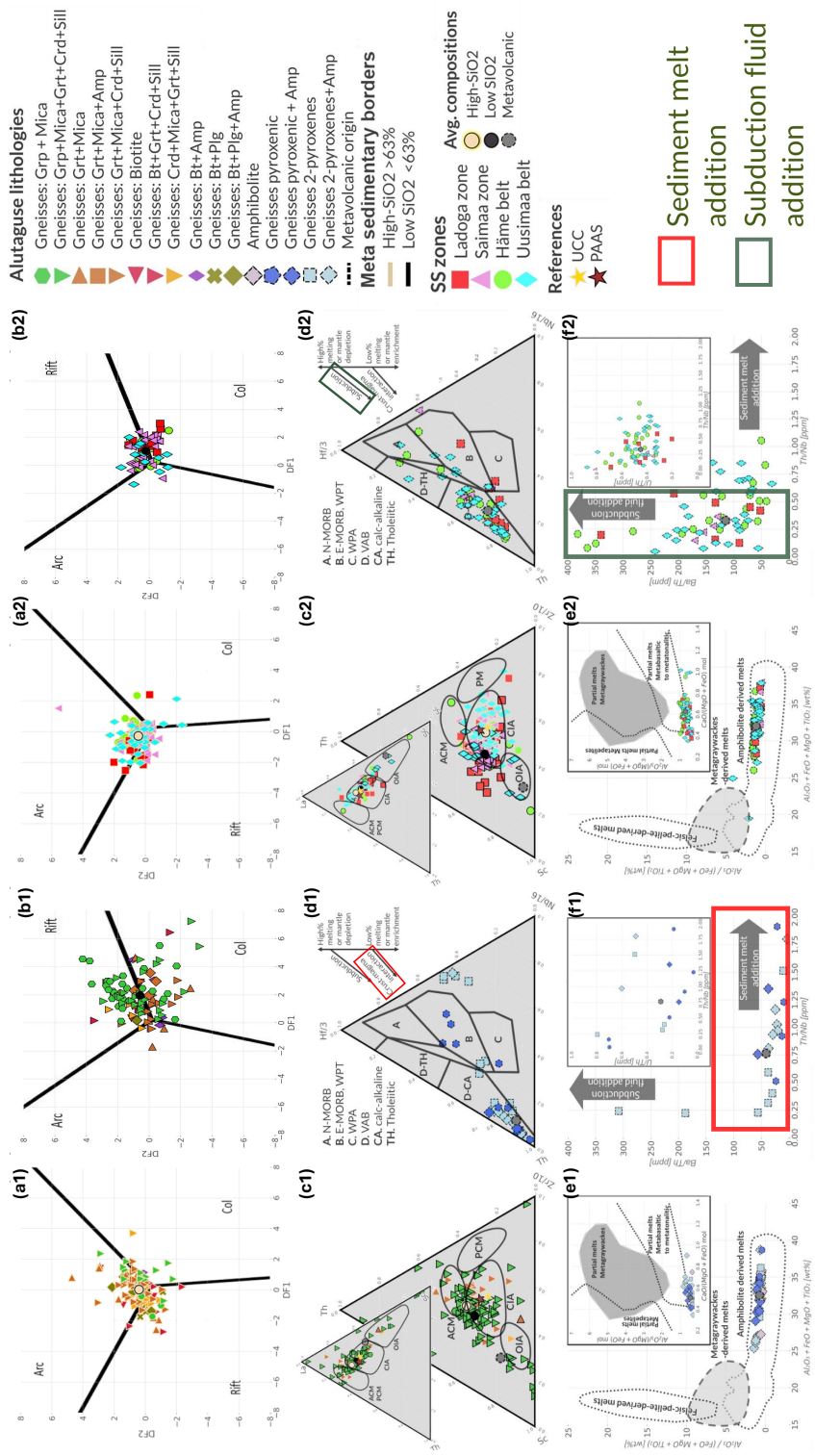


Figure 10: Geochemical discrimination diagrams comparing (1) Alutaguse and (2) Southern Svecofennian (SS) zones: (a1–a2) Discriminant Function plots (DF1 vs. DF2) for High-SiO₂ metasediments; (b1–b2) DF1 vs. DF2 for Low-SiO₂ metasediments, both after Verma and Armstrong-Altrin (2013) [140].; (c1–c2) Ternary diagrams showing La–Th–Sc (inset) and Th–Sc–Zr/10 (background) after Bhatia and Crook (1986) [151].; (d1–d2) Hf/3–Th–Nb/16 (after Wood 1980 [155]) triangular plots for metavolcanic units.; (e1–e2) Binary plots of Al₂O₃/(FeO + MgO + TiO₂) vs. Al₂O₃+FeO + MgO + TiO₂, after Patiño Douce (1999) [156], with inset showing Al/(Mg+Fe)_[mol] vs. Ca/(Mg+Fe)_[mol] discrimination from Altherr et al. (2000) [157].; (f1–f2) Ba/Th and U/Th vs. Th/Nb after Saccani et al. (2018) [158] and Ye et al. (2018) [159], highlight subduction influence in SS samples and the dominance of sediment-derived melts in Alutaguse, consistent with a back-arc rifting setting.

In the metavolcanics, Alutaguse samples show the clearest separation, with higher Fe₂O₃, MgO, CaO and SO₃, lower Al₂O₃, Na₂O, K₂O and P₂O₅, the highest transition-metal and Pb concentrations of all domains (Table 2), and strong LREE dominance with low HREE (Fig.9d). The Uusimaa–Häme–Saimaa metavolcanics exhibit more moderate Fe–Mg–S, higher Al–Na–K, and higher Zr–Hf–Y ratios (Table 2; Fig.9c), with stronger HREE contributions—particularly in Saimaa and Häme—and REE patterns showing LREE-enrichment in Ladoga metavolcanics and HREE-enrichment in Saimaa. Though CaO and MnO are highest in the Alutaguse and Uusimaa metavolcanic systems (Table 1; Fig.6). The Alutaguse positive Eu anomaly may relate to high-temperature hydrothermal processes under reducing conditions, although additional samples are required to confirm this interpretation (Fig.9d; Table 2).

Metavolcanic tectonic discrimination diagrams (e.g., Hf/3–Th–Nb/16) confirm compressional arc settings for all metavolcanics (Figs.10d1,d2). Amphibolite-derived melt signatures dominate the metavolcanic datasets, supported by oxide systematics (Al/(Mg+Fe) vs. Ca/(Mg+Fe)) and experimental petrology [156, 157] (Figs.10e1,e2).

The metavolcanic rocks are dominantly tholeiitic (high Fe and Ti) and metaluminous (A/CNK < 1) (Table 1). Alutaguse and Uusimaa samples show low La/Yb and high Zr/Nb ratios, consistent with partial melting of a depleted asthenospheric mantle source, whereas Ladoga metavolcanics show higher La/Yb and lower Zr/Nb (Table 2), suggesting a deeper melt source with slab-derived input (*Manuscript II*). Alutaguse is further distinguished by elevated Th/Nb and depleted Ba/Th and U/Th, indicative of asthenospheric control (Fig.10f1). In contrast, the SS domains exhibit higher Ba/Th, U/Th, and Nb/Zr, reflecting subduction sources (Fig.10f2).

5.2.1 Geodynamic Implication

The evolution of the SO has been interpreted through three principal models. One model proposes 1.90–1.88 Ga accretion of arc complexes to the Karelian craton, marking micro-continent–continent amalgamation [8, 10, 37, 53, 58, 100, 160]. A second model, based on tectonic switching theory, invokes successive subduction episodes and arc migration [8, 37, 53, 101, 102, 105]. Both frameworks imply widespread 1.92–1.87 Ga arc magmatism extending from the Skellefte district (Sweden) to the Finnish Svecofennian belts (Fig.2b). A third model emphasises pre-1.91 Ga evolution, interpreting the SO as a single long-lived arc with continuous 1.90–1.88 Ga magmatism forming the CFAC and SS domains, followed by oroclinal buckling at 1.87–1.86 Ga [27, 51]. These interpretations converge into two end-member scenarios: multiple arcs generated by double subduction and rifting, or a single linear arc later deformed. The pronounced geochemical contrasts between the SS and Alutaguse metasedimentary and metavolcanic units (Tables 1 & 2; Fig.9) support the multiple arc collision explanation, indicating that these belts represent distinct

tectonic units.

In this research, the Svecofennian evolution recorded across the Alutaguse–(Tallinn)–Uusimaa–Southern Svecofennian domains is best explained by a double-subduction arc-back-arc model, rather than by a single long-lived linear arc. As synthesised in Figures 11a,b 1–5, early convergence at ~ 1.92 – 1.90 Ga initiated arc magmatism along both Karelian- and Bergslagen-facing margins, with earlier juvenile arc development in the Uusimaa–North Estonian sector (Fig. 11a,b 1). Subsequent slab retreat and asymmetric convergence promoted back-arc rifting by ~ 1.90 – 1.89 Ga, generating mantle-enriched metavolcanics and immature sedimentation in Alutaguse and Uusimaa, while arc magmatism persisted within the CFAC belts (Fig. 11a,b 2). Progressive collision between the Bothnian, Keitele, and Bergslagen microcontinents at 1.89 – 1.87 Ga resulted in lithospheric thickening, delamination, renewed magmatism, and oroclinal tightening of the SS composite arc system (e.g., Häme belt) (Fig. 11a,b 3). Final suturing at ~ 1.87 – 1.86 Ga closed the remaining oceanic tracts and was followed by a short-lived extensional pulse marked by within-plate magmatism and mantle upwelling [53, 102] (Fig. 11a,b 4). Subsequent low pressure and high temperature metamorphism and late Svecofennian granite-migmatite (LSGM) development between 1.85 and 1.79 Ga reflect post-collisional collapse and reactivation of inherited extensional structures (Fig. 11a,b 5). The coherent Ca–Mn–Mg geochemical signatures shared by Alutaguse and Uusimaa metasediments and metavolcanics, together with their contrasted behaviour relative to Ladoga, Saimaa, and Häme, place Alutaguse within a Bergslagen-facing back-arc system genetically linked to the Uusimaa–Tallinn belt, rather than with the other SS zones.

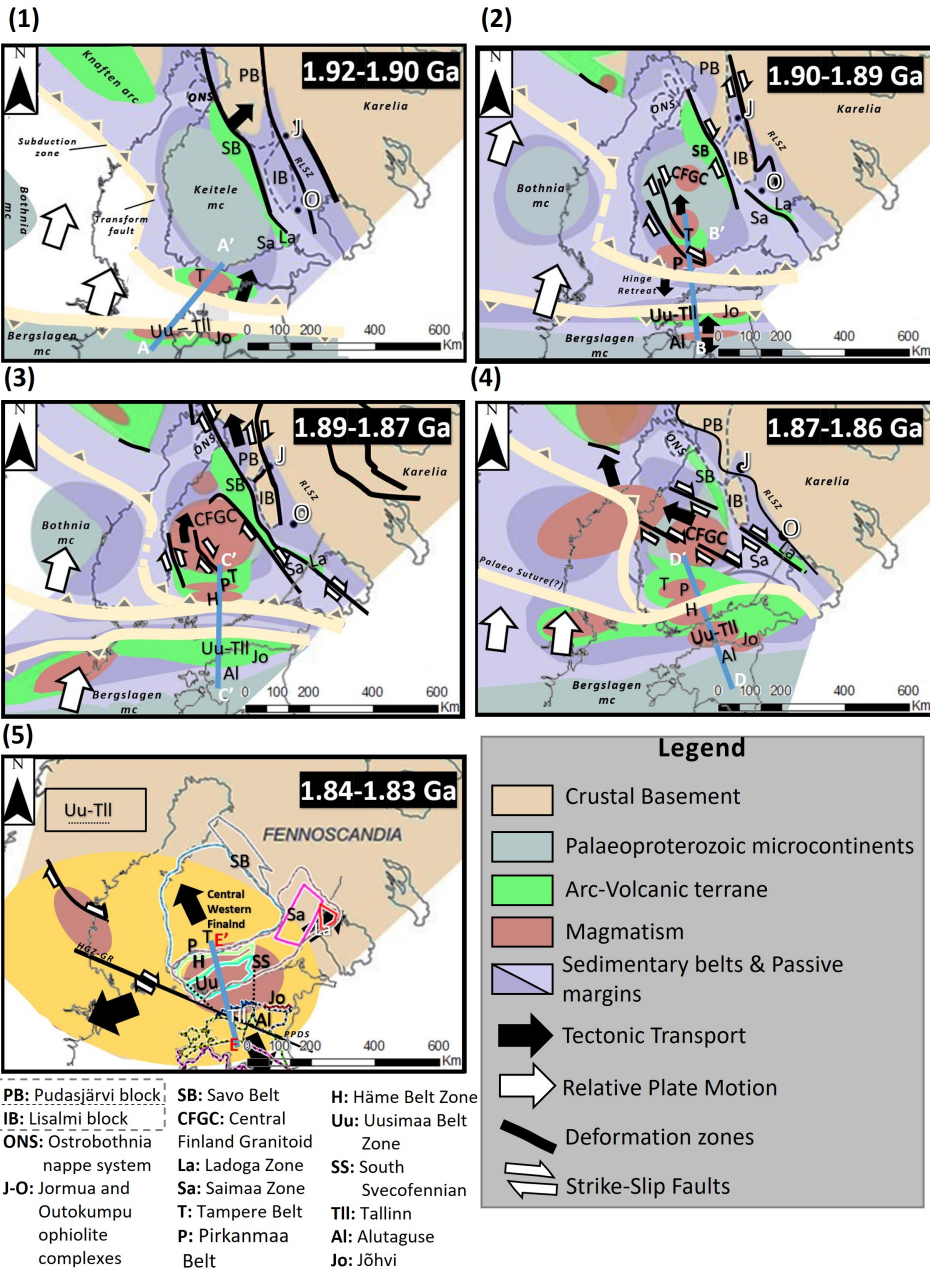
5.2.2 North Eastern Estonian Potential-Field Analysis and Exploration Implications

The EGT national potential-field datasets for NE Estonia (Fig. 3) were processed using the same workflow as the regional analysis with global models (Figs. 2d1–7), including filtering, spectral separation, and residual–regional partitioning. Regional and residual fields for the Alutaguse and Jõhvi domains, together with residual derivatives (Fig. 12), sharpen anomaly edges and delineate prospective zones. The strongest geophysical responses coincide with known metalliferous targets, including magnetite-rich rocks at Jõhvi and sulphide–graphite bearing gneisses in the Uljaste, Haljala, and Assamalla areas (Fig. 2c) [28, 40, 71].

Multiple models have been advanced for mineralisation associated with the Jõhvi Magnetic anomaly [28, 40, 41]. Early views invoked skarn or metasomatic origins in interlayered skarn–biotite–hornfels sequences cut by granite veins. The prevailing interpretation, however, is metamorphic overprinting of volcanic–sedimentary protoliths [40, 41, 71]. The chemistry of the magnetite supports this interpretation. The Jõhvi magnetites indicate that the iron formation units have undergone multiple phases of deformation, and the Jõhvi ore complex may have formed through repeated metasomatic events, skarnification, and recrystallization in contact with chemically reactive lithologies such as marbles. The influence of a granite intrusion should not be disregarded [40, 41]. Alternatively, they may represent recrystallized carbonate-facies iron formation deposits characterized by distinctly lower Ni and higher Mn contents. In addition, their elevated Ti+V levels—unusual for typical skarns—plot in porphyry-type fields close to the skarn–IOCG boundary on Ti+V vs. Ca+Al+Mn discrimination diagrams from Soesoo et al. (2021) [40].

Plado et al. (2020) [71] provide geological insight into the Jõhvi magnetic anomaly in NE Estonia based on a high-resolution ground magnetic survey acquired using a G-856AX proton-precession magnetometer. Their analysis shows that three prominent ground magnetic peaks, with maximum amplitudes of 19.290 nT (western), 15.880 nT (eastern), and

(a)



(b)

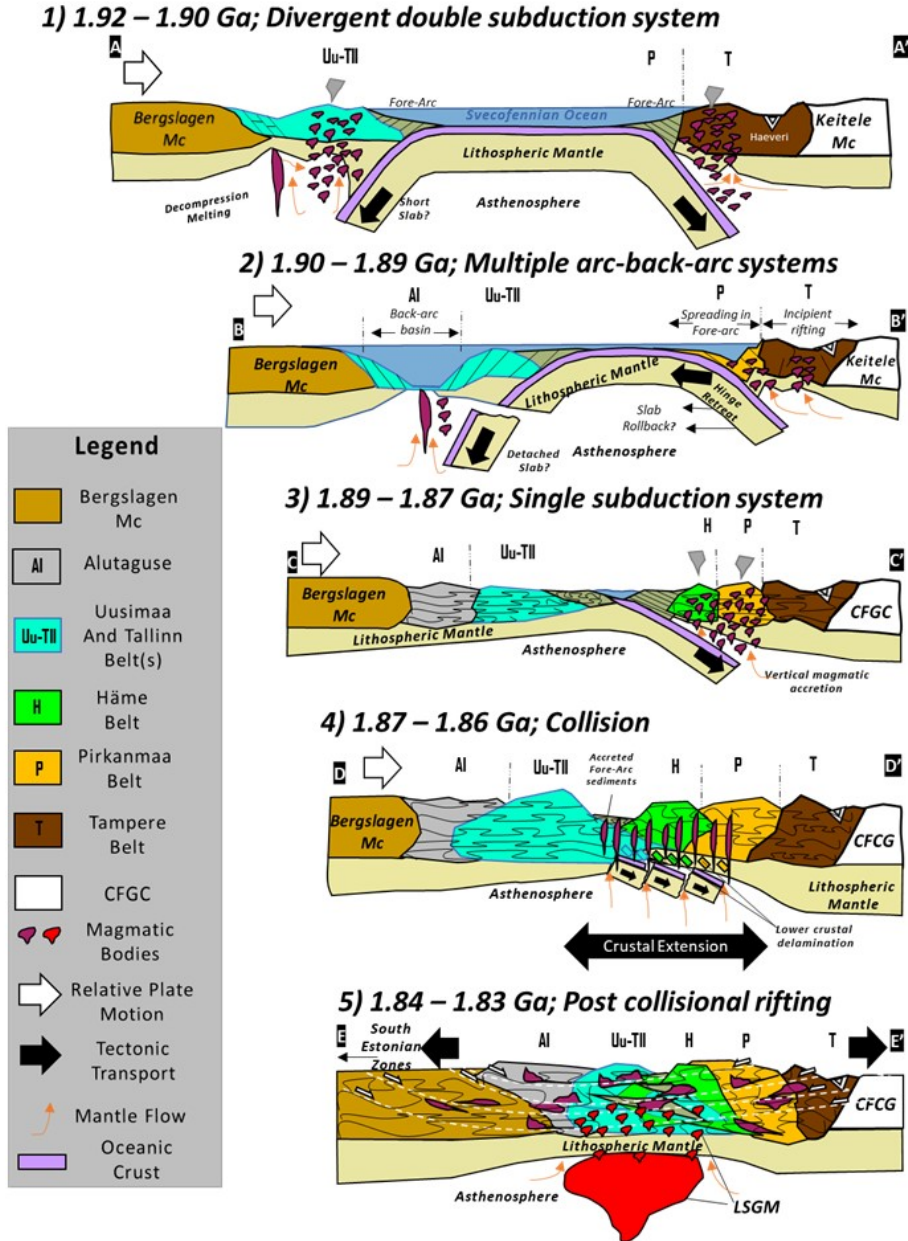


Figure 11: Schematic Geodynamic Model in (a)XY (Modified from Nironen (2017) [10], after Kotova and Podkovyrov (2014) [67]) and (b) YZ (Based on Kukkonen and Lauri (2009) and Kara et al. (2021) [37, 53]) sections of the evolution of Fennoscandia from: 1) 1.92-1.90 Ga; 2) 1.90-1.89 Ga; 3) 1.89-1.87 Ga; 4) 1.87-1.86 Ga, and 5) 1.84-1.83 Ga. The model is a not-to-scale conceptual representation designed to illustrate key dynamics rather than depict precise physical dimensions. XY cross-sections are indicated in Figure 11a with the blue line, respectively.

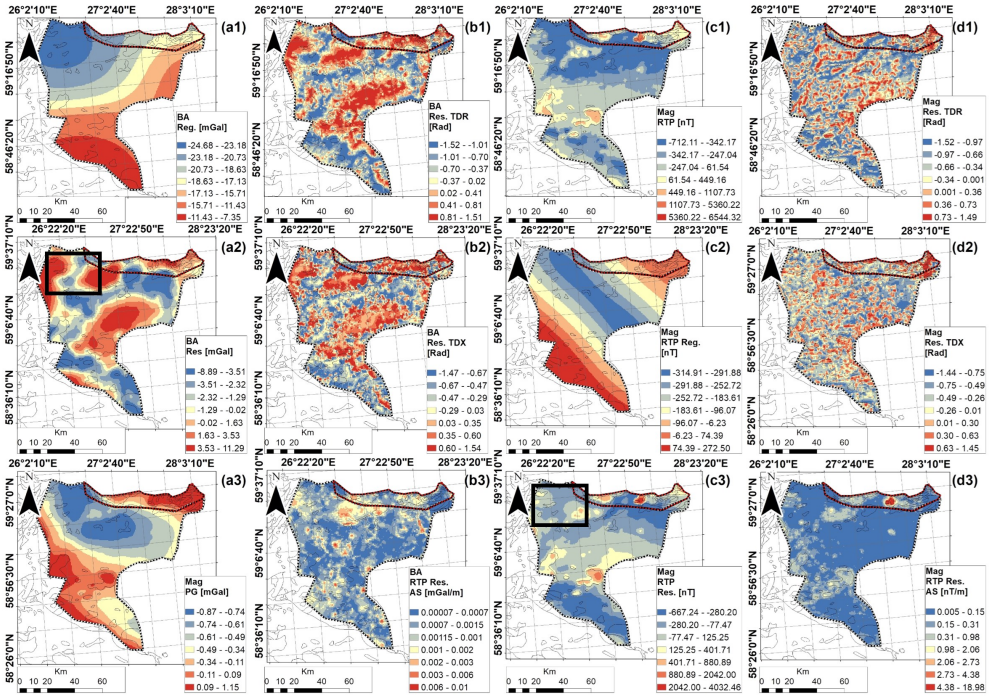


Figure 12: Geophysical regional and residual potential maps of NE Estonia (cut at 0.01 cycles/km). (a) Bouguer maps: (a1) regional component, (a2) residual anomalies, (a3) pseudo-gravity transformation derived from the RTP magnetic map.; (b) Bouguer residual derivatives: (b1) TDR, (b2) TDX, (b3) AS.; (c) Magnetic Reduce-To-Pole (RTP) maps: (c1) RTP magnetic anomaly, (c2) regional field, (c3) residual anomaly.; (d) RTP residual derivatives: (d1) TDR, (d2) TDX, (d3) AS. A black box in the residual plots (a2-c3) shows the zone used for further inversion in Figure 13.

8.080 nT (northern), are primarily produced by strong remanent magnetisation oriented sub-vertically and directed downward. The geometry of the anomalies is consistent with tabular, elliptic cylindrical sources, whose orientation aligns with the dip of the iron-ore formation. The dominance of remanent over induced magnetisation, by approximately an order of magnitude, indicates a high abundance of fine-grained magnetite ($< 1 \mu\text{m}$) within the magnetite quartzites, highlighting the need for further detailed (magneto) mineralogical investigations of the Jõhvi ore system.

As part of the exploration strategy developed in this dissertation, residual potential fields were analysed over a focused area encompassing the Uljaste, Haljala, and Assamalla zones (Figs.13a, 13b) (Manuscript III). A cross-gradient joint inversion using the SimPEG module was applied to gravity and magnetic anomaly data, generating 30 slices (15 in the X direction and 15 in the Y direction). These inversions estimated density (g/cc) and magnetic susceptibility (SI), using background values of 1×10^{-3} for both. Results up to 10 km depth are shown in Figures 13c-f. The mineralized areas—identified through sulphur-rich core samples (Fig.5)—correspond to Haljala (X_{2-4} , Y_{12-15}), Assamalla (X_{2-6} , Y_{1-4}), and Uljaste (X_{12-15} , Y_{8-11}). As documented for the Jõhvi area [71], the investigated target zones are characterised by tabular to elliptic bodies producing strong positive magnetic anomalies coupled with negative density contrasts. Areas with subdued magnetic responses are interpreted as graphite-rich gneisses with effectively diamagnetic behaviour, likely formed

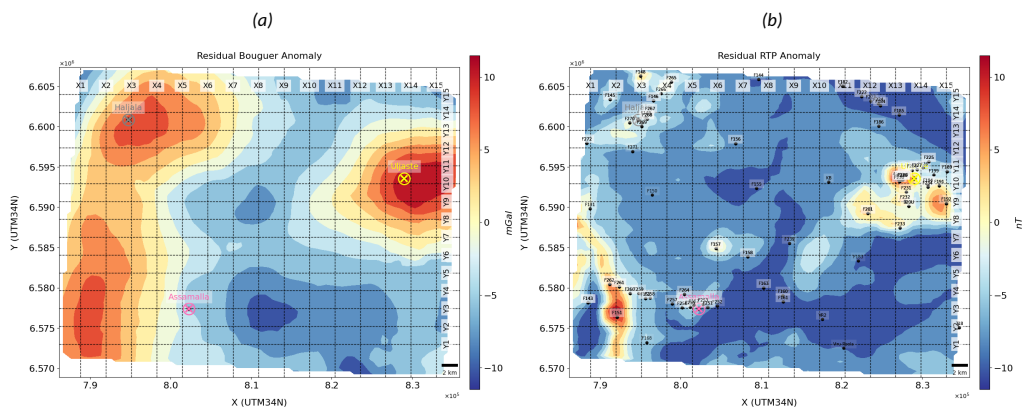


Figure 13: Residual Bouguer and RTP SimPEG cross-gradient joint inversion results for the target mineralized zones of Alutaguse, displayed in 3x5 grids. (a) Bouguer data and (b) RTP data (Black dots indicate EGT drill-cores). The inversion models include fifteen slices per XY direction, presented as follows: (c) Bouguer data (X direction), (d) RTP data (X direction), (e) Bouguer data (Y direction), and (f) RTP data (Y direction), providing a detailed representation of spatial density and magnetic susceptibility variations up to 10 Km. Continue on the next page.

during consolidation of the back-arc basin. By analogy, metalliferous zones in Alutaguse are inferred to be associated with tabular bodies and/or hydrothermal-skarn-related enrichment, comparable to the mineralised systems recognised at Jöehvi [40, 41, 68, 71].

Within the tectonic framework adopted here, the Alutaguse domain is interpreted as a Paleoproterozoic back-arc basin subsequently trapped during convergence and amalgamation of the Bergslagen–Livonia microcontinent with the Central Fennoscandian Granulite Complex (CFGK) (Figs.2,11). Its predominantly metavolcanic and metasedimentary assemblage [23, 24, 28] exhibits a geochemical signature distinct from adjacent SS domains and highly favourable for polymetallic exploration.

High-SiO₂ Alutaguse metasediments are systematically enriched in Fe₂O₃, MgO, SO₃, and K₂O and depleted in Al₂O₃ and Na₂O relative to Uusimaa, reflecting enhanced mafic detrital input and basin-scale fluid circulation rather than purely felsic arc sediment supply. These units show elevated transition metals (V, Cr, Co, Ni, Cu, Zn), Ba–Sr enrichment, high ΣREE, strong LREE dominance, and subdued Eu anomalies, distinguishing them from the more compositionally mature Häme and Saimaa arc-front sediments and from the quartz-rich, REE-poor Ladoga endmember.

Low-SiO₂ Alutaguse metasediments sharpen these contrasts and represent the most metal-fertile component of the basin fill, with the lowest SiO₂ and Al₂O₃, the highest Fe₂O₃, MgO, and SO₃, broad CaO–MgO ranges, and high base-metal concentrations (Ni 385 ppm, Cu 1060 ppm, Zn 4100 ppm, Pb 1430 ppm), far exceeding all SS equivalents. Elevated ΣREE with pronounced HREE dominance indicates felsic-volcanic detritus superimposed on a mafic-rich basin substrate.

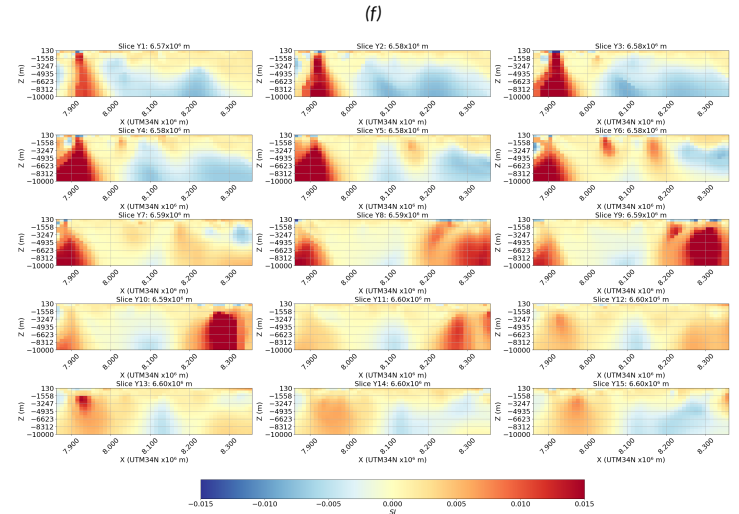
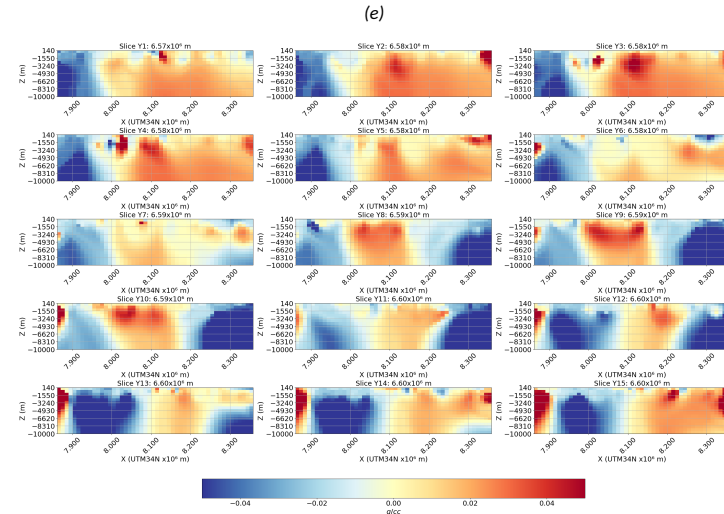
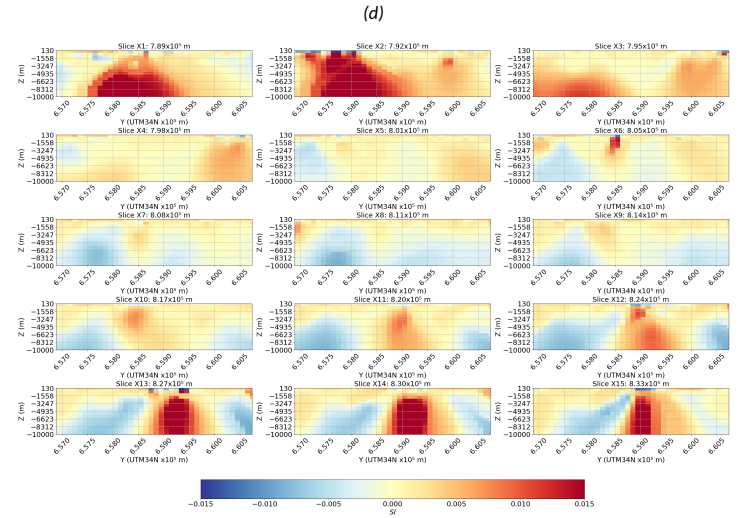
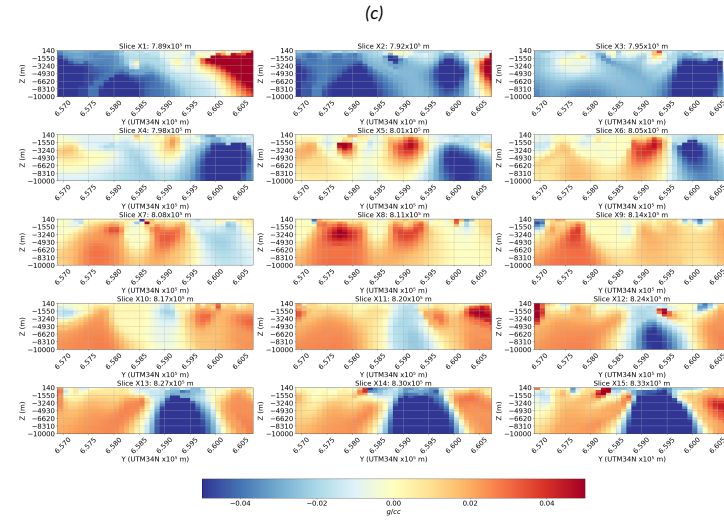


Figure 13

Alutaguse metavolcanics further reinforce this exploration potential: they define a tholeiitic, Fe-Ti-rich and metaluminous suite enriched in TiO_2 , Fe_2O_3 , MgO , and CaO and depleted in Al_2O_3 , Na_2O , K_2O , and P_2O_5 , consistent with juvenile magmatism in an extensional arc-proximal to back-arc setting. Trace-element ratios (low La/Yb, high Zr/Nb, elevated Th/Nb, depleted Ba/Th and U/Th) indicate derivation from a hot, depleted asthenospheric mantle with limited slab input, while exceptionally high base-metal contents (Ni 210 ppm, Cu 2230 ppm, Zn 2650 ppm, Pb 4030 ppm). Collectively, these elemental distributions highlight Alutaguse as a structurally controlled, metal-fertile back-arc system with strong potential for polymetallic and hydrothermal mineralisation, although elevated concentrations alone do not imply economic ore grades.

Regionally, the Bergslagen ore province is characterised by a close spatial association between Zn-Pb-Ag-(Cu-Au) sulphide mineralisation and magnetite-rich iron oxide deposits [13, 101]. Typical deposit styles include banded iron formations, magnetite-bearing calc-silicate and manganiferous skarns, carbonate-hosted iron ores, apatite-bearing iron ores, stratiform Zn-Pb-Ag-(Cu-Au) sulphides, and tungsten skarns, forming a continuum of related mineral systems [41, 69, 70, 161]. The southern Uusimaa belt (1.90–1.88 Ga), notably the Aijala-Orijärvi district, forms the southeastern continuation of Bergslagen and comprises felsic-dominated volcanic-sedimentary successions with abundant carbonates, polymetallic sulphides, and iron oxide mineralisation [8, 40, 41, 61, 62, 100, 162]. Magnetite-quartz-Fe-silicate iron formations at Jussarö (Finland) and Nyhamn (Gotland, Sweden) closely resemble the Jöhhvi magnetite-rich gneisses, supporting an origin from iron-rich chemical sediments later modified by deformation, metamorphism, and local skarnification; intrusion-related processes remain plausible given the presence of surrounding granitic veins and charnockites [40, 41].

Recently, EGT scanning with a Geotek MSCL-XYZ core 5–15 cm line spacing along selected sections enabled the recognition of key element associations in the Tallinn and Alutaguse corridor [68]: Ni-Co-Cr, Ti-V-Fe, Mo-W-Bi, Sn-Zn-Cd (In), Cu-Ni (PGM), Nb-Y-P (REE), K-Sn-Rb-Ga (Li), and As-Sb-Bi-W-Se-Sn (Au-Ag). The Ni-Co-Cr and Cu-Ni groupings are largely associated with mafic-ultramafic intrusions; Ti-V-Fe is focused on the Haljala area; Mo-W-Bi is present in both the Haljala and Tallinn structural zones; and Sn-Zn-Cd anomalies are concentrated in the Uljaste and Viru-Nigula areas [68]. The remaining associations (Nb-Y-P, K-Sn-Rb-Ga, As-Sb-Bi-W-Se-Sn) extend across several lithologies, indicating diverse magmatic and hydrothermal controls [40, 41, 68]. MSCL-XYZ measurements demonstrate a higher degree of heterogeneity than implied by older descriptions [25, 26, 116, 118], revealing sharp variations in rock characteristics and metal concentrations [68]. This directly enhances exploration targeting by pinpointing previously unrecognised mineralised intervals and correlating them with distinct geophysical anomalies and lithological units. To properly interpret these signals, additional detailed laboratory-based geochemical and petrographic-mineralogical investigations are necessary.

5.3 Geophysical and Petrogenetic Insights from Märjamaa and Kloostri Rapakivi

5.3.1 Geophysical Results

Spectral separation isolates the long-wavelength regional field (Figs. 14a1,c2) and enhances short-wavelength residuals that sharpen the signatures of individual plutonic phases (Figs. 14a2,c3). The RTP magnetic map (Fig. 14c1) depicts a narrow, composite magnetic fabric at the regional scale, while local anomalies align with the mapped tripartite granitoid phases. The pseudo-gravity RTP transform emphasizes lateral density contrasts (Fig. 14a3), outlining Märjamaa Phase I as the most Fe-rich and dense [95, 96, 122, 163]. Residual Bouguer and RTP maps (Figs. 14a2,c3) define three concentric domains matching Phases I–III: Phase I shows coincident gravity and magnetic highs; Phase II yields paired lows; Phase III produces a Bouguer low with a local magnetic high on the NW margin.

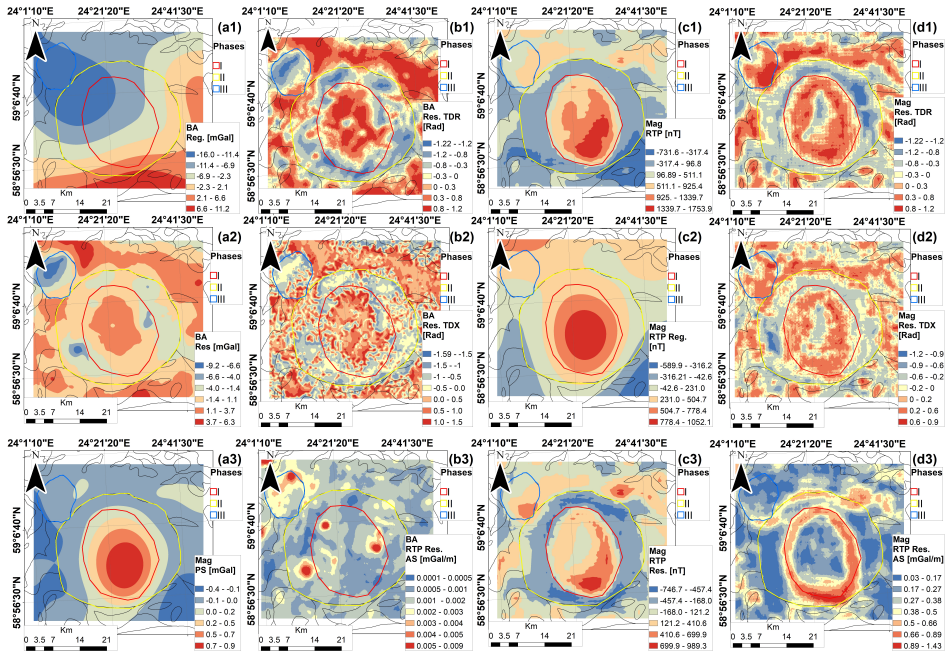


Figure 14: Geophysical regional and residual potential maps (cut at 0.05 cycles/km) of Märjamaa and Kloostri rapakivi bodies. (a) Bouguer maps: (a1) regional component, (a2) residual anomalies, (a3) pseudo-gravity transformation derived from the RTP magnetic map.; (b) Bouguer residual derivatives: (b1) Tilt Derivative Ratio (TDR), (b2) Tilt Angle Derivative (TDX), (b3) Analytic Signal (AS). (c) Magnetic Reduce-To-Pole (RTP) maps: (c1) RTP magnetic anomaly, (c2) regional field, (c3) residual anomaly.; (d) RTP residual derivatives: (d1) TDR, (d2) TDX, (d3) AS.

Residual derivatives of Bouguer and RTP data (Figs. 14b1–b3,d1–d3) resolve the internal architecture of the Märjamaa–Kloostri granitoids. TDR and TDX define concentric–radial patterns consistent with tripartite zoning and ring structures (Figs. 14b1–b2,d1–d2). The analytic signal (AS) [164] further sharpens these contrasts by delineating dyke-like lineations in the residual Bouguer field (Fig. 14b3) and rim-focused responses outlining Phase I and the northern Phase II host rock contact in the residual RTP field (Fig. 14d3).

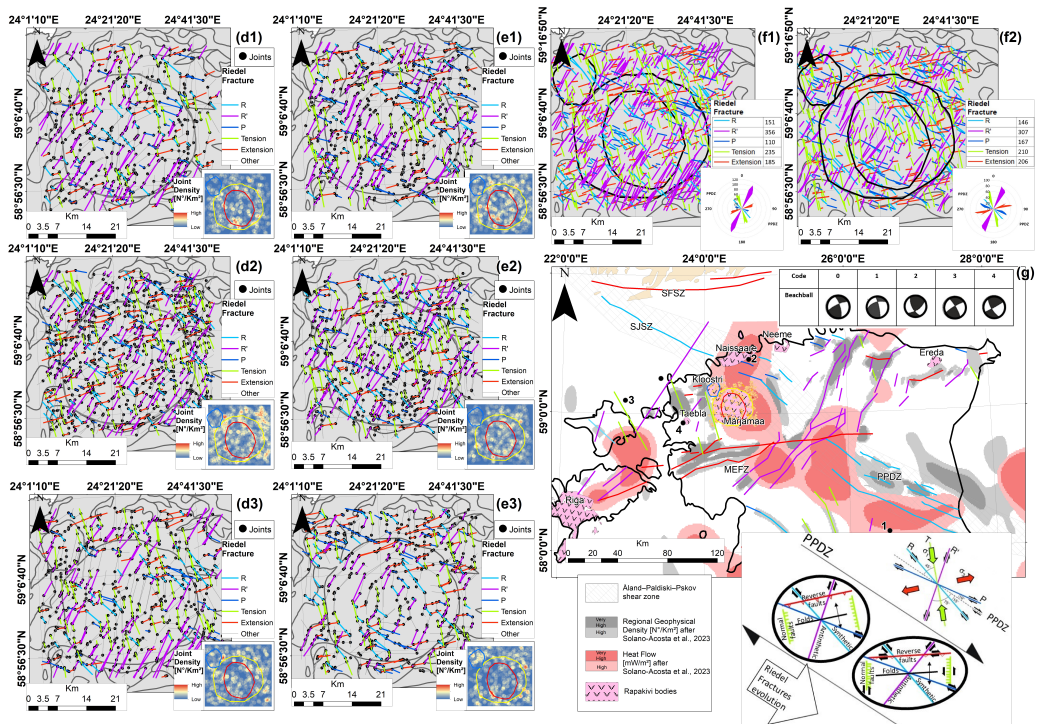
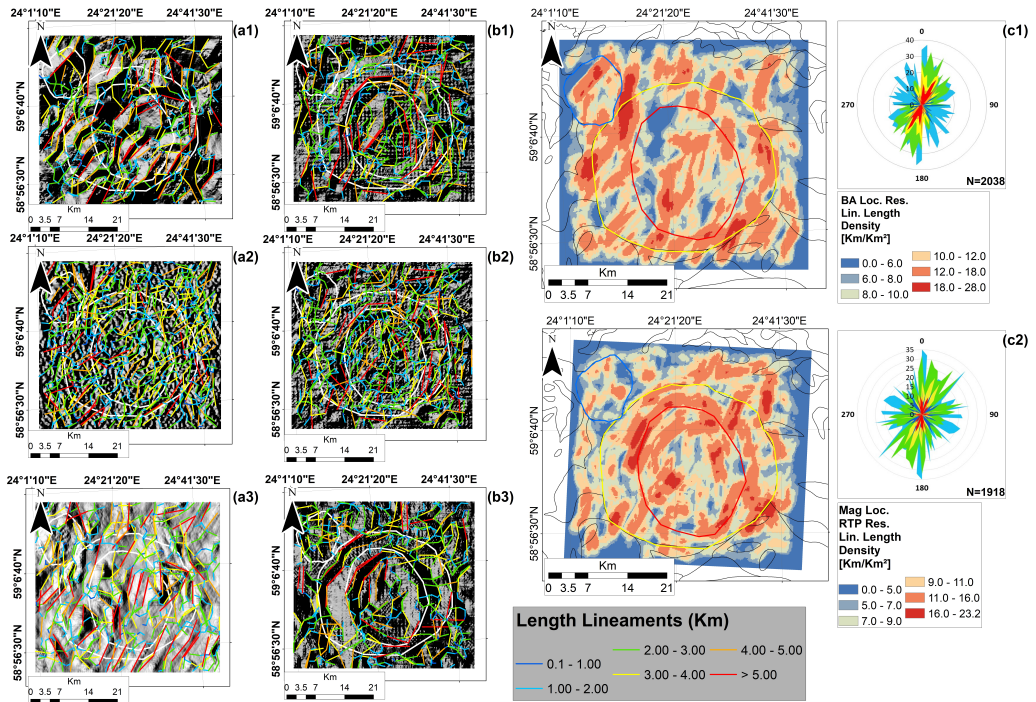


Figure 15: The lineament analysis of the potential residual data was conducted (Publication III), with residual derivative lineaments organised by length: **(a)** Bouguer (1): TDR, (2): TDX, (3): AS) and **(b)** RTP ((1): TDR, (2): TDX, (3): AS). Hillshade maps correspond to an azimuthal illumination of 0°.; **(c)** Length density maps and rose diagrams were categorized by lineament lengths for the complete residual datasets: (1) Bouguer and (2) RTP.; Residual lineaments were further organised into Riedel fractures for each derivative dataset: **(d)** Bouguer (1): TDR, (2): TDX, (3): AS) and **(e)** RTP (1): TDR, (2): TDX, (3): AS), including the lineament joints and joint density distribution maps for each dataset.; **(f)** Merged lineaments for each dataset were classified by Riedel shears, excluding lineaments not associated with Riedel structures, with rose diagrams for (1) Bouguer and (2) RTP.; **(g)** Estonian rapakivi bodies were overlaid on zones with the highest High Flow values and the highest density of tectonic features derived from WGM-12 and EMAG2v3 potential lineaments and EGT basement faults. The Märjamaa and Kloostri areas showed the highest density of potential lineament joints. Basement faults and lineaments were categorized based on their orientation relative to the dextral PPDZ maximum horizontal stress (σ_H of 115°–135°, 295°–315°) in central Fennoscandia [31, 92]. The angle relative to σ_H and the expected mode of movement were represented as follows: blue indicates (transfer (P) and normal (R), purple indicates 90° (reverse - R'), green indicates 30°–40° (sinistral strike-slip (T), and red indicates 30°–40° (dextral strike-slip (σ_3). The epicentres of earthquakes analysed by Soosalu et al. (2022) [31] are shown as black dots labeled with a number (code), each with its respective focal mechanism, highlighting the transpressive-transpressive arrangement due to the Åland-Paldiski-Pskov shear zone.

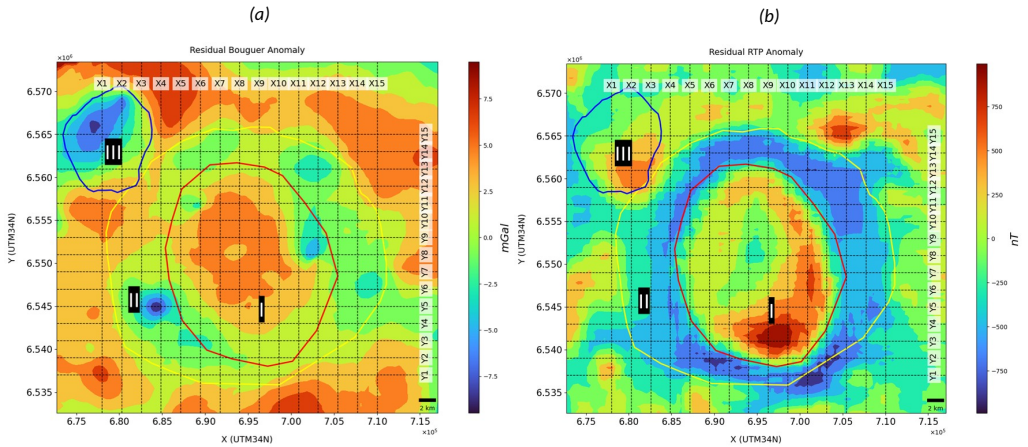


Figure 16: Residual Bouguer and RTP SimPEG cross-gradient joint inversion results for the study area, displayed in 3x5 grids. **(a)** Bouguer data and **(b)** RTP data. The inversion models include fifteen slices per XY direction, presented as follows: **(c)** Bouguer data (X direction), **(d)** RTP data (X direction), **(e)** Bouguer data (Y direction), and **(f)** RTP data (Y direction), providing a detailed representation of spatial density and magnetic susceptibility variations, highlighting the Märjamaa pluton and its Kloostri satellite tripartite phases with interpretative dashed grey lines. Continue on the next page.

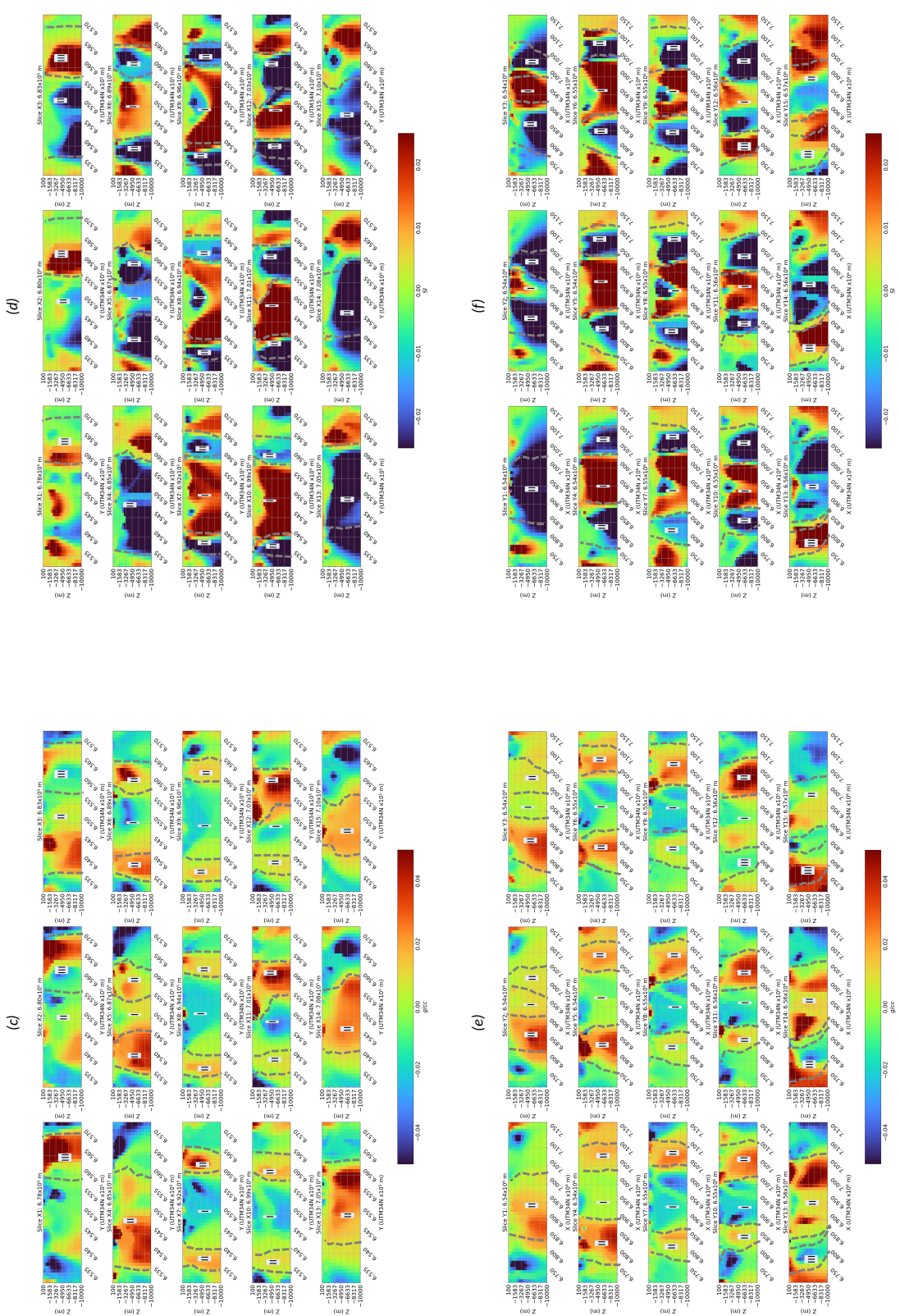


Figure 16

Automated lineament extraction from TDR, TDX, and AS grids yields dense clusters around the Märjamaa phases and the Kloostri body, with length–density maxima along Phase II margins and in the NW sector (Figs.15a–c). Rose diagrams show two dominant sets: short (0–2 km) NW–SE to NNE–SSW trends and longer (>2 km) NE–SW trends (Figs.15c1,c2). Interpreted in a dextral Riedel framework, R, R', P, tensile and extensional (σ_3) sets are recognized, with joint-density peaks near Phase II boundaries and within the PPDZ σ_H -parallel corridor—the principal fault structure [31, 92, 93]. The prevalence of antithetic R' fractures, followed by tensile/extensional elements, is consistent with pull-apart shearing (Figs.15f–g) [137, 165–168]. Within Phase II, the RTP residual map exhibits a *porphyroclastic dextral-flow* fabric, indicating a ductile-to-brittle overprint (Fig.14c3).

The 10 km-depth SimPEG cross-gradient inversion of residual Bouguer (Figs.16c,e) and RTP magnetic data (Figs.16d,f) resolves three geophysically distinct zones corresponding to the mapped phases. Phase I shows a negative density anomaly with high susceptibility; Phase II exhibits a positive density contrast and lower susceptibility; Phase III is positive in both petrophysical parameters. Sections X_{6–8} and Y_{5–8} image a central Phase I body flanked by Phase II, with sharp contrasts atop Phase I and susceptibility increasing with depth; X₈ and Y₈ highlight a focused density high and a conical magnetic structure suggestive of localised downward collapse. Linear/cylindrical features at Phase I–II boundaries (X_{4–5,11–12}; Y_{2–3,11–12}) may represent conduits, dike-like bodies, or local heterogeneities. Phase III (X_{1–3}; Y_{13–15}) appears smaller, with a vertical susceptibility profile akin to Phase I. The to-10 km inversion reproduces the main residual-field features and indicates that the dominant density–susceptibility contrasts agree with UC attenuation trends and CPD constraints. Deeper limits and host-rock interactions will require broader-scale potential-field datasets and/or seismic profiling.

5.3.2 Petrogenetic Analysis

Recent analyses of rapakivi intrusions within the Estonian basement [169] (*Manuscript III*) show that major-element indices (Fig. 17) discriminate the Estonian rapakivi bodies from the Wiborg complex. The mainland Estonian suites display predominantly oxidised geochemical signatures, whereas the Wiborg and Riga intrusions are characterised by more reduced compositions. Among the analysed intrusions, the Märjamaa–Kloostri system exhibits the highest Ti, Fe, Mg, Ca, and P concentrations, reflecting its comparatively less evolved and more oxide-rich character [169].

Major-element geochemical classification of the Märjamaa–Kloostri rapakivi complex follows the defined magmatic phase boundaries (Fig. 4c1). Phase-specific average compositions and compositional ranges are summarised in Table 3, while systematic geochemical trends are illustrated in Figure 7. SiO₂ increases with magmatic evolution: Phase I 66.69% (63.13–74.30%), Phase II 69.76% (66.02–73.66%), and Phase III 71.59% (69.84–75.34%). Al₂O₃ slightly declines: Phase I 13.51% (12.56–14.21%), Phase II 13.31% (11.81–15.78%), and Phase III 13.16% (12.19–13.89%), within the 11–14 wt.% range typical for A-type granites [170], though up to 17% has been reported [20, 171]. TiO₂ decreases from 0.79% (0.28–1.31%) in Phase I to 0.33% (0.16–0.49%) in Phase III, FeO* from 4.80% (3.43–5.96%) to 2.41% (1.84–3.07%), and MnO from 0.09% (0.07–0.12%) to 0.06% (0.04–0.08%). The Mg# is non-linear: Phase I 19.55 (14.11–25.16), Phase II 14.61 (8.21–21.65), and Phase III 18.29 (12.48–24.11). Alkaline oxides record fractionation, with CaO decreasing from 3.10% (1.79–4.54%) in Phase I to 1.94% (1.32–2.56%) in Phase III, Na₂O increasing to 3.34% (2.89–3.87%), and K₂O peaking in Phase II at 6.23% (5.61–6.91%). Incompatible components contract: P₂O₅ decreases from 0.32% (0.01–0.49%) in Phase I to 0.04% (0.01–0.08%) in Phase III, while SO₃ remains near 0.22% (0.10–0.35%). LOI

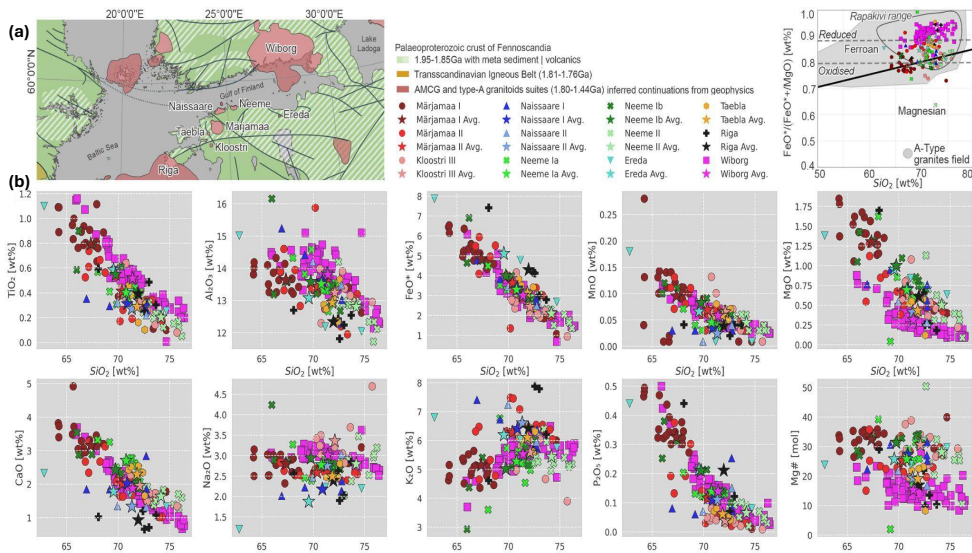


Figure 17: **(a)** Close-up of the Finnish-Estonian rapakivi bodies analysed, within Fennoscandia. **(b)** Hacker plots of Wiborg suite rapakivi bodies and the Riga body. Estonian rapakivi samples were analysed by magmatic generation (see Klein et al. (1994) [95]). Estonian samples come from Kivisilla et al. (1999) [116], and Wiborg body data extracted from the GTK Rock Geochemical Database [128]. Analysis and figure modified from Solano-Acosta et al. (2025) [169] (Manuscript III).

also declines from Phase I to III, consistent with differentiation and volatile loss.

Normative mineral proportions reflect systematic evolution from Phase I to III. Quartz increases from 23.44% (18.86 – 29.89%) in Phase I to 26.19% (19.37 – 33.16%) in Phase II and 26.79% (22.11 – 30.94%) in Phase III. Feldspars show contrasting behaviour: albite rises steadily from 23.83% (19.72 – 27.11%) and 21.96% (20.36 – 24.08%) in Phases I–II to 28.48% (23.91 – 39.75%) in Phase III, while anorthite declines from 9.39% (2.36 – 14.29%) in Phase I to 4.54% (2.31 – 7.14%) in Phase III; orthoclase peaks in Phase II at 37.23% (27.61 – 44.29%), compared to 30.53% (25.87 – 37.54%) in Phase I and 33.09% (23.07 – 37.22%) in Phase III. Oxide phases decrease systematically: magnetite drops from 3.21% (0.80 – 4.21%) in Phase I to 2.56% (1.04 – 3.94%) in Phase II and 1.75% (0.68 – 2.83%) in Phase III, with ilmenite following a similar trend (1.51% (0.19 – 2.13%), 0.79% (0.33 – 1.41%), 0.63% (0.15 – 1.22%)). Phosphates also diminish, with apatite falling from 0.76% (0.02 – 1.15%) in Phase I to 0.27% (0.05 – 0.87%) in Phase II and 0.09% (0.02 – 0.19%) in Phase III.

TAS [172] (Fig.18a) and $An - Ab - Or$ [173, 174] (Fig.18b) plots, showing Märjamaa and Kloostri samples evolve from quartz-monzonites to granodiorites (Phase I) to more granitic fields (Phases II and III), consistent with Fennoscandian rapakivi suites trends [175–177]. Alkali content ($Na_2O + K_2O$) increases progressively: 7.88% (6.62 – 9.16%) in Phase I, 8.80% (7.23 – 10.11%) in Phase II, and 8.89% (7.92 – 9.69%) in Phase III, matching Finnish rapakivi trends (Fig.18a).

Samples range from alkali-calcic to alkali (MALI > 3.8), with ASI mostly < 1.0, indicating a dominantly metaluminous character. Phase I shows MALI 4.76 (2.85 – 7.76), ASI 0.88 (0.73 – 1.00); Phase II, 6.79 (3.98 – 8.48), 0.92 (0.81 – 1.05); and Phase III, 6.95 (5.29 – 7.63), 0.88 (0.81 – 1.00), reflecting increasing alkalinity and metaluminous to marginally peraluminous compositions (Table 3), unlike Finnish rapakivi granites, which range from

metaluminous to peraluminous [176, 178]. Most samples fall within the ferroan A-type granite field (Fig. 18c), overlapping with typical Finnish rapakivi suites [20, 179, 180].

A-type granites in Laurentia-Baltica are grouped into ilmenite-, magnetite-, and peraluminous two-mica types [170, 181, 182]. In Laurentia, magnetite-series typically yield $\text{FeO}^*/(\text{FeO}^* + \text{MgO}) = 0.80\text{--}0.88$, whereas ilmenite-series exceed 0.88 under reduced conditions [182, 183]. Finnish rapakivi granites span 0.79–1.00 [20, 176, 178], linking them to reduced A-types though their relation to oxidised variants remains debated [46, 50, 170]. The Märjamaa-Kloostri intrusions plot dominantly in the oxidised field (Fig. 18d), with Phase I 0.73–0.85 (avg. 0.79), Phase II 0.76–0.91 (avg. 0.84), and Phase III 0.74–0.85 (avg. 0.80) [170, 179, 180]. FeO^*/MgO ratios further trace melt evolution [50, 184], with Phase I 3.79 (2.68–5.79) suggesting limited fractionation, Phase II 5.63 (3.20–10.06) indicating peak fractionation, and Phase III 4.19 (2.80–5.70) reflecting moderate to borderline fractionation.

Experimental data indicate that A-type granites form at low pressure (≤ 5 kbar) through hybridisation of calc-alkaline granitoids with high-Al olivine tholeiites or metagraywackes, leaving plagioclase and orthopyroxene residues [156, 170, 185]. The Märjamaa-Kloostri samples plot along this low-pressure trend in the $\text{CaO}/\text{Al}_2\text{O}_3$ vs. $\text{CaO} + \text{Al}_2\text{O}_3$ diagram (Fig. 18e), consistent with Finnish rapakivi granites, whereas high-pressure (12–15 kbar) melts yield clinopyroxene-rich residues and plot with calc-alkaline granitoids [170].

Duan et al. (2022) [186] critiqued zircon-saturation thermometry for rapakivi studies (e.g., Grabarczyk et al. (2023) [20]), noting limited Zr control on melt compositions, and instead proposed formulas for granitoid temperatures: $T_{\text{SiO}_2} [^\circ\text{C}] = -14.16 \times \text{SiO}_2 [\text{wt.}\%] + 1723$ and $T_{\text{MgO}} [^\circ\text{C}] = 887.6 \times (\text{MgO} [\text{wt.}\%])^{0.0989}$. These models suggest decreasing temperatures from Phase I (778.66°C, range 670.98–829.06°C) to Phase II (735.18°C, range 680.01–788.09°C) and Phase III (709.26°C, range 656.18–734.12°C), with T_{MgO} estimates consistently $\sim 100^\circ\text{C}$ lower (Table 3).

The multidimensional discrimination tectonic diagram for acid rocks, based on the analysis by Verma et al. (2013) [131], has been well-established in A-type complexes [187] (Fig. 18f). This setting is defined by two functions, $DF1$ and $DF2$ (Table 3). The analyses show a plot mostly within the continental rift domain, and some in the collision domain.

Correlation matrices (Fig. 18g–i) highlight systematic mineral–element associations in the three phases. In Phase I (Fig. 18g), a strong oxide–phosphate cluster (magnetite–ilmenite–apatite) shows positive intercorrelations, with quartz negatively correlated to these phases, reflecting silica enrichment at the expense of Fe–Ti–P minerals; feldspars are decoupled, with orthoclase negatively correlated to anorthite and oxides, and albite weakly negative overall. In Phase II (Fig. 18h), correlations weaken, though the oxide–phosphate cluster persists; quartz remains negatively associated with oxides, while anorthite correlates positively with apatite and ilmenite, and orthoclase negatively, marking Ca–alkali partitioning; albite shows variable links, including some positive ties to anorthite and apatite. In Phase III (Fig. 18i), quartz correlates positively with feldspars and negatively with oxides, indicating silica–alkali enrichment; albite and anorthite are strongly correlated with quartz but negative to oxides, while orthoclase shifts to positive correlations with oxides and apatite, suggesting late-stage alkali coupling with Fe–Ti–P phases; magnetite and ilmenite remain tightly linked, consistent with an oxidised residual melt.

Indices based on major elements provide key constraints on A-type granite classification, reflecting crystallization history and magma sources [46, 50, 179]. These suites are broadly divided into reduced and oxidised types, controlled by $f\text{O}_2$, H_2O content, and pressure [20, 170, 181, 183, 190]. Oxidised magmas typically derive from hydrous quartz feldspathic crust, forming magnetite-bearing assemblages near the NNO buffer [20], as

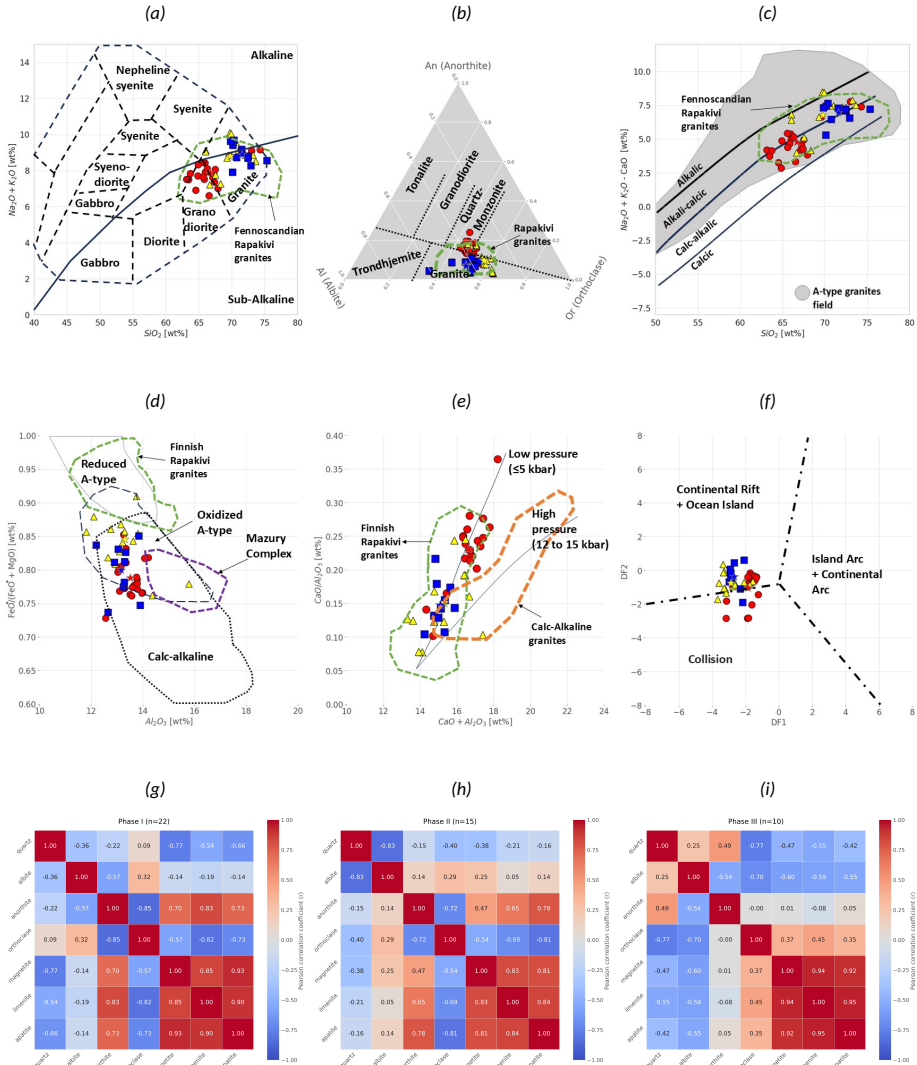


Figure 18: Geochemical plots from the major elemental data for the Märjamaa and Kloostri rapakivi: **(a)** $\text{K}_2\text{O} + \text{Na}_2\text{O}$ vs. SiO_2 diagram from Cox et al. (1979) [188] after Kosunen (1999) [175].; **(b)** (a) An Ab Or diagram from Barker (1979) [173] with rapakivi classification after Luo et al. (2024) [174]; **(c)** $\text{Na}_2\text{O} + \text{K}_2\text{O} - \text{CaO}$ vs. SiO_2 diagram after Frost et al. (2001) [179].; **(d)** $\text{FeO}^*/(\text{FeO}^* + \text{MgO})$ vs. Al_2O_3 classification plot of calc-alkaline and A-type granites, including fields for reduced and oxidized subtypes, based on Dall'Agnol and de Oliveira (2007) [170].; **(e)** $\text{CaO}/\text{Al}_2\text{O}_3$ vs. $\text{CaO} + \text{Al}_2\text{O}_3$ for pressure ranges in A-type granites, after Patiño Douce (1997, 1999) [156, 185] modified from Dall'Agnol and de Oliveira (2007) [170]. Sources of Fennoscandian rapakivi plutons are based on the work of Dall'Agnol and de Oliveira (2007) [170] and Grabarczyk et al. (2023) [20], utilizing results from Ramo and Haapala (1995, 2005) [176, 178] for the Wiborg area, Kosunen (1999) [175] for the Bodom and Obbnäs rapakivi plutons, Heinonen et al. (2010) [189] for the Ahvenisto body, and Sharkov (2010) [44] for the Salmi granitoids.; **(f)** Discriminant-function (DF1 and DF2 ; Manuscript I) multidimensional diagrams based on log-transformed major and trace element ratios for tectonic discrimination of acid granitic rocks, after Verma et al. (2013) [131]; Comparative analysis of mineral correlation matrices across three phases: **(g)** Phase I matrix, **(h)** Phase II matrix, and **(i)** Phase III matrix.

reduced types reflect metasedimentary input or evolved tholeiites [170, 183, 191]. The Märjamaa–Kloostri; and mainland Estonian rapakivi intrusions (Fig.17), crystallised under oxidised conditions (Fig.18d), favouring biotite–hornblende stability [95, 122], and showing higher normative magnetite contents than ilmenite, in contrast to reduced Wiborg, Bodom, Obbnäs, and Ahvenisto granitoids [175, 176, 178]. Comparisons with the Mazury Complex highlight their shared oxidised character but also differences: Mazury granites display elevated alumina from partial melting of granodiorite at mid- to lower-crustal depths, whereas Märjamaa–Kloostri bodies evolved within the PPDZ corridor with abundant titanite, apatite, and fluorine phases enhancing oxidation [20, 33, 95, 122, 171].

Recent trace-element and REE data from the Märjamaa–Kloostri intrusion [163, 169, 192] were reorganised according to the tripartite magmatic phases (*Manuscript III*), showing that Phase I is strongly enriched in trace elements (Fig.19a) and REE (Fig.19b), with a progressive decrease toward Phases II and III, whereas the host Biotite–Amphibole Gneiss is notably depleted when compared to the granitoid bodies. The samples also display marked LREE enrichment compared to the Finnish rapakivi samples [176, 178]. The Märjamaa–Kloostri intrusion stands out among Fennoscandian rapakivi bodies due to its high to very high total REE contents, reaching values up to 3600 ppm [163, 192]. Moreover, the trace-element systematics reproduce the similar tectono-magmatic inferences shown in Fig.18f. The Nb–Y–Ce diagram (Fig.19c) indicates an A₂-type granite affinity, typically linked to post-collisional or post-orogenic settings and derived from mixed sources such as continental crust, island-arc basalts, or remelted crustal material [20, 176, 178]. Likewise, the Rb vs. Nb+Y discrimination plot (Fig.19d) reinforces the within-plate A-type character of the intrusion [20, 174, 176, 178, 193] (e.g., Fig.18f).

5.3.3 Geodynamic Features

The Märjamaa–Kloostri complex constrains A-type granite emplacement in the Fennoscandian Shield. Despite compositional variability, A-type granitoids share petrogenetic controls tied to local *P–T* conditions and tectonic regime [11, 44, 45, 50]. Across Fennoscandia, rapakivi bodies record multistage melt generation and ascent at differing depths [47, 83, 194]. At Märjamaa–Kloostri, structural and geophysical evidence favours transtensional emplacement (Fig.15), with no preserved volcanic cover [33, 35, 36]. Below, competing A-type granite emplacement models are summarised and evaluated against integrated gravity, magnetic, and lineament data from Märjamaa–Kloostri.

Three schematic frameworks capture A-type magmatism. Bonin (2007) [50] envisaged a vertically zoned system: ultradeep mantle melts (30–40 km, > 10 kbar) generate mafic cumulates (*ol + cpx + opx ± Ca–Ti–amph*), mid-crustal reservoirs (15–20 km, 5–7 kbar) host rapakivi batholiths, and shallow levels (1–4 km) form felsic ring complexes (*af + pl + cpx + ap + zr ± Ca–amph*). Sharkov (2010) [44] proposed plume-driven, transcrustal staging beneath thick sialic crust, with anorthosite–gabbro–norite cumulates underlying buoyant silicic magmas that ascend in pulses. Johansson (2023) [45] argued for long-lived basaltic underplating that thermally drives partial melting, fractional crystallisation, and assimilation within a single evolving lower-crustal column. All invoke mantle input, but differ in their thermal regimes, crustal architectures, and plume conditions. The Salminen et al. (2021) [11] superswell hypothesis supports both plume-rich [44] and columnar evolution [45] scenarios while remaining compatible with intraplate settings [50].

These ideas align with the planetary-scale link between plumes and supercontinent breakup [74]. Plume-related doming and rifting weaken the lithosphere; tomography indicates stagnant slabs in the mantle transition zone (MTZ) and high-velocity anomalies above the core–mantle boundary (CMB), which feed superplumes and superswells.

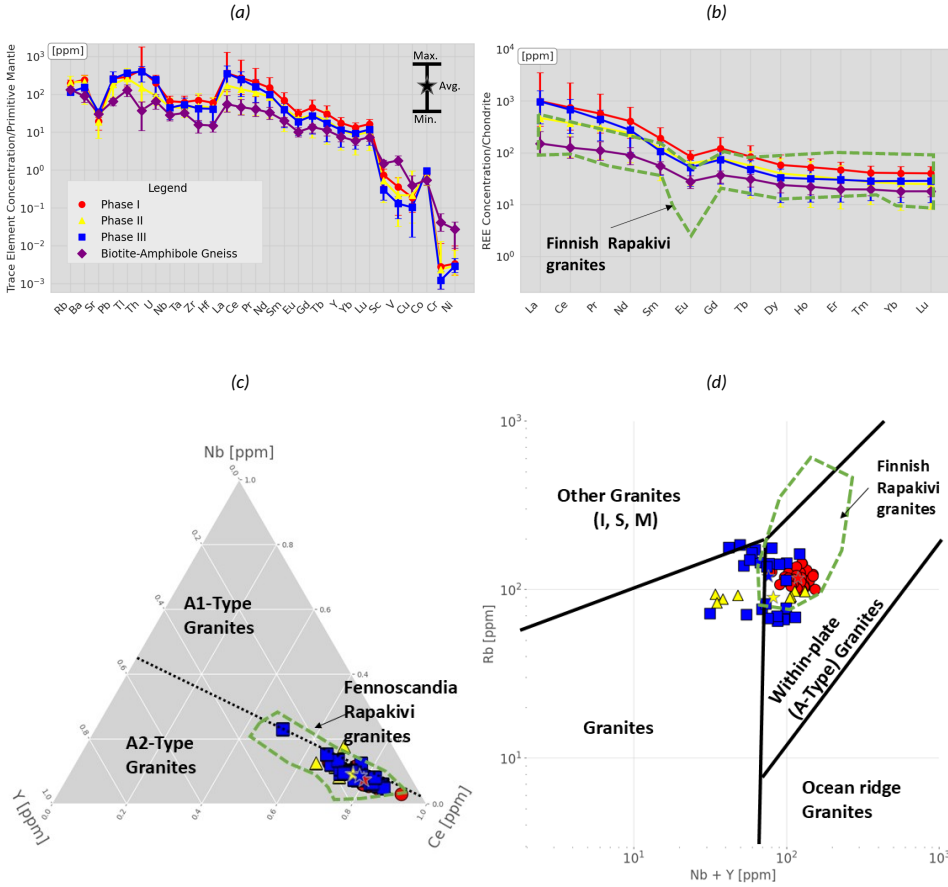


Figure 19: Trace-element and REE geochemical diagrams for the Märijamaa-Kloostri samples. Dataset taken from Potagin (2024) [163]. Subfigures (Modified from Solano-Acosta et al. (2025) [169]) enhance: (a) multi-element trace patterns, (b) chondrite-normalised REE patterns, (c) Nb–Y–Ce systematics, and (d) Rb vs. Nb+Y discrimination fields. Sources of Fennoscandian rapakivi plutons are based on the work of Rämö and Haapala (1995, 2005) [176, 178] and Grabarczyk et al. (2023) [20].

Primary, secondary, and tertiary plumes may operate at different mantle depths, with modern Africa– Pacific superswells analogous to a low-latitude Nuna superswell (Fig.21a) [11, 74]. Consequent upwelling, transient thinning, and mafic underplating favour anorogenic, low- H_2O magmatism [74]. Therefore, regionally, the Fennoscandian rapakivi intrusions are mid-shallow crustal AMCG expressions associated with mantle processes driven by a superswell beneath a thermally insulated, low-latitude Nuna [11, 195, 196] (Fig.21a). AMCG magmatism—including rapakivi granites and massif anorthosites—fits Nuna breakup, with superswell-induced perturbation of the SCLM, partial melting, and extension [44, 45, 47, 50, 74].

Metallogeny commonly accompanies A-type magmatism (SEDEX, IOCG, Sn, stratiform Cu–Au–U during 1.8–1.1 Ga; [74]); fluid–structure interactions along shear zones can focus HFSE and chalcophile metals [47, 50, 74, 137, 168]. At Märijämaa, a NE-trending magnetic highs over the host lithologies (Fig.14d3) spatially coincides with SimPEG derived density and susceptibility contrasts (X_{12-15} , Y_{15} ; Figs. 16c–f), delineating a structurally controlled subsurface target, possibly associated with metalliferous anomalies (*Publication III & Manuscript III*). Also, a phase-resolved analysis of lineament distributions (Fig.20) suggests the presence of internal fracture or fault networks, which may have acted as pathways for vein development or mineralisation [86, 190, 197, 198]. Phase I is characterised by the longest (> 5 km) and most abundant lineaments (Figs.20a1,b1), particularly within the RTP residual dataset. Phase II exhibits a dominant NE–SW trend for intermediate-length lineaments (2–3 km), together with a subordinate NW–SE trend for shorter lineaments (0.1–2 km) (Figs.20a2,b2). This pattern likely reflects mechanical interaction with the host rocks and dextral kinematics along the PPDZ, potentially linked to a porphyroclastic dextral-flow fabric (Fig. 14c3). In contrast, the Kloostri body (Phase III) (Figs. 20a3,b3) displays the shortest and least abundant lineaments, predominantly 1–3 km in length, with only a few longer structures transecting the intrusion.

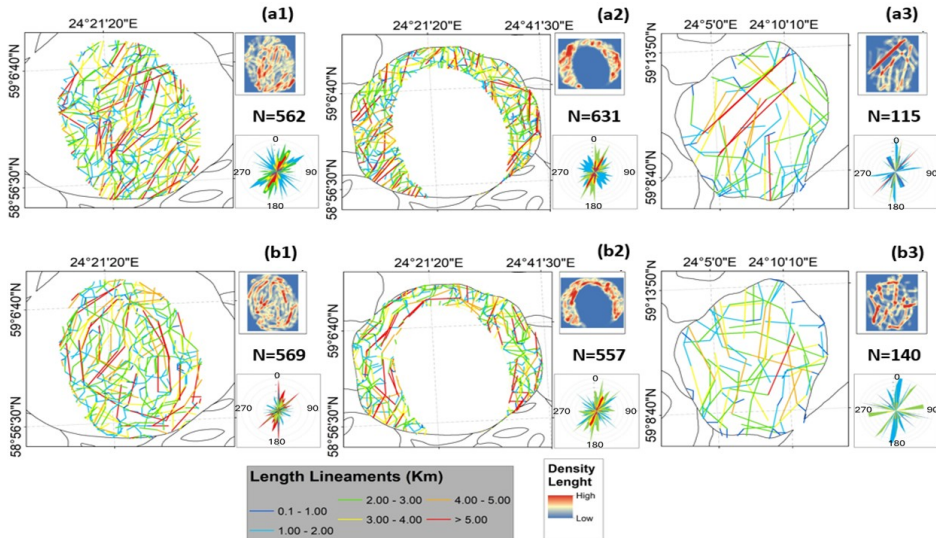


Figure 20: Residual potential-field lineament distribution by phase for (a) Bouguer and (b) RTP magnetic data: (1) Phase I, (2) Phase II, and (3) Phase III (Kloostri).

NE–SW SO compression established NW–SE to NNW–SSE shear corridors later reactivated during Mesoproterozoic extension; structural deflections near rapakivi massifs and

present Rg-wave observations indicate ongoing oblique strike-slip [8, 14, 31, 32, 84, 91, 92]. Within this oblique-extensional setting, late Wiborg Suite intrusions (including Märjamaa-Kloostri) formed tabular, sheet-like plutons along dilatational ramps imaged on seismic profiles (Figs. 15g, 2b) [32, 42, 78–83, 91, 94]. During N–S extension at 1.65–1.60 Ga, reactivation of the SFSZ (Fig. 15g) and PMSZ (Porkkala–Mäntsälä Shear Zone) overprinted older E–W fabrics and enabled emplacement of the ~1.64 Ga Bodom and Obbnäs stocks [52, 86, 94, 175, 199]. Along the Åland–PPDZ corridor, joint-density peaks, NW–SE high-density lineaments, and shear-wave-splitting fast polarisations [93] parallel to the corridor indicate focused deformation and heat flow (*Publications I & III*). Mapped Riedel extensional/tensional sets (Fig. 15f–g) record reactivation of SO fabrics during Nuna breakup, consistent with pull-apart kinematics and coeval AMCG magmatism and basin formation [8, 15, 52, 74, 86, 200], and Major-geochemical results (Fig. 18f).

Geophysical inversion and petrology constrain emplacement depths. Post-magmatic uplift/erosion likely removed several kilometers of cover [80, 201], but the mechanisms of thinning remain debated. Puura and Flodén (1999) [80] proposed that mantle and crustal uplift played a role alongside erosion, while their later work in Puura and Flodén (2000) [42] emphasized differential erosion and tilting. Sharkov (2010) [44] estimated a stripping of 16 to 20 kilometers at 4.5 to 5 kbar of metamorphic pressure. Reported depths for AMCG-rapakivi systems span 7–20 km (2–6 kbar): Salmi (2–3 kbar), Korosten (3–4 kbar; pegmatites 2.6–3.0 kbar), Wiborg (2.5–5.4 kbar), Mazury (~4 kbar), and Riga (5–7 kbar; 14–17 km) [20, 33, 36, 44, 198, 202]. For Märjamaa-Kloostri, Buddington–Barth geothermometry (550–650°C) and amphibole Al (5.2–6.5 kbar) suggest ~17–22 km [95].

Basement tilting of ~100 m across the Märjamaa-Kloostri area (Fig. 1d) implies asymmetric denudation and block tilting [28, 39]. Host amphibolites reached ~700°C and 4.5–5.5 kbar [28], consistent with proposed regional pressures [44]. The Conrad lies at ~17–19 km beneath Märjamaa-Kloostri (Fig. 2d2); allowing ~20 km of erosion places the original Conrad near ~40 km, matching Sharkov (2010) [44] SIAL–SIMA reconstructions. Previous Estonian Potential-field profiles [23] indicate comparable structural levels for the intrusions, with the present Märjamaa base at ~10 km, consistent with CPD (Fig. 2d3) and Upper Continuation (UC) depths (*Publication III*). Accounting for ~20 km of post-emplacement erosion implies original depths of ~20–30 km, in line with Klein et al. (1994) [95] thermobarometric results.

Many Fennoscandian rapakivi plutons (Vehmaa, Ahvenisto, Märjamaa-Kloostri, Mårdsjö, Wiborg-Kymi) display circular planforms and inward-dipping listric faults diagnostic of piston-cauldron subsidence, with limited brecciation indicating low differential stress [81–83, 95, 123, 194, 203]. While the caldera-cauldron boundary is blurred, calderas typically form at ≤7 km and erupt, whereas piston-cauldrons occur at 7–30 km and need not erupt [82, 83, 121, 204–209]. In Estonia, erosion has probably removed any ignimbrites [25, 33, 42], and pressure determinations [44, 95] classify Märjamaa and Kloostri as piston-cauldron type intrusions.

The proto-Phase I magma chamber likely developed at ~15–30 km depth as a steep-sided, ovoid body with a vertical extent of ~15 km and a lateral footprint of 40 × 25 km (Fig. 21b) [206, 210]. Transtensional pull-apart shearing along the PPDZ triggered piston-style floor collapse, producing ring fractures and a proto-piston block that accommodated magma accumulation and chamber expansion [81, 204, 205, 208, 211]. UC modelling (*Publication III*) and SimPEG inversion profiles (X_{9-10} , Y_{13-14} ; Fig. 16) identify a southeastern feeder conduit, indicating vertical magma ascent and a central recharge zone. This chamber acted as the main reservoir sustaining a broadly symmetrical intrusion.

The transition from the proto-Phase I chamber (Fig. 21b) to piston-cauldron subsi-

dence (Fig.21c) was driven by continued transtensional shearing along the PPDZ (Fig.15g), which generated concentric and radial fractures, promoting roof collapse. Magma initially ascended vertically along fractures but later shifted into subhorizontal, inward-directed flow [168, 204, 210–213]. SimPEG cross-sections (X_{7-9} , Y_{7-9} ; Fig.16) reveal a cylindrical central structure with a positive density and susceptibility gradient at ~ 3 km depth, extending ~ 4 km in thickness. This anomaly, interpreted as a paleo-conduit or syenitic residual core, corroborates piston-style collapse and supports the possibility of ignimbrite discharges. Progressive tectonic extension along the PPDZ reduced chamber pressure, initiating sagging of the roof and subsidence along ring faults, thereby inducing radial stress, decompression, magma ascent, and potential ignimbrite eruptions [82, 204, 206, 210].

Subsequent piecemeal roof collapse facilitated melt exchange between upper and lower chamber zones [207, 208, 210, 211], allowing discrete magma pulses to coalesce into a consolidated Phase I body. Phase II intruded peripherally through PPDZ-related fractures (Fig.21d) and is characterised by paired negative Bouguer and magnetic anomalies, consistent with advanced fractional crystallisation and crustal assimilation [95, 96, 122].

The final leucogranitic Phase III represents the last subsidence stage [95, 96], emplaced asymmetrically as a trapdoor-style body on the NW margin of the intrusion (Fig.21e) [206, 210]. This phase likely reflects chamber depressurisation and enhanced piston-cauldron dynamics, remobilising residual felsic melts and focusing magma ascent along NW-oriented conduits.

Potential data demonstrate distinct petrophysical differences between the Märjamaa–Kloostri phases, supporting a piston-cauldron-style emplacement along the PPDZ. Phase I, a ferroan melanocratic granodiorite, is magnetite-rich and denser, producing coincident gravity and magnetic highs ($\rho \approx 2.71$ g/cm³; Fe–Ti oxides $\sim 3\%$) [95, 96, 122]. Phase II, a quartz–K-feldspar–rich ring, is lighter ($\rho \approx 2.66$), magnetite-poor, and produces paired lows [95, 96]. Phase III, a late Na-rich trachytic leucogranite, is the least dense ($\rho \approx 2.59$ – 2.63) [95, 96, 122], forming a gravity low and a local NW magnetic high linked to superposition with Phase I's collapse structure or late NW conduits. Geochemically, the sequence reflects progressive differentiation of an oxidised A-type system: Phase I shows low SiO₂ and high TiO₂, FeO*, P₂O₅ with abundant oxides–phosphates, and REE–HFSE content (Fig.19); Phase II marks peak K₂O and orthoclase with declining oxides; and Phase III records the highest SiO₂ and albite with minimal oxides and lower crystallisation temperatures. These trends define a multi-stage, oxidised A-type system formed by fractional crystallisation under PPDZ-controlled transtension.

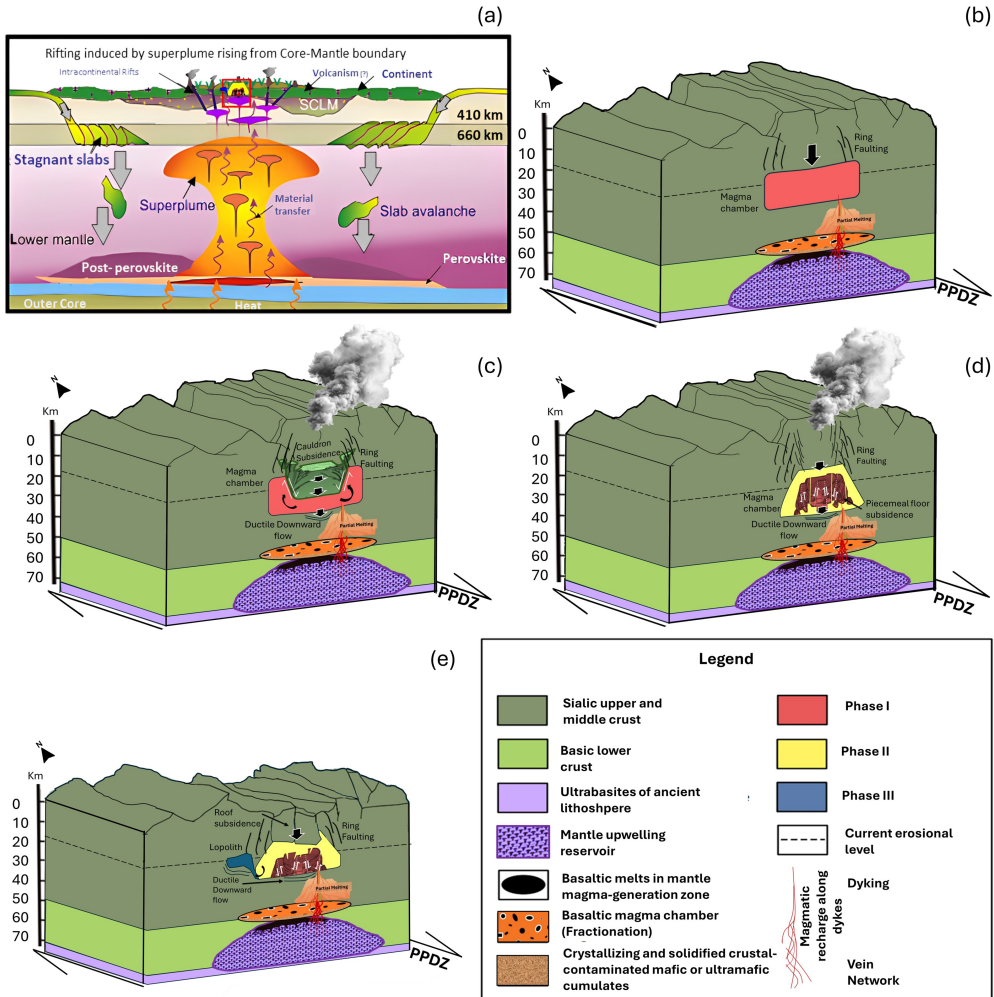


Figure 21: The geodynamic model for the emplacement system of the Märjamaa-Kloostri rapakivi system is presented as follows: (a) Rifting induced by superplume rise from the core-mantle boundary, fueled by 'slab graveyards', exemplifying the superswell plume model for Fennoscandian rapakivi bodies as suggested by Salminen et al. (2021) [11]. A block model, adapted from Pirajno and Santosh (2015) [74], references the structural crustal position of the Märjamaa and Kloostri bodies within a red square. Interpretive block diagrams (note: the shapes and dimensions of the magma chamber and collapsing blocks are approximations and not to scale) illustrate the emplacement mechanism, inferred magma flow paths, and associated strain patterns in the Märjamaa and Kloostri rapakivi granitoids. The tripartite emplacement is described as follows: (b) Proto-Phase I chamber: a steep-sided, tabular intrusion formed by magma ascent through a southeast conduit during PPDZ-related extension.; (c) Piston-cauldron subsidence: roof collapse along ring faults due to magma decompression, inducing vertical and convergent subhorizontal flow.; (d) Phase I-II development: Phase I consolidates through block assimilation, while Phase II intrudes peripherally and cross-cuts earlier structures during piecemeal collapse; (e) Phase III emplacement, a satellite wedge-shaped body formed by asymmetric (trapdoor) subsidence, reflecting late-stage intrusion during crustal relaxation. The legend for the Märjamaa and Kloostri figures b to e is approximate in scale, with the principal purpose being to enhance the geodynamic understanding of the emplacement system.

6 Conclusions

This dissertation integrates multi-scale geophysical and geochemical datasets to refine models of crustal architecture, tectonic evolution, and magmatic processes in the concealed Precambrian basement of Estonia. The thesis establishes a coherent framework for tectonic assembly along Baltica's Palaeoproterozoic Svecofennian margin, demonstrates the role of inherited shear-zone fabrics in structuring the crust, and enhances how Mesoproterozoic reactivation modified lithospheric configuration in this understudied sector of Fennoscandia.

Regional integration of WGM-12 gravity, EMAG2v3 magnetic, and EGT national potential field datasets delineates six tectono-petrological basement domains with distinct petrophysical signatures. Spectral and derivative analyses resolve the main crustal layers, constrain their thicknesses and geometries, and map NW-SE geophysical lineaments that intersect the dominant NE-SW Svecofennian collisional grain. These intersection zones coincide with elevated heat flow and shallow Curie Point Depth (CPD) near the PPDZ and rapakivi granitoids, localized crustal thinning, and lithospheric segmentation. Collectively, the results validate the effectiveness of coupled gravity-magnetic interpretation for recovering deep structural inheritance and domain-scale architecture beneath Phanerozoic cover (*Publication I*).

Whole-rock geochemical constraints from the Alutaguse zone indicate a primary tectonic affinity with the Uusimaa-Tallinn belt and support its interpretation as a rifted back-arc domain predating Bergslagen accretion. The metasedimentary rocks span a wide compositional range from silica-poor to silica-rich varieties, reflecting variable contributions from mafic-intermediate and felsic sources related to rifting and arc-derived sedimentation. Metavolcanic rocks in Alutaguse display predominantly mantle-derived, tholeiitic signatures, in contrast to the calc-alkaline, subduction-modified magmatism characteristic of the Southern Svecofennian belts. Elevated Ca-Mg-Mn contents, tectonic discrimination diagrams, and key elemental ratios place Alutaguse and Uusimaa within a coupled arc-back-arc framework, consistent with a double-subduction system culminating in Bergslagen accretion at ~1.87–1.86 Ga. This interpretation strengthens the genetic linkage between Alutaguse and Uusimaa and is further supported by significant Pb-Zn-Cu signatures comparable to mineralization documented in the Bergslagen and Orijärvi zones (*Publication II & Manuscript II*).

At the local scale, 3D gravity and magnetic inversion models over the principal Alutaguse exploration targets (Assamalla, Haljala, and Uljaste) resolve laterally extensive, tabular subsurface bodies characterised by sharp contrasts in density and magnetic susceptibility. These bodies define stratiform to sheet-like geometries embedded within the Paleoproterozoic basement and highlight pronounced petrophysical heterogeneity within the Alutaguse back-arc assemblage. The inversion results constrain depth extent, internal architecture, and structural controls, providing a geophysically robust basis for delineating prospective zones and linking basement structure to metallogenic potential (*Manuscript III*).

The Märjamaa-Kloostri rapakivi granitoids are interpreted as an oxidised A₂-type suite emplaced under transtensional conditions along the PPDZ, linked to Mesoproterozoic superearth activity. Structural, geophysical, and geochemical evidence indicates that magma ascent and emplacement were strongly controlled by reactivation of Svecofennian shear-zone architecture along the Åland-PPDZ corridor, facilitated by mantle heat input, mafic underplating, and crustal extension between 1.67 and 1.41 Ga. The three-phase magmatic evolution is expressed by progressive enrichment in Si and alkalis and depletion in Fe, trace elements, and REE: (i) a ferroan, magnetite-rich Phase I forming a steep-sided tabular body

with high trace-element and REE concentrations; (ii) a transitional Phase II associated with piston–cauldron subsidence and feldspar-dominated differentiation; and (iii) a late, Na-rich leucogranitic Phase III emplaced during crustal relaxation. These results demonstrate how Mesoproterozoic extension exploited inherited Svecofennian fabrics to generate a mechanically and chemically zoned rapakivi system (*Publication III & Manuscript I*).

Overall, the thesis presents a geodynamic synthesis of the Precambrian basement in Estonia, explicitly linking lithospheric segmentation, Svecofennian structural inheritance, and Mesoproterozoic magmatism into a unified evolutionary model. By coupling regional and local-scale potential-field architecture with petrogenetic and tectonic interpretations, the dissertation clarifies the development of Estonian Fennoscandian crust and establishes a foundation for future 3D geodynamic simulations and metallogenic targeting in northern Estonia and the broader concealed Baltic shield (*Manuscript III*).

List of Figures

1	Geological framework of the Estonian Paleozoic units.	11
2	Geological Proterozoic setting of the study zone, highlighting key features of the dissertation.	13
3	Alutaguse and Jõhvi geophysical datasets (Bouguer and magnetic anomalies)	21
4	Geophysical datasets and drill core distribution of the Märjamaa–Kloostri pluton.	22
5	Alutaguse domain drill core locations and representative PPX images.	29
6	Boxplots and Harker diagrams of Alutaguse and SS (South Svecofennian) samples.	34
7	Major element distributions of tripartite rapakivi granite phases.	36
8	Lineament density maps and heat flow from residual potential field models.	37
9	Bivariate average geochemical trends of Alutaguse and SS (South Svecofennian) zones.	39
10	Geochemical discrimination diagrams of Alutaguse and SS (South Svecofennian) zones	41
11	Schematic geodynamic evolution model of Fennoscandia (South FinalInd-North Estonia).	44
12	Regional and residual potential-field maps of NE Estonia.	45
13	SimPEG cross-gradient joint inversion results for Alutaguse target zones	46
14	Regional and residual potential-field maps of the Märjamaa–Kloostri rapakivi bodies.	49
15	Residual potential-field lineament analysis and tectonic classification in Estonia.	51
16	SimPEG joint inversion results highlighting the Märjamaa–Kloostri system. .	51
17	Finnish–Estonian rapakivi bodies and Harker plots of the Wiborg suite.	54
18	Major-element geochemical classification diagrams of the Märjamaa–Kloostri rapakivi.	56
19	Trace-element and REE diagrams of Märjamaa–Kloostri rapakivi samples. ..	58
20	Residual lineament distribution by phase from Bouguer and RTP data.	59
21	Geodynamic emplacement model of the Märjamaa–Kloostri rapakivi system.	62

List of Tables

1	Major element averages and ranges for Alutaguse and South Svecofennian (SS) lithologies.	25
2	Trace and REE element averages of Alutaguse and South Svecofennian (SS) lithologies.	28
3	Major element data and geochemical parameters of the Märjamaa rapakivi samples.	32

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Abstract

Crustal Structure, Tectonics, and Magmatic Evolution of the Estonian Proterozoic Basement: Geophysical and Geochemical Insights

The Estonian Precambrian crystalline basement, concealed beneath ~100–900 m of Phanerozoic cover, forms part of the Fennoscandian Shield within the Proterozoic continent Baltica. Owing to limited exposure, its crustal architecture, tectonic evolution, and magmatic history remain incompletely understood. This dissertation investigates how inherited Paleoproterozoic crustal structures influenced subsequent lithospheric segmentation and Mesoproterozoic magmatism in the Estonian basement.

The study integrates gravity and magnetic potential-field modelling with whole-rock geochemistry and petrological analysis to address three principal objectives: (1) to define the crustal architecture and lithospheric segmentation of Estonia using major geophysical discontinuities; (2) to evaluate whether the Alutaguse zone represents a continuation of the Southern Svecofennian system or a distinct tectonic entity; and (3) to determine how Svecofennian structural fabrics controlled the emplacement, geometry, and geochemical evolution of the Märjamaa–Kloostri rapakivi granitoids along the Paldiski–Pskov Deformation Zone.

Processing and modelling of Bouguer gravity and magnetic anomaly datasets reveal a Moho depth of approximately 60 km and a Conrad discontinuity at ~18 km. Spectral separation and derivative filtering delineate six petrological basement domains, highlighting dominant NW–SE trending lineaments that are subparallel to the Paldiski–Pskov Deformation Zone and oblique to the Fenno-Sarmatian collision front. Elevated magnetic-derived heat-flow estimates coincide with this deformation zone and Mesoproterozoic rapakivi intrusions, indicating lithospheric weakening and reactivation.

Whole-rock geochemical data from Paleoproterozoic metasedimentary and metavolcanic rocks in the Alutaguse and Southern Svecofennian domains indicate contrasting source characteristics and tectonic settings. High-SiO metasediments record felsic, arc-related provenance, whereas low-SiO types reflect mafic to intermediate rift-related sources. Trace-element systematics and elemental ratios distinguish asthenospheric mantle input in Alutaguse from subduction-modified magmatism in the Southern Svecofennian domain. These results support the interpretation of the Alutaguse zone as a 1.90–1.89 Ga back-arc basin related to the Tallinn–Uusimaa belt, with closure during accretion to proto-Fennoscandia between 1.89 and 1.87 Ga.

Integrated geophysical inversion and geochemical modelling of the ~1.62 Ga Märjamaa–Kloostri rapakivi granitoids reveal a multiphase intrusion emplaced along reactivated Svecofennian shear zones. Their internal architecture reflects transtensional pull-apart dynamics, with successive magma pulses forming piston-cauldron and trapdoor-style structures. Geochemical trends indicate progressive differentiation under oxidising conditions, consistent with A-type AMCG magmatism driven by asthenospheric upwelling during Nuna breakup.

Overall, this dissertation demonstrates that Paleoproterozoic crustal inheritance exerted a fundamental control on Mesoproterozoic magmatism and lithospheric evolution in Estonia. The results provide a coherent geodynamic framework for the concealed Fennoscandian basement, with direct implications for regional tectonic reconstructions and metallogenic exploration.

Keywords: Estonian Precambrian basement; crustal architecture; Paleoproterozoic back-arc basin; rapakivi granitoids; potential-field inversion; whole-rock geochemistry; Fennoscandian Shield.

Kokkuvõte

Eesti proterosoilise maakoore ehitus, tektoonika ja magmaliste protsesside areng: geofüüsikaline ja geokeemiline vaade

Eesti eelkambriumi vanusega kristalne aluskord, mida katab 100–900 m paksune paleosoi-kumi vanusega settekiivimite kate, moodustab osa Fennoskandia kilbist – proterosoikumi-aegsest mandrist Baltika.

Piiratud paljandumise tõttu on Eesti maakoore arhitektuur, tektooniline areng ja mag-maatiline ajalugu endiselt puudulikult mõistetud. Käesolev dissertatsioon uurib, kuidas päritud paleoproterosoikumi maakoore struktuurid mõjutasid hilisemat litosfääri segmen-teerumist ja mesoproterosoikumi magmatismi Eesti aluskorras.

Uuring ühendab gravitatsiooni- ja magnetiliste potentsiaalväljade modelleerimise ko-gukivimi geokeemia ja petroloogilise analüüsiga, et saavutada kolm peamist eesmärki: (1) määratleda Eesti maakoore arhitektuur ja litosfääri segmenteerumine peamiste geo-füüsikaliste katkestuste alusel; (2) hinnata, kas Alutaguse tsoon kujutab endast Lõuna-Svekofennia süsteemi jätku või eraldiseisvat tektoonilist üksust; ning (3) selgitada, kuidas Svekofennia struktuurid kontrollisid Märjamaa–Kloostri rapakivi granitoidide paigutust, geomeetriat ja geokeemilist arengut Paldiski–Pihkva deformatsioonivööndi lähedal.

Bouguer' gravitatsiooni- ja magnetanomaaliade andmestike töötlemine ja modelleeri-mine viitavad ligikaudu 60 km sügavusel paiknevale Moho piirile ning umbes 18 km süga-vusel asuvalle Conradi katkestusele. Spektraalne eraldamine ja tuletiste filtreerimine eris-tavad kuus petroloogilist aluskorra domeeni ning toovad esile domineerivad loode–kagu suunalised lineamendid, mis on peaaegu paralleelsed Paldiski–Pihkva deformatsioonivöön-diga ja nurga all Fenno-Sarmatia kokkupõrkerinde suhtes. Magnetiliselt tuletatud kõrgen-datud soojusvoo hinnangud langevad kokku selle deformatsioonivööndi ja mesoprotero-soikumi rapakivi intrusioonidega, viidates litosfääri nõrgenemisele ja taasaktiveerumisele.

Paleoproterosoikumi metasedimentaarse ja metavulkaaniliste kivimite kogukivimi geokeemilised andmed Alutaguse ja Lõuna-Svekofennia domeenides näitavad selgelt eri-nevaid magma algallikaid ja tektoonilist keskkonda.

Metasedimentaarsed kivimid moodustavad laia geokeemilise spektri ulatudes kõrge ränisisaldusega kuni madala ränisisaldusega kivimiteni, mis tõenäoliselt peegeldavad eri-nevusi ka magmade algaalikes.

Mikroelementide süsteemaatika ja elemendilised suhted eristavad Alutaguse astenos-fäärse vahevöö sisendit Lõuna-Svekofennia domeeni subduktsioonist mõjutatud magma-tismist. Need tulemused toetavad Alutaguse tsooni tõlgendamist kui 1.90–1.89 Ga va-nust saarkaarte-tagust basseini, mis on seotud Tallinna–Uusimaa vööndiga ning mis sulgus proto-Fennoskandia akretsiooni käigus ajavahemikus 1.89–1.87 Ga.

Ligikaudu 1.62 Ga vanuseliste Märjamaa–Kloostri rapakivi granitoidide integreeritud geofüüsikaline inversioon ja geokeemiline modelleerimine viitavad mitmefaasilisele intru-sioonile, mis paikneb piki taasaktiveeritud Svekofennia nihkevööndeid. Rabakivigraniiti-de sisemine arhitektuur peegeldab transtensionaalset tõmberežiimi, milles järjestikused magma intrusioonid kujundasid *piston* – *cauldron* ja trapdoor tüüpi struktuure.

Geokeemilised trendid osutavad oksüdeerivates tingimustes toimuvale progresseeru-vale magmalisele diferentseerumisele, mis on kooskõlas A-tüüpi AMCG magmatismiga ning mida toetas astenosfääri ülesvool Nuna superkontinendi lagunemise ajal.

Kokkuvõttes näitab dissertatsioon, et paleoproterosoikumi maakoore areng on aval-danud olulist kontrolli ka mesoproterosoikumi-aegse magmatismi ja litosfääri evolutsiooni üle Eesti aluskorras. Tulemused pakuvad sidusat geodünaamilist raamistikku vähe-uuritud

Fennoskandia aluskorra mõistmiseks ning omavad otsest tähtsust piirkondlike tektooniliste rekonstruktsioonide ja metallogeensete uuringute jaoks.

Märksõnad: Eesti eelkambriumi aluskord; maakoore arhitektuur; paleoproterosoikumiaegne saarkaarte-tagune bassein; rapakivi granitoidid; potentsiaalivälja inversioon; kogu-kivimi geokeemia; Fennoskandia kilp.

Appendixes

Appendix I

J D Solano-Acosta, A Soesoo, and R Hints. New insights into the crustal structure across Estonia using satellite potential fields derived from WGM-2012 gravity data and EMAG2v3 magnetic data. *Tectonophysics*, 846:229656, 2023. [doi:10.1016/j.tecto.2022.229656](https://doi.org/10.1016/j.tecto.2022.229656)



New insights of the crustal structure across Estonia using satellite potential fields derived from WGM-2012 gravity data and EMAG2v3 magnetic data

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ABSTRACT

Global WGM-12 gravity and EMAG2v3 magnetic models were used to present new information of the Estonian crust. The Estonian Precambrian crystalline basement, composed of Paleo-Meso Proterozoic metamorphic and igneous rocks, is covered by a Paleozoic sedimentary rock deposit 100–800 m thick. The present work aims to map structural patterns and recognise physical layers of the crust over Estonia using global potential models. The gravimetric data was used to identify the depth of the Moho and Conrad discontinuities. The magnetic data has been processed to calculate the Curie point depth (CPD), which was then used to estimate heat flow (HF) values in the study area. The depth of the Moho indicates an average value of 60 km, while the average depth of the Conrad discontinuity is around 18 km. The depth of the CPD reveals an average value of 15 km. Different derivative techniques were applied over the residual part of both potential models, to delineate and map geophysical lineaments. The subsurface of Estonia has been divided into six petrological-structural zones: Tallinn, Alutaguse, Jõhvi, West-Estonian, Tapa and South-Estonian. Geophysical lineaments in each zone subsurface division delineates a NW-SE trend, perpendicular to the NE-SW oriented Fenno-Sarmatian collision. To visualise the anomalous crustal variations over Estonia, different profiles showing varying values of potential fields, CPD and heat flow are presented, especially around the Precambrian Rapakivi granitic plutons and in the Paldiski-Pskov deformation zone, due to the high contrast observed by mapping the different results over these areas.

1. Introduction

Gravity and magnetic potential methods are useful geophysical techniques for identifying and mapping subsurface geological structures, and are commonly used to delineate lineaments and deep crustal layers (Forsberg and Olesen (2010); Hu et al. (2015); Altinoğlu et al. (2015); Essa et al. (2018); Khazri and Gabtni (2018); Bba et al. (2019); Dilalos et al. (2019); Osagie et al. (2021); Peredo et al. (2021); Omietimi et al. (2021); Dumais et al. (2021)). In Estonia previous geophysical researches have been carried out (Ankudinov et al. (1994); All et al. (2004); Oja et al. (2019); Soesoo et al. (2020); Plado et al. (2020)), however some of the gravity and magnetic data corresponds to data of the last century (Maasik (1959); All et al. (2004)), and there are no recent publications of potential field trends, or crust structure, not to mention the lack of geophysical analysis derived from global models based on satellite data (i.e., WGM 2012 gravity model (Bonvalot et al. (2012)) and EMAG2v3 magnetic model (Meyer et al. (2017))).

The gravity and magnetic surveys are a non-destructive remote sensing geophysical methods to determine rock physical changes in the

subsurface geology (i.e. density and remanent magnetisation of materials). As a result, they are a sum of regional and residual potential anomalies (Petit et al. (2002)). The Bouguer anomaly (BA) represents the response of any density change below the surface (Karcol et al. (2017)), and has been widely used to understand subsurface structures and trends, as well as depth-to-basement and its structures, and sedimentary basin infills (Olesen et al. (2010); Nasuti et al. (2012); Fairhead (2012); Karcol et al. (2017); Khazri and Gabtni (2018); Bba et al. (2019); Dilalos et al. (2019); Omietimi et al. (2021)).

Induced and remnant magnetisation of the crust and upper mantle are the primary sources of crustal magnetic fields. The difference in magnetic mineral concentration is a good indicator of the structure and composition of the crust (Hope and Eaton (2002); Aitken and Betts (2008)). Magnetic anomaly maps provide insight into the subsurface structure and composition of the Earth's crust (Nakanishi et al. (1992); Golynsky et al. (2002); Purucker and Whaler (2007); Quintero et al. (2019)). Over continental areas, magnetic anomalies illuminate geologic, tectonic, and geothermal evolution of crust and lithosphere (Milligan et al. (2003); Hemant and Maus (2005); Aitken and Betts

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(2008); Blaich et al. (2009); Fairhead (2012); Li et al. (2017)).

Gravity and Magnetic measurements from satellites, ships and aircraft were used to create the World Gravity Model 2012 (WGM-12) Bouguer model (Bonvalot et al. (2012)) and the EMAG2v3 magnetic model 2017 (Meyer et al. (2017)). Both potential models were selected for use in this research because they have proven to be suitable for detecting lateral variations and crustal structures and discontinuities (Osagie et al. (2021); Omietimi et al. (2021); Saada et al. (2021)). Potential field data have the advantage of providing essentially continuous coverage from onshore and offshore (Olesen et al. (2010); Nasuti et al. (2012); Dumais et al. (2021)). Comprehensive analysis of potential data allows the isolation of signals and, eventually, the mapping of deep crustal structures (Likkason (2011)).

The main goal of this work is to map the magnetic and gravity

anomalies, the physico-mechanical layers of the Estonian crust (i.e. Curie depth point, Moho and Conrad discontinuities, and shallow anomalies), and geophysical lineaments. Potential anomaly processing and matching filters are used in this research, allowing the shallow and deep geophysical responses of these structures and domains to be compared. The geophysical modelling and interpretations provided here will help to better understand the physical characterisation of the Precambrian Estonian basement and the possible overprinting of the Svecofennian orogeny, possibly during Fenno-Sarmatian collision (1.82–1.80 Ga) contributing with the debate on tectonic amalgamation theories (All et al. (2004); Bogdanova et al. (2006, 2015)).

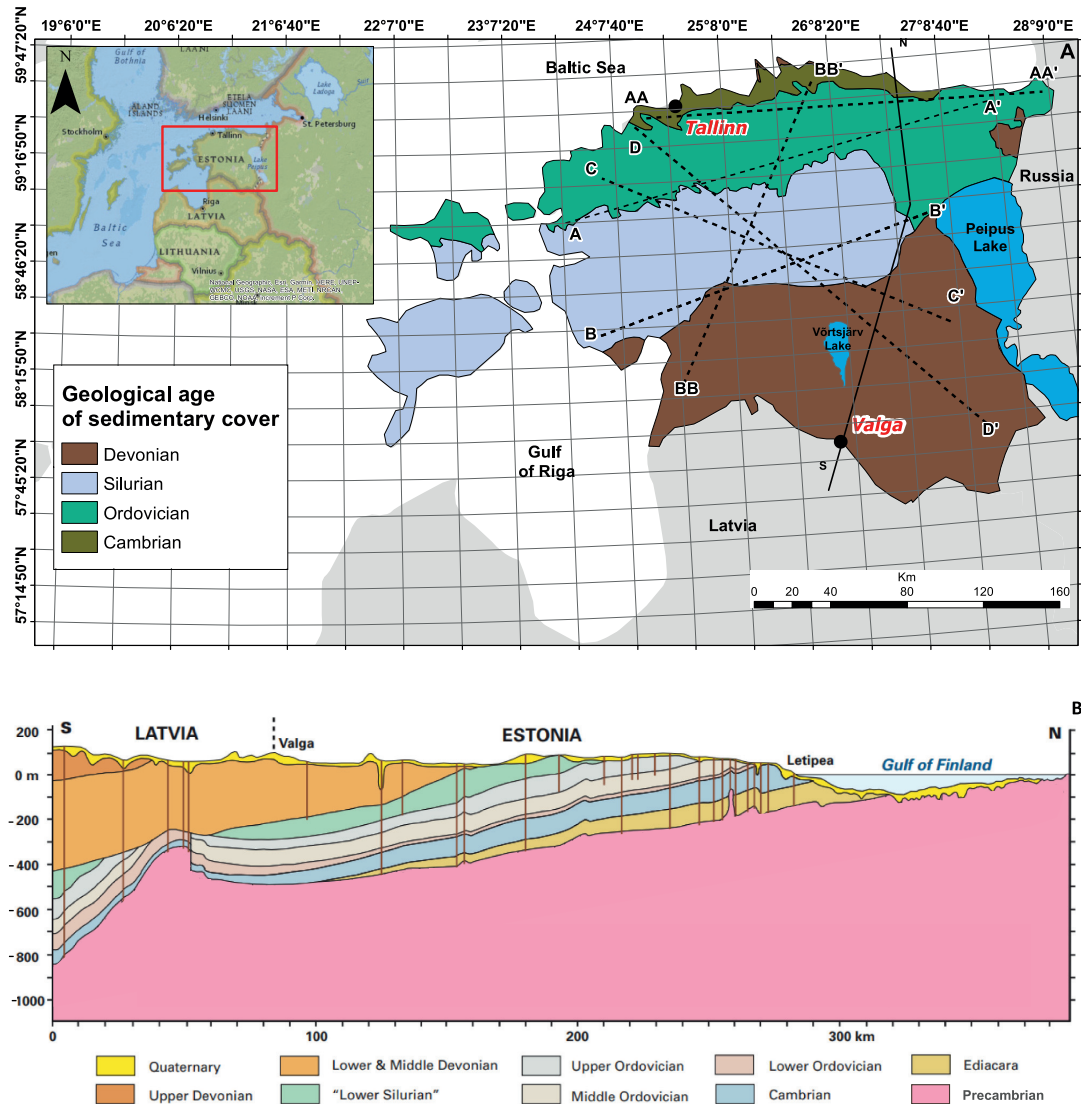


Fig. 1. Geology schemes of Estonia. A) Generalised map with the sedimentary cover of Estonia showing outcrop areas of lower and middle Paleozoic rocks (after Puura and Vaher (1997)). Dash lines correspond to the profiles used to analysed potential data (Fig. 16), and continued line corresponds to the cross-section profile showed in Fig. 1-B; B) Geological cross-section profile of Estonia modified from Bauert et al. (2015). Vertical lines correspond to drill cores.

2. Geological setting

Estonia is located on the southern slope of the Fennoscandian Shield, comprising Lower Paleozoic (Cambrian to Devonian) sedimentary strata with a thickness ranging from 100 m (in the north) to 800–900 m overlay the crystalline rocks (in the south and SW Estonia), where the thickness of the sedimentary layer increases southwards. The sedimentary bedrocks are composed of shallow marine sandstones, mudstones, and carbonates. The crystalline rocks developed around 1.9–1.5 Ga, the sedimentary bedrocks approximately 550–360 Ma, and the uppermost Quaternary layers originated during the glacial period (Kirs et al. (2009)). The Estonian sedimentary cover is depicted in the Fig. 1.

The initial thickness of the sedimentary cover was likely higher, but the rocks were later subjected to extensive erosion (Kirsimäe et al. (1999)). The resulting topography displays cores of several uplands that can be observed in modern topography, and the depressions of lakes Peipus and Võrtsjärv and the Gulf of Finland (Tavas (1997); Tuuling (2017); Ani and Meidla (2020)).

In terms of structure, Estonia is located in the centre region of the Baltica paleocontinent and is part of the East European craton, lying for the most part within the boundaries of the southern slope of the Fennoscandian shield (Raukas and Teedumäe (1997); Soesoo et al., 2020; Bogdanova et al., 2015); Fig. 2). In contrast to the majority of the Fennoscandian shield region, the patterns of the Palaeoproterozoic crust beneath the platform cover in Belarus, Estonia, Latvia, Lithuania, and Poland are marked by a prominent array of alternating belt-shaped tectonic domains made up of granulite or amphibolite facies rocks (Bogdanova et al. (2015)). Many of these belts, which formed relatively

late in the history of the Svecofennian orogeny, have been thought to reflect crust heaped tectonically during the collision of Fennoscandia and Volgo-Sarmatia, known colloquially as the Fenno-Sarmatian collision, which occurred about 1.82–1.80 Ga (Bogdanova et al. (2006)). Typically, metamorphic belts span structurally defined tectonic domains and terranes that each had their unique pre-collisional Palaeoproterozoic histories and are mostly limited by zones of substantial shearing and faulting (Skridlaite and Motuza (2001); Taran and Bogdanova (2001); Puura et al. (2004); Soesoo et al. (2004); Bogdanova et al. (2006, 2015)).

Precambrian rocks do not outcrop anywhere in Estonia but are present in the subsurface, so drill core material, geophysical and geochemical methods have been used for geological reconstructions (Soesoo et al. (2020); Fig. 2). The main features of the regional crustal structure were formed during the Svecofennian orogeny (Bogdanova et al. (2015)). Estonia is situated on a thick crust (Puura and Vaher (1997)). Most of the crystalline basement comprises the Palaeoproterozoic (1.8–1.9 Ga) Svecofennian orogenic complex, subjected to high-grade metamorphism. The gneisses of amphibolite facies in northern Estonia represent an extension of the major structural units of southern Finland. Tallinn, Alutaguse, Jõhvi, West-Estonian, Tapa, and South-Estonian structural-petrological zones constitute the division of the basement (Soesoo et al. (2004, 2020)) (Fig. 2). These zones differ in terms of rock composition, origin, geophysical features, and metamorphism degree (Bogdanova et al. (2015)). The structural domains of northern and north-eastern Estonia are separated from those of the west and south by the almost 30 km wide, NW-trending (Åland)–Paldiski-Pskov deformation zone (Known as PPDZ); made up of two major sub-

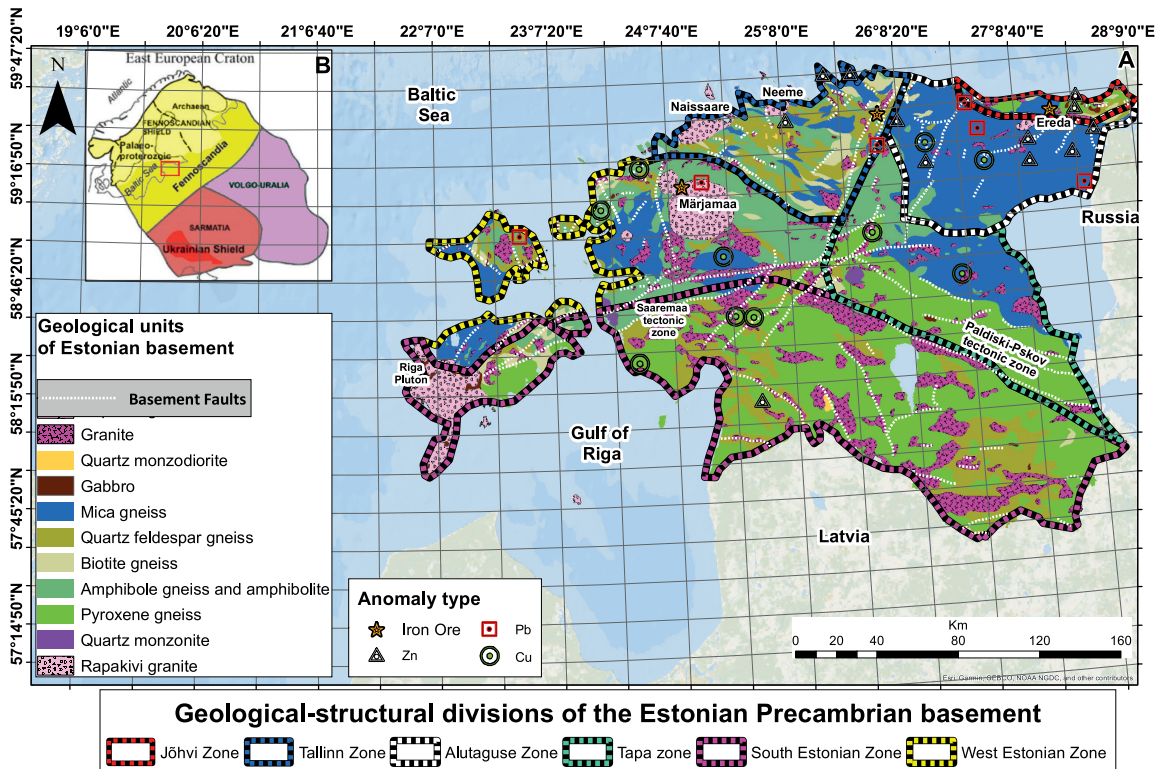


Fig. 2. A) Geological map scheme of the Precambrian basement of Estonia, showing geochemical anomalies according to Soesoo et al. (2020). The geological data can be downloaded from the Geoportal of the Land Board of the Republic of Estonia (<https://geoportaal.maaamet.ee/eng/Spatial-Data/Geological-Data-p317.html>); B) Crustal segments of the East European craton (modified after Gorbachev and Bogdanova (1993)).

parallel shear zones that dip between 65° to 75° SSW, and the E-W striking Saaremaa deformation zones (Fig. 2), have significant deformation, and contains slivers of both granulite and amphibolite-facies rocks (Soesoo et al. (2020)), mirroring the rocks in the southern portion of the Bergslagen tectonic domain in Sweden (All et al. (2004)). Tallinn, Tapa, and Alutaguse Precambrian divisions in NNE Estonia, are composed of metasedimentary and meta-volcanic rocks of approximately 1.90 Ga age, which closely mirror the rocks of neighbouring belts in southern and western Finland, as well as those in central Sweden (Bogdanova et al. (2015)). The Tallinn domain seems to belong to a volcanic arc that continues into the Finnish Uusimaa belt (Kähkönen (2005)). The metasedimentary Alutaguse area might be part of the enormous Kalevian-age peripheral basin that stretches from St. Petersburg in Russia to Lake Ladoga in the east. Turbidite deposition in the basin is believed to have occurred between 1.96 and 1.90 Ga (Lahtinen et al. (2010); Melezhik and Hanski (2013)). Nonetheless, the dates of deposition and metamorphism of the Alutaguse metasedimentary sequence are unknown and must be determined before its tectonic context can be determined (Bogdanova et al. (2015)). A thick sequence of alternating Fe- and S-rich garnet pyroxene bearing quartzites, high-Al garnet-cordierite-sillimanite gneisses, and Ca-rich and Ca-poor pyroxene-amphibole and biotite gneisses may be found in the distinct Jõhvi domain. These rocks suffer migmatization when they were intruded by a granite about 1.8 Ga, with similar ages of late- and/or post-orogenic granites in southern Finland (Nironen (2005); Kurhila et al. (2011)). Multiple high-metal-concentration geochemical anomalies have been discovered in the Estonian basement (Fig. 2) (Soesoo et al. (2020)).

3. Data and methodology

3.1. Data

There are various global gravity models with spatial resolutions in the range of 10×10 km, such as EIGEN-6C4 (Förste et al. (2014)), EGM 2008 (Pavlis et al. (2012)), and XGM 2019 (Zingerle et al. (2020)). These models have been used in a variety of gravity studies, including lineaments extraction (Kumar et al. (2019); Sahoo and Pal (2019); Chouhan (2020)), tectonic features extraction (Kumar et al. (2020); Sahoo and Pal (2021)), mineral exploration (Rani et al. (2020)), and density modelling (Andersen et al. (2010); Kahveci et al. (2019); Chouhan (2020); Kassa et al. (2021)). Nevertheless, the WGM-12 model is the first set of global gravity anomaly maps to use a realistic Earth model to analyse the contribution of most surface masses, including the atmosphere, land, oceans, inland seas, ice caps, and ice shelves (Balmino et al. (2012); Bonvalot et al., 2012).

The WGM-12 is a combined gravity dataset that includes data from satellites such as the Challenging Minisatellite Payload (CHAMP), Gravity Recovery and Climate Experiment (GRACE), EGM 2008, Gravity Field and Steady-State Ocean Circulation Explorer (GOCE), DTU 10 global gravity field, Global Topography 30 arc-second (GTOPO30), and in-situ gravity data (Bonvalot et al. (2012)). Those data sets are combined using the iterative least-squares approach, with a maximum spherical harmonic coefficient of up to 2190° and order 2159° (Balmino et al. (2012); Bonvalot et al. (2012)). The spatial coverage of the WGM-12 gravity data is averaged over 2.5 arc-min by 2.5 arc-min (Balmino et al. (2012)). Bonvalot et al. (2012) used the World Geodetic System 84 (WGS 84) as a datum, with a reference density of 2.67 g/cm^3 for crustal

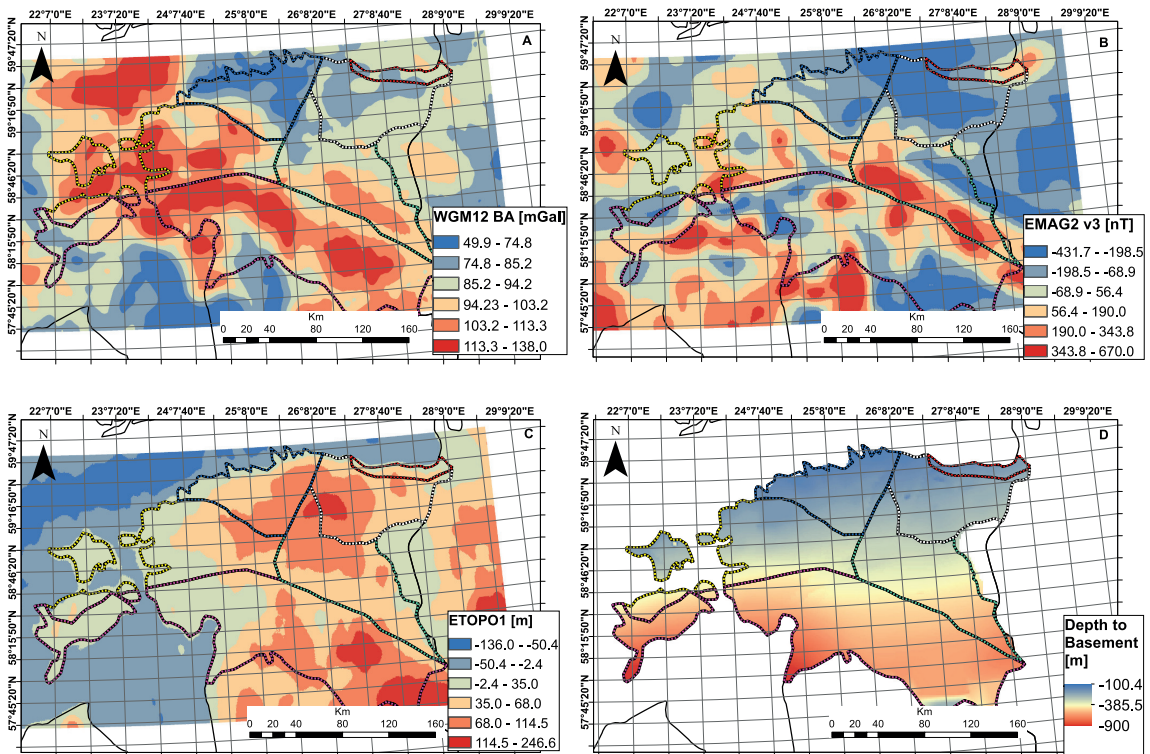


Fig. 3. Base data used in this research. A) EGM12 Bouguer anomaly (CBA); B) EMAG2v3 Magnetic anomaly; C) ETOPO1 Height altitude; D) Depth to the Basement from drill cores.

rocks. We chose WGM-12 for this research because it has already been used effectively by other research groups to explore basement depth and structures (Peredo et al. (2021); Omietimi et al. (2021); Saada et al. (2021)). The WGM-12 Bouguer model of Estonia (Fig. 3-A) was gathered from the International Gravimetric Bureau (BGI) gravity database (<http://bgi.obs-mip.fr/data-products/grids-and-models/wgm2012-global-model/>). Magnetic anomalies are variations (higher or lower) of the mean magnetic field for a given location. The interaction of the geomagnetic field with rocks in the Earth's crust causes magnetic anomalies to be observed during aeromagnetic surveys (AMS) (Reeves (2005)). The magnetic characteristics of these rocks are due to induction by an external magnetic field or to natural remnant magnetisation (NRM) caused by the existence of a previous external inducing field. The magnetic susceptibility of the rock influences the degree of magnetisation in the presence of the earth's magnetic field (Reeves (2005); Likkason (2011); Osagie et al., 2021). The anomalies observed with AMS are mainly attributed to the presence of naturally occurring magnetite (Fe_3O_4), given its significant remanent magnetisation and its high magnetic susceptibilities. The contribution of other minerals to the magnetic anomalies detected during AMS is minor on a global scale and is often overlooked such as pyrrhotite ($Fe_{1-x}S$) (Reeves (1985); Grant (1985a, 1985b)). About 1.5% of the earth's crust has ferromagnetic magnetite, being the most essential parameter in geophysical mapping (Clark and Emerson (1991)).

EMAG2v3 is the newest version of the EMAG2 that is an Earth Magnetic Anomaly grid at 2 arc-min resolution, which dataset was constructed from a combination of airborne, marine and satellite magnetic dataset of CIRES (Cooperative Institute for Research in Environmental Sciences) (Maus et al. (2009)). The current EMAG2v3 includes more than 11.5 million new ship and aircraft tracking data, as well as several new or updated pre-compiled dataset grids. Whereas the previous EMAG2 2009 version interpolated anomalies at non-existent data locations using known or idealised local geology, the new EMAG2v3 is based solely on the data provided. Consequently, EMAG2v3 adequately portrays locations with missing data and better illustrates the complexity of these anomalies (Meyer et al. (2017)). total field magnetic anomaly (TFMA) EMAG2V3 data (Fig. 3-B) was downloaded from the National Oceanic and Atmospheric Administration (NOAA) geomagnetism database (<https://www.ngdc.noaa.gov/geomag/emag2.html>). It comprises almost 50 million additional data points added in recent years to NCEI's Geophysical Database System (GEODAS). The satellite lithosphere magnetic field model, MF7, was used for the longest wavelengths (≥300 km) (Maus et al. (2009)). For all maritime and coastal locations, this grid is given at 4 km elevation above the geoid as well as at sea level (Meyer et al. (2017)). The inclusion of more than 50 million new magnetic data points, the incorporation of a new global magnetic model (MF7), and the use of no a priori information on oceanic magnetic anomalies are all improvements over the previous magnetic model (Meyer et al. (2017)). In addition, the stripe magnetic pattern was appropriately reduced in this EMAG2 version, although magnetic anomalies in the continental areas show a similar pattern (Osagie et al. (2021)).

Elevation data (Fig. 3-C) were obtained from the ETOPO1 global relief model that integrates all available land topography and bathymetry data into a 1 arc-min grid (Amante and Eakins (2009)). The depth to the basement (Fig. 3-D) was taken from contour lines with distance values from the surface to the basement, values which were obtained from drill core data, and which can be downloaded from the website of the Geoportal of the Land Board of the Republic of Estonia (<https://geoportal.maaamet.ee/eng/Spatial-Data/Geological-Data-p317.html>). For all data sets processing and quantitative interpretation we used the Oasis Montaj (Geosoft®; Geosoft, 2004) software and ArcMap v10.8 for data integration and qualitative and quantitative interpretations.

For all data processing the projection used was that of the Universal Transverse Mercator coordinate system in zone 34 of the northern hemisphere (UTM zone 34°N). It is noteworthy that this study focuses

mainly on the theoretical aspects and sensitivity analysis of the potential models, mapping the different structures of the Estonian crust.

3.2. Gravity and magnetic data processing

The flowchart (Fig. 4) illustrates the entire process followed in this study. To utilise the aforesaid data to understand the nature of the Estonian crust, regional-residual anomaly separation techniques must be used to extract gravity and magnetic anomalies attributable to crustal density and/or magnetic susceptibility sources (Mickus et al. (1991)). Longer wavelength anomalies are often induced by deeper and larger sources, whereas shorter wavelength anomalies are typically caused by shallower sources (Lowrie and Fichtner (2020); Saada et al., 2021). The regional anomalies were estimated using derivatives, wavelength, and phase match filtering (Lowrie and Fichtner (2020)). In this research a wavelength-cut was done by using the High/Low pass filter of MAGMAP tool of Geosoft Oasis Montaj software (Geosoft (2004)), as will be fully discussed subsequently. The residual grids for gravity/magnetic data were obtained by removing the regional field from the filtered data (Osagie et al. (2021)), and allow to recognise shallower crustal density sources (Saada et al. (2021)), meanwhile the regional represent deeper sources (i.e., Moho discontinuity) (Hu et al. (2015)). To obtain the values of deep distance of the Moho and Conrad discontinuities, as well as the shallowest recorded signals, the Complete Bouguer Anomaly (CBA) data was used, as it will be explained in the next sections. For the magnetic data, the reduction-to-pole (RTP) transformation was implemented prior to the separation of the regional and residual field.

Only at the magnetic poles or at the equator does a symmetric TFMA appear in a symmetric magnetised body. However, at various latitudes between the extremely high polar zones and the very low equatorial regions, the magnetised body will show an asymmetric TFMA – horizontally offset with respect to the body (Aina (1986)). The RTP transformation converts the complex mid-latitude asymmetric TFMA to the polar symmetric counterpart with zero horizontal displacement (Blakely (1996)). The RTP anomaly is the theoretical value recorded at the geomagnetic pole as a result of vertically magnetised material. The RTP approach presupposes that the magnetisation direction is consistent across the causative body and that the geomagnetic field is homogeneous throughout the region of inquiry, which may not be true in broad areas (Osagie et al. (2021); Blakely (1996)). The RTP transformation helps to centre the peaks of TFMA over their sources, and it was performed within the Oasis Montaj package (Geosoft (2004)). Simplifying, the RTP correction is often applied to the magnetic anomalies to obtain corrected maps of magnetic anomaly values induced by the inclination of 90° and the declination of 0°. For further understanding of the RTP algorithm see Baranov and Naudy (1964). The inclination and declination angles used for the application of the RTP are 72.531° and 9.394° respectively, and were obtained according to the International Geomagnetic Reference Field (IGRF) (Finlay et al. (2010)) for a latitude of 58.3°N and a longitude of 25.6°E, a coordinate that is approximately the midpoint of the study region. Fig. 6-B1 present the result of the RTP.

The wavelength filtering method was used by applying the High/Low pass filtering. To highlight the short wavelength anomalies induced by upper crustal sources, a high-pass filter was developed using a phase match filter (a sort of Fourier Transformed based wavelength filter) and applied to the Complete Bouguer Anomaly (CBA) data and the RTP anomaly data (Phillips (2001)). When modelling the observed anomalies of various source layers at different depths indicated by the spectral characteristics of the data, phase-matching filtering splits the data into different wavelength components (Phillips (2001)). The radial average power spectrum of the CBA data (Fig. 5-A) displays a threshold wavenumber value of 0.007 cycles/km, which is equivalent to source layer with a depth of 36.01 km. The lower wavenumbers (<0.007 cycles/km) represent the deeper or regional gravity anomalies whereas, the higher wavenumbers (>0.007 cycles/km) reflect the shallower or residual gravity anomalies. Regarding the radial average power spectrum of the

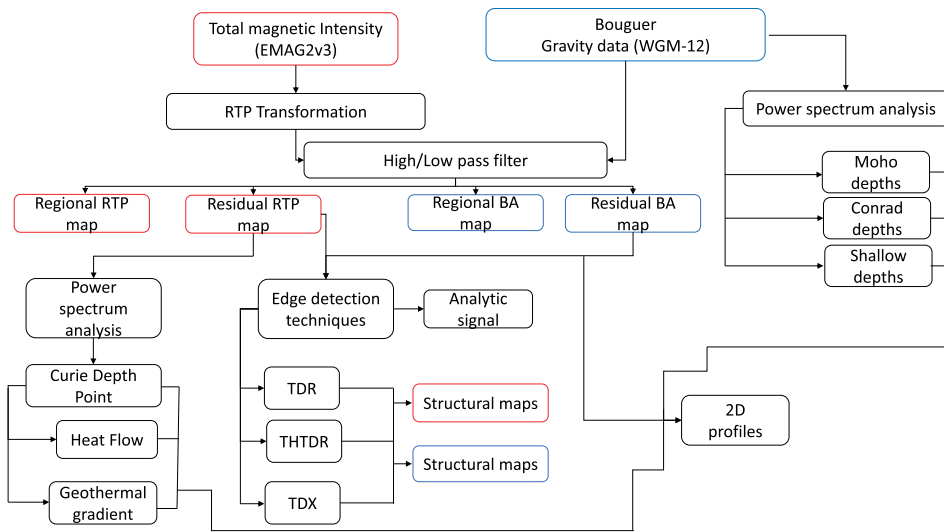


Fig. 4. The process used to get the relevant maps depicted in a flowchart.

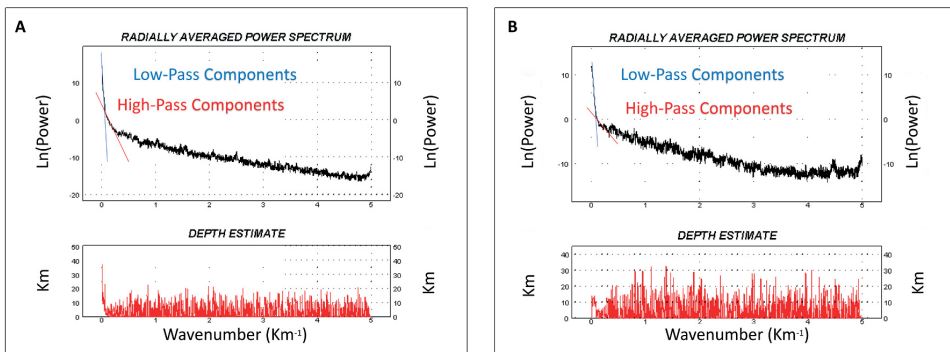


Fig. 5. A) The radial averaged power spectrum of the CBA data used to create residual and regional gravity anomaly maps.; B) The radial averaged power spectrum of the RTP data used to create residual and regional magnetic anomaly maps.

RTP data (Fig. 5-B) displays a threshold wavenumber value of 0.009 cycles/km, which is

equivalent to source layer with a depth of 15.13 km, value which is generally similar to the Curie Depth Point (CPD) (Saada (2016a)). By removing the regional trends, we are eliminating the effects of the core, mantle magnetosphere and induced fields that the data can present (Berguig et al. (2016); Quintero et al. (2019); Ngoh et al. (2020)). The lower wavenumbers (<0.009 cycles/km) represent the regional magnetic anomalies whereas, the higher wavenumbers (>0.009 cycles/km) reflect the residual magnetic anomalies. That depths are just an approximation to all the source depths, where the automated depth determination methods must have depth-weighting in order to image the sources at their correct depths (Fedi and Pilkington (2012); Saada et al. (2021)). The regional and residual potential anomaly maps can be observed in the Fig. 6.

We also used the pseudo-gravity (PSG) transformation to improve the regional component in magnetic anomalies (Fig. 6-A3) (Blakely (1996); Nasuti et al. (2012)). Baranov (1957) was the first to described an application of Poisson's relation in which the TFMA is converted into the gravity anomaly that would be observed if the magnetisation

distribution were to be replaced with an identical density distribution (i. e., is a constant throughout the source and magnetisation is induced and remanence is negligible). Here, the PSG transform was applied to the RTP grid using the Fast Fourier Transformation (FFT) filter package available in Oasis montaj (Pratt and Shi (2004)).

4. Theory and results

4.1. Moho and Conrad discontinuities from WGM-12 Bouguer anomaly

A two-dimensional collection of data indicating a potential field, such as gravity, can be viewed as a superposition of fields caused by a subsurface distribution of sources (Tselentis et al. (1988)). The technique of computing and understanding the spectrum of potential field data is known as spectral analysis. The spectral depth technique is based on the idea that a potential field measured at the surface may be thought of as an integral of anomaly signatures from all depths. The application of the power spectrum method to potential field data was proposed by Bhattacharyya, 1966; and the determination of the anomalous body depth was given by Spector and Grant (1970). The CBA data may be

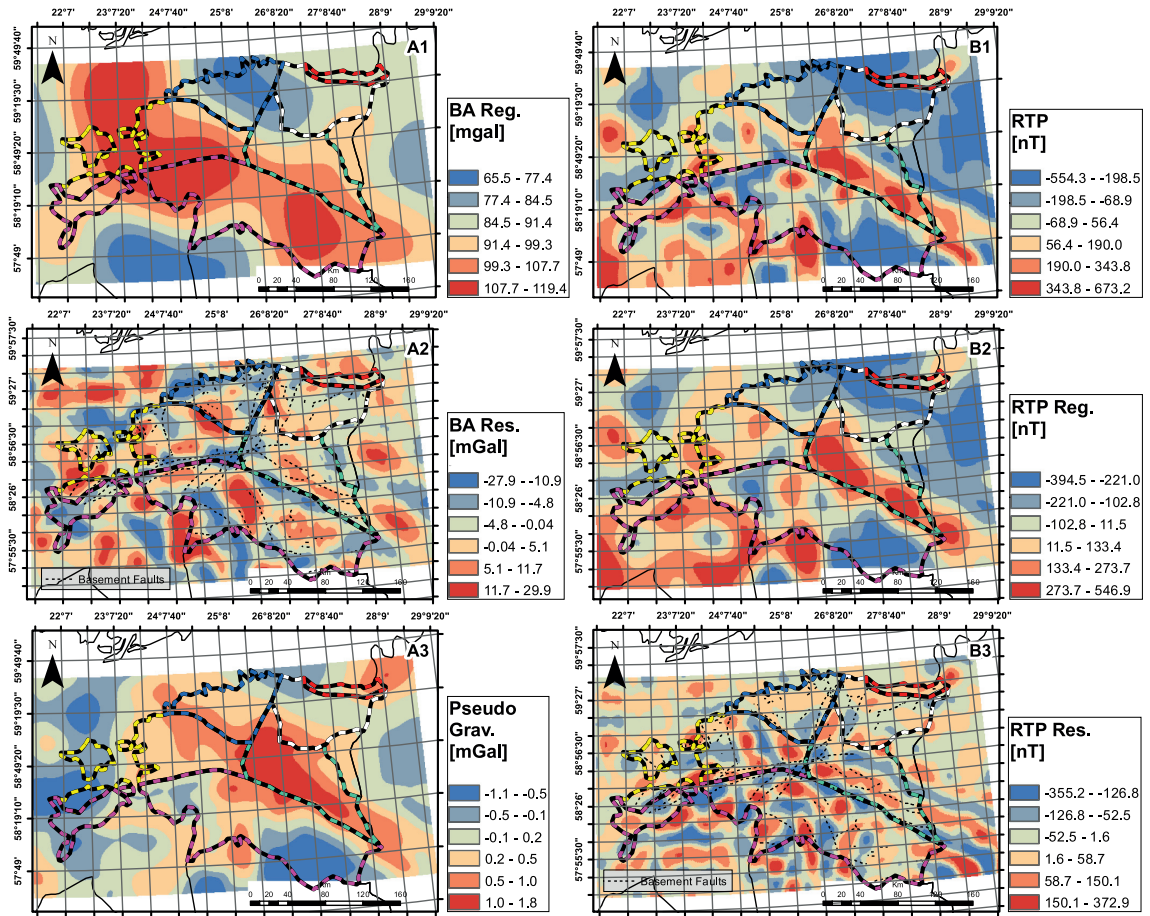


Fig. 6. Transform map results of the potential data of Estonia. A1) Low-pass filtered gravity (regional) anomaly map.; A2) High-pass gravity (residual) anomaly map.; A3) Pseudo-gravity map obtained from the RTP results.; B1) Reduction-to-Pole (RTP) magnetic anomaly map.; B2) Low-pass filtered RTP magnetic (regional) anomaly map.; B3) High-pass filtered RTP magnetic (residual) anomaly map. .

represented as a two-dimensional Fourier series with distinct frequencies that characterise the anomalies using Fourier processing. We use [Spector and Grant \(1970\)](#) technique to calculate the depth of causal bodies using a piece-wise linear relationship between the logarithm of the squared spectral amplitude and the wavenumber ([Shin et al. \(2009\)](#)), in where its assumed that the earth is represented as a collection of rectangular, vertically sided parallelepipeds with varying depth, breadth, and thickness ([Bhattacharyya \(1966\)](#); [Spector and Grant \(1970\)](#)), as a result, linear approaches, such as two-dimensional filtering, should be used to separate the individual components based on their wavenumber spectrum ([Tselentis et al. \(1988\)](#); [Kumar et al., 2016](#)).

The power spectra technique predicted that, for large samples, the logarithm of the gravitational field power spectrum of a mono-pole source versus wavenumber (cycles/km) can show a linear connection. The slope of the straight line is proportional to the depth to the top of the relevant body creating the gravitational anomaly ([Spector and Grant \(1970\)](#); [Hahn et al. \(1976\)](#); [Sanchez-Rojas \(2012\)](#)). The average depth of the set h is half the gradient in the logarithmic radial spectra for this type of model. The depth factor for this set of anomalies is the exponent if h is the mean depth of the layer ($2hk$) ([Peredo et al. \(2021\)](#); [Ekwok et al.](#)

(2021)). Consequently, the logarithmic plot of the radial spectrum shows a straight line with a slope of $-2h$. As a result, the ensemble mean depth is given as:

$$h = -\frac{m}{2} \quad (1)$$

where m is the slope. Eq.1 can be used directly to compute h if wavenumber is expressed in cycles/km ([Peredo et al. \(2021\)](#); [Ekwok et al. \(2021\)](#)), and the corresponding relation is expressed as:

$$m = 4\pi h \quad (2)$$

The power spectrum analysis was conducted through the 2D FFT of the WGM-12 CBA gravity field in each block (Fig. 7). So, The spectral determination of layers depths was based on the following procedures; division into 40 spectral grids of $150 \times 150 \text{ km}^2$ (Fig. 8-A), generation of radial energy spectrum, plots of log energy values versus frequency and the estimation of the depth to sources and interpolation surface plot. $150 \times 150 \text{ km}^2$ block size was chosen because it exposes the steepest, shallowest and intermediate spectral wavenumber components of the spectral curve without dramatic overlaps ([Tselentis et al. \(1988\)](#); [Emishaw et al. \(2017\)](#); [Ngalamo et al. \(2018\)](#); [Ch et al. \(2020\)](#)).

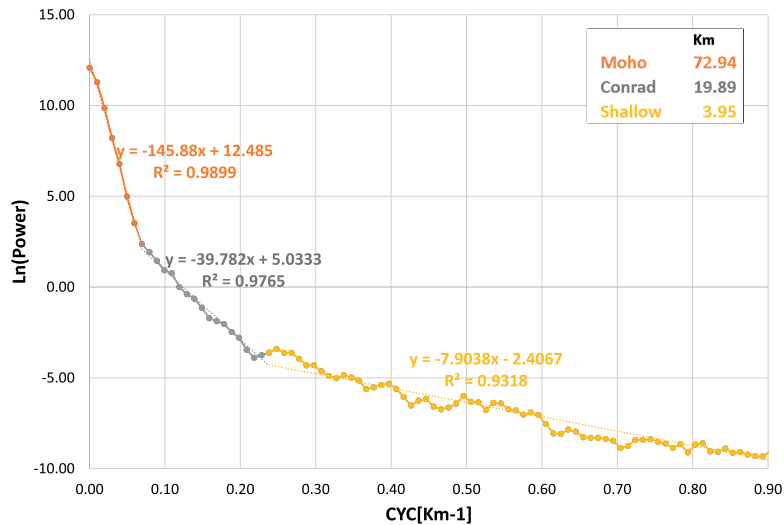


Fig. 7. Example of the radial average power spectrum of CBA gravity data for the block 1, showing three defined segments (i.e., Moho discontinuity, Conrad discontinuity and shallow signal recorded) due to the decay of power with respect to frequency. ₂

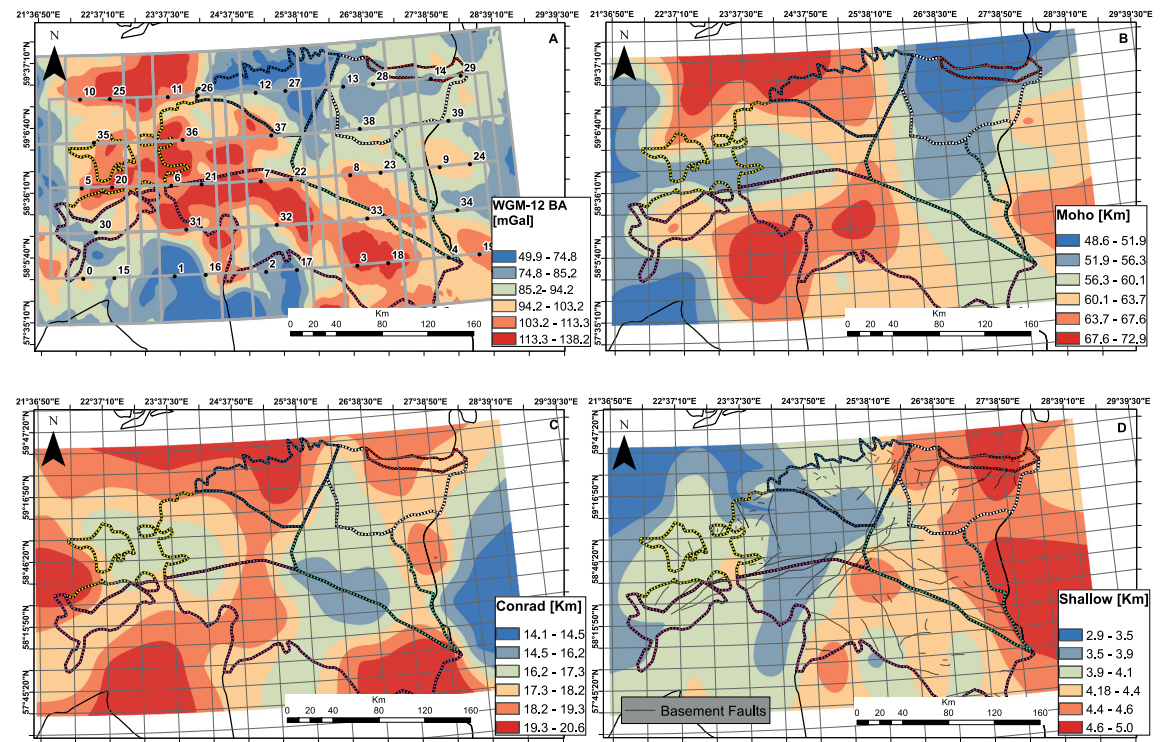


Fig. 8. Map results of the layer discontinuities obtained after the power spectra analyses of the Bouguer WGM-12 model. A) 150 × 150 km² block grids over the CBA WGM-12 which were analysed individual, to obtain the depth results of the Moho and Conrad discontinuities, and the depths of the shallowest records of the signal (Table 1); B) Moho discontinuity depth map; C) Conrad discontinuity depth map; D) shallowest depths of the recorded spectrum map. ₂

Power spectrum analysis was used in conjunction with 2D spectral analysis to determine depths for abrupt density contrasts that develop with depth, such as the Moho and Conrad discontinuities, where density values rapidly shift (Kumar et al. (2016)). The crustal discontinuities might be deciphered using this approach if the data spread is broad enough (Kumar et al. (2016)). Several studies have demonstrated the effectiveness of this strategy by giving data on the ensemble average depths of randomly dispersed anomalous causal sources occurring below the plane of observations (Curtis and Jain (1975); Bose and Sengupta (1984); Tselentis et al. (1988); Carbó et al. (2005); Bansal et al. (2006); Kumar et al. (2016); Sahoo and Pal (2021)).

The average radial power spectrum of each block was obtained by using the MAGMAP tool of Geosoft Oasis Montaj software, and the grids were extended by 10% using the greatest entropy approach to remove edge effects. The power spectra were obtained using FFT. Since the dataset was two-dimensional, radial averaging of the power spectrum was used to produce a one-dimensional representation (Tselentis et al. (1988); Kumar et al. (2016)). Carbó et al. (2005) found that analyses of gravity anomalies of longer wavelengths (λ ; ~ 20–360 km) would help in understanding deeper crustal inhomogeneities (Conrad, Moho, sub-Moho, etc.) in a study area. Results of the power spectrum analysis for each block showed three defined segments (Fig. 7), each being associated to the top of a definite density layer. The deepest segment is linked to the top of Moho discontinuity. The slope of the second segment is related to Conrad discontinuity. The third segment defines the depth to shallowest anomalies of the basement rocks recorded. From the computed values (Table 1), the depth of the Moho and Conrad discontinuities, and the shallowest density basement anomalies were plotted (Fig. 8).

From the power spectra of each grid, three segments can be depicted (Fig. 7), each being associated to the top of a definite density layer: (i) 48.63–68.94 km, (ii) 14.12–20.68 km and (iii) 2.95–5.05 km. Of these the deeper layer correlates with the Moho discontinuity (Fig. 8-B). Intermediate depth is interpreted to be the depth of the top of the Conrad discontinuity (Fig. 8-C). The shallowest depths correspond to the shallowest signals that were registered from the WGM-12 CBA model (Fig. 8-D).

4.2. Calculation of the Curie point depth, geothermal gradient and heat flow

The CPD is defined as the depth at which the rocks of the Earth's crust reach their Curie temperature, which corresponds to the temperature at which the geological material composing the lithosphere undergoes a transition from the ferromagnetic to the paramagnetic phase (Okubo et al. (1985); Tanaka et al., 1999). Rock magnetisation depends on composition (amount of magnetic minerals) and temperature. Recent calculations of the thickness of the magnetised crust show that there may be two constraints corresponding to changes in vertical composition and/or temperatures at which rocks lose their ferromagnetic capabilities (Núñez Demarco et al. (2021)). The magnetic characteristics of rocks are mostly governed by ferromagnetic minerals, as most minerals are paramagnetic or diamagnetic with extremely low magnetic susceptibilities (Núñez Demarco et al. (2021)). The titanomagnetite solid solution series is the primary carrier of magnetism. Other minerals, such as hematite, pyrrhotite, and iron and nickel alloys, are essential in geological circumstances, only when space is limited; yet, it is still a point of debate (Kletetschka et al. (2002); McEnroe et al. (2004)). It is considered that the main magnetic minerals in the lower crust are magnetite or titanomagnetites (Frost and Shive (1986); Hunt et al. (1995)).

The magnetic crust is that part of the crust whose magnetisation limit is the Curie point isotherm (i.e. CPD) (Aboud et al. (2014)). Under certain conditions, the Curie depth can be used as a temperature data point to constrain deep thermal gradients (Beardmore et al. (2001)). So, the CPD can be predicted when the basal depth of the magnetic body has been determined, being a significant parameter for understanding the lithosphere's thermal structure (e.g., heat flow and geothermal gradient), rheological behaviour, and geological and tectonic development (Bhattacharyya and Morley (1965); Gaudreau et al. (2019); Saada et al., 2021; Herrera et al. (2021)).

CPD can be calculated by two methods: one using geothermal data (the geothermal method) and one using magnetic data (the magnetic approach) (Bhattacharyya and Morley (1965); Spector and Grant (1970); Maus et al. (1997); Salem et al. (2014); Gaudreau et al. (2019); Saada et al. (2021); Herrera et al. (2021)). Geothermal methods are based on utilisation of geothermal data such as a heat flow (HF) and/or vertical geothermal gradient (GG), heat production, and heat

Table 1
Depth estimates of the layers for the 40 spectral blocks from the FFT analysis of the WGM-12 BA.

Quadrant	Moho (Km)	Conrad (Km)	Shallow (Km)	Long (UTM34N)	Lat (UTM34N)	Quadrant	Conrad (Km)	Shallow (Km)	Long (UTM34N)	Lat (UTM34N)	Moho (Km)
0	52.21	17.60	3.94	22.24	57.93	20	17.46	4.00	22.70	58.64	60.73
1	72.94	19.89	3.95	23.59	57.93	21	16.43	3.74	24.04	58.64	57.01
2	60.15	17.66	4.65	24.94	57.93	22	16.21	4.68	25.39	58.64	66.33
3	60.95	19.08	4.39	26.28	57.93	23	16.47	4.53	26.74	58.64	55.29
4	58.34	20.26	4.67	27.63	57.93	24	15.04	4.84	28.09	58.64	59.18
5	56.77	20.69	4.09	22.24	58.64	25	18.16	4.11	22.70	59.33	69.15
6	54.89	17.54	4.17	23.59	58.64	26	18.57	4.18	24.04	59.33	68.74
7	60.34	18.74	3.99	24.94	58.64	27	20.26	3.92	25.39	59.33	66.69
8	54.43	16.44	4.27	26.28	58.64	28	18.94	4.31	26.74	59.33	48.64
9	59.70	19.30	4.81	27.63	58.64	29	17.72	4.38	28.09	59.33	55.21
10	51.00	18.52	2.96	22.24	59.33	30	17.35	3.90	22.44	58.29	63.85
11	64.23	18.87	3.68	23.59	59.33	31	18.55	3.96	23.79	58.29	70.54
12	66.20	19.15	4.05	24.94	59.33	32	19.13	4.24	25.13	58.29	69.81
13	51.47	16.42	4.69	26.28	59.33	33	15.04	4.24	26.48	58.29	58.23
14	50.90	18.02	5.05	27.63	59.33	34	14.15	5.01	27.83	58.29	67.18
15	50.04	19.86	4.14	22.70	57.93	35	16.85	4.08	22.44	58.99	64.25
16	69.08	18.26	3.83	24.04	57.93	36	17.04	4.10	23.79	58.99	57.03
17	63.07	16.71	3.85	25.39	57.93	37	18.64	3.56	25.13	58.99	62.84
18	59.80	19.26	4.01	26.74	57.93	38	16.46	4.40	26.48	58.99	54.22
19	63.21	14.13	4.77	28.09	57.93	39	17.26	4.55	27.83	58.99	63.79

conduction (Eppelbaum and Pilchin (2006)). The challenges of using geothermal methods are always related to the inability to take direct measurements of geothermal parameters at great depths (Tanaka et al. (1999); Aboud et al. (2014); Gao et al. (2015)). Therefore, magnetic techniques that consider CPD as the depth to the bottom of a thick magnetic layer (Z_b) are widely used (Quintero et al. (2019)). Calculating the CPD from magnetic anomaly maps typically entails first calculating the power spectrum of magnetic data in a defined area, then calculating the radial average of the power spectrum, and finally fitting the experimental curve with a theoretical curve that is directly proportional to the depth to the magnetic source (Núñez Demarco et al. (2021)). This approach may be used to produce a single regional average depth from the complete data set, or the examined area can be separated into smaller zones (blocks) (Shuey et al. (1977); Blakely (1988)). The primary drawback of spectrum approaches is that tiny scale CPD changes are difficult to detect, and determining the depth to the bottom of a magnetic source is a function of the region investigated (Hussein et al. (2013)). In order to be certain of the determined depth, the breadth of the region investigated must be at least three to four times the actual depth to the bottom of a magnetic source (Saada et al. (2021)), being the optimal window size to estimate Z_b should be carefully selected. The maximum CPD is limited by $2\pi/L$, L being the window size length (Quintero et al. (2019); Campos-Enríquez et al. (2019)). Windows of $100 \times 100 \text{ km}^2$ and $150 \times 150 \text{ km}^2$ were used due to they have been providing good Z_b resolution (Núñez Demarco et al. (2021); Saada et al. (2021)) (Fig. 10-A). Both window sizes were used to avoid possible depth uncertainties.

Numerical tests done by Okubo et al., 1985 and Zhou and Thybo (1998) showed that CPD results present minimal differences between results obtained with and without applying reduction to the pole. In this research, regional trends were removed from the RTP magnetic anomaly map to calculate the CPD, as it was suggested by Aigbogun and Olorunsola (2018), Kasidi (2019), Quintero et al. (2019), Ngho et al. (2020),

among others, due to the bottom of the magnetised crust determined from the spectral analysis of residual magnetic anomalies is generally interpreted as the level of the Curie point isotherm (Tanaka et al. (1999)). To carry out the spectral analysis (Fig. 9), RTP residual data was divided into overlapping blocks of $100 \times 100 \text{ km}^2$ and $150 \times 150 \text{ km}^2$ (Fig. 10-A and Tables 2 and 3), which were used to obtain the depths to top boundary (Z_t) and Centroid (Z_o). MAGMAP tool of Geosoft Oasis Montaj was used to obtain the average radial power spectrum of each window. Each block grids were expanded 10% by the maximum entropy method to eliminate edge effects. FFT was used to obtain the power spectra. The method of estimating Z_o of a parallelepiped constant magnetisation was originally proposed by Bhattacharyya and Leu, 1975 and further developed by Okubo et al. (1985) and Tanaka et al. (1999), assuming random magnetisation and employing the logarithm of the Fourier power spectrum. Tanaka et al. (1999) constructed the radial average power-density spectra of the whole field magnetic anomaly using Blakely (1996) simplified calculus as is shown in eqn.3.

$$P(\kappa) = \Phi_{RT}(\kappa) = A_1 e^{-2|\kappa|Z_t} (1 - e^{-|\kappa|(Z_b - Z_t)})^2 \quad (3)$$

In the eqn.3, A is a constant and $P(\kappa)$ is the radial average power spectrum, and Z_b and Z_t are respectively the depths to the bottom and to the top of the magnetic body ensemble. Many methods have been proposed to calculate Z_b , including the spectral peak method (Spector and Grant (1970)), forward modelling of the spectral peak method (Idárraga-García and Vargas (2018)), and the centroid method (Bhattacharyya and Leu (1975); Okubo et al. (1985); Tanaka et al. (1999)), which is used in this study. In the centroid method, after determining Z_t and the Z_o of the magnetic source, Z_b is computed. By simplifying eqn.3, the centroid depth of the magnetic source can be calculated from the large wavelengths spectrum zone (Bhattacharyya and Leu (1975); Okubo et al. (1985); Tanaka et al. (1999); Quintero et al. (2019); Núñez Demarco et al. (2021)), as is shown in the eqn.4.

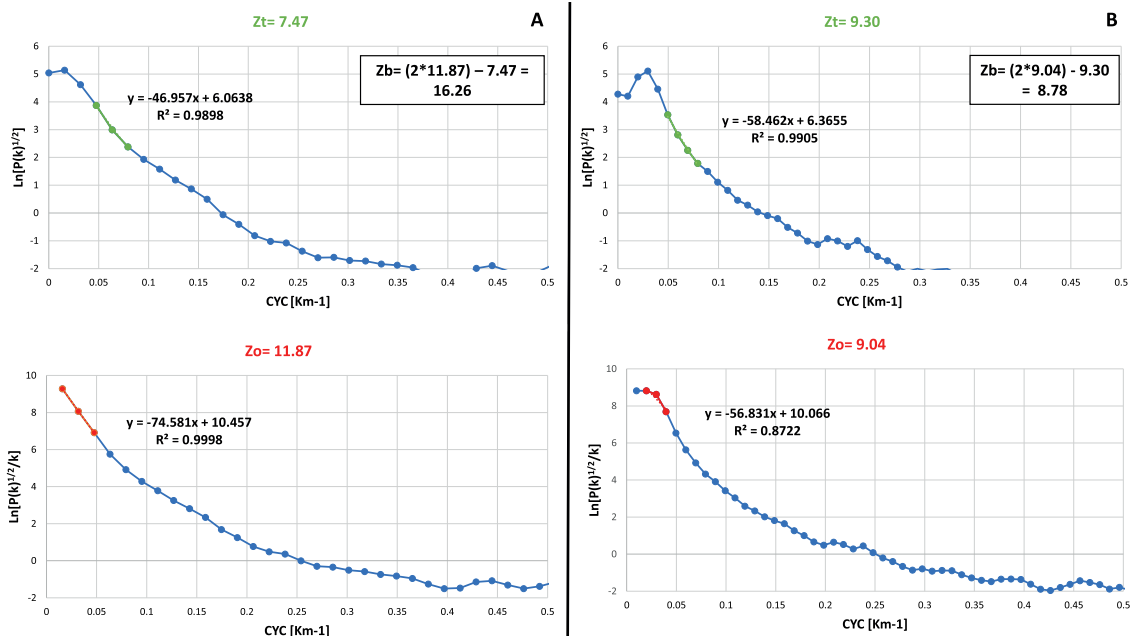


Fig. 9. Examples of the power spectrum analysis to estimate the depth to the base of magnetic sources (CPD). A) Representative $100 \times 100 \text{ km}^2$ block (block 34; Table 2); B) Representative $150 \times 150 \text{ km}^2$ block (block 14; Table 3). The depth Z_t to the top of the magnetic body, the depth Z_o to the centre of the magnetic body and the depth Z_b to the bottom of the magnetic body.

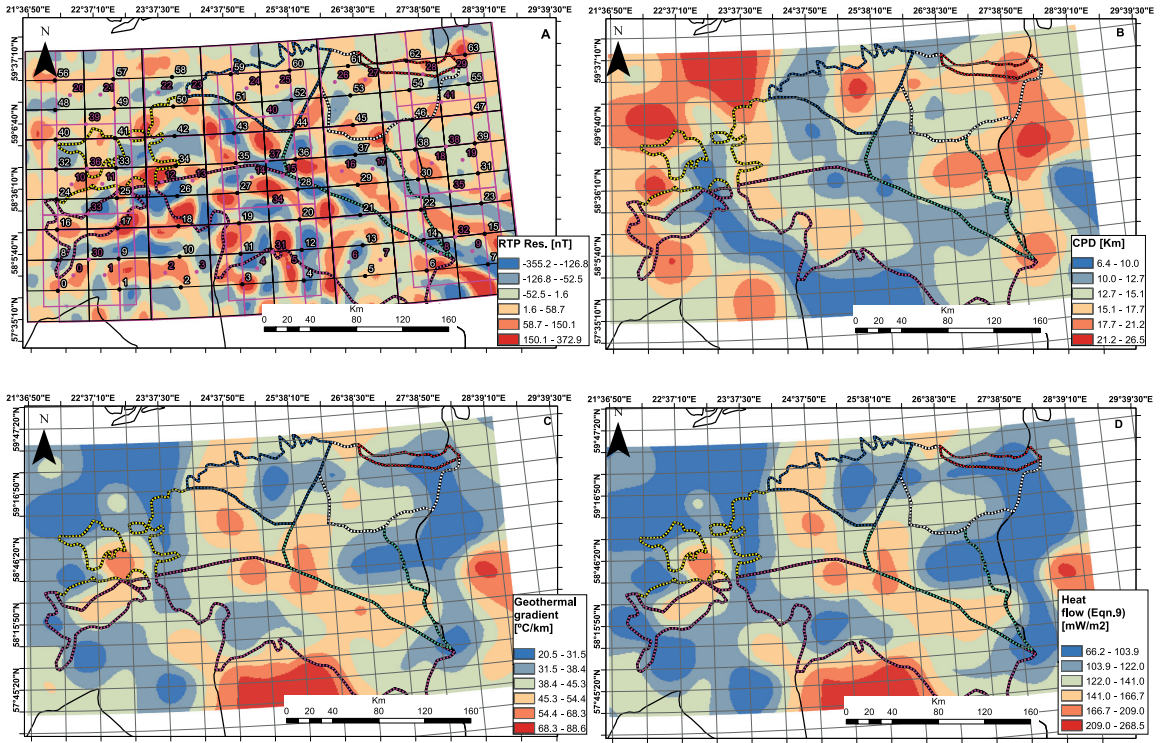


Fig. 10. Map results of the CPD, Geothermal gradient (GG) and Heat flow (HF). A) RTP residual map showing the overlapping squared blocks ($100 \times 100 \text{ km}^2$ in black grids and $150 \times 150 \text{ km}^2$ in purple grids) to estimate the CPD. Each block is label by a centred dot; B) CPD map estimated from the power spectrum analysis of the RTP residual magnetic data.; C) GG map estimated from the CPD analysis.; D) HF map estimated from the typical CPD analysis (Eq.9). . **(For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)**

$$\ln\left(\frac{P(\kappa)^{1/2}}{\kappa}\right) = A_2 - |\kappa|Z_o \quad (4)$$

where A_2 is a constant. For the high wavenumber region, eqn.3 can be reduced to (Spector and Grant (1970); Bhattacharyya and Leu (1975); Okubo et al. (1985)):

$$\ln\left(P(\kappa)^{1/2}\right) = A_3 - |\kappa|Z_t \quad (5)$$

where A_3 is a constant, and Z_t is the magnetic source top depth. In that way, once Z_o and Z_t are calculated from eqns.4 and 5, depth to the bottom is given in the eqn.6.

$$Z_b = 2Z_o - Z_t \quad (6)$$

It's important to mention that if the wavenumber is expressed in cycles/km, or if frequencies are expressed in $1/\text{km}$, eqns.4 and 5 present a $Z = \text{slope}/2\pi$ (Okubo et al. (1985); Núñez Demarco et al. (2021)). To select the optimal wavenumber ranges to recover the centroid (Z_o) and top (Z_t) depths, the researches of Núñez Demarco et al. (2021) and Xu et al. (2021) were used, as they present an appropriate compilation of different wavenumbers used in previous works. In this research the wavenumber ranges between the upper and lower limit of wavenumber selected for straight-line fitting to calculate Z_t and Z_o are specified as 0.04–0.08 and 0.01–0.04 cycles/km, respectively. Fig. 10-B present the Z_b results (CPD map). CPD values vary between 6.4 and 26.5 Km with an average of 15.5 km.

The GG is based on the approximate temperature at which the

magnetite becomes paramagnetic, being 580°C . Then, the GG for each block was determined by dividing the Curie isothermal point (i.e. 580°C) by the CPD. GG map can be observed in Fig. 10-C.

HF is the transfer of energy (heat) from the inner Earth to its surface. The HF may be computed over a large region as a result of evaluating the CPD. The one-dimensional conductive HF equation with a constant temperature gradient was used to estimate the HF of the crust for the Curie depths. The HF value is expressed by Fourier's law with the following formula:

$$q = kc \frac{dT}{dZ} \quad (7)$$

where q is the HF and kc is the coefficient of thermal conductivity. In eqn.7, it is assumed that the direction of the temperature variation is vertical and the temperature gradient dT/dZ is constant. Tanaka et al. (1999) calculated the Curie temperature (θ) from the CPD (Z_b) and the GG (dT/dZ) as is indicated in the eqn.8.

$$\theta = \left[\frac{dT}{dZ}\right] Z_b \quad (8)$$

In Eqns.7 and 8, a relationship can be determined between the Z_b and the HF (q) as (eq.9):

$$q = kc \left[\frac{\theta}{Z_b}\right] \quad (9)$$

The CPD is inversely proportional to the HF in this equation (Tanaka et al. (1999); Quintero et al. (2019); Saada et al. (2021)). GG of the rocks

Table 2
Results of estimating CPD, geothermal gradient (GG) and heat flow (HF) for each block of 100 x 100 km². Each point represents a range of Te values according to Pérez-Gussinyé et al. (2004)).

Block100	Zt [km]	Zo [km]	CPD [km]	Block100										Zt [km]	Zo [km]	CPD [km]	Block100										Zt [km]	Zo [km]	CPD [km]
				GG [580 °C/ km]	HF Eq. 9 [mW/m ²]	HF Eq. 10 [mW/m ²]	HF Eq. 11 [mW/m ²]	HF Eq. 12 [mW/m ²]	Te [km]	Long (UTM34N)	Lat (UTM34N)	GG [580 °C/ km]	HF Eq. 9 [mW/m ²]				HF Eq. 10 [mW/m ²]	HF Eq. 11 [mW/m ²]	HF Eq. 12 [mW/m ²]	Te [km]	Long (UTM34N)	Lat (UTM34N)							
0	8.57	10.79	13.01	44.56	133.70	84.96	122.42	124.92	92–105	22.01	57.81	32	6.47	11.26	16.05	36.13	108.39	81.33	112.14	115.16	105–115	22.01	58.76						
1	8.63	11.22	13.80	42.03	126.10	84.02	129.26	131.59	92–105	22.91	57.81	33	8.98	8.93	8.89	65.25	195.76	89.88	196.76	197.82	105–115	22.91	58.76						
2	7.36	13.74	20.13	28.81	86.44	76.46	91.66	95.99	92–105	23.81	57.81	34	7.47	11.87	16.27	35.66	106.97	81.07	110.84	113.91	92–105	23.81	58.76						
3	9.41	8.87	8.33	69.62	208.85	90.55	208.10	209.06	115–130	24.71	57.81	35	8.11	10.76	13.42	43.23	129.69	84.48	133.02	135.24	92–105	24.71	58.76						
4	9.16	7.82	6.48	89.51	268.54	92.76	261.49	262.04	115–130	25.60	57.81	36	8.68	9.51	10.34	56.09	168.28	88.15	170.05	171.46	92–105	25.60	58.76						
5	7.58	10.86	14.14	41.03	123.08	83.62	126.40	128.83	92–105	26.50	57.81	37	7.47	11.95	16.42	35.32	105.96	80.89	109.72	112.86	105–115	26.50	58.76						
6	7.06	10.87	14.68	39.50	118.50	82.96	121.89	124.48	92–105	27.40	57.81	38	6.66	12.32	17.97	32.27	96.81	79.03	100.84	104.57	105–115	27.40	58.76						
7	9.18	10.69	12.19	47.57	142.70	85.94	145.23	147.11	92–105	28.30	57.81	39	6.89	13.50	20.12	28.83	86.49	76.47	92.44	96.73	105–115	28.30	58.76						
8	8.59	11.38	14.16	40.96	122.89	83.59	126.15	128.59	105–115	22.01	58.05	40	7.10	13.62	20.15	28.79	86.37	76.44	91.20	95.56	105–115	22.01	58.99						
9	8.65	12.39	16.13	35.96	107.89	81.24	111.74	114.77	92–105	22.91	58.05	41	7.28	12.13	16.98	34.16	102.48	80.22	107.21	110.50	92–105	22.91	58.99						
10	8.12	9.54	10.96	52.92	158.77	87.41	160.70	162.27	92–105	23.81	58.05	42	7.76	12.38	17.00	34.13	102.38	80.20	106.44	109.75	92–105	23.81	58.99						
11	7.75	11.96	16.18	35.85	107.54	81.18	112.01	115.03	115–130	24.71	58.05	43	9.90	9.78	9.67	59.99	179.97	88.95	181.50	182.75	82–92	24.71	58.99						
12	8.69	11.61	14.54	39.89	119.67	83.13	123.85	126.38	115–130	25.60	58.05	44	8.01	10.28	12.54	46.24	138.73	85.52	141.51	143.48	92–105	25.60	58.99						
13	6.74	9.82	12.90	44.95	134.86	85.09	137.75	139.82	92–105	26.50	58.05	45	9.20	11.22	13.25	43.77	131.31	84.67	134.31	136.47	105–115	26.50	58.99						
14	7.85	11.55	15.24	38.05	114.14	82.29	117.64	120.41	92–105	27.40	58.05	46	7.08	10.97	14.86	39.02	117.06	82.75	120.50	123.15	105–115	27.40	58.99						
15	7.04	10.87	14.70	39.46	118.39	82.95	121.90	124.48	92–105	28.30	58.05	47	6.27	12.28	18.29	31.72	95.16	78.66	99.48	103.22	105–115	28.30	58.99						
16	6.35	12.96	19.56	29.65	88.94	77.13	93.57	97.74	105–115	22.01	58.29	48	6.14	14.46	22.78	25.46	76.39	73.29	81.69	87.00	105–115	22.01	59.22						
17	8.54	11.79	15.05	38.54	115.61	82.53	119.16	121.85	92–105	22.91	58.29	49	8.23	11.01	13.80	42.02	126.07	84.02	129.21	131.54	92–105	22.91	59.22						
18	9.39	11.88	14.38	40.33	121.00	83.33	124.49	126.97	92–105	23.81	58.29	50	6.63	12.17	17.72	32.74	98.22	79.34	102.63	106.16	92–105	23.81	59.22						
19	7.33	12.37	17.42	33.30	99.89	79.70	104.04	107.48	115–130	24.71	58.29	51	9.05	10.44	11.83	49.01	147.02	86.37	149.60	151.38	82–92	24.71	59.22						
20	7.43	10.38	13.32	43.54	130.62	84.59	133.46	135.65	92–105	25.60	58.29	52	9.04	12.16	15.29	37.92	113.77	82.23	116.86	119.71	92–105	25.60	59.22						
21	7.98	9.25	10.52	55.15	165.46	87.94	167.43	168.88	105–115	26.50	58.29	53	9.75	11.15	12.55	46.23	138.68	85.52	141.10	143.11	105–115	26.50	59.22						
22	8.23	9.83	11.42	50.79	152.36	86.86	154.71	156.38	105–115	27.40	58.29	54	7.22	11.37	15.53	37.35	112.04	81.95	115.53	118.37	105–115	27.40	59.22						
23	7.29	9.25	11.22	51.70	155.09	87.10	157.10	158.74	105–115	28.30	58.29	55	7.57	11.94	16.31	35.57	106.71	81.03	110.61	113.70	105–115	28.30	59.22						
24	7.73	11.68	15.63	37.10	111.30	81.83	114.99	117.87	105–115	22.01	58.52	56	9.92	12.15	14.38	40.34	121.01	83.33	124.36	126.86	92–105	22.01	59.45						
25	9.71	10.37	11.04	52.54	157.60	87.31	159.77	161.34	105–115	22.91	58.52	57	5.30	13.85	22.41	25.88	77.65	73.74	82.79	87.98	92–105	22.91	59.45						
26	8.48	12.46	16.43	35.30	105.89	80.88	109.84	112.96	92–105	23.81	58.52	58	6.40	13.91	21.42	27.07	81.22	74.91	86.55	91.35	82–92	23.81	59.45						
27	7.89	10.25	12.61	45.98	137.94	85.43	140.71	142.72	92–105	24.71	58.52	59	8.00	9.54	11.08	52.33	157.00	87.26	159.22	160.81	74–82	24.71	59.45						
28	8.91	9.22	9.54	60.80	182.41	89.11	183.60	184.82	92–105	25.60	58.52	60	7.99	11.60	15.22	38.11	114.34	82.33	117.82	120.60	82–92	25.60	59.45						
29	7.42	12.48	17.54	33.07	99.21	79.55	103.84	107.30	105–115	26.50	58.52	61	7.99	10.12	12.26	47.32	141.95	85.86	144.43	146.34	105–115	26.50	59.45						
30	7.52	12.26	17.00	34.12	102.36	80.20	106.54	109.84	105–115	27.40	58.52	62	7.66	11.82	15.97	36.32	108.95	81.43	112.71	115.69	105–115	27.40	59.45						
31	7.60	7.67	7.74	74.90	224.71	91.25	223.17	224.05	105–115	28.30	58.52	63	9.92	12.15	14.38	40.34	121.01	83.33	124.37	126.86	105–115	28.30	59.45						
Mean				14.58	42.38				127.13			83.09				130.03				132.66									
Min				6.48	25.46				76.39			73.29				81.69				87.00									
Max				22.78	89.51				268.54			92.76				261.49				262.04									

Table 3
Results of estimating CPD, geothermal gradient (GG) and heat flow (HF) for each block of 150 x 150 km². Each point represents a range of Te values according to Pérez-Gussinyé et al. (2004).

Block 150	Zt [Km]	Zo [km]	CPD [Km]	GG [580 °C/ Km]	HF.Eq. 9 [mW/ m2]	HF. Eq. 10 [mW/ m2]	HF. Eq. 11 [mW/ m2]	HF. Eq. 12 [mW/ m2]	Te [Km]	Long	Lat	Block 150	Zt [Km]	Zo [km]	CPD [Km]	GG [580 °C/ Km]	HF.Eq. 9 [mW/ m2]	HF. Eq. 10 [mW/ m2]	HF. Eq. 11 [mW/ m2]	HF. Eq. 12 [mW/ m2]	Te [Km]	Long	Lat
0	8.81	13.38	17.96	32.30	96.89	79.05	101.33	104.94	92–105	22.24	57.93	21	8.36	13.70	19.05	30.45	91.36	77.75	95.93	99.91	92–105	22.70	59.33
1	9.38	13.64	17.90	32.40	97.20	79.12	101.48	105.08	92–105	22.70	57.92	22	7.35	16.82	26.28	22.07	66.20	69.11	71.96	78.63	92–105	23.59	59.33
2	9.02	12.94	16.87	34.39	103.17	80.36	107.29	110.55	92–105	23.59	57.93	23	8.94	10.96	12.98	44.68	134.03	85.00	136.71	138.83	92–105	24.05	59.33
3	10.42	11.44	12.45	46.57	139.71	85.63	142.51	144.46	92–105	24.05	57.92	24	8.94	10.82	12.70	45.67	137.00	85.33	139.23	141.32	74–82	24.94	59.33
4	9.40	10.65	11.91	48.71	146.13	86.28	150.89	152.67	115–130	24.94	57.93	25	6.94	15.32	23.69	24.48	73.44	72.20	78.81	84.48	74–82	25.40	59.33
5	10.13	11.00	11.86	48.91	146.74	86.34	151.69	153.50	115–130	25.40	57.92	26	9.61	14.07	18.54	31.29	93.88	78.36	98.47	102.27	82–92	26.28	59.33
6	9.82	11.17	12.51	46.35	139.05	85.56	141.84	143.80	92–105	26.28	57.93	27	8.76	12.83	16.90	34.32	102.97	80.32	107.13	110.40	105–115	26.75	59.33
7	11.14	12.16	13.18	44.10	132.32	84.76	135.01	137.16	92–105	26.74	57.92	28	7.80	14.87	21.94	26.43	79.30	74.29	84.42	89.43	105–115	27.63	59.33
8	9.07	14.25	19.44	29.84	89.53	77.29	94.17	98.29	92–105	27.63	57.93	29	10.76	14.30	17.85	32.49	97.46	79.18	101.88	105.45	105–115	28.09	59.33
9	8.57	13.34	18.11	32.03	96.09	78.87	101.03	104.65	92–105	28.09	57.92	30	8.68	15.08	21.48	27.00	81.01	74.85	86.01	90.86	92–105	22.48	58.05
10	9.01	17.29	25.57	22.68	68.05	69.96	73.73	80.11	105–115	22.24	58.64	31	9.79	11.00	12.21	47.50	142.51	85.92	144.98	146.87	115–130	25.17	58.05
11	10.84	13.32	15.80	36.71	110.12	81.63	116.80	119.57	105–115	22.70	58.63	32	7.84	13.26	18.68	31.04	93.13	78.19	97.52	101.39	92–105	27.87	58.05
12	10.06	13.84	17.63	32.90	98.71	79.45	102.93	106.44	92–105	23.59	58.64	33	9.65	11.83	14.01	41.41	124.24	83.77	162.10	163.65	105–115	22.48	58.75
13	10.01	11.87	13.73	42.26	126.77	84.11	129.91	132.22	92–105	24.05	58.63	34	9.63	11.25	12.88	45.04	135.12	85.12	119.59	122.27	92–105	25.17	58.75
14	9.30	9.04	8.79	66.02	198.06	90.01	198.94	199.99	92–105	24.94	58.64	35	8.00	15.60	23.21	24.99	74.98	72.79	103.47	107.06	105–115	27.87	58.75
15	9.03	10.66	12.29	47.18	141.54	85.82	144.27	146.17	92–105	25.40	58.63	36	4.86	14.33	23.81	24.36	73.07	72.06	126.58	129.09	105–115	22.48	59.10
16	9.49	12.99	16.48	35.18	105.55	80.81	109.70	112.83	92–105	26.28	58.64	37	9.55	12.28	15.01	38.65	115.94	82.58	138.66	140.72	92–105	25.17	58.40
17	8.24	14.27	20.30	28.57	85.70	76.25	90.48	94.91	105–115	26.74	58.63	38	9.68	14.96	20.24	28.65	85.96	76.33	80.27	85.76	105–115	27.87	59.10
18	7.50	13.32	19.14	30.31	90.93	77.65	95.33	99.40	105–115	27.63	58.64	39	8.04	12.62	17.19	33.73	101.20	79.97	78.49	84.20	92–105	27.87	58.40
19	7.64	12.05	16.46	35.24	105.73	80.85	112.45	115.58	105–115	28.09	58.64	40	10.20	10.53	10.87	53.37	160.12	87.52	109.61	112.75	82–92	22.48	58.40
20	6.53	13.04	19.55	29.66	88.98	77.15	93.79	97.94	105–115	22.24	59.33	41	8.67	12.58	16.50	35.15	105.45	80.79	90.72	95.13	105–115	25.17	59.10
					CPD	GG			HF. Eq. 9 [mW/m2]			HF. Eq. 10 [mW/m2]			HF. Eq. 11 [mW/m2]			HF. Eq. 12 [mW/m2]					
Mean					17.00	36.31			108.94			80.20			113.05			116.45					
Min					8.79	22.07			66.20			69.11			71.96			78.63					
Max					26.28	66.02			198.06			90.01			198.94			199.99					

in the location is another component that influences the predicted HF. Areas with deeper Curie isotherms are correlated with areas of intermediate-to-low HF values, and areas exhibiting shallower magnetic bottom depths corresponds to areas of higher HF values (Idárraga-García and Vargas (2018)). According to Li et al. (2017), the thermal conductivity (kc) present average values of $2.0\text{W}/(\text{m}^\circ\text{C})$ for oceanic crust and $2.5\text{W}/(\text{m}^\circ\text{C})$ for continental crust. Nevertheless, Lillie (1999) suggest that values normally range between 1.3 and $3.3\text{W}/(\text{m}^\circ\text{C})$ for granites to $2.5 - 5.0\text{W}/(\text{m}^\circ\text{C})$ for metamorphic rocks. In this paper we use $kc = 3.0\text{W}/(\text{m}^\circ\text{C})$ as it was suggested by Joeleht and Kukkonen (1998) to be the value of granulite facies rocks in the Precambrian basement of Finland and Estonia. HF map can be observed in Fig. 10-D.

In this research, different methods were implemented to obtain different HF results. Vargas et al. (2015) produced an empirical equation that links HF and CPD, which was also employed by Idárraga-García and Vargas (2018). This relation is defined as follows (eqn.10):

$$q = (-1.1943)Z_b + 100.5 \quad (10)$$

where q is in mW/m^2 and Z_b in km. Vargas et al. (2015) combined CPD data and heat flow rates from 12 different continents, including their own, to develop a consistent equation. However, these authors cautioned that this linear expression should be used with caution, since the CPD can change laterally owing to crustal composition and the huge data ranges obtained from the literature might contain considerable errors (Fig. 11-A).

Other authors suggest other methods to find HF values of a region by using the CPD results (Wen et al. (2019); Speranza et al. (2016); Wang and Li, 2015), considering a 1D heat conductive transport model, CPDs and that the HF values should be inversely correlated, as follows (eqn.11):

$$q = kc \frac{T_c - T_o}{Z_b - Z_o} + h_r^2 H_o \frac{e^{-Z_b/h_r} - e^{-Z_o/h_r}}{Z_b - Z_o} + h_r H_o e^{-Z_o/h_r} \quad (11)$$

where q is HF at the surface, kc is the coefficient of thermal conductivity, H_o is the radiogenic heat production rate at the surface, h_r is the characteristic drop-off of heat production, T_o is the temperature at the surface elevation Z_o , and T_c is the Curie temperature. We assumed for our zone $T_o = 6.4^\circ\text{C}$, $T_c = 580^\circ\text{C}$, $Z_o = 0$, $h_r = 10\text{km}$, $kc = 3\text{W}/(\text{m}^\circ\text{C})$, and $H_o = 1\mu\text{W}/\text{m}^3$, that are standard values routinely considered in the continental crust (Wang and Li (2015); Andrés et al. (2018)) (Fig. 11-B). Andrés et al. (2018) calculated a HF for from the CPD map assuming the steady state condition. The surface HF of a layer is composed of two components: background heat and heat generation inside the layer due to radiogenic decay. According to Fourier's law, the steady-state 1-D surface HF with constant heat production ($H_o = 1\mu\text{W}/\text{m}^3$), constant thermal conductivity ($kc = 3\text{W}/[\text{m}^\circ\text{C}]$), surface temperature ($T_o = 6.4^\circ\text{C}$), and temperature at CPD ($T_{CPD} = 580^\circ\text{C}$) is expressed in the eqn.12. The results are shown in the Fig. 11-C.

$$q = \frac{H_o(CPD)}{2} + \frac{(T_{CPD} - T_o)kc}{CPD} \quad (12)$$

4.3. Edge and lineaments detection methods of residual potential data

In this research, different derivative techniques (e.g., TDR, HDTDR, TDX and Analytic signal (AS)) have been applied on the study zone to locate the boundaries and lineaments of the potential sources. Moreover, by using advanced imaging techniques of derivatives maps it can be obtained structural trends of lineaments and fault systems across Estonia.

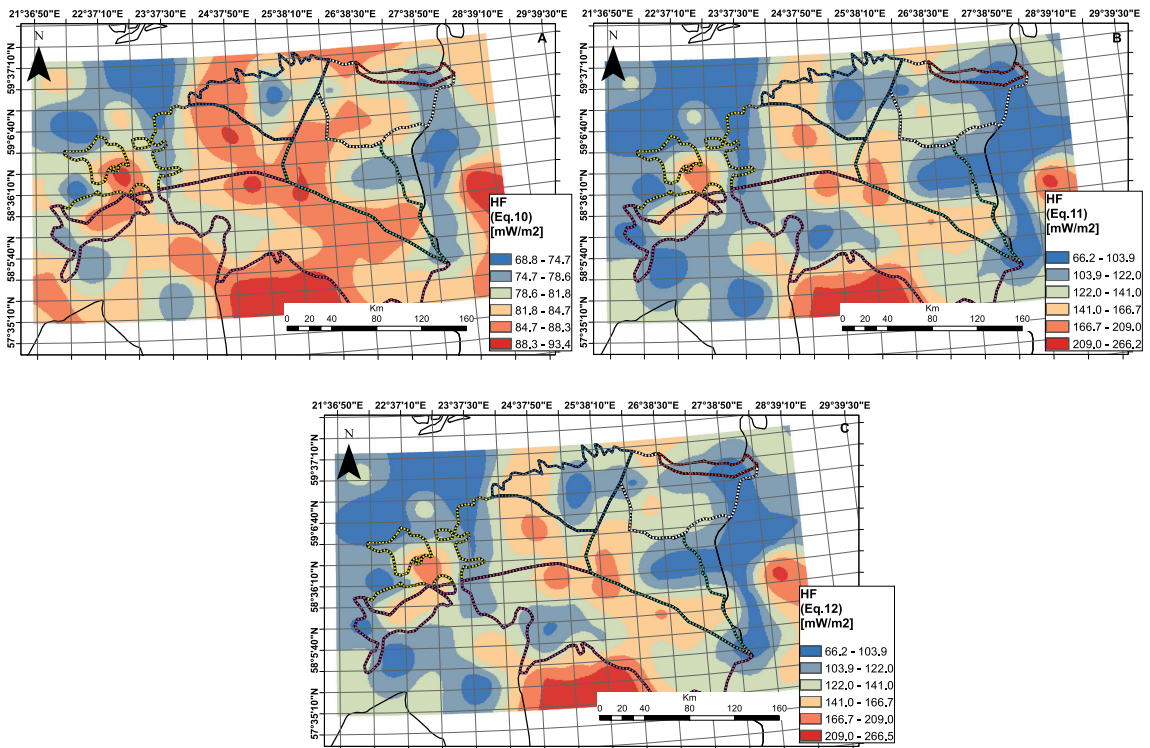


Fig. 11. Heat flow (HF) maps derived from different methods. A) HF map from eq.10.; B) HF map from eq.11.; C) HF map from eq.12. _

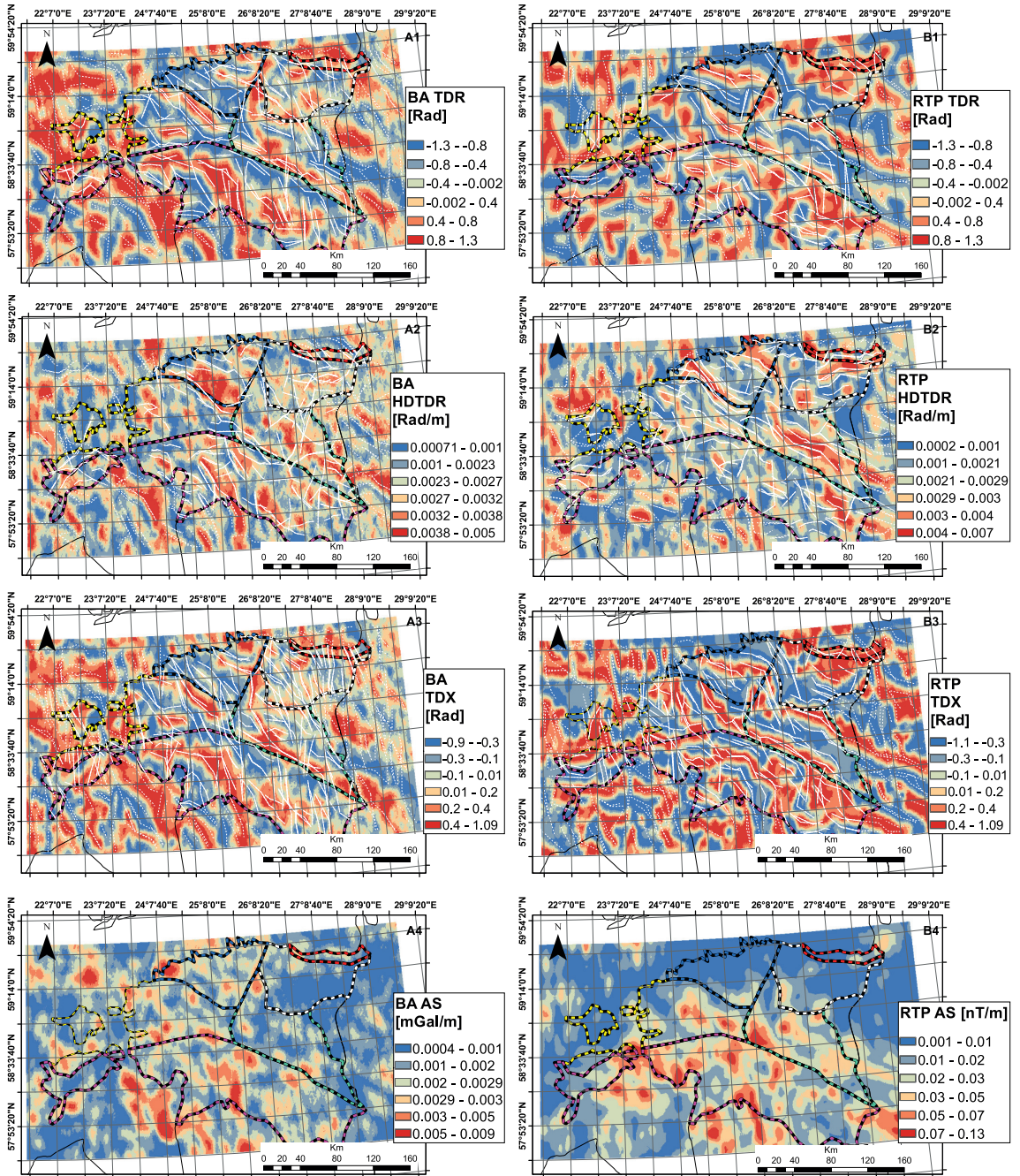


Fig. 12. Derivative technique maps and their lineaments from potential data from A) Residual Bouguer anomalies and B) Residual magnetic anomalies.; 1) TDR.; 2) HDTDR.; 3) TDX.; 4) AS. The dashed lineaments are the ones outside the Estonian Precambrian basement divisions. _

Tilt Derivative (*TDR*) and its total horizontal derivative (*HDTDR*) are imaging techniques used to study shallow basement features and mineral exploration prospects (Geosoft (2004)). *TDR* is employed in potential field pictures to enhance features and pinpoint causal body edges. The *TDR* is calculated as follows (eqn.13):

$$TDR = \tan^{-1} \left(\frac{\partial f / \partial z}{\sqrt{(\partial f / \partial x)^2 + (\partial f / \partial y)^2}} \right) \quad (13)$$

where f is the potential field, while, $\partial f / \partial x$, $\partial f / \partial y$ and $\partial f / \partial z$ are the first derivatives of the field f in the x , y and z directions, respectively. The amplitudes of *TDR* range between $-\pi/2$ and $\pi/2$ [Rad] or -90° to 90° irrespective of the amplitude of the vertical derivative or the absolute value of the overall horizontal gradient Salem et al. (2007). The *TDR* generates a zero value at or near the source edges and may thus be used to trace the contour of the edges (Saada (2016b)). Positive values are right over the sources, whereas negative values are below them (Mazur (2014)).

The square root of the sum of the square derivatives of the tilt angle in the x and y axes is *HDTDR* and is calculated as is shown in eqn.14.

$$HDTDR = \sqrt{\left(\frac{\partial TDR}{\partial x} \right)^2 + \left(\frac{\partial TDR}{\partial y} \right)^2} \quad (14)$$

HDTDR form sharper and better-defined maximum centered on the body edges. Another advantage of this independence is that it will provide useful magnetic responses for substances with either induced or remnant magnetisation, or a mixture of the two. *HDTDR* is in units of Rad/distance.

Cooper and Cowan (2006) established the Horizontal Tilt angle derivative (*TDX*) by utilising the total horizontal derivative and the absolute value of the vertical derivative, as follows (eqn.15):

$$TDX = \tan^{-1} \left(\frac{\sqrt{(\partial f / \partial x)^2 + (\partial f / \partial y)^2}}{|\partial f / \partial z|} \right) \quad (15)$$

TDX is mathematically discontinuous, approaching $\pi/2$ when tilt angle approaches 0 from positive side and $-\pi/2$ when tilt angle approaches 0 from negative side. *TDX* has a much sharper gradient over the edges compared to the original *TDR*. Both the methods *TDR* and *TDX* show a contrast variation along the sharper boundary.

The AS, also known as the total gradient, is a popular filter that creates bell-shaped anomalies across magnetic materials (Nabighian (1972)). AS is a complicated quantity, the amplitude of which is provided by the eqn.16.

$$AS = \sqrt{(\partial f / \partial x)^2 + (\partial f / \partial y)^2 + (\partial f / \partial z)^2} \quad (16)$$

The success of the AS method is since the position and depth of sources can be determined with only a few assumptions about the nature of the source body, which is often believed to be a 2D source (e.g. a step, a contact, a horizontal cylinder or a dike). The maximum values of AS are very useful for delineating edges of magnetic sources because of the amplitude of the analytical signal peaks over magnetic sources. However, if more than one magnetic/density source is present, shallow sources dominate the analytical signal. As *HDTDR* derivative AS not depend on the inclination of the source (Saada (2016b)). Results of derivative techniques for the residual potential data can be observed in the Fig. 12. A population of geophysical lineaments is a mappable linear or curvilinear feature over a physical layer that may be a manifestation of a fault, a magnitude shift, an intrusive body, or another linear zone of weakness (Radaideh et al. (2016); Acosta et al. (2020)). Following the development of geological mapping using potential models, lineament mapping has gained popularity (Osagie et al. (2021)). A lineament is formed by the intersection of two sub-planar surfaces. Exploration of the

Earth's surface yields maps which show lineaments caused by the intersection of near-vertical geological boundaries with the ground (Hall (1986)).

To identify geophysical lineaments, linear features from the *TDR*, *HDTDR* and *TDX* derivatives grids were extracted. Hillshade techniques are widely used in tectonically active regions in order to delineate linear features with preferential trends (Radaideh et al. (2016); Prabhakaran and Jawahar Raj (2018); Acosta et al. (2020)). Hillshade images reproduce a lighting effect from a starting point with certain values of altitude and azimuth of the light, and the linear features with the highest amplification and contrast are those perpendicular to the direction of the light source (Abdullah et al. (2010)). To extract lineaments with varied orientations, the hillshade tool in ArcMAP 10.8 was used to produce eight hill-shaded raster images for each tilt derivative grid. The first shaded relief image was illuminated from the north (i.e., the solar azimuth was 0°), with a solar elevation of 45° (ArcGIS default); for the other seven relief images, an identical solar elevation was used along with seven other illumination directions, including: 45° , 90° , 135° , 180° , 225° , 270° and 315° . Depending on their direction regarding the simulated light source, azimuth illumination and tilt values tend to obscure or accentuate morphological characteristics in a hillshaded image. The MDOW (multi-directional oblique weighted) shaded relief approach was then used to construct two multi-illumination hillshaded pictures from combinations of the individual shaded relief images. These combination images highlight linear features with a variety of directions, which would not be visible if only a single shaded image was used (Radaideh et al. (2016); Acosta et al. (2020)). The first four shaded relief images were superimposed and combined to produce one image with multi-illumination directions (0° , 45° , 90° , 135° – Combination 1), and the four remaining images (180° , 225° , 270° and 315° – Combination 2) were processed in the same way.

Each combination of the tilt derivatives multi-illumination grids were uploaded to the PCI Geomatica 2012 program, and a LINE method was used to each dataset to extract linear features (i.e., lineaments). The LINE module requires three stages: (1) edge detection, (2) picture thresholding, and (3) curve extraction. Each dataset yields a vector layer of polylines as a result (Radaideh et al. (2016)). The operation of the algorithm requires a group of parameters for its execution: RADI (Filter Radius) = 10: GTHR (Edge Gradient Threshold) = 10: LTHR (Curve Length Threshold) = 30: FTHR (Line Fitting Threshold) = 7: ATHR (Angular Difference Threshold) = 7: and DTHR (Linking Distance Threshold) = 70. The choice and understanding of each parameter were based on previous works (Abdullah et al. (2010); Radaideh et al. (2016); Kusák and Krbcová (2017); Salui (2018); Acosta et al. (2020)). Each vector layer was imported into ArcGIS 10.8 to be processed and analysed. The numerical lengths and directions of each set of lineaments were determined using the ArcMap10.8 COGO tool.

From the merge of all the geophysical lineaments, for each residual potential model, two density maps of lineaments versus their length were developed (Fig. 13-A & 13-B), in order to understand their spatial distribution and highlight tectonic features (i.e. Fracture zones). A density map of the basement faults versus their spatial distribution was also produced (13-C). Areas with high lineament density could represent areas of high deformation, igneous intrusion geometries and/or regional faults (Hall (1986); Acosta et al. (2020); Omietimi et al. (2021); Osagie et al. (2021)). The clearest trend of the geophysical lineaments with the faults occurs near the Märjamaa rapakivi pluton and over the PPDZ, that in its southeastern path divides the basement into areas of amphibolite and granulite facies of metamorphism, in the NNE and SSW, respectively (All et al. (2004)). In the Fig. 13-D, areas with high HF results (In purple tones) were compared with areas with higher lineament densities (In orange to red), showing again a similar spatial relationship between those parameters, particularly around the Märjamaa rapakivi pluton and the PPDZ. For each basements division zone, a group of geophysical lineaments rose diagrams were created to determine the dominant azimuth direction (Fig.14). A general NW-SE trend is observed for all the

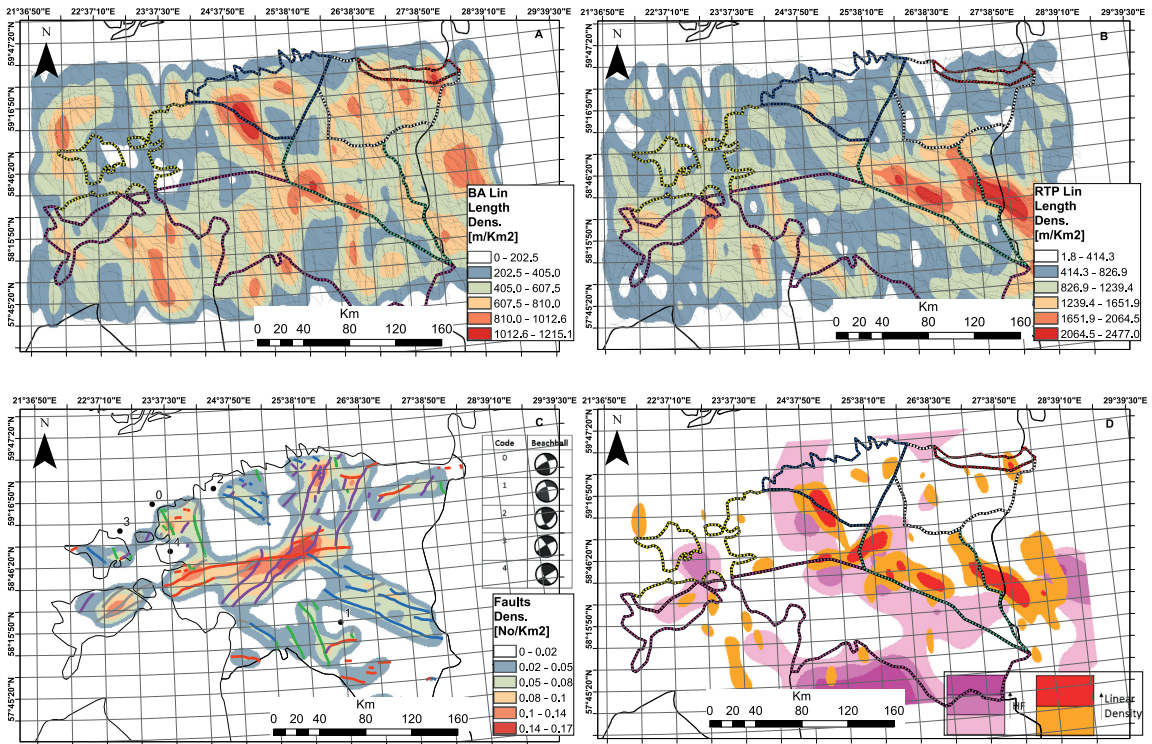


Fig. 13. Density maps of the lineaments obtained from the residual part of the potential models. A) Length density map of the gravimetric lineaments.; B) Length density map of the magnetic lineaments.; C) Spatial density map of the Precambrian faults. Basement faults are coloured according to their orientation regarding the maximum horizontal stress (σ_H of 115° – 135°) in central Fennoscandia (from Korja et al. (2019)). The angle to σ_H and the optimal orientation of the anticipated mode of movement: blue indicates 0° , transfer and normal; purple indicates 90° , reverse; green indicates 30° – 40° , sinistral strike slip; red indicates 30° – 40° , dextral strike slip; grey indicates other. The epicenters of the earthquakes analysed by Soosalu et al. (2022) (See Table 2 in Soosalu et al. (2022)) are shown as a black dot labelled with a number (code"), each with its respective focal mechanisms.; D) Comparison between the zones with highest HF values (from Eqns.9,11,12) and the zones with highest density of tectonic features (i.e. Lineaments and basement faults...). (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

lineaments, however the gravity lineaments tend to present a NNW-SSE trend, while the magnetic lineaments enhance a WNW-ESE trend.

5. Discussion and interpretation

5.1. Gravity, magnetic and topography data

By observing the WGM-12 CBA (Fig. 3-A) and the TFMA EMAG2v3 (Fig. 3-B), the highest potential values are located across the Saaremaa tectonic zone and PPDZ to the NW zone. When comparing the potential models with the ETOPO1 topography (Fig. 3-C), it is observed that highest potential values are not correlated with areas with highest topography zone, which is normal since Estonia is a flat country that alternates between uplands and plateau-like regions, lowlands, depressions, and valleys. There is not a clear correlation between the Depth to the basement map (Fig. 3-D) with the potential results, however, the highest potential values are mostly observed in the southern area, where the basement is at a greater depth. The lowest regional potential values of the are observed in the Tallinn and Alataguse regions, characterise by amphibolite facies (All et al. (2004)). Tapa division of granulite facies rocks, produce notorious positive gravity and magnetic anomalies, contrasting the Tallinn and Alataguse zones.

The PSG map (Fig. 6-A3), can be compared directly with measurements of the gravity field (Nasuti et al. (2012)). It is observed that the

highest values are also presented in defined tectonic zones, presenting a similar trend as the CBA (Fig. 3-A). One of the most notable differences can be seen in the Jõhvi zone, in where in the PSG map present high values, which is not observable in the CBA map. Other notable differences are observed on the islands of Saaremaa and Hiiumaa, where the CBA values are high. Nonetheless, when comparing the PSG map with the CBA and RTP maps, a well spatial correlation of positive anomalies are observed throughout the territory, specially over the PPDZ and the Märjama rapakivi pluton. After applying a low/high pass filter to both CBA and TFMA data, the shallowest sources could be highlighted and defined. From BA residual (Fig. 6-A2) and RTP residual (Fig. 6-B3) maps, it can be seen that high potential values are presented in the areas that in where microcline and rapakivi granites are located, as well as in the Saaremaa tectonic zone and PPDZ. The NE border of the PPDZ is marked by a system of discordant negative magnetic and gravity anomalies suggesting the existence of a fault belt reworked by intrusive processes (All et al. (2004)). The W–E striking Saaremaa tectonic zone crosscuts the PPDZ, and it is well expressed as a linear belt of positive anomalies of the residuals gravity and magnetic fields.

The residual values observed in the Tallinn and Alataguse divisions enhance some zones with high values, coinciding with zones with metal anomalies, specially for the Alataguse zone. The gravity field in the Alataguse zone is complicated by several relative highs, produced by blocks of high-grade metamorphic rocks (Jõhvi zone and Uljaste

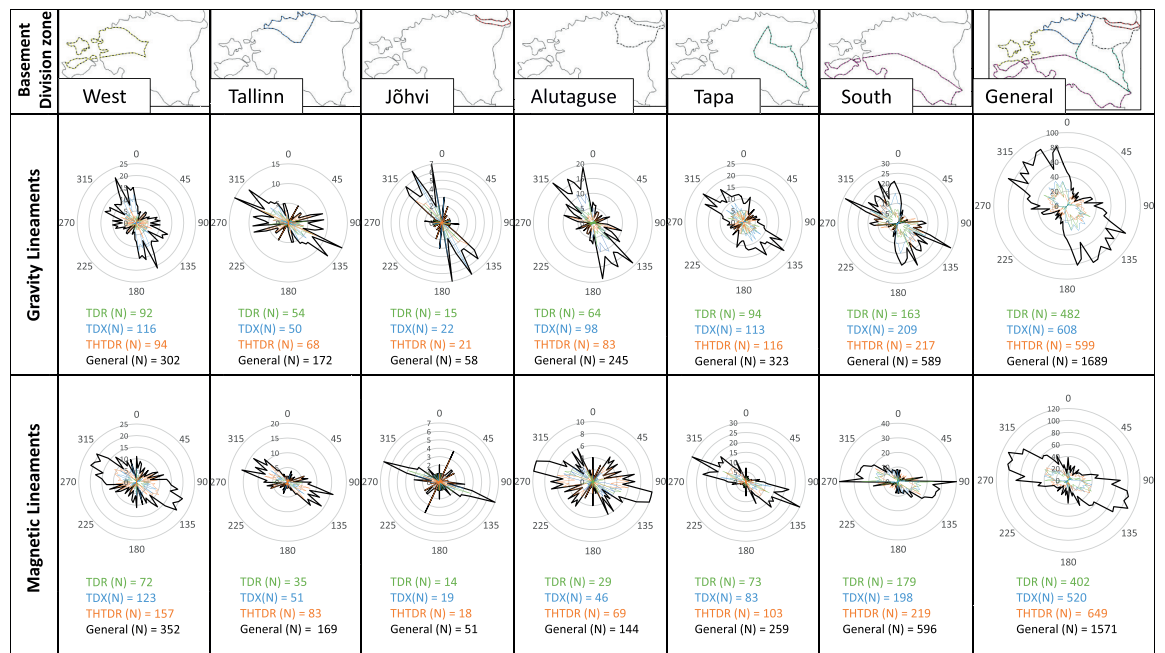


Fig. 14. Rose diagrams from potential subdivided into the geological-structural divisions of the Estonian Precambrian basement. . (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

anomaly; All et al. (2004); Soesoo et al. (2021)). The magnetite quartzites, discovered in the Jõhvi zone, are manifested by the sharpest magnetic anomaly, with not production of any significant gravity anomaly. The AS of the BA residual (Fig. 12-A4) has a fairly spread distribution of positive values across Estonia, with a slightly higher trend of positive values in the west, also the Märjamaa and Naissaare rapakivi granites are well enhanced, like the Riga pluton. The AS map of the residual RTP anomalies (Fig. 12-B4), shown that the positive values occur mainly in central and southern Estonia. The AS maps for both residual potential fields show positive values for the Jõhvi region.

5.2. Lithosphere layers

The morphology map of the Moho discontinuity under Estonia is shown in Fig. 8-B. Several models have been done for Europe, in where in the values for Estonia presents a range of values of 55–60 Km (Grad et al. (2009)), 45–58 km (Molinari and Morelli (2011)), 48–50 Km (Polkowski et al. (2013)). Kaikkonen et al. (2000) present a Moho model for Fennoscandian, in where their depth values for Estonia ranges from 48–56 km. All et al. (2004) used deep seismic sounding data and geophysical modelling, that shows that the Estonian crust is substantially over thickened up to 55–65 km. Our results are consistent with the mentioned models but suggest that the Moho is a little bit deeper than the previous results with values between 48–72 km, with an average depth of 60 km (Table 1). Grad et al. (2014) present a Moho model for the Fennoscandian shield with depths between 35–70 km, where the highest values are observed in central and NE Latvia, and in central and southern Finland, suggesting a consistency with our results. Also, as in the previous investigations, the NE region of Estonia (Alutaguse and Jõhvi areas) presents the shallowest Moho depths, meanwhile, the deepest values are found in the Tallinn zone and over the southern part of Estonia. Tectonic stacking of crustal sheets was a late Svecofennian orogenic process responsible for the main specific features of the basement of south-western Estonia, including its thickness (All et al. (2004)).

The lithosphere's effective elastic thickness (T_e) is a proxy for its strength, comparable to the thickness of a hypothetical elastic beam that would bend in the same way as the lithosphere under the same applied stresses. According to the thin plate concept, the lithosphere is expected to deform elastically in response to ancient stresses. The flexural rigidity of the lithosphere is allowed to vary spatially and to have density variations that are considered as internal loads (Tassara et al. (2007)). Pérez-Gussinyé et al., 2004 present a model of the effective T_e from the Fennoscandian Shield using spectral methods, in where the values from Estonia ranges from 74 to 130 km, estimated from coherence analysis of topography and Bouguer anomaly data (Fig. 15-A). Due to T_e exceeds crustal thickness in this area, the mantle must contribute significantly to the total strength (Pérez-Gussinyé et al. (2004)), indicating that the stability of cratons is due not only to the chemical buoyancy of their root, but also, and importantly, to the strength of the root (Pérez-Gussinyé and Watts (2005)). T_e values in continental areas are primarily related to the combined effects of rheological and thermal heterogeneity's, which controls the coupling or decoupling conditions of the lithospheric layers, as demonstrated in Tesauro et al. (2012), where the lithosphere of the Fennoscandian region, and other Precambrian shields present layers coupled.

Seismic anisotropy—or shear wave splitting—has proven to be a valuable technique to better understand lithosphere-asthenosphere interaction. A delay time between the fast and the slow components will accumulate, in proportion to the length of their path, as well as the strength of anisotropy of the medium (Crampin and Peacock (2008)). Media with wider and larger cracks will have a longer time delay (>0.7s) than a media with small or even closed cracks (Wüestefeld (2007)), being also generally longer in a thick lithosphere. A material has seismic anisotropy when elastic waves move at different velocities in different directions through it. Under the subcontinental upper mantle – lower crust conditions, the fast polarisation is expected to align with the direction of the deformation (Wüestefeld (2007)). The fast wave founded by Wüestefeld (2007) in Estonia denotes a NW-SE direction fast

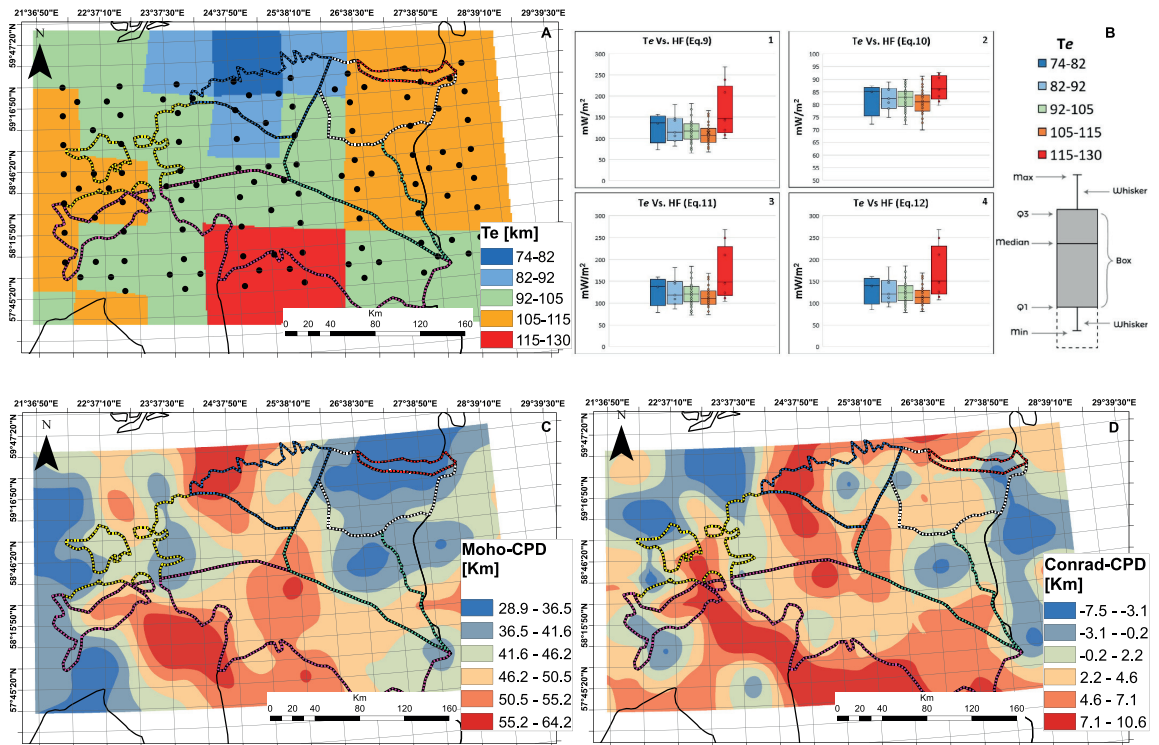


Fig. 15. A) T_e map after Pérez-Gussinyé et al. (2004). The points represent the centred dot of each block used for the CPD calculation; therefore, each point represent a HF value.; B) Box-Plots of the HF values obtained with the Eqns.9–12 versus T_e values, plotting according to the value of each HF point within each T_e polygon category (Tables 2 and 3). HF values obtained from: 1) Eq.9, 2) Eq.10, 3) Eq.11, and 4) Eq.12.; C) Map showing the differences between the Moho depths and the CPD.; D) Map showing the differences between the Conrad depths and the CPD. .

polarisation, with a time delay of ≈ 1 s, attributed to the thick lithosphere of the Baltic shield, as our results also suggest. Estonia basement has been proposed to be result of the convergence of Precambrian separate tectonic unit (Bogdanova et al. (2015)), therefore, parallelism between the fast polarisation direction; associated with mantle processes, and crustal magnetic lineaments suggests vertically coherent deformation and/or coupling of different lithospheric layers (Wüsterfeld (2007)).

Conrad discontinuity has been interpreted as Dessai (2020): (i) a seismic velocity boundary which separates the upper crust from the lower crust, (ii) a transitional layer between the continental and oceanic crust and (iii) a boundary between the felsic rocks such as; granite etc., (silica, aluminum, SIAL) and mafic lithologies such as; basalt etc., (silica, magnesium, SIMA). No recent work has been found with measurements of the Conrad discontinuity in the Baltic countries, in fact the only work published to date, where the depth of the Conrad discontinuity is estimated, is found in the research by Ankudinov et al. (1994), which, from seismic profiles, was able to determine that this discontinuity occurs east of Estonia at a depth of 20 to 25 km. Our results are quite similar to those found by Ankudinov et al. (1994), with a depth from 14.1 to 20.7 km, with a mean value of 17.8 km (Table 1). Our results show (Fig. 8-C) that regions where the deepest values of the Conrad discontinuity occurred are under the Tallinn, Jõhvi and SE of the southern zones. Also, some of the shallowest values of this discontinuity are observed over the PPDZ.

Li et al. (2017) obtain the first global model of CPD from the inversion of the EMAG2 magnetic model, showing that the CPD measures are larger in young oceanic lithosphere of age ≈ 30 Ma, but are smaller in

older lithospheres, so small CPD have been found in most of ancient continental cratons, indicating that deep thermal reactivations generated discrepancies between CPD and estimated surface HF, in fact their results for the Baltic countries present a range between 5 to 25 km, similar to those found in this research for Estonia (6.4–26.5 km). No previous research on Estonian CPD, or Fennoscandia, has been found in the literature. Our CPD results over Estonia show that higher depths can be found in the Jõhvi and Tallinn areas and at the boundary of the western terrestrial border. The lowest CPD values are found in the PPDZ and near the Märjamaa rapakivi pluton. The GG (Fig. 10-C) has higher values in areas where the CDP is lower.

HF in ancient and stable continental cratons may differ from subsurface thermal structures due to extended evolutionary histories and late-stage thermal disturbances. There may be significant delays between changes in asthenosphere temperature and observable impacts reaching the surface in thick continental lithosphere. The contribution of shallow radioactive particles to surface heat transport is likewise greater in continental domains Li et al. (2017). Our HF results (Figs.10-D and 11) show higher values in areas where the CDP is lower. It is observed that the lowest range of HF values is given by using the empirical equation proposed by Vargas et al., 2015 (eqn.10; Fig. 11-A), probably because of its empirical assumption, which has only proved its usefulness in the regional HF calculation of the South American continent. The HF obtained from the eqn.9 (Fig. 10-D), eqn.11 (Fig. 11-B), and eqn.12 (Fig. 11-C) present similar trends (Tabs.2 and 3), but the results obtained from the eqns.11 and 12 (Fig. 11-B and C) depict a better the zones with higher and lower HF values, by having more

variables in its calculation.

High- T_e regions generally correlate with areas of high thermal thickness, derived from HF estimates (Audet and Bürgmann (2011)), indicate that T_e is high in regions where the lithosphere is thick and cold (Pérez-Gussinyé and Watts (2005)), therefore, the resistance to HF is dependent of the temperature and the fractures of the crust (Audet and Bürgmann (2011)). Also, magnetic data suggest that the lower mantle is electrically conductive, implying large numbers of free electrons, can also transfer heat, increasing the HF depending on the tectonic and rheological characteristics of the crust (Beardsmore et al. (2001)). Deflections at the Moho boundary are useful to reveal the crustal deformation in different tectonic regimes concerning the crustal stresses, so T_e values indicate the areas where the Moho should be deep (Pérez-Gussinyé et al. (2004); Pérez-Gussinyé and Watts (2005); Tesauro et al. (2012)). Fig. 15-B shows a relationship between the HF values obtained and the T_e zones determined by Pérez-Gussinyé et al. (2004), where it is highlighted that as the T_e value increases, the values of HF tend to increase, together with the range of the HF values, while the areas with lower T_e seem to present lower HF values, and lower range of HF values. This pattern is not that notorious when using the HF values obtained with eqn.10, possibly related to its empirical calculation (Vargas et al. (2015)). The fact that the zones with the highest HF values are spatially related to the zones of highest lineament density (Fig. 13-D), may indicate that these zones are areas where crustal stresses are higher (Tassara et al. (2007)), due to compressional and transtensional regimes (i.e. the Svecofennian orogeny; Puura et al., 2004; Wüestefeld (2007); Bogdanova et al. (2015)), leading to fracture zones where heat flow would be facilitated. By comparing areas with higher HF (Fig. 11) with areas with higher Moho depth (Fig. 8-B) certain spatial correlation can be observed, particularly in the southern part of Estonia, where both values are the highest, while in NE Estonia the HF values are lower, as well as the Moho depths. Over the Märjamaa rapakivi pluton is observed a positive HF anomaly, and a transition to a shallower Moho depth. To understand the relationship between the crust-mantle interface and the CPD for Estonia, we compared our results of the Moho depth with the CPD (Fig. 15-B). The Moho-Curie difference indicates magnetised uppermost mantle and serves as a proxy for mantle temperature (Wang and Li (2015); Herrera et al. (2021)). Since all our values are positive, due to the CPD layer is notoriously shallower than the Moho, it seems that mantle magnetisation does not play the most important role in the study area, and the magnetisation is here mostly related to it geological crust processes.

Although it appears to have relationships with some geological interfaces, the Curie surface is not a stable geological contact. The calculation findings in numerous places show that the Curie surface is mostly around the Conrad discontinuity (Wen et al. (2019)). Fig. 15-C presents the depth differences between the Conrad discontinuity and the CPD, where the positive values represent the areas where the Conrad discontinuity is slightly deeper. Although some differences can be observed between these two depths, it can be seen that the variations between the depth of the two layers are not too high, with an average difference value of $\sim \pm 3$ km.

Shallowest depths recorded from analyse the spectra of the CBA (Fig. 8-D) in this research have an average of 4.2 km. The values show an increase in the depth of the sources from west to east, where a markedly higher value is observed at the Jõhvi area. There is no correlation with the depth of the crystalline basement (Fig. 3-D), which shows an increase in depth from north to south. The results found show that the shallowest recorded depths of the CBA are slightly deeper when amphibolitic facies are observed (Fig. 2).

According to Korja et al. (2002) the conductance in Log Siemens layers of Estonia can be divided in three layers: (i) 0–10 km is $1.3 - 2.5\text{Log}(S)$ for most of the area but the zone of the in where is the Saaremaa Island with $2.5 - 3.5\text{Log}(S)$; (ii) 10–30 km $0.5 - 1.3\text{Log}(S)$ for the Western and Tallinn zones, $1.3 - 2.5\text{Log}(S)$ and for most of the Estonian zone the conductance is $1.3 - 2.5\text{Log}(S)$. (iii) 30–60 km Similar values as

for the (ii) layer, but for the Jõhvi area with a conductance of $2.5 - 3.5\text{Log}(S)$, and an increase in Tallinn with values of $1.3 - 2.5\text{Log}(S)$ (See Fig. 9 of Korja et al. (2002) research). In this sense, the conductance of the basement and the shallow sources present the conductance of the (i) layer, the Conrad and the CPD layers the conductance of the (ii) layer, and in the Moho a conductance of the (iii) layer.

5.3. Structural interpretation

Soosalu et al. (2022) analysed source parameters of five Estonian earthquakes (Fig. 13-C) by utilising the permanent seismic network augmented with several temporary local stations. By joint analysis of Estonian and Finnish waveform data they showed that the earthquakes are confined to the upper crust with a predominantly strike-slip failure mechanism. The low-magnitude earthquake (1.2–4.5) events studied by Soosalu et al. (2022), seems to be located in the upper crust, being expressions of a predominantly left-lateral strike slip movement on NW-SE to NNW-SSE subvertical faults; as well as occurs in southern Finland, in concert with the overall stress field of northern Europe (Korja et al. (2019)).

In southern Finland, from 1.85–1.84 Ga, much of the regional deformation split into regional folds and crustal-scale ductile shear zones along the Fennoscandian shield, evidencing kilometer-wide sinistral shear structures extending in a NW-SE direction across the Åland and Turku archipelagos (Soesoo et al. (2004); Torvela and Ehlers, 2010). Some of these shear zones follow crustal discontinuities that have been interpreted as accreted terrane boundaries (Torvela and Ehlers (2010)). Different researches suggest that the basement of Estonia and southern Finland was formed during the Svecofennian orogeny (Puura et al. (2004); Bogdanova et al. (2006, 2015); Soesoo et al. (2021)), derived from the collision and amalgamation of crustal units by compression movements in a NE-SW direction (σ_1), where the contact plane of the amalgamated structures presented a strike perpendicular to the compression direction (σ_H : NW-SE; (Fig. 13-C)) (Korja et al. (2019)), which subsequently allowed the generation of NW-SE to NNW-SSE sinistral strike-slip fault systems, as well as associated fracture systems and deformation zones (Korja et al. (2019); Soosalu et al. (2022)). It seems that the detected geophysical lineaments from different derivative techniques (TDR, HDTDR, TDX; Fig. 12), depict the tectonic setting mentioned, showing that the geophysical lineaments present the same NW-SE preferential trend, as does the direction of fast polarisation of the shear wave (Wüestefeld (2007)), suggesting that the potential lineaments presented in this research mostly present a tectonic significance, which may be associated with the fact that the highest density of longest lineaments tend to be located around the basement faults occurrences (Fig. 13).

5.4. 2D profiles

To enhance the results obtained, six profiles with the different results were constructed along the Estonian mainland borders (Fig. 16). The location of the profiles can be observed in the Fig. 1. In these profiles we look for contrasts and trends in potential residual signatures, Moho, and Conrad discontinuities, CPD, HF and GG; also including topography and depth to basement, as basement lithology varies; also taking into account geochemical anomalies.

Märjamaa rapakivi plutons (Fig. 16-A and C) for example are characterised for present high residual magnetic values enables to notice how the Märjamaa pluton has a rather high magnetic anomaly, which may be associated with the presence of iron ores. Also as was expected, BA values decrease when granite bodies are found. Yet, one thing that can be observed in the profiles when crossing the Märjamaa granite rapakivi body, is the slight tendency for the residual values of the BA to increase in W-E direction. The HF values tend to increase when the granitic bodies are found in most cases.

PPDZ (Fig. 16-B and D) presents noticeable peaks in the residual

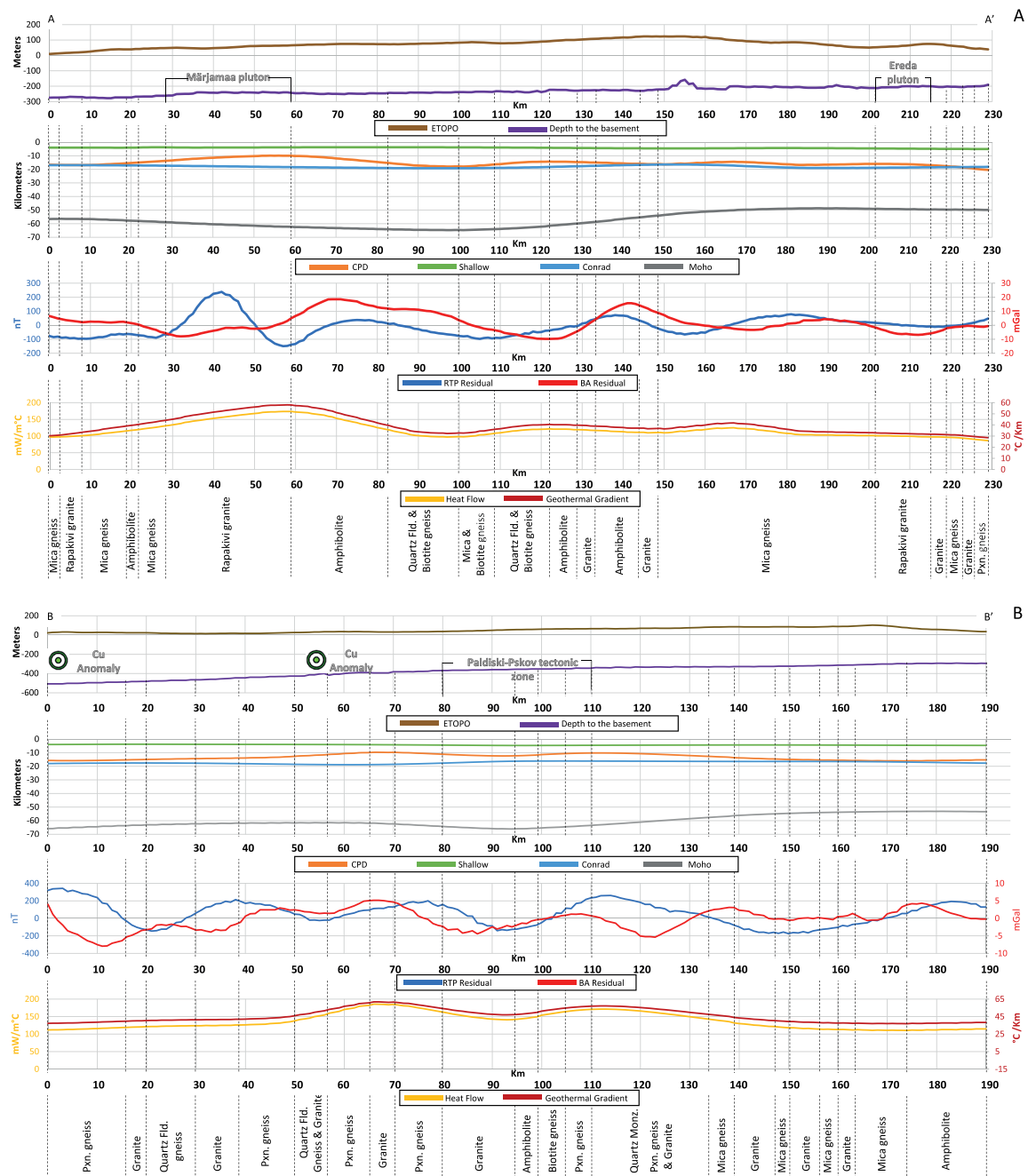


Fig. 16. Profiles across Estonia showing the relations of the topography, crustal layers, residual potential anomalies, and heat flow across the basement lithology. Fig. 1 presents where the profiles are located. A) A-A' profile.; B) B-B' profile.; C) C-C' profile.; D) D-D' profile.; E) AA-AA' profile.; F) BB-BB' profile. .

magnetic and Bouguer anomalies, as well as slightly higher HF values. Some of the areas where geochemical anomalies are found can be B-B' (Fig. 16-B), with a high value in the magnetic anomaly where the Cu anomaly occurs. The AA-AA' profile has the highest number of

geochemical anomalies, where peaks in the Bouguer anomaly are observed in the Zn and Pb anomalies, while a peak in residual magnetic values is observed in the area where the iron ore is hosted, in the NE of Estonia.

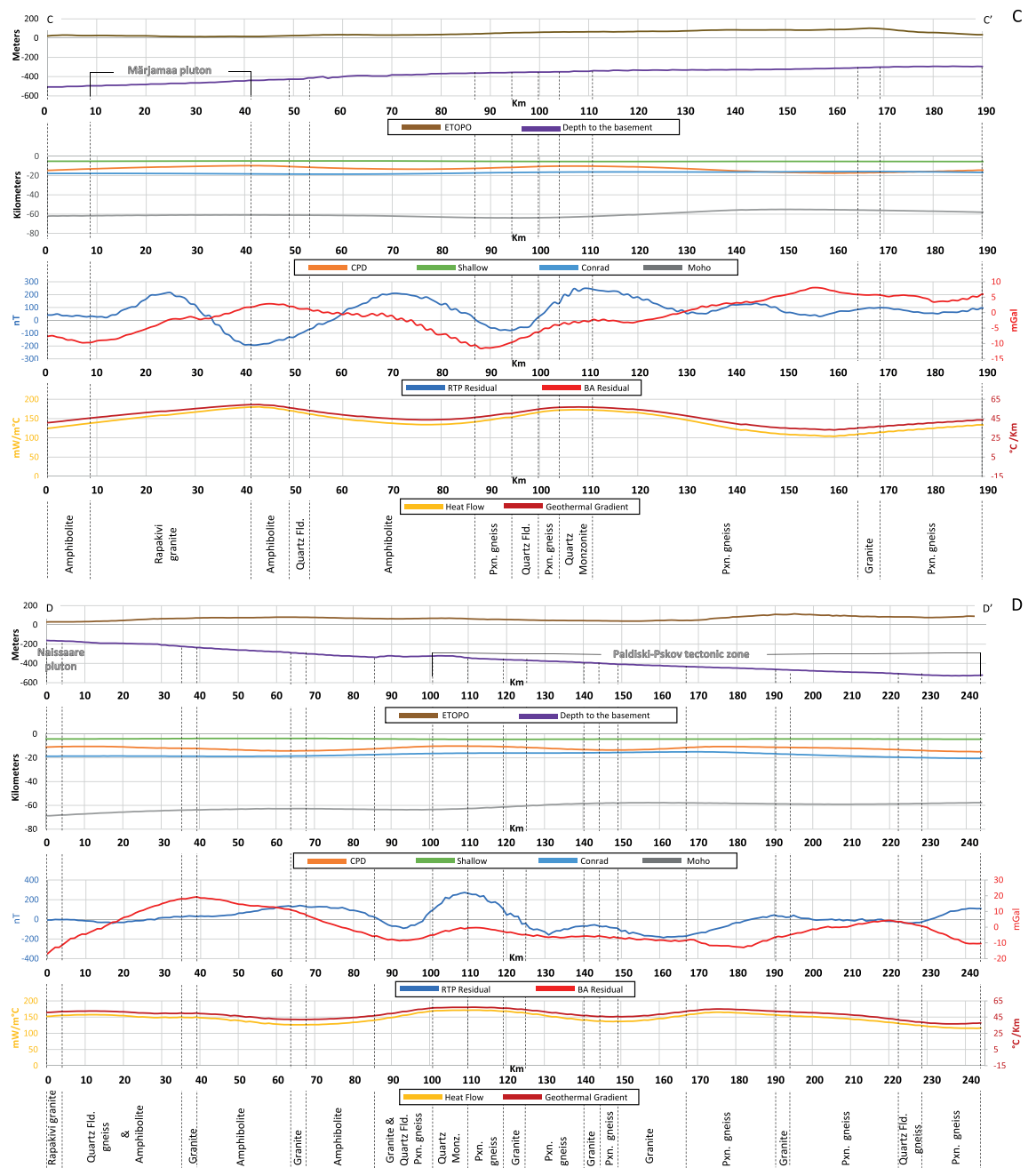


Fig. 16. (continued).

Residual magnetic anomaly values are also evident when looking at the quartz monzonite in central Estonia, which is often associated with high mineralisation content in copper deposits. Also, that anomaly is observed in the profile BB-BB' (Fig. 16-B), which may be associated to large faults or possible metalliferous ores.

6. Conclusions and recommendations

The different processes done over the high-resolution potential data sets were mapped to understand better the tectonic structure of the Estonian crust. The following results can be highlighted from this analysis:

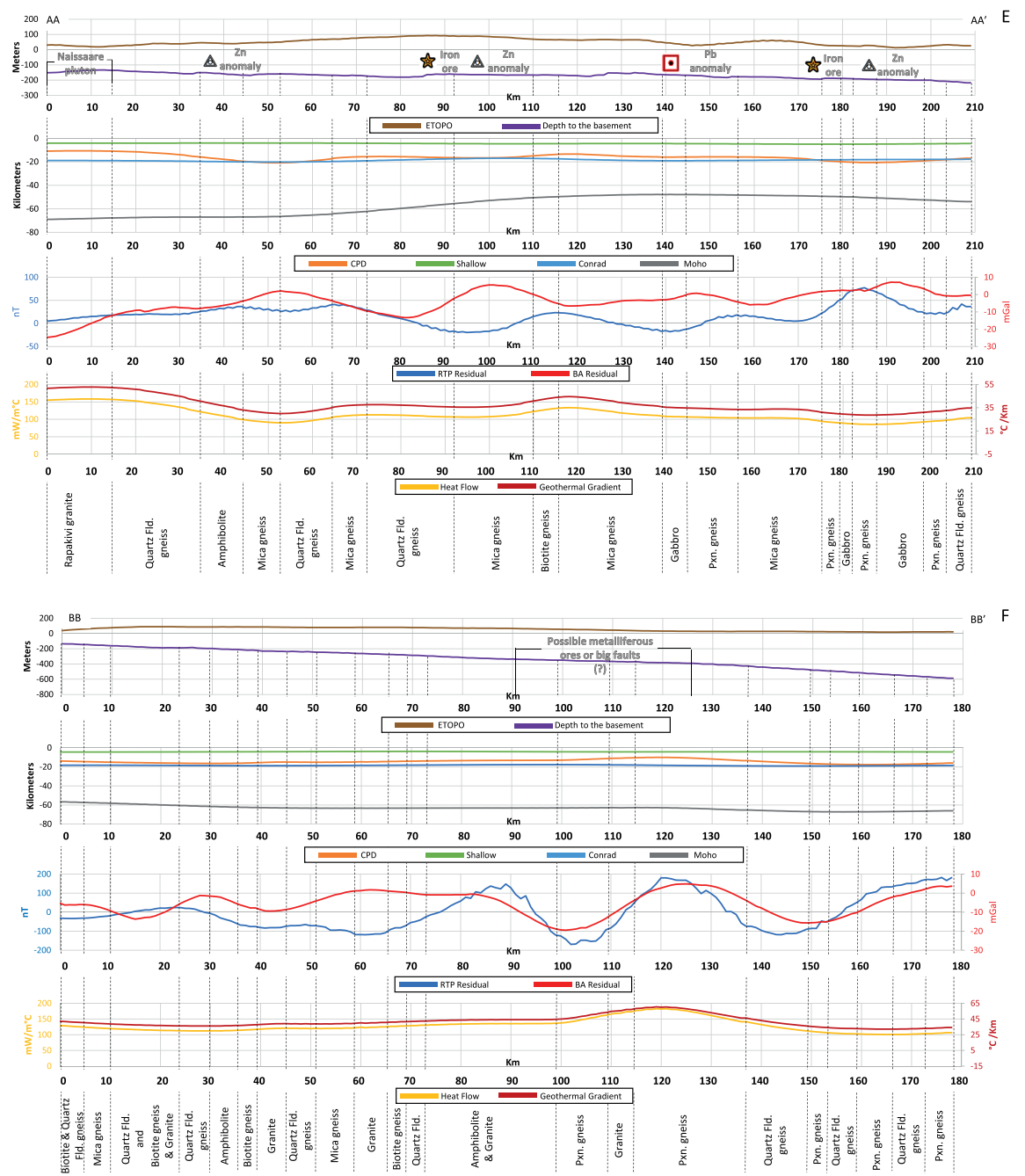


Fig. 16. (continued).

- It has been discovered that using residual data, the Tallinn and Altaguse areas enhance several zones with high values, which match with zones with metal anomalies, particularly in the Altaguse zone. Jõhvi zone has high and increased residual potential values, which are most likely related to the existence of iron ores.
- Detailed power spectrum analysis of CBA map has brought out shallow sources depth map (an average depth of 4.21 km), Conrad discontinuity map (an average depth of 17.8 km) and the Mohorovičić discontinuity map (an average depth of 60.45 km) in the study region.

- Previous research on elastic thickness and seismic anisotropy backs up the crust thickness results (Moho boundary).
- CPD values are slightly shallow (at a depth like the Conrad discontinuity). The lowest CPD values are found in the PPDZ and near the Märjamaa rapakivi pluton, in contrast to the HF, which has its highest values in these zones.
- Heat Flow tends to increase around the zones where the T_e and the Moho discontinuity are deeper, and possible fractured zones arise (i. e., deformation zones).
- The Moho-Curie difference shows that crust magnetisation seems to be more significant rather than mantle magnetisation.
- Predominant NW-SE lineament are tectonic expressions that supports the theory that claims that Estonian basement is formed due to the convergence of tectonic blocks convergence during the Svecofennian orogeny.

For future research, we recommend, to compare the results observed in this research with values obtain from local potential data. Also, the comparison of the HF values should be contrasted from the HF values obtained from drill cores. It could be interesting in generate a 3D-modelling over the Märjamaa rapakivi pluton and in the PPDZ with potential models, involving petrophysical analysis. Detail geophysical modelling in zones with high-contrast potential values and recognized metalliferous anomalies should be developed as well.

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CRedit authorship contribution statement

Juan David Solano-Acosta: Conceptualization, Methodology, Formal analysis, Investigation, Resources, Writing – original draft, Visualization. **Alvar Soesoo:** Conceptualization, Investigation, Supervision, Writing – review & editing, Project administration. **Rutt Hints:** Conceptualization, Investigation, Supervision, Writing – review & editing, Project administration.

Declaration of Competing Interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

Data availability

Data will be made available on request.

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Appendix II

J D Solano-Acosta, A Soesoo, and R Hints. Geochemistry, provenance, and tectonic setting of Paleoproterozoic metasedimentary and metavolcanic units of the Estonian Alutaguse region, eastern Fennoscandia. *Estonian Journal of Earth Sciences*, 74:61–82, 2025. [doi:10.3176/earth.2025.05](https://doi.org/10.3176/earth.2025.05)



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Geochemistry, provenance, and tectonic setting of Paleoproterozoic metasedimentary and metavolcanic units of the Estonian Alutaguse region, eastern Fennoscandia

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ABSTRACT

This research focuses on the geochemical analysis of Paleoproterozoic metasedimentary and metavolcanic units in the Alutaguse region of northern Estonia, shedding light on the geodynamic evolution during the Svecofennian orogeny in eastern Fennoscandia. The metasedimentary units consist of micaceous gneisses ($\pm$ Grt $\pm$ Crd $\pm$ Sil), and the metavolcanic units include amphibolites and pyroxenic gneisses. Geochemical analyses utilized both historical and new whole-rock geochemical data. Weathering indices indicated their applicability for provenance studies and tectonic setting analyses. Metasediments are classified by their silica content: high-SiO₂ (>63 wt%) metasediments resemble litharenites, implying higher maturity and felsic origins akin to the upper continental crust reference; low-SiO₂ (<63 wt%) metasediments align with graywackes and shales, indicative of mafic to intermediate origins, similar to the post-Archean Australian shale, with TiO₂-Ni suggesting sedimentary trends. Discriminant tectonic parameters associated these metasedimentary groups with a continental rift domain. Total alkali-silica classified the metavolcanics as subalkaline units. Geochemical ratios, such as La/Yb vs. Zr/Nb and La/Sm vs. Sm/Yb, crossing the spinel-lherzolite trend, were closest to the primitive mantle reference. The Th/Nb and Th/Zr ratios revealed asthenospheric mantle origins for the basaltic magma sources in Alutaguse. Tectonic settings derived from Y/15-La/10-Nb/8 and TiO₂-10(MnO)-10(P₂O₅) ratios suggested a predominant oceanic arc affinity. It is proposed here that the Alutaguse structural zone developed as the back-arc of the Tallinn-Uusimaa belt(s), following the accretion of the Bergslagen microcontinent at 1.9–1.87 Ga, concluding with the closure of the paleo-Svecofennian ocean.

1. Introduction

The Estonian Paleoproterozoic crystalline basement in Baltica is not fully understood due to the predominance of Lower Paleozoic strata concealing it, with insights primarily derived from drill cores and geophysics (Puura and Huhma 1993; Kivisilla et al. 1999; Skridlaite and Motuza 2001; Soesoo et al. 2004; All et al. 2004; Bogdanova et al. 2015; Soesoo et al. 2020; Nirgi and Soesoo 2021; Solano-Acosta et al. 2023). Despite limited surface exposure, studies indicate high-grade Paleoproterozoic metavolcanic and metasedimentary rocks in northern Estonia, segmented principally into the Tallinn, Alutaguse, and Jõhvi structural zones (Fig. 1). These zones possess 1.92–1.88 Ga rocks, akin to those found in southwestern Finland and central Sweden's Bergslagen zone (Kirs et al. 2009; Bogdanova et al. 2015; Soesoo et al. 2020).

The Alutaguse zone has been described as a folded metasedimentary basin, formed post-closure of the Tallinn volcanic belt's back-arc, but it still lacks comprehensive geochemical and geochronological studies, making its evolutionary model unclear and its genesis widely debated (Kivisilla et al. 1999; Kirs et al. 2009; Bogdanova et al. 2015; Soesoo et al. 2020). Geophysical and isotopic studies suggest that the Tallinn zone is an accreted island arc belt, potentially extending into the Finnish southern Svecofennian Uusimaa belt domain across the Gulf of Finland seabed,

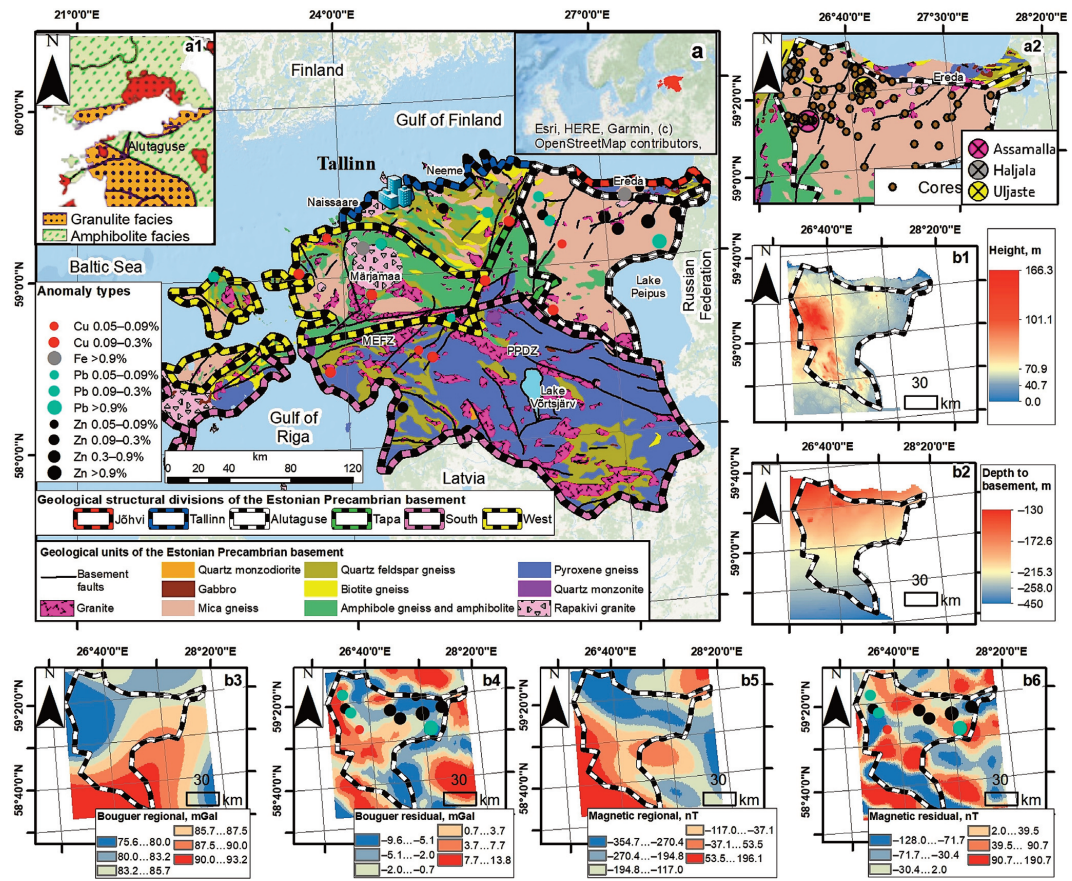


Fig. 1. a – geological schematic map of the Precambrian basement of Estonia, showing sulfide anomalies according to Soesoo et al. (2020); a1 – inset illustrating the distribution of granulite- and amphibolite-facies metamorphic rocks, with reddish polygons corresponding to rapakivi lithologies, after Bogdanova et al. (2015); the anomaly types represent the most prominent metalliferous anomalies; a2 – zoomed-in geological basement map of the Alutaguse zone, depicting the locations of cores and prominent metallogenic regions. b – geophysical maps of the Alutaguse zone, depicting the 10 × 10 m topography (b1), depth to basement (b2), Bouguer gravitational anomaly (b3), residual Bouguer anomaly (b4), regional magnetic anomaly (b5), and residual magnetic anomaly (b6). Residual maps present the locations of metalliferous anomalies. Graphs b3–b6 are modified after Solano-Acosta et al. (2023). The geological, topographical, and depth-to-basement data can be downloaded from the Geoportal of the Estonian Land Board. Abbreviations: MEFZ – Middle-Estonian fault zone, PPDZ – Åland–Paldiski–Pskov deformation zone.

implying that the Alutaguse zone may represent a back-arc extension of the Tallinn–Uusimaa belt(s) (Puura et al. 1983; Petersell and Levchenkov 1994; All et al. 2004; Kirs et al. 2009), with metamorphic aluminosilicates dated to approximately 1.88 Ga (Kähkönen 2005; Bogdanova et al. 2015; Nironen 2017; Lahtinen et al. 2017; Kara et al. 2021).

The major, trace, and rare earth elements (REE) in metasediment and metavolcanic rocks provide insights into their origin, weathering processes, and tectonic settings, enhancing our understanding of crustal geochemistry (Bhatia and Crook 1986; McLennan et al. 1995; Han et al. 2019). These elements elucidate the contributions of felsic and mafic sources to crustal evolution and rock origin and provenance (Sifeta et al. 2005). Additionally, geochemical studies on Precambrian metasedimentary and metavolcanic units have been crucial in understanding ancient lithological arrangements within

tectonic and geodynamic contexts (Lahtinen et al. 2002; Sifeta et al. 2005; Lahtinen et al. 2010; El-Bialy 2013; Chen et al. 2014; de Carvalho Mendes et al. 2021).

This research uses new and historical whole-rock data to present an updated geochemical analysis of the metasedimentary and metavolcanic rocks in the Alutaguse zone. The study focuses on analyzing major and trace elements, as well as REE components, to determine the composition and origin of the source rocks. The goal is to identify geochemical trends for weathering, depositional features, and tectonic settings, which can provide insights into the geodynamic evolution that shaped the Estonian Alutaguse zone. These patterns can provide insights into the processes that have shaped this area. Understanding these patterns enhances our knowledge of the geological development of the Svecofennian orogeny and the Proterozoic history of Estonia's basement within the broader

context of Fennoscandia (Kähkönen 2005; Kirs et al. 2009; Baltybaev 2013; Bogdanova et al. 2015; Nironen 2017; Soesoo et al. 2020; Pesonen et al. 2021; Kara 2021; Lahtinen et al. 2022).

2. Geological setting

The Estonian Paleoproterozoic crystalline basement, part of the 1.9–1.7 Ga old Svecofennian domain, features metasedimentary and metavolcanic rocks that underwent amphibolite-to-granulite metamorphism, with some retrograde modifications, associated with contemporaneous mafic to intermediate intrusions (Puura et al. 1983; Puura and Huhma 1993; Kivisilla et al. 1999; Soesoo et al. 2004; All et al. 2004; Kirs et al. 2009). Between 1.6–1.4 Ga, rapakivi granite plutons intruded the consolidated basement (All et al. 2004; Soesoo et al. 2020; Fig. 1a). Estonia's structural framework is outlined by the 30 km-wide northwest-trending Åland–PPDZ–MEFZ (Åland–Paldiski–Pskov deformation zone – Middle-Estonian fault zone) deformation zone, with major shear zones dipping 65–75° SSW, separating northern regions of amphibolite and granulite facies from predominantly granulitic southern regions (Puura et al. 1983; Soesoo et al. 2006; Kirs et al. 2009; Bogdanova et al. 2015; Soesoo et al. 2020; Solano-Acosta et al. 2023).

Geological categorization of Precambrian basement rocks in Estonia identifies three main age groups: (1) the oldest in northern Estonia, which includes amphibolite-facies metavolcanic rocks dated at 1918 ± 10 Ma (Petersell and Levchenkov 1994); (2) southern Estonia's granulitic metavolcanic rocks and tonalites in the Tapa zone, dated at 1832 ± 22 , 1827 ± 7 , and 1824 ± 26 Ma, and magnetite-rich gneisses in the Jõhvi zone with ages spanning 1874 ± 18 to 1789 ± 19 Ma (Soesoo et al. 2004, 2006; Kirs et al. 2009; Bogdanova et al. 2015; Soesoo et al. 2020); and (3) the youngest group, which includes 1.6 Ga rapakivi granite plutons and associated mafic and felsic rocks (Kirs and Petersell 1994; Soesoo and Hade 2012; Rämö et al. 2014). Notably, no dating analyses are reported for Alutaguse lithologies.

Northern Estonia's Svecofennian orogeny units host metamorphosed and migmatized amphibolite-facies rocks, divided into the Tallinn zone (northwest), which contains amphibolites and biotite-amphibole gneisses, and the Alutaguse zone (northeast), which presents high-alumina gneisses, amphibolites, and biotite-amphibole gneisses (Kivisilla et al. 1999; Soesoo et al. 2004; Kirs et al. 2009; Bogdanova et al. 2015). Additionally, the Jõhvi zone, located to the northeast of Alutaguse, contains thick Fe- and S-rich quartzites, high-Al garnet-cordierite-sillimanite ($\pm$ Grt $\pm$ Crd $\pm$ Sil) gneisses, and Ca-rich and -poor pyroxene-amphibole-biotite gneisses, and has experienced granulite facies metamorphism (Bogdanova et al. 2015; Soesoo et al. 2020; Nirgi and Soesoo 2021).

The Alutaguse zone is characterized by high-temperature amphibolite-facies conditions (Fig. 1a1), with pressures estimated at around 3–5 kbar (Puura et al. 1983). It prominently features metapelites, Al-rich mica gneisses, graphite gneisses, and biotite-plagioclase gneisses, along with some metavolcanic pyroxenitic gneiss sequences and felsic/mafic intrusions

(Kivisilla et al. 1999; Puura et al. 2004; Kirs et al. 2009; Bogdanova et al. 2015; Soesoo et al. 2020). The metasedimentary Alutaguse zone is considered to be part of the large Kalevian-age (1.9–1.8 Ga) marginal basin that extends to the vicinity of St. Petersburg and Novgorod in Russia, and farther east to Lake Ladoga. However, the ages of deposition and metamorphism of the Alutaguse metasedimentary sequence are still unknown and must be determined before its tectonic setting can be established (All et al. 2004; Bogdanova et al. 2015). Our research will focus on the Estonian Alutaguse section (Fig. 1).

Contrastingly, the Tallinn zone covers mafic amphibolite-facies sequences with amphibole gneisses and magnetite quartzites (Kivisilla et al. 1999). Research by Klein (1986) revealed the average mineral contents of different rocks in the Tallinn and Alutaguse zones, calculated from drill core data. The results showed that rocks with Al-rich garnet make up 25.4% in Tallinn and 90.45% in Alutaguse, while biotite and plagioclase gneisses account for 24.4% and 1%, respectively. Combinations such as Bi-Pl and Bi-Hbl-Pl comprise 50.2% and 6.1%, respectively. The Estonian–Latvian Granulite Belt, situated in Estonia's west and south, consists of charnockitized amphibolites and biotite-feldspar gneisses at 5–6 kbar (All et al. 2004; Soesoo et al. 2006; Bogdanova et al. 2015).

The Estonian Alutaguse region is overlain by a low-altitude landscape, with the highest point reaching up to 166 m (Fig. 1b1), underlain by Precambrian lithologies that extend to depths ranging from 130 m in the north to 450 m in the south (Fig. 1b2). Furthermore, the Alutaguse zone is characterized by low gravity and magnetic field values (Fig. 1b3, 5), yet notable metalliferous anomalies for elements such as Cu, Pb, and Zn have been recognized in the northern part. These anomalies are geographically related to positive residual potential anomalies (Soesoo et al. 2020; Solano-Acosta et al. 2023; Fig. 1b4, 6). Such anomalies are especially prominent in the Uljaste, Assamalla, and Haljala localities, which have been linked to sulfide-graphite gneisses and quartzites (Kivisilla et al. 1999; All et al. 2004; Soesoo et al. 2020; Figs 1a2 and S1).

3. Materials and methods

Thirteen metasedimentary and three metavolcanic rock samples were collected. These were subjected to whole-rock chemical analyses to evaluate their major, trace, and REE compositions and provenance signatures. X-ray fluorescence (XRF, $n = 14$) and inductively coupled plasma optical emission spectrometry (ICP-OES, $n = 2$) were used to determine the major elemental composition. Trace and REEs were analyzed via inductively coupled plasma mass spectrometry (ICP-MS). Sample preparation involved powdering fresh samples to a grain size of less than 200 meshes using an agate mill. The powdered samples were dried at 110 °C for over 24 hours to determine major element abundances. At the Tallinn University of Technology laboratory, an XRF analysis was performed on major element concentrations from the Alutaguse samples. The samples were transformed into glass beads using a fusion method that involved a 1:1 mix of lithium

tetra- and metaborate for macroelements in XRF procedures. Every bead consisted of 1 gram of the sample combined with 10 grams of borate, followed by fusion. The loss on ignition (LOI) value was established based on the total weight loss after igniting at 1000 °C for two hours. The XRF measurements were calibrated with the SPECplus software (Malvern Panalytical, Netherlands), with validation ensured through the GeoPT international proficiency tests. At Origin Analytical Limited (UK), the ICP-MS and ICP-OES analyses followed the ISO 9001:2015 standards. ICP-MS and ICP-OES adhered to the GeoPT program calibration, which aims for international rock standard characterization. The trace and major element concentrations were analyzed with the ICP techniques using a SCIEX ELAN 6000 ICP-MS by PerkinElmer (USA).

In the last century, detailed geochemical measurements of the Estonian Precambrian basement lithologies were conducted using traditional silicate analysis techniques, known as wet chemistry, at the laboratory of the Geological Survey of Estonia (EGT), providing valuable information on whole-rock major element geochemistry. All these results were compiled by Kivisilla et al. (1999) into a large dataset, exposing major elemental concentrations (SiO₂, TiO₂, Al₂O₃, Fe₂O_{3tot}, MnO, MgO, CaO, Na₂O, K₂O, P₂O₅, SO₃, LOI; wt%) of the Estonian basement units. Here, we used the major element data from Kivisilla et al. (1999) for the metasedimentary and metavolcanic samples from the Alutaguse zone (n = 229; Table S1). The complete dataset from Kivisilla et al. (1999) is available in the Estonian geoscience literature database, as part of the SARV geological information system (<https://kirjandus.geoloogia.info/reference/21247>).

The EGT has recently published geochemical trace element data from multiple drill cores in the Uljaste zone (Fig. 1a2). This dataset is presented in the supplementary material and primarily includes trace elements. Major elements, such as silica, and most REEs, except for La and Ce, are notably lacking. We have incorporated these new data into our comprehensive Alutaguse elemental analysis (n = 149; Table S2). The complete Alutaguse trace element dataset is available for download through eMaapõu, an Estonian geological data service managed by institutions such as the EGT, Estonian Land Board, and universities (<https://geoloogia.info/analysis?analysisQ=Uljaste&page=1&itemsPerPage=25&method=26>).

The complete geochemical dataset with the proper ID, core, coordinates, depth, and references for the analyzed lithological units is available in the supplementary material.

4. Geochemical data and processing

The comprehensive analyses of major, trace, and REEs in sixteen newly examined Alutaguse samples are detailed in Tables 1 and 2, respectively. These new major element data were integrated with the existing dataset from Kivisilla et al. (1999) for an enhanced comparative analysis. Similarly, the newly acquired trace element data were integrated and analyzed alongside the information provided by the Estonian Geological Service, utilizing their lithological sampling

Table 1. Major bulk-rock element data of the 16 novel Alutaguse metasedimentary and metavolcanic samples

	Drill core	Analytic method	ID-code	Lithology	SiO ₂	TiO ₂	Al ₂ O ₃	Fe ₂ O _{3tot}	MnO	MgO	CaO	Na ₂ O	K ₂ O	P ₂ O ₅	SO ₃	LOI	Sum
Metasediments	High-SiO ₂	F268 XRF	F2683025	Gamet-bearing mica gneisses ± Crd ± Sil	74.81	0.47	10.96	3.46	0.03	1.64	1.78	1.86	2.39	0.02	0.03	1.32	98.78
		F199 XRF	F1992940	Graphite-bearing mica gneisses ± Grt ± Crd ± Sil	65.64	0.83	16.35	4.03	0.04	2.42	0.20	0.72	6.12	0.06	0.06	2.63	99.10
	Low-SiO ₂	F268 XRF	F2682485	Gamet-bearing mica gneisses ± Crd ± Sil	58.50	0.69	17.41	9.67	0.23	3.59	1.52	1.41	5.65	0.02	0.04	1.06	98.80
		F189 XRF	F1893010	Graphite-bearing mica gneisses ± Grt ± Crd ± Sil	57.18	1.16	11.80	11.85	0.10	4.53	2.61	1.42	2.41	0.20	0.12	5.56	98.94
		F189 XRF	F1893089	Graphite-bearing mica gneisses ± Grt ± Crd ± Sil	56.27	0.69	15.83	10.25	0.12	4.74	3.30	1.82	2.15	0.04	0.08	3.49	98.78
		F152 XRF	F1523492	Graphite-bearing mica gneisses ± Grt ± Crd ± Sil	55.25	0.97	15.03	11.49	0.05	4.35	0.46	0.70	6.29	0.02	0.11	4.26	98.99
		F189 XRF	F1893050	Graphite-bearing mica gneisses ± Grt ± Crd ± Sil	52.16	1.21	13.39	12.95	0.11	4.98	2.41	1.37	3.15	0.21	0.25	6.53	98.73
		F255 XRF	F2553994	Graphite-bearing mica gneisses	51.99	0.53	11.49	16.39	0.10	5.52	0.37	0.52	2.02	0.08	0.31	9.97	99.29
		F152 XRF	F1523600	Graphite-bearing mica gneisses	51.89	0.72	12.66	16.97	0.03	3.31	1.32	1.65	4.74	0.04	0.04	5.56	98.93
		F189 ICP-OES	F1893041	Graphite-bearing mica gneisses ± Grt ± Crd ± Sil	51.86	0.90	13.76	13.49	0.10	3.49	2.09	1.37	2.69	0.15	5.75	—	95.64
		F189 XRF	F1893328	Graphite-bearing mica gneisses ± Grt ± Crd ± Sil	46.37	0.80	16.00	15.57	0.29	4.21	3.09	1.91	3.90	0.06	0.15	6.73	99.09
		F268 XRF	F2682450	Graphite-bearing mica gneisses ± Grt ± Crd ± Sil	44.76	2.33	13.86	10.91	0.28	4.11	6.42	0.07	1.98	0.22	1.74	12.55	99.23
		F189 ICP-OES	F1893070	Graphite-bearing mica gneisses ± Grt ± Crd ± Sil	44.57	1.06	14.26	18.44	0.15	5.37	2.47	1.42	2.73	0.14	9.32	—	99.92
		F189 XRF	F1893458	2-Pyroxene gneisses	45.77	1.05	12.53	13.04	0.23	6.93	9.11	0.22	0.56	0.09	1.87	8.49	99.90
		F268 XRF	F2683260	2-Pyroxene gneisses	52.01	2.25	13.32	11.44	0.18	10.11	2.76	0.91	4.38	0.08	0.11	1.61	99.17
		F255 XRF	F2553248	Pyroxene gneisses	51.45	1.14	14.68	12.71	0.29	7.77	9.84	0.53	0.44	0.10	0.42	0.38	99.74

2-Pyroxene gneisses

Table 2. Trace and REE bulk-rock data of the 16 novel Alutaguse metasedimentary and metavolcanic samples (presented in two parts)

	ID code	Be	V	Cr	Co	Ni	Cu	Zn	Ga	Rb	Sc	Sr	Y	Zr	Nb	Mo	Sn	Cs
Metasediments	High-SiO ₂	F2683025	1.24	58.48	60.73	105.11	36.68	21.23	90.58	16.41	96.12	7.46	167.90	8.32	151.36	13.04	2.64	0.66
		F1992940	1.29	122.37	129.65	76.90	30.37	27.51	116.38	21.89	185.82	7.93	112.52	19.61	227.50	15.62	0.33	0.51
		F2682485	0.98	115.22	117.90	67.64	33.79	8.36	115.84	18.67	156.58	35.14	196.02	82.47	300.37	13.08	0.84	0.35
		F1893010	2.56	294.94	131.38	83.25	93.88	134.77	406.79	17.79	117.35	13.99	99.83	25.02	167.62	19.44	18.28	0.87
		F1893089	7.17	239.15	255.66	82.89	72.17	265.58	224.55	20.65	69.48	19.63	128.90	29.50	131.99	8.79	11.57	0.61
Metasediments		F1523492	1.83	206.87	138.47	76.83	50.60	401.43	831.34	23.78	214.83	10.87	87.77	4.22	266.42	25.44	3.55	0.56
		F1893050	2.01	342.43	160.35	70.36	87.50	128.81	572.05	19.89	155.84	17.23	112.01	28.76	181.96	22.64	10.76	1.33
		F2553994	4.32	214.97	87.86	52.74	36.97	41.48	316.60	17.95	144.32	17.02	25.99	20.48	112.81	10.81	13.04	2.26
		F1523600	0.93	233.65	123.38	73.87	74.09	578.78	183.68	19.85	175.12	16.25	111.14	15.67	159.98	21.82	8.40	1.73
		F1893041	3.41	312.29	122.99	72.10	111.05	146.88	438.49	18.79	107.93	15.73	123.22	29.48	155.64	17.27	19.78	1.13
Metavolcanics		F1893328	1.35	245.79	147.55	60.74	104.18	99.65	270.07	19.59	103.19	27.39	144.46	48.90	146.22	10.92	11.99	0.67
		F2682450	13.48	512.33	87.85	66.16	43.28	54.18	206.71	20.18	44.07	45.94	49.25	33.99	145.31	8.82	1.62	0.47
		F1893070	3.15	288.55	178.71	62.21	125.11	154.26	235.84	21.46	120.36	33.41	117.05	61.73	156.22	16.00	19.02	0.83
		F1893458	1.31	282.53	62.91	53.89	30.56	167.07	377.37	15.42	16.94	39.34	173.58	23.64	64.00	4.94	1.65	0.87
		F2683260	0.97	436.00	938.45	72.97	134.52	25.94	689.20	26.05	224.00	48.57	82.61	13.97	119.56	29.24	0.83	2.04
	F2553248	1.15	310.30	92.62	67.76	33.66	54.98	258.79	18.53	15.77	47.19	270.00	29.19	76.18	4.01	2.10	0.59	0.94

	ID code	Ba	La	Ce	Pr	Nd	Sm	Eu	Gd	Tb	Dy	Ho	Er	Tm	Yb	Lu	Hf	Ta	Pb	Th	U	
Metasediments	High-SiO ₂	F2683025	832.70	79.87	145.73	15.90	52.80	6.39	1.30	5.11	0.49	1.69	0.32	0.92	0.13	0.80	0.12	4.42	0.57	21.48	23.50	0.67
		F1992940	844.42	42.66	79.75	9.85	35.11	6.06	1.26	5.16	0.71	3.58	0.73	2.09	0.31	2.13	0.32	6.53	1.13	12.83	16.32	2.26
		F2682485	1707.24	27.63	46.20	4.97	17.07	3.28	1.74	5.34	1.33	10.66	2.94	9.59	1.61	11.85	1.89	8.58	0.60	32.88	5.44	0.70
		F1893010	471.79	38.25	72.91	9.32	34.76	6.67	1.49	6.26	0.91	4.63	0.93	2.42	0.35	2.29	0.35	4.54	1.74	2.78	10.61	7.27
		F1893089	382.53	48.08	92.49	11.68	43.52	7.93	1.68	7.29	0.98	5.09	1.15	3.27	0.51	3.70	0.63	3.89	0.84	7.40	17.32	3.52
Metasediments		F1523492	826.77	23.38	41.58	4.66	16.51	2.46	1.17	2.12	0.22	0.75	0.14	0.37	0.06	0.39	0.07	7.74	1.37	215.11	5.30	0.69
		F1893050	529.25	46.13	88.01	11.28	43.03	8.12	1.82	7.51	1.07	5.43	1.10	2.85	0.39	2.57	0.39	5.09	1.74	10.77	12.66	6.83
		F2553994	283.49	24.91	51.65	6.13	22.28	4.54	0.66	4.38	0.70	3.86	0.83	2.31	0.36	2.56	0.42	3.38	0.98	9.10	11.41	6.76
		F1523600	551.97	57.21	106.92	12.80	45.82	7.44	1.27	6.40	0.78	3.35	0.60	1.49	0.18	1.08	0.17	4.77	1.32	5.02	22.51	2.87
		F1893041	460.37	42.49	79.41	10.16	37.35	7.04	1.52	6.62	0.97	5.17	1.11	2.92	0.41	2.72	0.43	4.31	1.48	9.91	13.44	8.79
Metavolcanics		F1893328	1032.26	41.87	79.44	9.53	34.50	6.03	1.80	6.14	1.09	7.34	1.79	5.33	0.83	5.68	0.87	4.22	0.86	6.25	12.55	2.56
		F2682450	577.20	5.33	13.76	2.14	10.51	3.11	1.07	4.01	0.82	5.55	1.33	3.67	0.56	3.96	0.60	4.37	0.78	123.90	0.60	1.01
		F1893070	439.50	44.48	85.40	10.67	40.01	7.81	2.02	7.69	1.35	9.17	2.33	6.83	1.01	7.12	1.11	4.62	1.23	5.20	14.69	3.60
		F1893458	369.12	7.32	16.19	2.29	9.81	2.81	1.10	3.31	0.64	3.92	0.90	2.54	0.39	2.49	0.38	1.80	0.58	6.66	1.20	0.95
		F2683260	630.36	5.16	11.21	1.55	6.55	1.69	0.92	1.99	0.34	2.10	0.48	1.34	0.22	1.50	0.24	3.73	2.00	7.09	0.43	1.09
	F2553248	309.62	10.68	23.88	3.30	14.60	3.72	1.39	4.26	0.80	4.73	1.08	2.98	0.44	2.97	0.42	2.24	0.45	7.03	0.66	0.46	

classification system, which was aligned with the drilling core descriptions from the same source, accessible at <https://gis.ege.ee/portal/apps/dashboards/99f758ac4ef548f686b831adb3199378>. Consistency in classifying all analyzed units was ensured by adopting the system outlined by Kivisilla et al. (1999). Table S1 presents the average concentrations of major elements (wt%) found in Alutaguse metasediments, while Table S2 details the average concentrations of trace and REEs (ppm). Bulk-rock major element concentrations, excluding LOI, were normalized to 100% for the samples. Non-normalized LOI values were considered separately when necessary.

The Alutaguse samples were categorized based on their silica concentrations (Verma and Armstrong-Altrin 2013; Chen et al. 2014), distinguishing between high-SiO₂ (>63 wt%) and low-SiO₂ (<63 wt%; Fig. S2). This classification, following Verma and Armstrong-Altrin (2013), helped identify the discriminant functions of distinct tectonic settings and provenance characteristics.

The high-SiO₂ group includes lithologies such as biotite gneisses, garnet-bearing mica gneisses $\pm$ Crd $\pm$ Sil, cordierite-bearing mica gneisses $\pm$ Grt $\pm$ Sil, and graphite-bearing mica gneisses $\pm$ Grt $\pm$ Crd $\pm$ Sil. Conversely, the low-SiO₂ group comprises lithologies with similar mineralogical compositions but with silica values (χ SiO₂) <63 wt%, including biotite gneisses $\pm$ Grt $\pm$ Crd $\pm$ Sil, garnet-bearing mica gneisses $\pm$ Crd $\pm$ Sil, and graphite-bearing mica gneisses $\pm$ Grt $\pm$ Crd $\pm$ Sil. Due to the absence of major element data from the Estonian Geological Service, these units were theoretically classified based on their lithological classification and the preference for high or low silica content for each lithology (Table S2). Novel Alutaguse samples were also classified based on their silica content (Tables 1 and 2).

Metavolcanic rocks in the region, such as amphibolites and pyroxene gneisses, have not been further subclassified and are broadly categorized based on their amphibole and pyroxene content (Table S2). For a more detailed explanation of the lithological classification used in this study, please refer to the supplementary material.

5. Results

Metasedimentary sample values from the Alutaguse region were compared with established geochemical references, including the upper continental crust (UCC; Rudnick and Gao 2003) and post-Archean Australian shale (PAAS; Taylor and McLennan 1985; McLennan 2001). Metavolcanic samples were compared with the primitive mantle (PM) reference (Sun and McDonough 1989).

5.1. Geochemistry of major elements

Figure 2a illustrates the Al₂O₃ Harker binary plots (Harker 1909) for Alutaguse metasedimentary rocks, showing high-SiO₂ samples aligning with UCC, while low-SiO₂ samples resemble PAAS. TiO₂, MgO, and CaO concentrations are higher in low-SiO₂ metasediments. Additionally, the MgO Harker metavolcanic plots reveal higher Ti and Fe con-

centrations compared to the metasediments units (Fig. 2b), with a positive correlation between Ti, Fe, and Ca contents, especially in the 2-pyroxene gneiss samples.

According to geochemical lithological classifications, high-SiO₂ samples predominantly fall within the litharenite domain, while low-SiO₂ samples align with graywacke and shale zones, as indicated by Herron (1988; Fig. S3a) and Pettijohn et al. (1987; Fig. S3b) plots.

High-SiO₂ metasediments had an average SiO₂ content of 69.31%, ranging from 63.25% to 80.20%, suggesting minimal chemical alteration and a more uniform geochemical profile (Chen et al. 2014). This group (Fig. 3a) also displayed narrower ranges of oxides, such as Al₂O₃ (10.21% to 19.03%), Fe₂O₃ (0.00% to 14.42%), MnO (0.01% to 0.47%), and MgO (0.52% to 5.10%). The SO₃ content (0.00% to 4.61%) was insignificant. Conversely, low-SiO₂ samples (Fig. 3b) exhibited a moderate average SiO₂ content of 54.81%, with a broader range from 29.49% to 62.80%, suggesting more substantial chemical alteration and source variability, mostly mafic. This group showed broader oxide ranges, such as Al₂O₃ (7.59% to 28.07%) and Fe₂O₃ (5.56% to 32.59%). Other oxides, including MgO (0.67% to 8.56%) and MnO (0.01% to 1.01%), also presented more extensive ranges. SO₃ content was significant (0.00% to 23.09%), particularly over graphite mica gneisses (Table 1).

Using the total alkali-silica (TAS) classification by Le Bas et al. (1986), metavolcanic samples show low Na₂O, K₂O, and SiO₂ concentrations, aligning with the subalkaline series, which span from basaltic to andesite zones. Metavolcanic samples predominantly align with the ortho-amphibolite domain (Fig. S3c). A pronounced tholeiitic trend is observed on the AFM diagram (Fig. S3d), with most samples distributed across the high-Fe tholeiitic zone (HFT) in the Jensen (1976) plot (Fig. S3d). High Mg# values are observed in the Alutaguse metavolcanic samples, ranging from 34.87 to 68.40, with an average of 51.83, resembling the tholeiitic basalt magma fractionation trend (Casey et al. 2007).

Figure 3c presents major element data for metavolcanic rocks displayed on spider diagrams, normalized to the PM as specified by McDonough and Frey (1989). The data illustrate broader ranges in Na₂O and K₂O concentrations, particularly within the 2-pyroxene gneiss lithologies. Among the major oxides, MgO shows a depletion relative to the PM standard, while others, such as TiO₂ and K₂O, appear elevated.

Metavolcanic samples present an average SiO₂ content of 52.60%, ranging from 38.96% to 62.67%. The TiO₂, Al₂O₃, and Fe₂O₃ concentrations in the metavolcanic rocks are notably higher compared to the high-SiO₂ group, with averages of 1.20% (0.57% to 2.90%), 13.46% (8.99% to 17.54%), and 12.65% (7.14% to 18.65%), respectively. These values are similar to those found in the low-SiO₂ samples, emphasizing the enriched and diverse mineral content of the metavolcanic rocks. Notably, the rocks exhibit higher concentrations of MgO and CaO, with averages of 7.11% (3.60% to 13.13%) and 9.10% (2.83% to 13.16%), respectively, which exceed the values observed in both metasedimentary groups. Minor oxides, Na₂O and K₂O show averages of 1.15% and 1.16%

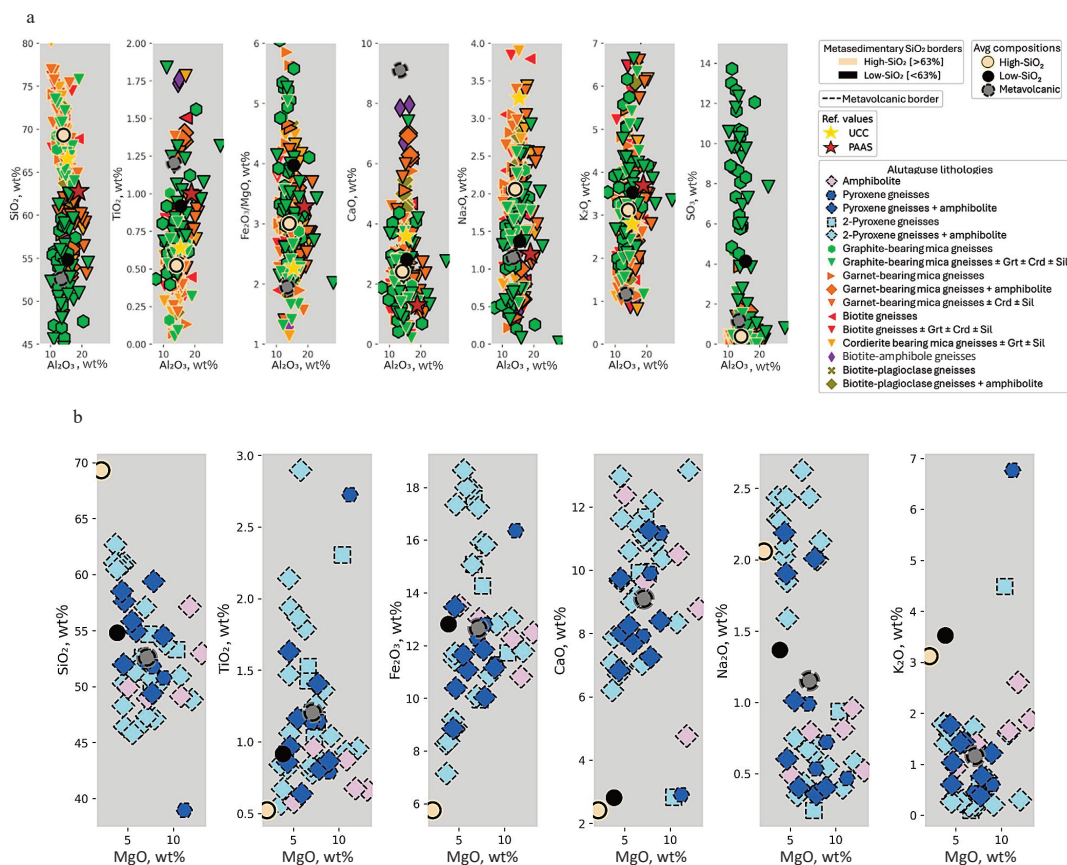


Fig. 2. Harker binary plots of Alutaguse samples: **a** – metasedimentary major oxides vs. Al₂O₃, with metasedimentary samples compared against upper continental crust (UCC) and post-Archean Australian shale (PAAS) references; **b** – metavolcanic major oxides vs. MgO, showing markers per lithology with borders representing sample groups: high-SiO₂ (wheat), low-SiO₂ (black), and metavolcanics (dashed grey). Average values for each group are plotted on both graphs.

(0.24% to 2.63% and 0.10% to 6.77%, respectively), while SO₃ averages at 1.17% (maximum value 7.15%).

5.2. Geochemistry of trace and rare earth elements

Transition elements, such as Cr, Ni, Sc, and V, are commonly found in mafic rocks and resist dispersion from secondary processes (Chen et al. 2014). High-SiO₂ units (Fig. 3d) have average concentrations of 85.21 ppm for Cr, 123.96 ppm for V, 50.66 ppm for Ni, and 14.20 ppm for Sc, with ranges of 16.00–203.00, 8.00–277.00, 2.50–195.00, and 3.70–36.00 ppm, respectively. On the other hand, low-SiO₂ rocks show slightly higher average values, with 101.42 ppm for Cr, 208.05 ppm for V, 120.14 ppm for Ni, and 15.6 ppm for Sc, alongside broader ranges of 14.00–432.00, 12.00–529.00, 8.30–385.00, and 0.60–55.40 ppm, respectively. The low-SiO₂ metasediments, in particular, exhibit a significant valley in Sc concentrations, especially evident in samples from graphite-bearing mica gneisses (Fig. 3e).

Metavolcanic rocks, however, exhibit the greatest variability and the highest average concentrations, especially for Cr and Sc. Cr ranges from 13.00 to 938.00 ppm, with an

average of 149.75 ppm; V ranges from 29.00 to 471.00 ppm, averaging 235.49 ppm; Ni ranges from 16.80 to 210.00 ppm, with an average of 88.44 ppm, and Sc ranges from 1.00 to 52.80 ppm, averaging 26.80 ppm. Metavolcanic rocks have higher Cr and V levels than high-SiO₂ and low-SiO₂ metasedimentary groups. Spider diagrams of PM-normalized trace element data for the Alutaguse metavolcanic rocks suggest significant depletion of Cr and Ni in the examined samples, alongside Sc and V averages that are normalized against the PM (Fig. 3f).

Regarding the large-ion lithophile elements (LILE), such as Ba, Rb, Pb, and Sr, the high-SiO₂ samples show average Ba, Rb, Pb, and Sr concentrations of 804.89, 110.54, 24.44, and 203.98 ppm, respectively. Their ranges extend from 190.00 to 1570.00, 59.40 to 202.00, 7.50 to 65.50, and 35.50 to 462.00 ppm, respectively. The low-SiO₂ group shows a lower average Ba concentration at 520.58 ppm but a higher Pb concentration at 101.59 ppm. The average Rb and Sr concentrations are similar to those of the high-SiO₂ group, at 110.63 ppm and 132.80 ppm, respectively. The ranges for these elements in the low-SiO₂ rocks are wider, especially

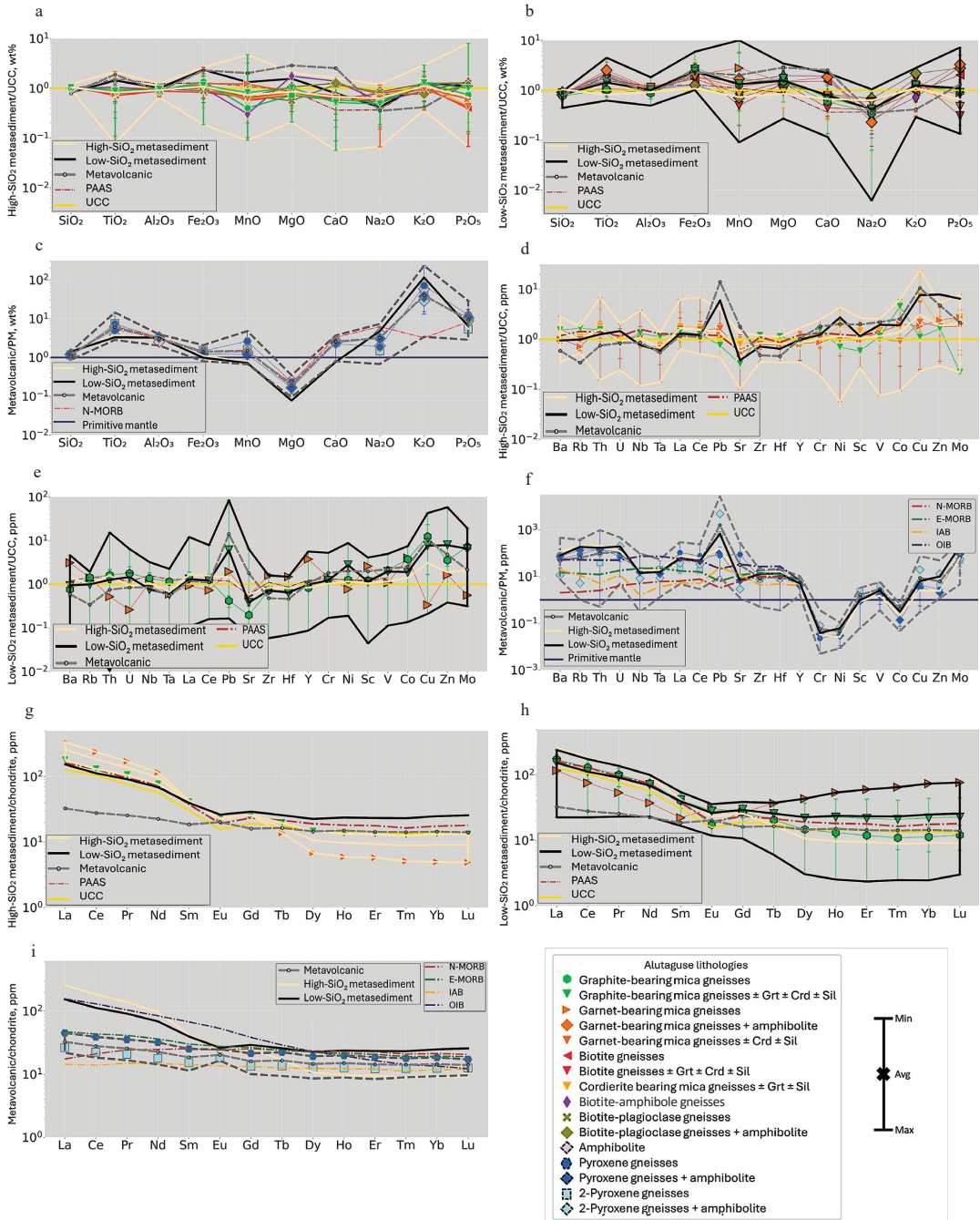


Fig. 3. Normalized spider plots of Alutaguse samples: **a, b** – upper continental crust (UCC)-normalized major element data for high-SiO₂ and low-SiO₂ metasediments (Rudnick and Gao 2003); **c** – primitive mantle (PM)-normalized major element data for metavolcanic samples (Sun and McDonough 1989); **d, e** – UCC-normalized trace element data for high-SiO₂ and low-SiO₂ metasediments; **f** – PM-normalized major element data for metavolcanic samples; chondrite-normalized (Sun and McDonough 1989) REE data for **g** – high-SiO₂ metasediments, **h** – low-SiO₂ metasediments, and **i** – metavolcanic samples. Abbreviations: N-MORB – normal mid-ocean ridge basalt, E-MORB – enriched mid-ocean ridge basalt, IAB – island arc basalt, OIB – oceanic island basalt.

for Ba (30.00–2540.00 ppm) and Sr (24.00–2230.00 ppm). Pb concentrations range up to 1430.00 ppm, particularly in the graphite gneiss samples.

Metavolcanic rocks exhibit the most significant variability in composition, particularly for trace elements, such as Pb and Sr. Compared to both the high-SiO₂ and low-SiO₂ groups, they display wider ranges across all analyzed LILEs. On average, the Ba, Rb, Pb, and Sr concentrations are 323.10, 37.93, 238.24, and 623.45 ppm, respectively. The ranges for these elements are remarkably wide, from 50.00 to 3030.00, 0.60 to 224.00, 1.40 to 4030.00, and 24.60 to 5530.00 ppm, respectively.

High field strength elements (HFSE), such as Nb, Ta, Zr, Hf, Th, and U, exhibit incompatibility during magma crystallization and anatexis processes. This leads to their preferential concentration in felsic rocks (Feng and Kerrich 1990; Han et al. 2019). In high-SiO₂ samples, the average concentrations of Nb, Ta, Zr, Hf, Th, and U are 12.69, 0.67, 159.74, 4.66, 16.79, and 2.45 ppm, respectively. The ranges for these elements are fairly broad, particularly for Nb (1.40–45.30 ppm) and Zr (67.20–227.50 ppm). Low-SiO₂ units show slightly lower average concentrations of Nb (9.20 ppm) and Ta (0.59 ppm) but similar levels of Zr (135.61 ppm) and Hf (3.74 ppm), with lower averages for Th (13.21 ppm) and U (3.93 ppm). The ranges for these elements are even broader than those of the high-SiO₂ group, with Zr ranging from 11.00 to 300.37 ppm, and Th and U exhibiting considerable variability (Th: 0.11–160.00 ppm; U: 0.30–17.20 ppm).

Alutaguse metavolcanic rocks display the widest range but intermediate average values for HFSEs. The average concentrations are 10.23 ppm for Nb, 0.55 ppm for Ta, and 90.75 ppm for Zr. The ranges are 0.20–62.20, 0.05–2.00, and 5.30–202.00 ppm, respectively. Hf concentrations range from 0.10 to 5.90 ppm, Th concentrations from 0.04 to 76.50 ppm, and U concentrations from 0.10 to 9.50 ppm, with respective averages of 2.66, 7.82, and 2.26 ppm. Metavolcanic rocks encompass a broader range of trace element concentrations and generally align more closely with the low-SiO₂ group in terms of average values, except for Ta and U, where they show lower averages (Table S2).

The REE analysis of the 16 new samples from Alutaguse (Table 2) reveals that the high-SiO₂ group (Fig. 3g) exhibits a higher total REE (Σ REE) average of 250.65 ppm, predominantly consisting of light REEs (LREE; La–Gd) at 243.48 ppm and comparatively lower heavy REEs (HREE; Tb–Lu) at 7.18 ppm. Σ REE ranges from 189.73 to 311.58 ppm, LREEs range from 179.85 to 307.10 ppm, and HREEs from 4.47 to 9.88 ppm. In contrast, the low-SiO₂ group (Fig. 3h) shows a lower Σ REE average of 174.94 ppm, but higher HREEs at 16.70 ppm, ranging from 2.00 to 39.87 ppm, and LREEs averaging 158.24 ppm, with a range of 39.93 to 237.86 ppm. Chondrite-normalized REE patterns for both high-SiO₂ (Fig. 3g) and low-SiO₂ samples (Fig. 3h) align closely with UCC and PAAS references, showing LREE enrichment, flat HREE profiles, no Ce anomalies, and similar negative Eu anomalies, whereas Alutaguse metavolcanic samples (Fig. 3i) exhibit null to slightly positive Eu anomalies (Table S2).

Metavolcanic samples exhibit the lowest REE concentrations, with an average Σ REE content of 54.88 ppm (35.29–75.26 ppm), LREEs averaging 44.58 ppm (29.07–61.84 ppm), and HREEs averaging 10.30 ppm (6.22–13.43 ppm). Overall, this pattern suggests that, although high-SiO₂ and low-SiO₂ groups have higher concentrations of REEs, particularly LREEs, the metavolcanic rocks, despite their lower Σ REE content, show a relatively higher proportion of HREEs (Fig. 3i).

6. Discussion

Multiple geological factors influence the geochemical composition of basin lithological units, including chemical weathering and alteration, as explored in both metasedimentary and metavolcanic contexts (Taylor and McLennan 1985; Gao and Wedepohl 1995; Cullers et al. 1997; Gao et al. 1999; Large et al. 2001; Sifeta et al. 2005; Karakaş and Güçtekin 2021). For metasediments, key determinants include: 1) transport and sedimentation sorting (McLennan et al. 1993; Cullers 1994), 2) diagenesis or metamorphism during burial (Fedo et al. 1995, 1996), and 3) sediment origin and deposition environment (Bhatia and Crook 1986; Roser and Korsch 1986). Thus, evaluating the impact of these factors on the chemical profiles of the Alutaguse samples is crucial before drawing petrogenetic conclusions. Subsequent sections delve into the geochemistry of both metasediments and metavolcanics, shedding light on their tectonic and compositional nuances (Bhatia and Crook 1986; Roser and Korsch 1986; McLennan et al. 1993; Cullers 1994; Fedo et al. 1995, 1996; Large et al. 2001; Sifeta et al. 2005; Bailie et al. 2011; Jian et al. 2013; Faisal et al. 2020; de Carvalho Mendes et al. 2021).

6.1. Weathering and alteration indices

During weathering, elements such as Na, K, and LILEs are depleted, while Al₂O₃, TiO₂, REEs, and HFSEs become enriched. Despite this, HFSEs and REEs exhibit limited changes due to their inherent immobility during weathering (McLennan 1989, 1993; Cullers et al. 1997). To determine the mineralogical and chemical changes that occur in the analyzed metasediments as a result of alteration, various chemical alteration indices are used, such as the chemical index of alteration (CIA; Nesbitt and Young 1982), the plagioclase index of alteration (PIA; Fedo et al. 1995), the chemical index of weathering (CIW; Harnois 1988), and the index of compositional variation (ICV; Cox et al. 1995). For metavolcanic samples, indices such as the Hashimoto alteration index (AI; Ishikawa et al. 1976), the chlorite-carbonate-pyrite index (CCPI; Large et al. 2001), the Parker weathering index (WIP; Parker 1970), and the sericitization index (SI; MacLean and Hoy 1991; Karakaş and Güçtekin 2021) were utilized.

High-SiO₂ metasediments in Alutaguse display CIA values ranging from 47.13 to 75.85, PIA values from 45.60 to 90.67, and CIW values between 55.05 and 94.43, suggesting moderate to intense weathering comparable to UCC averages. Conversely, low-SiO₂ metasediments show more intense weathering, with CIA values ranging from 52.03 to 88.38, PIA values from 53.52 to 99.73, and CIW values nearing

PAAS levels, between 60.99 and 99.76. ICV values reflect sediment maturity, with high-SiO₂ averaging 1.50 and low-SiO₂ averaging higher at 2.06. The Th/U ratio further underscores variations in weathering intensity between these groups, with high-SiO₂ averaging 8.68 and low-SiO₂ 4.33.

Metavolcanic samples exhibit hydrothermal alteration influence, demonstrated by relatively high alteration indices and analyses indicating metasomatic changes. These changes are characterized by shifts in mineral compositions and increased alteration intensity, particularly in the basalt/andesite zones of alteration box plots, suggesting profound hydrothermal influences.

The supplementary data further explore the weathering conditions of the analyzed metasedimentary and metavolcanic samples (Fig. S4).

6.2. Metasedimentary rocks

The Alutaguse metasedimentary samples, with a mid-weathered composite index (CIA <80) and chemical immaturity (ICV >1 and SiO₂/Al₂O₃ <6; Table S1), provide reliable data for provenance and tectonic analysis (El-Bialy 2013; Han et al. 2019). Their low LOI averages, below 4% (Table S1), indicate minimal secondary alterations (Han et al. 2019). However, the high LOI values in graphite-bearing mica gneisses warrant cautious interpretation.

6.2.1. Provenance

Al₂O₃/TiO₂ ratios (Fig. S5a), commonly used in rock origin studies, reflect the source rock compositions in sandstones and mudstones. This ratio is consistent between silts, shales, and their sources, as Ti predominantly exists in clay minerals or as ilmenite inclusions, making it a reliable marker for igneous source rocks. Ratios of 3–16, 8–21, and 21–70 indicate mafic, intermediate, and felsic sources, respectively (Hayashi et al. 1997). The Alutaguse high-SiO₂ group averages a ratio of 35.18 (9.90–270.40), while the low-SiO₂ group averages 19.66 (5.17–34.16). Low-SiO₂ metasediments seem to present mafic to intermediate sources closer to PAAS (18.9), contrary to high-SiO₂ samples with higher affinity within UCC (24.06), suggesting felsic sources.

Utilizing the discriminant plot by Roser and Korsch (1988; Fig. 4a), high-SiO₂ metasediments from Alutaguse exhibit a transition from intermediate to felsic origins, whereas low-SiO₂ samples vary from mafic to intermediate, with graphite-bearing mica gneisses predominantly positioned in the mafic zone. The mobility of lithophile elements K and Rb during diagenesis and low-grade metamorphism aids in tracing igneous sources, as demonstrated in the K₂O vs. Rb plots (Fig. S5b), indicating felsic to intermediate origins for these metasediments.

The TiO₂–Ni plot (Fig. 4b) reveals that high-SiO₂ samples predominantly indicate felsic origins, whereas low-SiO₂ samples suggest sedimentary trends from mafic sources. In contrast, transition elements, such as Cr, Sc, Ni, Co, and V, which are typically concentrated in mafic minerals, including pyroxene and olivine, are found to enrich sedimentary rocks derived from mafic igneous sources (Cullers et al. 1997; El-Bialy 2013; Chen et al. 2014). During magma differentiation,

felsic rocks maintain higher concentrations of HFSEs, such as Zr, Hf, Th, and U, displaying stability against diagenetic and metamorphic alterations. This stability makes them effective provenance indicators (Feng and Kerrich 1990; Armstrong-Altrin et al. 2004). Table S2 shows that low-SiO₂ samples are characterized by lower HFSE levels, suggesting a mafic source, whereas high-SiO₂ samples exhibit higher HFSE concentrations, likely obscured by quartz dilution (Chen et al. 2014), reflecting a significant felsic source influence (Table S2).

Elemental ratios, such as La/Sc and Th/Sc, effectively distinguish between mafic and felsic sources (Table 3). High-SiO₂ metasediments align with felsic source signatures, while low-SiO₂ samples show elevated ratios indicative of an intermediate source. Th and Zr, predominantly found in felsic rocks due to their incompatibility in igneous processes, contrast with Sc, which is present in early-forming mafic minerals, such as olivine and pyroxene (McLennan and Taylor 1991). The ratios of Th/Sc and Zr/Sc for Alutaguse metasediments (Fig. S5c) generally trace the igneous differentiation path from andesite to granite, with low-SiO₂ samples showing lower ratios indicative of basaltic influences – a trend supported by the binary plots of La/Th vs. Hf (Fig. S5d) and La/Sc vs. Co/Th (Fig. 4c).

REEs, with their low partition coefficients between water and rock, readily transfer from source rocks to clasts, maintaining stability through weathering, transport, diagenesis, and medium-grade metamorphism, which makes them robust provenance indicators (Chaudhuri and Cullers 1979; McLennan 1989; Gao and Wedepohl 1995; Cullers et al. 1997). While mafic igneous rocks typically exhibit low REE concentrations without significant negative Eu anomalies, felsic rocks display higher REE concentrations with pronounced negative Eu anomalies (Cullers et al. 1997). Chondrite-normalized REE patterns for Alutaguse metasediments (Fig. 3g, h) show substantial LREE to HREE fractionation, indicative of predominantly intermediate to felsic source contributions (Chen et al. 2014). These patterns, featuring marked negative Eu anomalies, suggest extensive feldspar fractionation within their parent rocks (Han et al. 2019) and are consistent with a composition rich in quartz and chlorite (McLennan 1993). The negative Eu anomaly typically aligns with differentiated silicic sources, similar to granitic origins (Condie 1993; McLennan 1993; Gao and Wedepohl 1995; Gu et al. 2002). Eu/Eu* values, indicating differentiation, are slightly higher in high-SiO₂ samples, averaging 0.52 (ranging from 0.27 to 0.76), compared to low-SiO₂ samples at 0.43 (ranging from 0.22 to 0.77), as shown in Table S2.

6.2.2. Sorting, recycling and maturation

Sorting during sedimentary transport significantly influences the mineralogical and chemical characteristics of sediments. Textural maturity, evaluated through grain sizes, morphologies, and mineralogical and geochemical profiles, offers insights into the sorting process (McLennan et al. 1993). Lower SiO₂/Al₂O₃ ratios are indicative of minimal sedimentary sorting (Taylor and McLennan 1985; Rudnick and Gao 2003; Chen et al. 2014). High-SiO₂ specimens exhibit

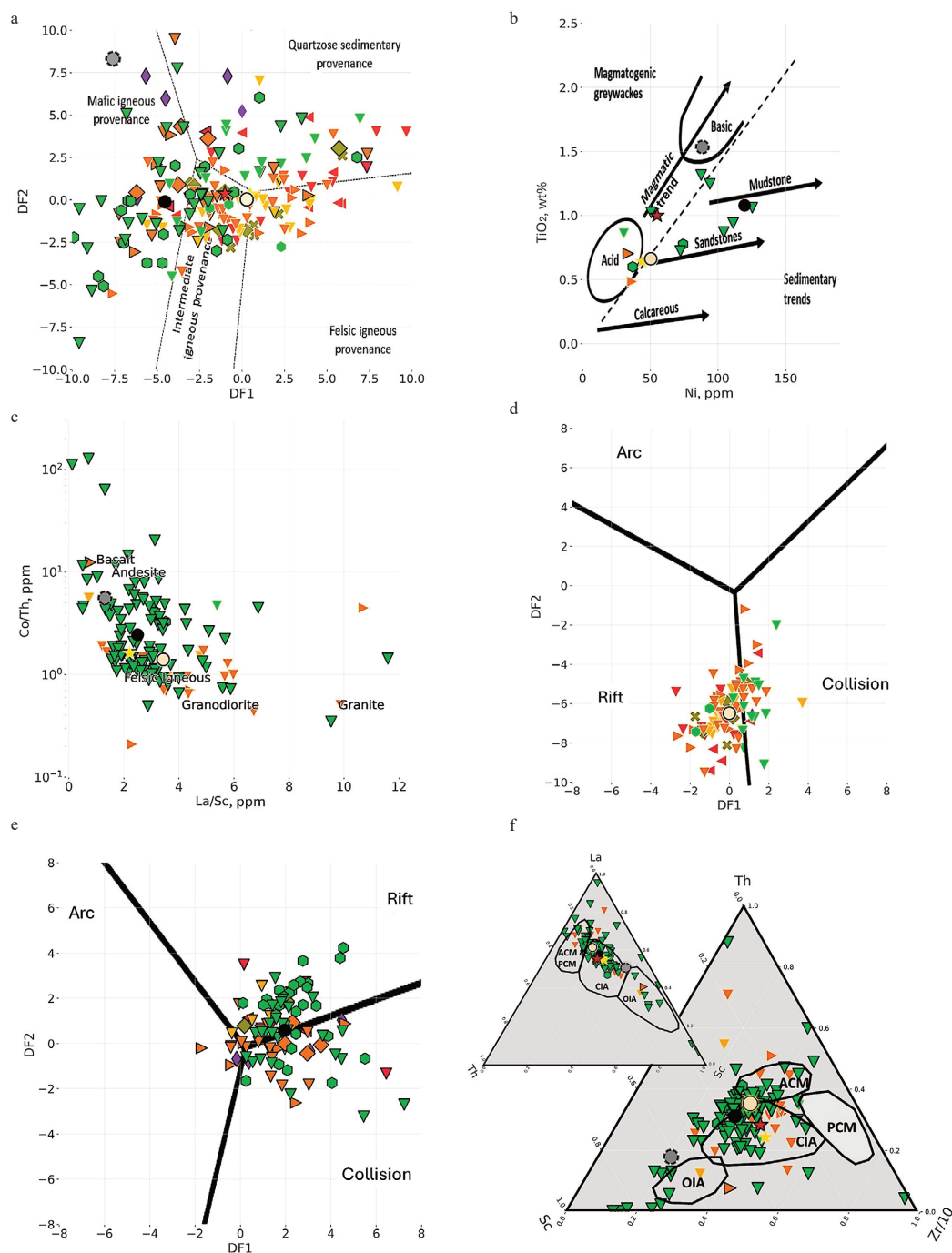


Fig. 4. Source and tectonic setting plots of Alutaguse metasedimentary samples: **a** – discriminant function plot of provenance signatures using major elements, with the accompanying equations (after Roser and Korsch 1988): $\text{DF1} = 30.638 \text{ TiO}_2/\text{Al}_2\text{O}_3 - 12.541 \text{ Fe}_2\text{O}_3/\text{Al}_2\text{O}_3 + 7.329 \text{ MgO}/\text{Al}_2\text{O}_3 + 12.031 \text{ Na}_2\text{O}/\text{Al}_2\text{O}_3 + 35.402 \text{ K}_2\text{O}/\text{Al}_2\text{O}_3 - 6.382$ vs. $\text{DF2} = 36.500 \text{ TiO}_2/\text{Al}_2\text{O}_3 - 10.879 \text{ Fe}_2\text{O}_3/\text{Al}_2\text{O}_3 + 30.875 \text{ MgO}/\text{Al}_2\text{O}_3 - 5.404 \text{ Na}_2\text{O}/\text{Al}_2\text{O}_3 + 11.112 \text{ K}_2\text{O}/\text{Al}_2\text{O}_3 - 3.89$; **b** – TiO_2 vs. Ni plot (after Floyd et al. 1989); **c** – Co/Th vs. La/Sc plot (after Gu et al. 2002); **d**, **e** – tectonic classification of high- SiO_2 and low- SiO_2 metasediments (see Verma and Armstrong-Altrin 2013 for further details on DF1 and DF2 formulas for both groups); **f** – ternary plots of La–Th–Sc (inset) and Th–Sc–Zr/TiO2 (back) (after Bhatia and Crook 1986). Abbreviations: ACM – active continental margin, PCM – passive continental margin, CIA – continental island arc, OIA – oceanic island arc. Sample marker legend as in Figure 2a.

Table 3. Range of elemental ratios for the Alutaguse metasedimentary units compared to the ratios of the upper continental crust (after Rudnick and Gao 2003) and sediments derived from felsic and mafic rocks (after Armstrong-Altrin et al. 2004). Values for Alutaguse are organized as “min–max; avg”. High-SiO₂ graphite-bearing mica gneisses ± Grt ± Crd ± Sil lithology represent only one sample (Table S2)

Sample type	La/Sc	La/Co	Th/Sc	Th/Co	Cr/Th
Range of sediments from felsic sources	2.50–16.3	1.80–13.8	0.84–20.05	0.67–19.4	4.00–15.00
Range of sediments from mafic sources	0.43–0.86	0.14–0.38	0.05–0.22	0.04–1.40	25.00–500.00
UCC	2.20	1.79	0.77	0.61	7.53
PAAS	2.39		0.91		7.53
Garnet-bearing mica gneisses ± Crd ± Sil	1.20–9.87; 4.28	1.55–13.81; 4.11	0.41–3.74; 1.33	0.53–2.29; 1.10	1.59–16.99; 6.86
Garnet-bearing mica gneisses	2.03–10.71; 4.86	0.76–10.98; 4.48	0.69–3.15; 1.56	0.22–4.78; 1.75	2.58–8.35; 4.86
Cordierite-bearing mica gneisses ± Grt ± Sil	0.72–4.77; 3.12	0.38–3.54; 2.44	0.22–1.95; 1.09	0.18–1.09; 0.81	1.76–6.67; 5.28
Graphite-bearing mica gneisses ± Grt ± Crd ± Sil	5.38	0.55	2.06	0.21	7.94
Total	0.72–10.71; 4.19	0.55–13.81; 3.71	0.22–3.74; 1.35	0.18–4.78; 1.11	1.59–16.99; 6.29
Low-SiO ₂ Graphite-bearing mica gneisses ± Grt ± Crd ± Sil	0.07–77.87; 4.36	0.03–21.75; 2.02	0.00–33.25; 1.43	0.00–9.64; 0.67	0.30–945.45; 32.18
Graphite-bearing mica gneisses	1.46–3.52; 2.49	0.47–0.77; 0.62	0.67–1.39; 1.03	0.22–0.30; 0.26	5.48–7.70; 6.59
Garnet-bearing mica gneisses	0.79	0.41	0.15	0.08	21.69
Total	0.07–77.87; 4.29	0.03–21.75; 1.98	0.00–33.25; 1.41	0.00–9.64; 0.65	0.30–945.45; 31.60

an average SiO₂/Al₂O₃ ratio of 5.02 wt%, suggesting greater sorting, compared to 3.65 wt% in their low-SiO₂ counterparts.

The analyzed metasediments display K₂O/Na₂O ratios above 1 wt%, indicating chemical immaturity (El-Bialy 2013), a feature more pronounced in low-SiO₂ samples, showing the highest values (Fig. S5e). Conversely, high-SiO₂ samples exhibit greater maturity, as indicated by higher Al₂O₃/TiO₂ ranges (Hayashi et al. 1997; Fig. S5a).

Provenance discrimination plots by Roser and Korsch (1988) reveal that high-SiO₂ samples fall within the quartzose sedimentary provenance field, suggesting mature, recycled

sediments with polycyclic quartzose detritus (Fig. 4a). In contrast, low-SiO₂ samples are derived from primary mafic-magmatic sources. Transport and recycling processes often result in higher CIW values. Sediments from extensive provenance areas display higher CIW/CIA ratios, indicative of longer transport distances and suggesting broader provenance and considerable travel before deposition (Gao et al. 1999; El-Bialy 2013). For the Alutaguse samples, high-SiO₂ specimens have an average CIW/CIA value of 1.17 mol, whereas low-SiO₂ samples average 1.20 mol, with all samples exhibiting CIW/CIA ratios greater than 1 mol, suggesting a long-distance provenance.

Sedimentary materials can be categorized by maturity and recycling into psammitic and pelitic types. Psammitic sediments are generally closer to their source, as indicated by lower maturity and coarser grains. In contrast, pelitic sediments, characterized by finer grains, suggest higher maturity due to extended transport and erosion. Using the 100 × TiO₂/Zr ratio (Garcia et al. 1994), values below 0.33 wt%/ppm are typically psammitic, whereas those above 0.33 wt%/ppm are pelitic. The SiO₂/Al₂O₃ ratio at 4.35 wt% is considered the threshold between these categories (Garcia et al. 1994; El-Bialy 2013). For the high-SiO₂ samples, the 100 × TiO₂/Zr ratio averages 0.35 wt%/ppm, with a SiO₂/Al₂O₃ average of 5.02 wt%, and ranges from 0.32 to 0.38 wt% and 3.49 to 7.80 wt%, respectively, clearly classifying them as psammitic. Conversely, low-SiO₂ samples display more varied characteristics, with averages of 0.67 wt%/ppm for 100 × TiO₂/Zr and 3.65 wt% for SiO₂/Al₂O₃, and ranges from 0.23 to 1.85 wt%/ppm and 1.54 to 5.16 wt%, respectively. These results indicate a mix of pelitic to psammitic properties, depending on the specific metrics considered, suggesting different levels of transport and sedimentary recycling.

Trace elements Zr, Th, and Sc are essential in assessing clastic rock provenance and recycling. The Th/Sc and Zr/Sc ratios, which indicate chemical differentiation and sediment recycling, respectively, show significant variability (McLennan et al. 1990). High-silica samples exhibit average Th/Sc and Zr/Sc ratios of 1.33 and 14.11 ppm, respectively, with ranges of 0.22–3.74 ppm for Th/Sc and 4.66–28.69 ppm for Zr/Sc. In contrast, low-silica samples display an average Th/Sc ratio of 1.40 and a Zr/Sc ratio of 16.97 ppm, but with extreme variability in their ranges (Th/Sc: 0.001–33.25 ppm; Zr/Sc: 1.22–423.33 ppm). The Th/Sc vs. Zr/Sc plot (Fig. S5c) reveals that both high-SiO₂ and low-SiO₂ metasedimentary averages align with UCC and PASS references, with no significant recycling.

The accumulation of heavy minerals, such as zircon, monazite, and/or allanite, during sediment transport contributes to the rise in normalized (Gd/Yb)_{cn} values, which typically range between 1.0 and 2.0 ppm for post-Archean sediments and most upper crust igneous rocks (McLennan and Taylor 1991). Notably, Alutaguse high-SiO₂ samples exhibit an average (Gd/Yb)_{cn} value of 4.58 ppm, while low-SiO₂ samples average 1.73 ppm, suggesting minimal heavy mineral fractionation and sediment recycling, with higher significance in high-SiO₂ samples. The significant negative Sr anomaly

in high-SiO₂ samples (Fig. 3d), compared to the low-SiO₂ counterpart (Fig. 3e), further underscores the limited recycled nature of these environments.

These observations underline the Alutaguse high-SiO₂ metasediments as highly sorted, reworked, and mature, in contrast to the geochemically immature low-SiO₂ samples.

6.2.3. Tectonic affinities

The tectonic environment of depositional basins and their geochemical characteristics are complexly related. The remaining sediments of derived sources can still provide valuable insights. However, studying tectonic settings requires caution, as sediments can cross boundaries (Chen et al. 2014). Several plots, initially designed for Phanerozoic clastic sediments, have been extended to Precambrian rock studies (El-Bialy 2013; Verma and Armstrong-Altrin 2013). However, relying solely on these plots can be misleading, as the elements used may be affected by processes such as sorting and mineral concentration (McLennan and Taylor 1991; El-Bialy 2013; Saccani 2015).

The elemental composition of sandstones, such as TiO₂, Al₂O₃, Fe₂O₃, and MgO concentrations, as well as Al₂O₃/SiO₂ ratios, varies across tectonic settings, transitioning from oceanic island arcs (OIA) to continental arcs, and further to active and passive continental margins (ACM, PCM; Bhatia 1983). Despite a wide range of values due to chemical mobility during weathering and diagenesis, the high TiO₂, Al₂O₃, Fe₂O₃, and MgO concentrations indicate that these sediments did not form in passive regimes (Bhatia and Crook 1986; McLennan and Taylor 1991).

Clastic sediments can be distinguished based on their K₂O/Na₂O and SiO₂ values, as shown by Roser and Korsch (1986; Fig. S5e), enabling differentiation between the PM, ACM, and ARC. Most Alutaguse metasediments align predominantly with the ACM. However, a few low-SiO₂ samples trend toward the ARC field. Some high-SiO₂ samples also overlap with the PM, but as Na and K are highly mobile, relying solely on Na₂O and K₂O for tectonic setting differentiation requires caution (El-Bialy 2013).

Verma and Armstrong-Altrin's (2013) discriminant-function-based multidimensional diagrams effectively distinguish between island/continental arcs, continental rifts, and collision settings, particularly for high-silica (SiO₂ > 63 wt%) and low-silica rocks (SiO₂ < 63 wt%). The high-silica metasediments (Fig. 4d) predominantly align with the continental rift zone, a trend that is also consistent in the low-silica samples (Fig. 4e). Nevertheless, some low-SiO₂ graphite-bearing gneisses plot within the collision zone, possibly due to alteration patterns (i.e., LOI; Table S1).

Trace elements, such as Th, Sc, La, and Zr, are stable in depositional environments and serve as effective indicators for identifying source rocks (Bhatia and Crook 1986; Roser and Korsch 1986; McLennan et al. 1993). Bhatia and Crook (1986) employed these elements in Th–La–Sc and Sc–Th–Zr/10 triangular plots (Fig. 4f) to distinguish between four tectonic settings. These diagrams show that most Alutaguse metasediments align with the CIA zone, corroborated by their positioning relative to UCC and PAAS references. Notably,

continental arcs and ACMs are similar, shaped by convergent plate dynamics, orogenic activity, and the evolution of subduction complexes (El-Bialy 2013).

The Alutaguse metasedimentary units display distinct geochemical signatures, such as a pronounced negative Eu anomaly, reduced Nb-Ta levels, dominant LREE patterns, and limited HREE fractionation (Fig. 3), suggesting a continental arc origin. However, as McLennan et al. (1990) note, specific geochemical markers do not conclusively determine tectonic settings, as the continental crust often exhibits arc-like characteristics. Nevertheless, the combined geochemical evidence supports a continental arc scenario for these units, with a potential back-arc context that is consistent with established tectonic models (All et al. 2004; Kirs et al. 2009; Bogdanova et al. 2015; Soesoo et al. 2020).

6.3. Metavolcanic rocks

6.3.1. The effects of shallow-level open-system processes

Shallow processes, such as crustal contamination, fractional crystallization, and post-magmatic alteration, can change the composition of mafic magma. Therefore, it is essential to consider these changes before using mafic igneous rocks to identify magma sources. For instance, the influence of crustal contamination on ascending magma is crucial, as it can cause chemical and isotopic variations in mafic magma. Typically, the continental crust exhibits enriched isotopic compositions and high SiO₂ content (Rudnick and Gao 2003).

The LOI values in the Alutaguse metavolcanic units range from 0.22 to 8.49 wt% (average 1.80 wt%; Table S1), indicating possible alterations caused by seawater or metamorphic fluids. This is further supported by carbonation patterns (Ma et al. 2021; Fig. S4d, e). Nevertheless, Figure S4e shows that the samples lie within or around the least-altered domain. The LOI values for 2-pyroxene gneiss samples are high (0.22–8.49 wt%; average 2.50 wt%). However, LOI values below 5 wt% generally do not show significant correlations with the abundances of fluid-mobile elements, such as Rb, Ba, Sr, U, Pb, K₂O, and particularly Zr and Nb, in mafic-ultramafic rocks (Pearce and Norry 1979; Saccani et al. 2018; Ma et al. 2021).

Bivariant Zr plots provide further insights into element mobility (Table S2). The effects of shallow-level open-system processes on the compositions of the studied metavolcanic rocks seem insignificant, providing valuable information on the composition of their magmatic origin and processes (Karakaş and Güçtekin 2021; Ma et al. 2021).

6.3.2. Crystallization and partial melting

The Alutaguse metavolcanic samples show tholeiitic tendencies (Fig. S3d). The predominantly subalkaline Zr underscores this distinction vs. Nb/Y ratios for most samples (Fig. S6a). These samples are also identified as metaluminous (Fig. 5a).

Major oxide relationships suggest fractional crystallization, influenced by ferromagnesian minerals (Yang et al. 2014; Ma et al. 2021; Fig. 2b). Metavolcanic units display MgO contents that are positively correlated with SiO₂ but inversely with Fe₂O₃ (Fig. 2b). Additionally, the Ni depletion

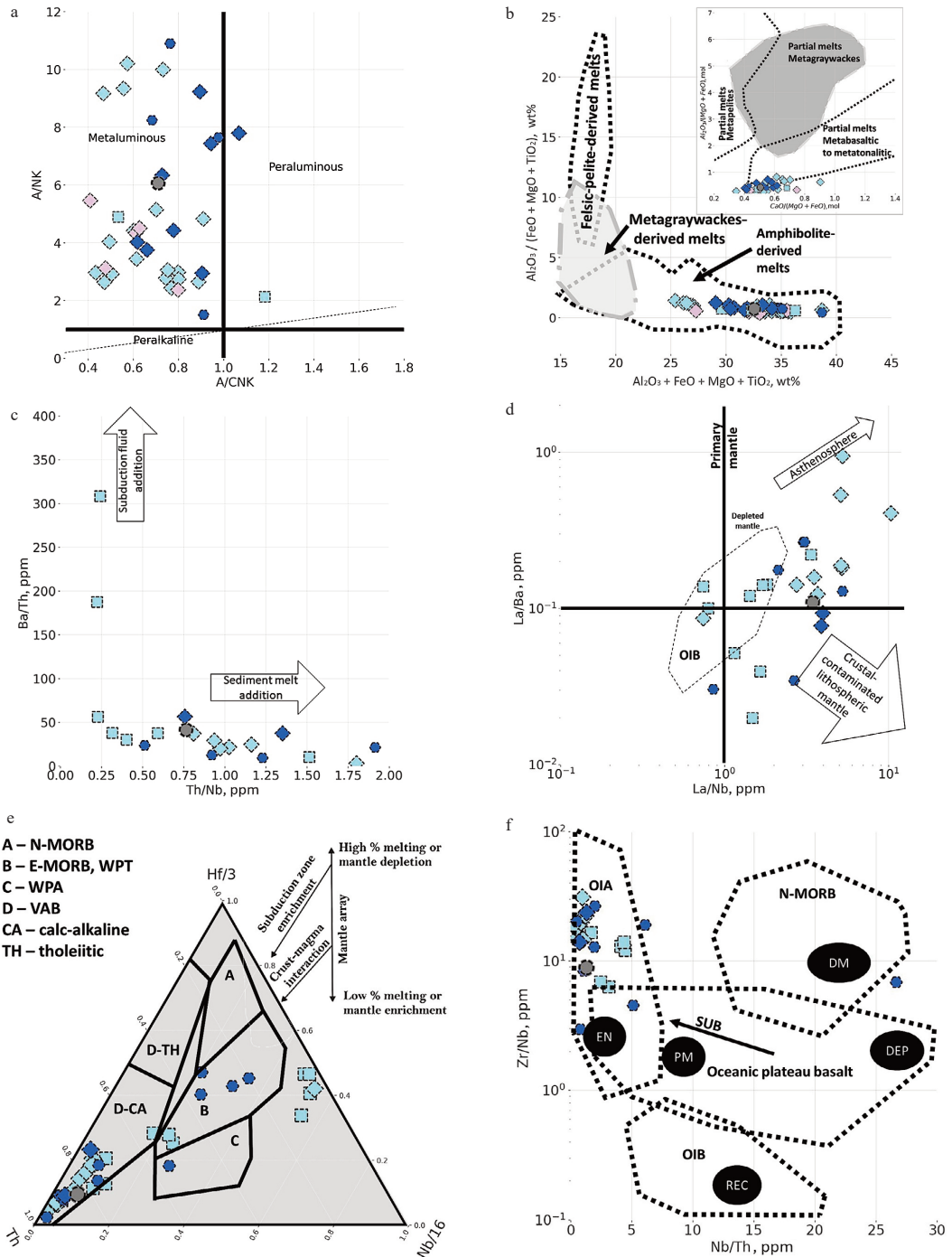


Fig. 5. Source and tectonic setting plots of Alutaguse metavolcanic samples: **a** – A/CNK (molar ratio of $Al_2O_3 / (CaO + Na_2O + K_2O)$) vs. A/NK (molar ratio of $Al_2O_3 / (Na_2O + K_2O)$) plot (after Shand 1943); **b** – $Al_2O_3 / (FeO + MgO + TiO_2)$ vs. $Al_2O_3 + FeO + MgO + TiO_2$ plot (after Patiño Douce 1999), with an inset of Altherr et al. (2000) discrimination diagram $Al / (Mg + Fe)$ vs. $Ca / (Mg/Fe)$ (mol); **c** – Ba/Th vs. Th/Nb plot (after Saccani et al. 2018); **d** – La/Ba vs. La/Nb plot (after Hart 1988); **e** – Hf/3–Th–Nb/16 triangular plot (after Wood 1980); **f** – Nb/Th vs. Zr/Nb plot (after Condie 2005). Sample marker legend as shown in Figure 2b. Abbreviations: OIB – ocean island basalt, N-MORB – normal mid-ocean ridge basalt, E-MORB – enriched mid-ocean ridge basalt, WPT – within-plate tholeiite, WPA – within-plate alkaline basalt, VAB – volcanic arc-basalt, OIA – oceanic island arc, SUB – subduction zone, EN – enriched mantle, PM – primitive mantle, DM – depleted mantle, DEP – deep mantle, REC – recycled material.

(Fig. 3f) points to olivine fractional crystallization (Ma et al. 2021). Stable $\text{CaO}/\text{Al}_2\text{O}_3$ ratios, despite fluctuating MgO levels, indicate minimal clinopyroxene fractionation (Ma et al. 2021; Fig. 2b). The negative correlation of MgO with Al_2O_3 , CaO , and Na_2O (Fig. 3c), along with minimal to slightly positive Eu anomalies (Eu/Eu^* : 0.34–0.55; average 0.44; Fig. 3i), implies limited plagioclase fractionation (Floyd et al. 1989; Faisal et al. 2020). Sr enrichment is due to its rejection by most magmatic minerals, such as pyroxenes, instead favoring its incorporation into plagioclases (Faisal et al. 2020; Ma et al. 2021). The negative correlation between TiO_2 and MgO contents (Fig. 3c) suggests that the fractional crystallization of Ti-bearing minerals is negligible (Ma et al. 2021).

Experimental studies by Patiño Douce (1999) reveal compositional variations in melts derived from pelites, graywackes, and amphibolites, marked by differing Al concentrations. These variations arise from the dehydration melting of source rocks with distinct mineral compositions, which is particularly relevant in the context of crustal contamination. Pelite-derived melts, for instance, exhibit higher aluminum concentrations compared to those derived from psammites, and significantly more than those from amphibolites. This trend is observable in the Alutaguse metavolcanic samples, suggesting a strong pelitic influence in the source material (Fig. 5b). Furthermore, the discrimination diagram by Altherr et al. (2000), which plots $\text{Al} / (\text{Mg} + \text{Fe})$ vs. $\text{Ca} / (\text{Mg}/\text{Fe})$ (mol), supports the interpretation that the Alutaguse metavolcanics are primarily derived from metabasaltic to metatonalitic partial melts, indicating that the Alutaguse mantle-derived magmas share a provenance similar to amphibolite-basaltic crustal material or may have interacted with crustal material of this composition, rather than originating from metagraywackes or felsic (meta)pelites (Fig. 5b inset).

Elevated Nb and reduced Zr levels indicate mantle sources that vary from depleted to transitional (Pearce et al. 1996). Specifically, Nb levels below 10 ppm and Zr levels below 200 ppm characterize depleted mantle, whereas higher concentrations suggest transitional to enriched sources. In Alutaguse, the average concentrations of Nb and Zr are 10.23 and 90.75 ppm, respectively, positioning the region's samples primarily on the cusp of the transitional and depleted mantle domains, with a tendency toward the latter, as illustrated in the Nb vs. Zr plot (Fig. S6a).

The Nb/Y and Zr/Y ratios, known for their reliability in tracing melting and crystallization processes, help in identifying magma sources using the Fitton et al. (1997) δNb formula, defined as $\delta\text{Nb} = \log(\text{Nb}/\text{Y}) + 1.74 - (1.92(\log(\text{Zr}/\text{Y}))$. A δNb value greater than 0 indicates an enriched mantle, while a value less than 0 suggests a depleted source. Alutaguse metavolcanics exhibit δNb values ranging from –0.63 to 1.44, with an average of 0.25, highlighting significant heterogeneity. These values generally indicate an enriched mantle source, except for the pyroxene gneisses + amphibolites, which trend toward a depleted mantle (Table S2). La/Yb vs. Zr/Nb and La/Sm vs. Sm/Yb plots effectively illustrate melting trajectories from spinel and garnet-lherzolite, demonstrating mantle depletion and enrichment trends

(Yang et al. 2014). The La/Yb vs. Zr/Nb plot (Fig. S6b) predominantly favors garnet-lherzolite melting patterns, while the La/Sm vs. Sm/Yb plot (Fig. S6c) shows samples closely following the spinel-lherzolite trend, approaching the PM reference. These patterns suggest that the parental magmas of the Alutaguse mafic metavolcanic rocks likely originated from high degrees of partial melting, approximately 30%, with a slight depletion trend. However, only three samples were analyzed using these relations (Table 2).

6.3.3. The nature of magma sources

The composition of mafic magmas is influenced by fractional crystallization and rock assimilation, which shape their elemental composition. However, trace element profiles, especially those of incompatible elements, are more reflective of the composition and melting degree of the mantle source. These profiles indicate distinct source characteristics specific to tectono-magmatic environments (Sifeta et al. 2005; Faisal et al. 2020; Ma et al. 2021). Determining magma sources and the degree of partial melting is pivotal and can be accomplished through the analysis of light REE and HFSE content and their ratios. Even amidst mineral accumulation or fractionation during mafic magmas' crystallization, LILE enrichment, HFSE depletion, and ratios of incompatible trace elements, such as Nb/Zr, Th/Zr, Ba/Th, U/Th, and Th/Nb, remain relatively consistent. These ratios are mainly influenced by the fractional crystallization of olivine, clinopyroxene, and plagioclase. Therefore, these elements arguably mirror the source's elemental ratios, even with moderate fractionation (Saccani et al. 2018; Wan et al. 2019; Ma et al. 2021).

The Alutaguse samples show elevated Th/Nb, Ba/Th (Fig. 5c), and U/Th ratios (Fig. S6d), as well as increased Th/Zr ratios (Fig. S6a), suggesting that their composition is primarily shaped by sedimentary melt processes rather than influence from subducted oceanic crust. The La/Ba vs. La/Nb plot (Fig. 5d) further supports the interpretation that the Alutaguse metavolcanics likely originate from an asthenospheric mantle source. This implies sedimentary melting activities (Fig. 5c) and indicates a mantle composition distinct from crust-contaminated sources, such as those found in the subduction-influenced southern Svecofennian domains of southern Finland (Lahtinen 2000; Kähkönen 2005; Kukkonen and Lauri 2009; Nironen 2017; Kara et al. 2021).

6.3.4. Tectonic setting implications

Debates on the tectonic origins of the analyzed metavolcanics have linked them to island arc collisions and rift mechanisms (Lahtinen 2000; Kirs et al. 2009; Bogdanova et al. 2015; Soesoo et al. 2020). HFSEs and HREEs, due to their stability, are critical for identifying the tectonic settings of extrusive rocks (Pearce et al. 1996; Sifeta et al. 2005; Saccani et al. 2018). Unlike typical mid-ocean ridge basalt (MORB) from asthenospheric mantle melting, which exhibits lower LREEs and LILEs, Alutaguse metavolcanics display an arc-like signature with notable HFSE depletion, suggesting subduction influences (Faisal et al. 2020; Ma et al. 2021). REE analysis shows that Alutaguse chondrite-normalized metavolcanics range from island arc basalts (IAB) to enriched mid-ocean

ridge basalt (E-MORB), significantly differing in LREE content from associated metasediments (Fig. 3i).

Geochemical plots, such as TiO_2 – $10(\text{MnO})$ – $10(\text{P}_2\text{O}_5)$ (Fig. S7a) and $\text{Y}/15$ – $\text{La}/10$ – $\text{Nb}/8$ (Fig. S7b), confirm that most samples from Alutaguse primarily fall within the island arc tholeiites (IAT) domain and the calc-alkaline basalts zone, indicative of a compressional arc setting. Furthermore, the $\text{Hf}/3$ – Th – $\text{Nb}/16$ plot (Fig. 5e) positions these samples in the calc-alkaline volcanic arc-basalt (VAB) zone, suggesting a trend of crustal-magma interactions. Additional plots, such as $\log(\text{Nb}/\text{Th})$ vs. $\log(\text{Y}/\text{La})$ (Fig. S7c) and Nb/Th vs. Zr/Nb (Fig. S6a), show the Alutaguse samples aligning mainly with the arc domain. These positions notably reflect influences from the IAB and an arc-enriched component.

The Zr/Y vs. Zr plot (Fig. S6a), which differentiates between continental and oceanic basalts with a threshold value of 3, shows that while the samples mostly occupy the continental domain, they extend into the oceanic domain, particularly evident in the 2-pyroxene gneiss samples with high Y values (Table S2). Discrimination plots by Saccani (2015) reveal back-arc basin basalt (BABB) affinities, aligning with the assimilation-fractional crystallization (AFC) trend and spanning the oceanic subduction setting domain + rifted margin (Figs S6e, S7d).

6.4. Geodynamic implications

Although data are sparse, compelling evidence details a rich geological and tectonic history of the southern Fennoscandian Shield. This includes a major accretionary and collisional event during the late Svecofennian era, around 1.9–1.8 Ga, that amalgamated Sarmatia, Fennoscandia, and Volgo–Uralia into a single landmass (Bogdanova et al. 2006; Bogdanova et al. 2015; Nironen 2017; Pesonen et al. 2021; Lahtinen et al. 2022). Paleomagnetic studies and geological data have revealed a 2000 km-wide oceanic basin existing around 1.9 Ga, featuring island arcs and microcontinents, such as the Uusimaa and Bergslagen belts, which underscore complex subduction and collisional dynamics persisting until about 1.7 Ga (Korja et al. 1993; Claesson et al. 2001; Korja et al. 2003; Skridlaite et al. 2003; Lahtinen et al. 2005). In northern Estonia's Tallinn zone, metapelitic and metavolcanic gneisses dating between 2.13 and 1.85 Ga, along with the Uusimaa belt's 1.92–1.91 Ga felsic metavolcanic rocks, indicate subduction-related Paleoproterozoic crustal growth (Petersell and Levchenkov 1994; Puura et al. 2004; Bogdanova et al. 2015; Soesoo et al. 2020). These features, which resonate with the Bergslagen region's similar geological formations, suggest a unified geological structure across these belts (All et al. 2004; Puura et al. 2004; Kirs et al. 2009; Stephens and Weihed 2020). Moreover, West Estonian metapelitic rocks display detrital zircon ages ranging from 1.97 to 1.90 Ga, indicating a maximum deposition age of 1.90 Ga. This suggests proximity to the Bergslagen microcontinent and adjacent Svecofennian arcs as potential primary sources for sedimentation in the West Estonian basin (Kähkönen 2005; Bogdanova et al. 2015).

During 1.92–1.90 Ga, the concurrent formation of the Uusimaa and Tampere belts is hypothesized to have origin-

ated from distinct subduction-arc systems, likely due to a relatively shorter slab subducting beneath Bergslagen, as indicated by the older ages of the arc belts (Kukkonen and Lauri 2009; Bogdanova et al. 2015; Nironen 2017; Kara 2021). This theory supports the model of a subduction system forming the Uusimaa belt, as suggested by Kukkonen and Lauri (2009), and maintains the proposed timeline for the Tallinn–Uusimaa belt formations around 1.92–1.90 Ga (Fig. 6a).

From 1.91 to 1.80 Ga, persistent lithospheric convergence hindered the gravitational collapse of the over-thickened crust. Evidence of this includes a 50 km-thick crust in central Finland and Estonia (Korja et al. 1993, 2003; Solano-Acosta et al. 2023). This era saw Sarmatia, Fennoscandia, and Volgo–Uralia merging, forming a vast landmass via accretionary and collisional dynamics. Prevalent arc-type magmatism around 1.90 to 1.89 Ga offers insights into the region's geodynamic evolution (Kähkönen 2005; Kara et al. 2021). The Bergslagen region reveals back-arc rifting events and associated granitoid magmatism and sedimentation until approximately 1.85 Ga. Geological correlations between the Swedish Skellefte district and the Tampere and Pirkanmaa belts highlight intricate magmatic and tectonic interactions (Allen et al. 1996; Beunk and Kuipers 2012; Bogdanova et al. 2015; Nironen 2017; Stephens and Weihed 2020; Kara et al. 2021).

Between 1.90 and 1.88 Ga, the Tallinn–Uusimaa belt(s) experienced sedimentation intermixed with volcanic activity, suggesting a shared origin with Alutaguse metasediments due to similar CaO and MnO enrichments (Kivisilla et al. 1999; Rasilainen et al. 2007; Kirs et al. 2009; Kukkonen and Lauri 2009; Bogdanova et al. 2015; Nironen 2017; Lahtinen et al. 2022; Fig. 6). The Alutaguse metavolcanics, influenced by asthenospheric magmatism (Fig. 5), contrast the southern Svecofennian Finnish domains, which show significant crustal interactions from subduction (Kukkonen and Lauri 2009; Kara 2021; Lahtinen et al. 2022). By 1.89 Ga, the Alutaguse zone was characterized as a back-arc area (Kirs et al. 2009; Bogdanova et al. 2015; Soesoo et al. 2020; Fig. 6b), followed by sedimentation in a back-arc rift system (Fig. 4). The 1.89 Ga Jõhvi units displayed a magmatic-magnetite peak, indicative of mantle uplift during Alutaguse rifting (Bogdanova et al. 2015; Nirgi and Soesoo 2021).

Figure 6c depicts the progressive collision of Bergslagen and its subsequent formations between 1.89–1.87 Ga (Bogdanova et al. 2015; Nironen 2017). In the South Estonian domain (SEG), during this same period, the crystallization of garnet–orthopyroxene granodiorite and subsequent deformation and migmatization by 1.86 Ga suggest an extensional back-arc scenario, likely initiating the Middle Estonian fault zone (MEFZ) and regional granulite-facies metamorphism (Kirs et al. 2009; Bogdanova et al. 2015; Soesoo et al. 2020). This phase aligns with broader regional tectonic processes, underscoring the dynamic geological evolution of these domains during the Proterozoic.

Between 1.87 and 1.86 Ga, the Bergslagen microcontinent collided with the Svecofennian arc, leading to the formation of calc-alkaline granitoids (Kähkönen 2005; Nironen 2017; Mikkola et al. 2018; Kara et al. 2021). Subsequent intra-orogenic sedimentary basins developed in southern Finland

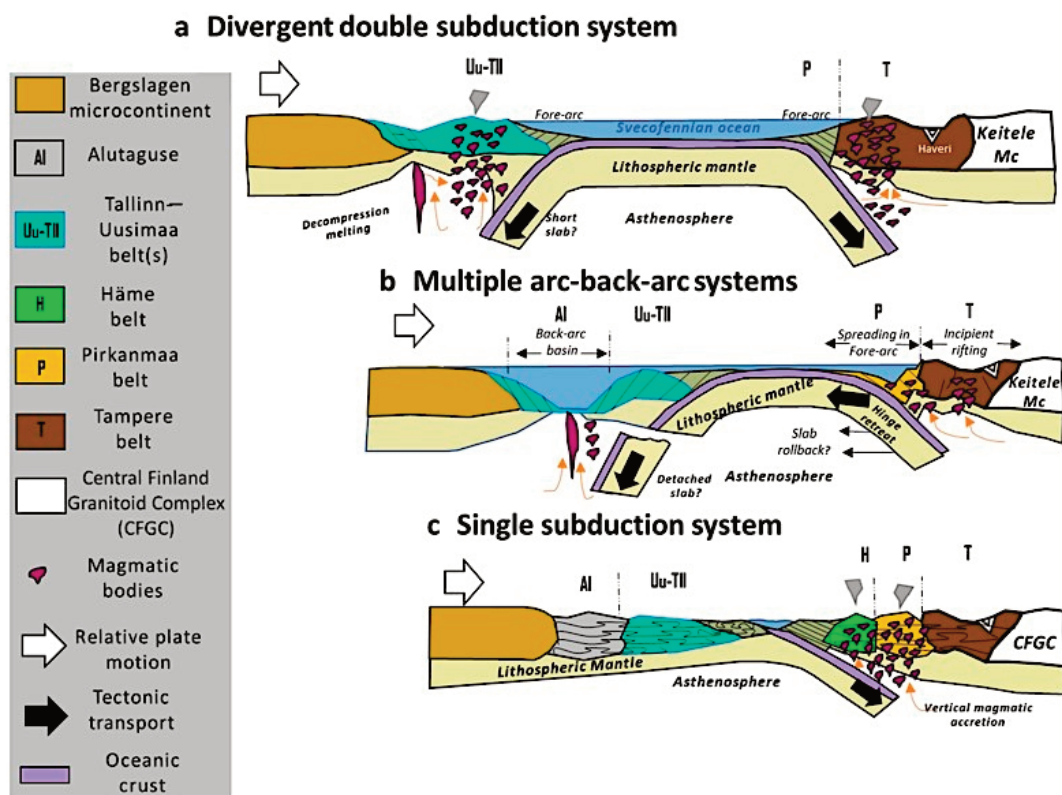


Fig. 6. Schematic geodynamic model (after Kukkonen and Lauri 2009; Kara et al. 2021) of the evolution of Fennoscandia during: **a** – 1.92–1.90 Ga, **b** – 1.90–1.89 Ga, **c** – 1.89–1.87 Ga. The model is a conceptual, not-to-scale representation, designed to illustrate key dynamics rather than precise physical dimensions.

between 1.86 and 1.83 Ga, following granitoid emplacement and preceding late-orogenic granites (Lahtinen et al. 2002, 2005, 2008, 2009; Lahtinen and Nironen 2010; Lahtinen et al. 2011; Nironen 2017). During this period, NW-directed convergence involved the Häme belt in the north and the Uusimaa belt in the south, culminating in the closure of the paleo-Svecofennian ocean (Kukkonen and Lauri 2009; Kara et al. 2021). This collision may be associated with collisional signals observed in a few metasediments, particularly those with low silica content (Fig. 4e). By 1.86 Ga, the geological similarities in the Häme and Uusimaa belts, as reflected in the trace elements of their metasediments, indicate a shared geodynamic evolution since then (Kukkonen and Lauri 2009; Nironen 2017; Kara et al. 2021).

7. Conclusions

This study has advanced the understanding of the metasedimentary and metavolcanic rocks in the Alutaguse zone by compiling historical geochemical data and integrating new samples, offering a refined perspective on the zone's geological evolution and its relation to the Svecofennian orogeny formations across Fennoscandia. The high-SiO₂ metasediments, resembling litharenites, show signs of extensive reworking and alignment with continental rift zones, indicative

of a dynamic geological history and a mature sedimentary environment. In contrast, the low-SiO₂ samples indicate a more complex, collisional tectonic setting with significant hydrothermal alterations, suggesting mafic to intermediate origins and limited sediment reworking. The metavolcanic units analyzed in this study are characterized by subalkaline, tholeiitic trends, metaluminous characteristics, and asthenospheric mantle origins, underscoring a compressional arc environment.

The study's findings suggest that the Alutaguse zone has rift origins and is genetically linked to the Uusimaa units. This supports a double subduction collision model for Fennoscandia's evolution around 1.92–1.87 Ga. Further investigation is needed to determine whether the Tallinn zone is affiliated with the Uusimaa belt or represents a distinct arc predating Uusimaa. The study emphasizes the importance of geochronological assessments, specifically U–Pb isotopic dating of zircon and garnet samples, to refine the provenance of metasedimentary rocks and explore Zn–Pb–Cu anomalies associated with rift-related deposits in the Bergslagen region. Additionally, detailed analysis of sulfurized gneisses could enhance the understanding of local metallogenesis in areas with significant metalliferous anomalies. Finally, detailed gravimetric and magnetic surveys across the Alutaguse zone are essential to map geophysical anomalies and their potential links to metal-bearing deposits.

Data availability statement

Data not already included in the paper and its supplementary materials will be made available upon request.

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Supplementary online data

Supplementary online data to this article can be found at <https://doi.org/10.3176/earth.2025.S05>. The supplementary material is designed to provide an in-depth analysis, including the complete analyzed geochemical dataset of the Alutaguse zone.

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Paleoproterosoikumi metasetendite ja metavulkaniitide geokeemia, päritolu ja tektooniline asend Alutaguse piirkonnas

Juan David Solano-Acosta, Alvar Soesoo ja Rutt Hints

Uuring keskendub Alutaguse piirkonna Paleoproterosoikumi metasetendite ja metavulkaniitide kogukivimi-proovide keemilistele analüüsidele, et mõista idapoolse Fennoskandia ala geodünaamilist arengut Svekofennia orogeneesi perioodil. Uuritud metasetendite üksused koosnevad vilgugneissidest ($\pm$ Grt $\pm$ Crd $\pm$ Sil), metavulkaniitid aga peamiselt amfiboliitidest ja pürokseengneissidest. Töös on kasutatud nii ajaloolisi kui ka uusi kogukivimi koostise keemilisi andmeid. Leitud murenemisindeksid näitavad uuringuala kivimite sobivust setete päritolu ja tektoonilise asendi analüüsiks. Metasetendid on uuringus klassifitseeritud ränisisalduse järgi. Suure SiO₂-sisaldusega (>63 mass%) metasetendid sarnanevad litoklastilistele areniididele, mis viitab suuremale küpsusele ja happelisele lähtematerjalile. Väikse SiO₂-sisaldusega (<63 mass%) metasetendite koostis sarnaneb grauvakkide ja savide omaga ning osutab pigem aluselise või keskmise koostisega lähtematerjalile. Tektoonilise diskriminantanalüüsi alusel saab metasetendeid seostada mandriliise riftistumisega. Alutaguse piirkonna metavulkaniidid klassifitseeruvad TAS-diagrammi alusel subleeliselisteks üksusteks. La/Yb vs. Zr/Nb ja La/Sm vs. Sm/Yb suhete põhjal on nende koostis kõlge lähedasem primitiivse vahevöö omale. Th/Nb ja Th/Zr suhted viitavad basaltse lähtemagma seotusele astenosfääriga, samas kui Y/15-La/10-Nb/8 ja TiO₂-10(MnO)-10(P₂O₅) suhetest tuletatud metavulkaniitide geokeemiline signaal viitab lähtemagma vulkaaniliste saarkaarte päritolule. Oletatavalt võis Alutaguse struktuurne vöönd kujuneda Tallinna-Uusimaa vööndi(te) saarkaartetaguse basseini pärast Bergslageni mikrokontinendi akretsiooni, 1,90–1,87 miljardit aastat tagasi, lõpetades Paleosvekofennia ookeani sulgumise.

Supplemental document of Geochemistry, provenance, and tectonic setting of Paleoproterozoic metasedimentary and metavolcanic units of the Alutaguse zone, Eastern Fennoscandia (Estonia).

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1. INTRODUCTION

This document is crafted to aid in the development of a comprehensive supplementary document for the article "Geochemistry, Provenance, and Tectonic Setting of Paleoproterozoic Metasedimentary and Metavolcanic Units of the Alutaguse Zone, Eastern Fennoscandia (Estonia)." Its purpose is to elaborate on the geochemical details mentioned in the main text. To ensure coherence and facilitate cross-referencing, the organization of this supplementary material will align with the chapter structure of the primary article, enabling readers to seamlessly locate and delve deeper into specific sections requiring further analysis.

4) GEOCHEMICAL DATA AND PROCESSING

Figure S1 outlines the drill cores and the zones where Alutaguse samples come from. In this study, we have standardised the lithological classification of all Alutaguse samples based on [1].

Some siliciclastic metasedimentary lithological classification within the Alutaguse match within both High and Low-SiO₂ groups. Therefore, Alutaguse samples will be examined principally through the silica group content, and the context of lithological classification will be mentioned if it is significant (Table S1). Nevertheless, for each lithological classification of the Alutaguse metasediments, the SiO₂ content distribution is predominantly, if not entirely, linked to one of the silica groups (Fig.S2a). The High-SiO₂ Alutaguse metasedimentary group comprises predominantly samples such as Biotite gneisses, Biotite gneisses ± Grt ± Crd ± Sil, Garnet bearing mica gneisses ± Crd ± Sil, Garnet bearing mica gneisses, Cordierite bearing mica gneisses ± Grt ± Sil, Graphite bearing mica gneisses ± Grt ± Crd ± Sil, Graphite bearing mica gneisses, Biotite-plagioclase gneisses + Amph, Biotite-plagioclase gneisses and Biotite-Amphibole gneisses.

Conversely, the Low-SiO₂ Alutaguse metasediments group includes the majority of samples classified as Biotite-plagioclase gneisses + Amph, Garnet bearing mica gneisses, Biotite gneisses ± Grt ± Crd ± Sil, Cordierite bearing mica gneisses ± Grt ± Sil, Biotite gneisses, Garnet bearing mica gneisses ± Crd ± Sil, Biotite-Amphibole gneisses, Garnet bearing mica gneisses + Amph, Graphite bearing mica gneisses ± Grt ± Crd ± Sil, and Graphite bearing mica gneisses.

Regarding metavolcanic samples, no further subgrouping was conducted. Alutaguse metavolcanics are amphibolites, pyroxene gneisses (±Amph), and 2-pyroxene gneisses (±Amph) (Fig.S2b).

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Estonian Journal of Earth Science

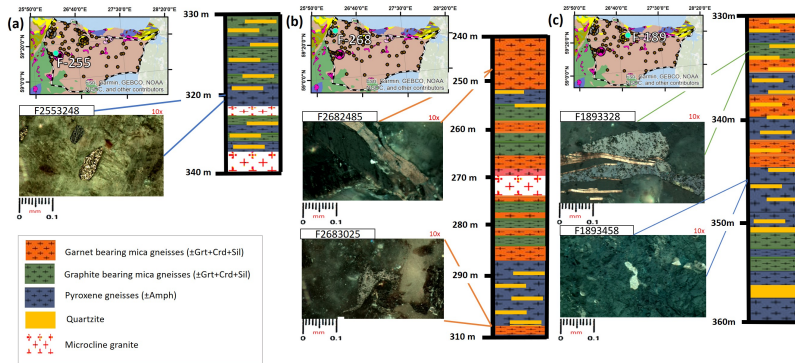


Fig. S1. Scheme of the Estonian basement drill cores, illustrating the geographic zones. A representative PPX-10x image of the analysed samples is also provided. The code in the upper-left corner of the PPX image corresponds to the samples listed in Table 1. The drill core localities are (a) Haljala, (b) Assamalla, (c) Uljaste. The Id-Number of each drill core is indicated within its respective map reference.

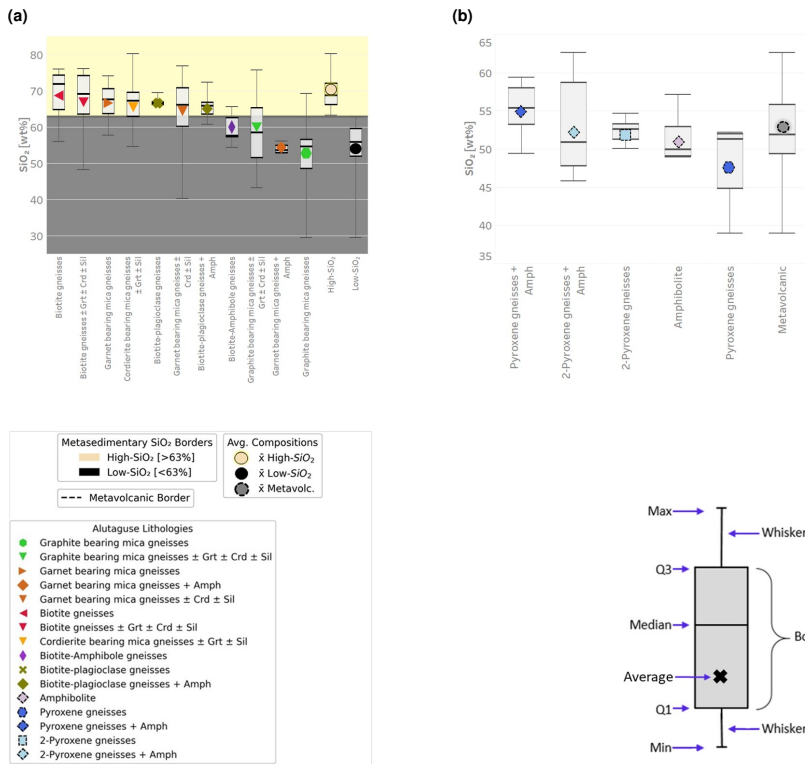


Fig. S2. Box-plot distributions of SiO_2 concentrations across lithological units in the Alutaguse zone: (a) metasedimentary and (b) metavolcanic rocks.

Metasedimentary units																																
Group	Classification samples	Num. Samples	SiO ₂	TiO ₂	Al ₂ O ₃	Fe ₂ O ₃	MnO	MgO	CaO	Na ₂ O	K ₂ O	P ₂ O ₅	SO ₃	LOI	K ₂ O/Na ₂ O	Al ₂ O ₃ /TiO ₂	SiO ₂ /Al ₂ O ₃	Na ₂ O/K ₂ O	CaO/Al ₂ O ₃	CIA	CN	ICV	PIA	CIW	CIW/CIA							
Alutaguse High-SiO ₂ metasedimentary samples	Biotite gneisses	8	72.09	0.46	13.45	3.98	0.07	2.04	2.15	2.58	2.94	0.04	0.20	1.08	1.23	30.91	5.61	1.03	0.16	55.43	0.07	1.50	56.98	64.02	1.16							
	Biotite gneisses ± Grt ± Crd ± Sil	13	70.37	0.50	13.41	5.80	0.06	1.80	2.04	1.91	3.66	0.08	0.37	1.61	2.93	28.80	5.42	0.68	0.15	57.59	0.06	1.50	62.40	69.88	1.21							
	Garnet bearing mica gneisses ± Crd ± Sil	24	69.95	0.56	14.31	5.61	0.06	2.10	2.27	2.02	2.72	0.06	0.33	1.38	1.49	29.33	5.03	0.86	0.17	60.82	0.06	1.43	64.43	69.50	1.14							
	Cordierite bearing mica gneisses ± Grt ± Sil	17	69.40	0.49	14.40	5.79	0.06	1.99	1.94	2.13	3.45	0.06	0.29	1.68	1.77	32.57	4.91	0.67	0.14	59.15	0.06	1.40	62.78	69.56	1.18							
	Garnet bearing mica gneisses	16	69.20	0.49	13.61	6.81	0.12	2.08	2.56	1.98	2.80	0.09	0.25	1.23	2.59	43.10	5.19	0.89	0.19	59.14	0.06	1.58	62.93	68.45	1.16							
	Graphite bearing mica gneisses ± Grt ± Crd ± Sil	15	68.70	0.55	14.67	5.68	0.08	2.55	1.85	1.58	3.60	0.13	0.61	1.65	3.01	50.73	4.78	0.63	0.13	62.61	0.05	1.45	69.00	75.14	1.20							
	Biotite-plagioclase gneisses	5	67.07	0.52	15.13	4.88	0.07	1.39	4.23	2.75	3.49	0.19	0.27	1.18	1.39	30.57	4.45	0.86	0.28	54.13	0.09	1.55	55.83	62.91	1.16							
	Graphite bearing mica gneisses	3	66.85	0.61	13.71	7.47	0.04	1.80	2.09	1.73	3.54	0.11	2.04	2.04	2.17	37.16	5.05	0.50	0.16	59.00	0.06	1.53	63.50	71.10	1.21							
	Biotite-plagioclase gneisses + Amph	6	66.57	0.53	14.85	5.71	0.07	2.63	4.56	2.47	2.15	0.17	0.31	1.11	0.88	32.55	4.49	1.22	0.31	58.78	0.08	1.73	60.77	64.73	1.10							
	Biotite-Amphibole gneisses	1	65.61	0.67	14.09	5.62	0.03	4.34	4.98	1.78	2.45	0.20	0.23	1.53	1.38	21.00	4.66	0.73	0.35	62.36	0.06	2.14	66.15	70.65	1.13							
	Total	108	69.31	0.52	14.16	5.75	0.07	2.11	2.42	2.06	3.12	0.09	0.39	1.44	2.03	35.18	5.02	0.81	0.17	59.25	0.06	1.50	63.09	69.17	1.17							
Alutaguse Low-SiO ₂ metasedimentary samples	Biotite-plagioclase gneisses + Amph	1	60.80	0.63	16.32	7.02	0.07	3.17	2.98	2.24	6.10	0.42	0.25	1.66	2.73	25.89	3.73	0.37	0.18	53.89	0.07	1.78	56.90	68.92	1.28							
	Garnet bearing mica gneisses	5	59.97	0.70	15.03	11.12	0.28	3.58	2.74	1.35	3.78	0.13	1.32	1.56	5.37	23.00	4.01	0.49	0.18	64.95	0.04	1.92	73.63	78.95	1.22							
	Cordierite bearing mica gneisses ± Grt ± Sil	6	59.64	0.92	17.90	11.53	0.08	2.35	2.09	1.93	3.18	0.07	0.30	1.91	1.82	21.95	3.36	0.69	0.12	67.62	0.06	1.40	72.15	76.37	1.14							
	Biotite gneisses	2	58.85	1.13	17.02	9.33	0.09	3.65	5.43	1.29	2.64	0.30	0.28	1.82	2.51	16.85	3.46	0.56	0.32	70.67	0.04	1.85	77.60	80.45	1.14							
	Garnet bearing mica gneisses ± Crd ± Sil	15	58.31	0.99	17.77	10.21	0.12	3.98	2.80	1.85	3.48	0.12	0.37	1.50	2.25	20.84	3.37	0.61	0.17	65.46	0.06	1.73	71.03	75.69	1.16							
	Biotite-Amphibole gneisses	4	57.98	1.32	14.87	10.19	0.15	4.86	6.96	0.86	2.04	0.44	0.33	1.78	3.69	13.80	3.91	0.56	0.47	74.86	0.03	2.53	82.53	84.51	1.13							
	Biotite gneisses ± Grt ± Crd ± Sil	3	56.65	1.43	15.65	14.54	0.05	3.43	1.60	1.31	4.82	0.05	0.48	1.20	9.90	16.03	3.63	0.57	0.10	65.48	0.04	1.99	76.75	82.74	1.28							
	Garnet bearing mica gneisses + Amph	4	54.02	1.62	16.36	12.35	0.17	3.67	6.60	0.75	3.10	0.49	0.86	1.32	4.49	10.70	3.30	0.26	0.40	73.76	0.02	2.21	84.19	86.97	1.18							
	Graphite bearing mica gneisses ± Grt ± Crd ± Sil	27	53.20	0.95	15.72	13.24	0.14	4.30	2.51	1.38	3.55	0.13	4.88	4.70	10.35	18.96	3.55	0.41	0.17	65.53	0.04	2.09	73.39	78.54	1.20							
	Graphite bearing mica gneisses	28	51.21	0.67	13.34	15.31	0.13	3.72	2.14	1.13	3.67	0.13	8.56	5.81	5.08	21.14	3.93	0.36	0.17	64.09	0.04	2.33	72.62	79.10	1.24							
	Total	96	54.81	0.92	15.50	12.82	0.13	3.84	2.81	1.37	3.53	0.16	4.13	3.71	5.90	19.66	3.65	0.46	0.19	65.97	0.04	2.06	73.64	78.86	1.20							
PAAS		62.80	1.00	18.90	7.23	0.11	2.20	1.30	1.20	3.70	0.16				3.08	18.90	3.32	0.32	0.07	70.38	0.04	1.05	79.05	82.72	1.18							
UCC		66.60	0.64	15.50	5.61	0.10	2.48	3.59	3.20	2.80	0.15				0.86	24.06	4.32	1.17	0.23	52.76	0.11	1.66	53.48	58.87	1.12							
Metavolcanic Units																																
Group	Classification samples	Num. Samples	SiO ₂	TiO ₂	Al ₂ O ₃	Fe ₂ O ₃	MnO	MgO	CaO	Na ₂ O	K ₂ O	P ₂ O ₅	SO ₃	LOI	K ₂ O/Na ₂ O	Al ₂ O ₃ /TiO ₂	SiO ₂ /Al ₂ O ₃	Na ₂ O/K ₂ O	CaO/Al ₂ O ₃	CIA	CN	ICV	PIA	CIW	CIW/CIA	WIP	AI	CCPI	SI	ACN/K	ANK	# Mg
Alutaguse metavolcanic samples	Pyroxene gneisses + Amph	8	55.27	1.04	14.77	11.06	0.16	6.18	8.43	1.11	1.08	0.23	0.67	1.59	1.65	15.56	3.78	1.50	0.58	76.53	0.04	2.93	80.47	81.60	1.07	39.00	43.23	87.87	0.53	0.82	5.74	51.84
	2-Pyroxene gneisses + Amph	20	52.66	1.26	13.44	12.98	0.19	6.39	9.67	1.53	0.83	0.18	0.88	1.58	0.74	12.67	3.97	3.41	0.74	71.39	0.05	3.63	73.38	74.66	1.05	42.23	39.00	87.27	0.34	0.66	5.63	48.88
	2-Pyroxene gneisses	5	52.40	1.49	12.88	13.72	0.22	7.69	8.70	0.53	1.37	0.11	0.90	2.50	2.05	9.33	4.09	1.28	0.68	81.70	0.02	3.79	86.31	88.25	1.09	39.01	49.40	91.74	0.57	0.75	9.59	52.67
	Amphibolite	5	51.61	0.75	11.13	12.42	0.18	9.62	9.21	0.72	1.69	0.21	2.46	2.15	2.40	14.99	4.77	0.45	0.84	72.79	0.02	4.91	79.63	82.65	1.14	49.16	52.59	89.73	0.70	0.58	3.96	58.93
	Pyroxene gneisses	4	48.43	1.45	14.56	13.18	0.35	8.76	7.98	0.68	2.06	0.26	2.29	2.03	4.14	12.67	3.32	1.18	0.53	77.32	0.02	3.56	83.43	86.95	1.16	49.48	54.63	89.89	0.54	0.83	7.07	56.66
	Total	42	52.60	1.20	13.46	12.65	0.20	7.11	9.10	1.15	1.16	0.19	1.17	1.80	1.59	13.10	3.98	2.23	0.69	74.33	0.04	3.66	77.97	79.72	1.08	42.75	44.15	88.46	0.46	0.71	6.06	51.83
PM		45.00	0.20	4.45	8.95	0.14	37.80	3.55	0.36	0.03	0.02				0.08	22.25	10.11	12.41	0.80													
N-MORB		49.62	1.26	14.99	12.27	0.19	8.21	11.65	2.25	0.09	0.17				0.04	11.94	3.31	24.24	0.78													
E-MORB		50.36	1.43	15.40	10.34	0.17	8.12	11.56	2.57	0.32	0.18				0.12	10.78	3.27	8.03	0.75													

Table S1. Average values of the major bulk-rock elemental data of the different metasedimentary and metavolcanic lithologies of Alutaguse. Major oxide values correspond to 100% normalised data. LOI correspond to original Non-normalized values. Fe₂O₃_{tot} = FeO_{tot} * 1.1113

Metasedimentary																																			
Units [ppm]																																			
Group	Classification samples	Num. Samples	V	Cr	Co	Ni	Cu	Zn	Rb	Sc	Sr	Y	Zr	Nb	Mo	Ba	La	Ce	Pr	Nd	Sm	Eu	Gd	Tb	Dy	Ho	Er	Tm	Yb	Lu	Hf	Ta	Pb	Th	U
Alutaguse Hypothetical High-SiO ₂ Metasedimentary samples	Garnet bearing mica gneisses ± Crd ± Sil	16	107.81	87.25	15.36	43.1	55.01	125.81	109.6	12.95	193.28	17.13	150.06	12.52	2.72	877.5	48.01	98.66													4.42	0.64	27.83	15.96	2.31
	Garnet bearing mica gneisses	4	144.12	73.18	34.2	47.7	45.66	171.14	80.11	13.71	177.6	17.76	164.09	13.58	4.16	558.17	50.57	102.38	15.9	52.8	6.39	1.3	5.11	0.49	1.69	0.32	0.92	0.13	0.8	0.12	4.68	0.6	24.04	17.35	2.59
	Cordierite bearing mica gneisses ± Grt ± Sil	5	159.8	79.4	30.24	81.26	184.16	129.4	122.82	19.84	277.62	22.38	173.7	11.96	4.16	762	53.94	110.38												5	0.73	16.2	19.1	2.86	
	Graphite bearing mica gneisses ± Grt ± Crd ± Sil	1	122.37	129.65	76.9	30.37	27.51	116.38	185.82	7.93	112.52	19.61	227.5	15.62	0.33	844.42	42.66	79.75	9.85	35.11	6.06	1.26	5.16	0.71	3.58	0.73	2.09	0.31	2.13	0.32	6.53	1.13	12.83	16.32	2.26
	Total	26	123.33	85.13	23.24	50.41	75.41	130.81	110.59	14.18	209.39	18.47	160.87	12.67	3.06	795.45	48.62	99.4	10.95	37.97	5.65	1.15	4.69	0.62	2.92	0.62	1.77	0.26	1.71	0.25	4.7	0.68	24.16	16.56	2.46
Alutaguse Hypothetical Low-SiO ₂ Metasedimentary samples	Graphite bearing mica gneisses ± Grt ± Crd ± Sil	104	208.63	101.18	31.38	122.21	190.05	566.4	109.24	15.4	133.43	20.96	134.01	9.03	9.76	511.15	39.01	77.41	8.68	32.53	6.15	1.57	5.95	0.93	5.39	1.24	3.46	0.52	3.55	0.56	3.69	0.57	104.07	13.21	3.94
	Graphite bearing mica gneisses	2	224.31	105.62	63.3	55.53	310.13	250.14	159.72	16.64	68.57	18.07	136.39	16.32	10.72	417.73	41.06	79.29	9.47	34.05	5.99	0.97	5.39	0.74	3.6	0.72	1.9	0.27	1.82	0.3	4.08	1.15	7.06	16.96	4.82
	Garnet bearing mica gneisses	1	115.22	117.9	67.64	33.79	8.36	115.84	156.58	35.14	196.02	82.47	300.37	13.08	0.84	1707.24	27.63	46.2	4.97	17.07	3.28	1.74	5.34	1.33	10.66	2.94	9.59	1.61	11.85	1.89	8.58	0.6	32.88	5.44	0.7
	Total	107	207.51	101.5	32.31	119.53	190.6	556.28	111.08	15.61	133.42	21.53	136.3	9.29	9.7	521.78	38.93	77.17	8.51	31.61	5.83	1.44	5.7	0.92	5.47	1.27	3.66	0.56	3.9	0.61	3.75	0.59	100.84	13.22	3.92
	PAAS	150	110		55				160	16	200	27	210	19		650	38.2	79.6	8.83	33.9	5.55	1.08	4.66	0.77	4.68	0.99	2.85	0.41	2.82	0.44	5	1.28	20	14.6	3.1
UCC			107	83	17	44	25	71	112	13.6	350	22	190	12	1.5	550	30	64	7.1	26	4.5	0.88	3.8	0.64	3.5	0.8	2.3	0.33	2.2	0.32	5.8	1	2.7	10.5	2.7
Chondrite			56	2650	500	10500	120	310	2.3	5.92	7.25	1.57	3.82	0.24	0.9	2.41	0.237	0.613	0.093	0.457	0.148	0.0563	0.199	0.0361	0.246	0.0546	0.16	0.0247	0.161	0.0246	0.103	0.0136	2.47	0.029	0.0074
Metavolcanic																																			
Units [ppm]																																			
Alutaguse Metavolcanic samples	2-Pyroxene gneisses + Amph	9	195.67	207	44.48	132.48	596.62	714.11	3.13	24.31	57.71	27	83.26	5.36	2.81	75.56	24.28	55.75													2.3	0.32	715.81	8.03	4.07
	2-Pyroxene gneisses	11	315.5	195.76	44.13	78.8	140.26	234.6	43.89	37.21	160.78	19.64	80	12.08	2.42	468.13	13.98	30.75	1.92	8.18	2.25	1.01	2.65	0.49	3.01	0.69	1.94	0.3	1.99	0.31	2.38	0.77	93.77	3.1	1.54
	Pyroxene gneisses + Amph	2	131	89.5	14.4	58.05	101.25	113	79.85	14.1	135.5	18	181.5	10.15	4.49	480	39.85	76.9													5.5	0.48	15.5	11.28	2.25
	Pyroxene gneisses	10	204.23	59.66	47.64	65.51	130.4	155.68	54.32	20.12	1739.15	22.56	91.18	12.6	4.22	354.96	67.74	125.85	3.3	14.6	3.72	1.39	4.26	0.8	4.73	1.08	2.98	0.44	2.97	0.42	2.7	0.52	11.89	12.14	1.44
	Total	32	235.49	149.75	43.47	88.44	263.09	337.2	37.93	26.8	623.45	22.52	90.75	10.23	3.22	323.1	35.29	70.38	2.38	10.32	2.74	1.14	3.19	0.59	3.58	0.82	2.29	0.35	2.32	0.35	2.66	0.55	238.24	7.82	2.26
PM			82	2625	105	1960	30	55	0.6	16.2	19.9	4.3	10.5	0.658	0.05	6.6	0.648	1.675	0.254	1.25	0.406	0.154	0.544	0.099	0.674	0.149	0.438	0.068	0.49	0.06	0.283	0.037	0.15	0.0795	0.0203

Continue next page

Metasedimentary Units																									
Group	Classification samples	Num. Samples	Th/Sc	Cr/Th	Nb/Ta	La/Nb	Th/La	Sm/Yb	Zr/Y	Th/Co	Th/U	Zr/Sc	La/Sc	La/Co	Sr/Ce	ΣREE	HREE	LREE	HREE/LREE	100*TiO ₂ /Zr	δNb	Eu*	(Gd/Yb) _{cn}	(La/Sm) _{cn}	(La/Yb) _{cn}
Alutaguse High-SiO ₂ Metasedimentary samples	Garnet bearing mica gneisses ± Crd ± Sil	16	1.33	6.86	18.33	4.8	0.31		10.21	1.1	8.2	14.43	4.28	4.11	0.01						-0.32				
	Garnet bearing mica gneisses	4	1.56	4.86	24.38	3.91	0.35	8	10.68	1.75	13.7	14.26	4.86	4.48	0.01	311.58	4.47	307.1	0.01	0.32	-0.28	0.27	4.81	4.8	36.09
	Cordierite bearing mica gneisses ± Grt ± Sil	5	1.09	5.28	16.27	4.34	0.34		8.27	0.81	6.49	10.05	3.12	2.44	0.01						-0.26				
	Graphite bearing mica gneisses ± Grt ± Crd ± Sil	1	2.06	7.94	13.81	2.73	0.38	2.85	11.6	0.21	7.23	28.69	5.38	0.55	0.01	189.73	9.88	179.85	0.05	0.38	-0.4	0.76	4.35	5.92	40.39
	Total	26	1.33	6.35	18.45	4.42	0.33	4.3	9.91	1.09	8.5	14.11	4.12	3.64	0.01	250.65	7.18	243.48	0.03	0.35	-0.3	0.52	4.58	5.36	38.24
Alutaguse Low-SiO ₂ Metasedimentary samples	Graphite bearing mica gneisses ± Grt ± Crd ± Sil	104	1.43	32.18	14.59	13.31	0.33	2.5	10.56	0.67	4.29	17.25	4.36	2.02	0.02	175.89	15.64	160.25	0.12	0.77	-0.42	0.45	1.69	3.6	12.66
	Graphite bearing mica gneisses	2	1.03	6.59	13.77	2.46	0.43	4.34	7.86	0.26	4.76	8.24	2.49	0.62	0.02	185.55	9.34	176.21	0.06	0.51	-0.01	0.37	1.62	3.55	8.22
	Garnet bearing mica gneisses	1	0.15	21.69	21.96	2.11	0.2	0.28	3.64	0.08	7.76	8.55	0.79	0.41	0.01	146.09	39.87	106.22	0.38	0.23	-0.14	0.35	2.21	3.58	11.33
Total		107	1.4	31.38	14.64	12.9	0.33	2.58	10.42	0.65	4.33	16.97	4.27	1.98	0.02	174.94	16.7	158.24	0.14	0.67	-0.41	0.43	1.73	3.59	11.74
PAAS			0.91	7.53	14.84	2.01	0.38	1.97	7.78		4.71	13.12	2.39		0.00	184.78	12.96	171.82	0.08	0.48	-0.12	0.63	9.20	4.30	1.34
UCC			0.77	7.90	12.00	2.50	0.35	2.05	8.64	0.62	3.89	13.97	2.21	1.76	0.09	146.37	10.09	136.28	0.07	0.34	-0.32	0.63	9.26	4.16	1.40
Metavolcanic units																									
Alutaguse Metavolcanic samples	2-Pyroxene gneisses + Amph	9	0.62	209.42	16.3	4.64	0.26		3.07	0.19	1.53	5.09	1.78	0.59	0.03						0.12				
	2-Pyroxene gneisses	11	0.17	688.46	14.82	1.36	0.18	1.13	4.91	0.14	1.36	3.29	0.67	0.53	0.07	44.69	8.73	35.96	0.24	1.87	0.34	0.5	0.98	2.98	3.5
	Pyroxene gneisses + Amph	2	0.82	9.17	23.03	3.93	0.27		10.96	0.98	5.72	12.86	2.85	4.31	0.01						-0.46				
	Pyroxene gneisses	10	1.35	119.37	22.11	66.09	0.14	1.25	5.84	0.39	5.82	8.26	27.6	1.81	0.01	75.26	13.43	61.84	0.22	1.5	0.41	0.34	1.96	4.39	13.63
	Total	32	0.71	333.44	18.02	22.67	0.2	1.17	5.06	0.28	3.07	5.95	9.54	1.18	0.04	54.88	10.3	44.58	0.23	1.74	0.25	0.44	1.3	3.45	6.88

Table S2. Average values of the Trace and REE elemental data and relations of the metasedimentary and metavolcanic samples of the Alutaguse zone.

5) RESULTS

5.1 GEOCHEMISTRY OF MAJOR ELEMENTS

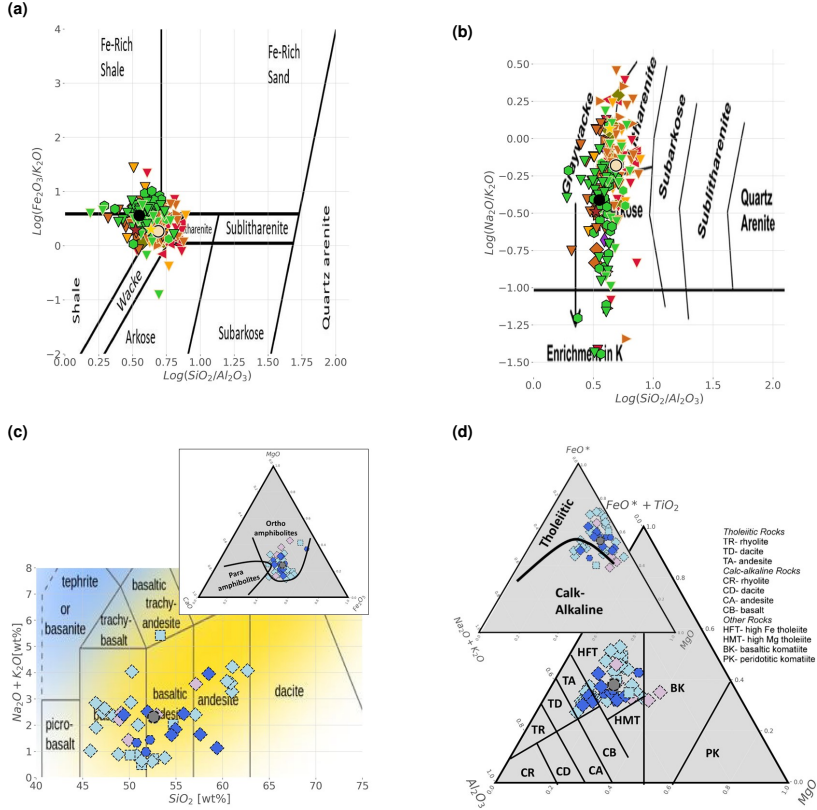


Fig. S3. Lithological classification plots for Alutaguse samples: (a) Classification of metasedimentary samples according to Herron (1988) [2] and (b) Pettijohn et al. (1987) [3]. (c) Classification of metavolcanic samples using the TAS diagram by Bas et al. (1986) [4], which includes distinctions between alkaline (blue) and sub-alkaline (yellow) compositions, complemented by a CaO-MgO-Fe₂O₃ triangular plot after Walker et al. (1959) [5]. (d) AFM triangular plots after Irvine and Baragar (1971) [6], with Jensen's (1976) [7] plot as inset. Legend as in Figure S2.

6) DISCUSSION

6.1) Weathering & Alteration indices

During weathering, elements like Na, K, and LILEs deplete, while Al₂O₃, TiO₂, REEs, and HFSEs enrich. Despite this, HFSEs and REEs exhibit limited changes due to their inherent immobility during weathering ([8–10]). To determine the mineralogical and chemical changes that occur in the analysed metasediments as a result of alteration, chemical alteration indices are used, like the Chemical Index of Alteration (CIA; [11]), Plagioclase Index of Alteration (PIA; [12]), Chemical Index of Weathering (CIW; [13]), and the Index of Compositional Variation (ICV; [14]). For metavolcanic samples, indices such as the Hashimoto Alteration Index (AI; [15]), Chlorite-Carbonate-Pyrite Index (CCPI; [16]), the Parker weathering index (WIP; [17]), and sericitisation index (SI; [18, 19]) were utilised.

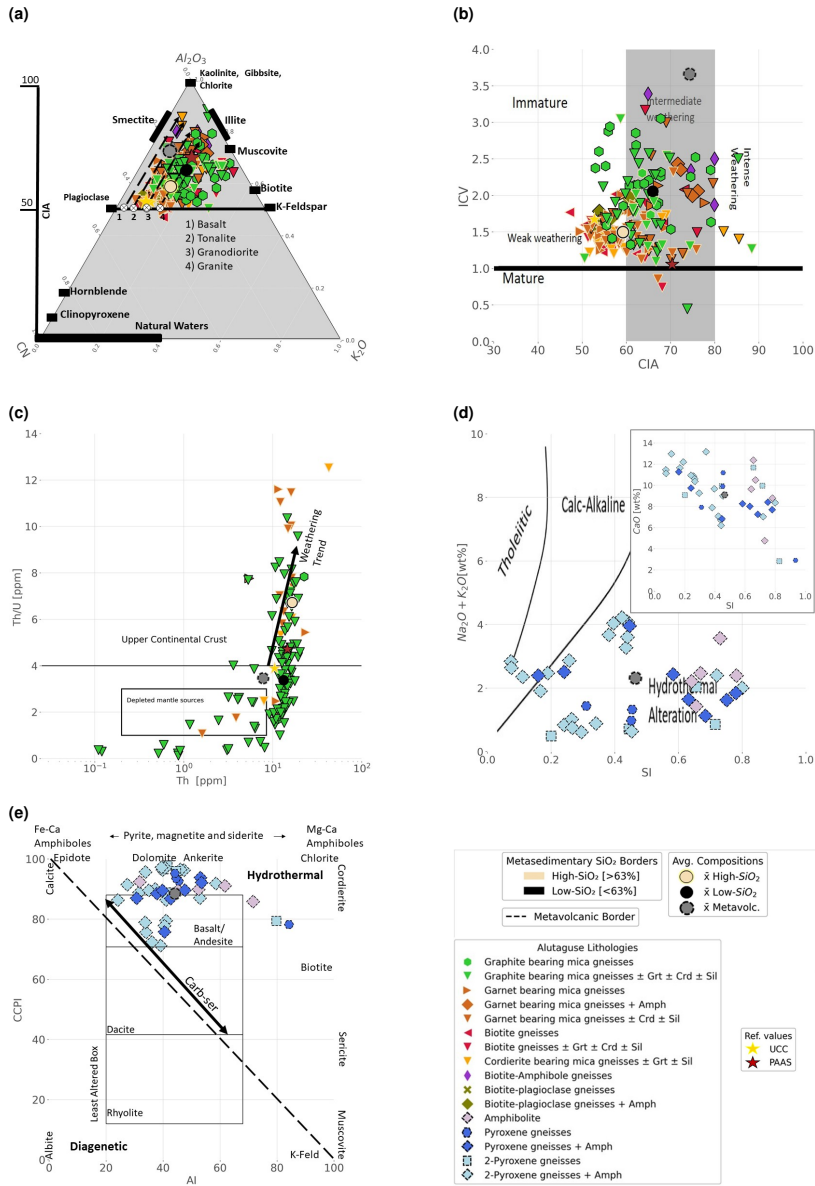


Fig. S4. Weathering and alteration discrimination plots for Alutaguse metasedimentary and metavolcanic samples: (a) A-CN-K [Al₂O₃–(CaO*+Na₂O)–K₂O] ternary (After Nesbitt and Young (1984, 1989) [20, 21]); (b) ICV vs. CIA (After Nesbitt and Young (1984); Cox et al. (1995) [14, 20]); (c) Th/U vs. Th (After McLennan (1993) [10]); (d) sericitisation index (SI) vs. K₂O+Na₂O for volcanic rocks with SI vs. CaO insets for metavolcanic rocks (After MacLean and Hoy (1991) [18]); and (e) CCPI vs. AI alteration box plots for metavolcanics (After Large et al. (2001) [16]).

The Chemical Index of Alteration (CIA) quantifies the transformation of feldspars into aluminous clays, serving as an indicator of weathering intensity and paleoclimatic conditions [10, 11, 22]. It is calculated as $CIA_{[mol]} = \left(\frac{Al_2O_3}{Al_2O_3 + CaO^* + Na_2O + K_2O} \right) \times 100$, where CaO^* represents the non-apatite silicate fraction of CaO [12]. This adjustment involves reducing CaO by the apatite-bound phosphorous [$CaO' = CaO - \left(\frac{10}{3} \times P_2O_5 \right)$], and equating CaO^* to either Na_2O or CaO' , whichever is lesser. Maximum weathering yields a CIA of 100, typical for kaolinite, with lower scores for feldspars [11, 12]. The Plagioclase Index of Alteration (PIA) assesses plagioclase weathering and ranges from 50 in fresh rocks to 100 in fully altered clays [12]. In Alutaguse, High-SiO₂ metasediments show CIA values between 47.13 and 75.85, and PIA between 45.60 and 90.67, aligning with UCC averages. Conversely, Low-SiO₂ samples show CIA values from 52.03 to 88.38 and PIA from 53.52 to 99.73, trending towards PAAS values (Table S1).

The A–CN–K plot ([20]) assesses weathering in provenance areas and elucidates source rock compositions. Alignment with the A–CN line indicates minimal weathering and the presence of fresh feldspars ([12, 20]). For Alutaguse (Fig. S4a), Low-SiO₂ samples veer towards the Al₂O₃ apex, suggesting more intense weathering, whereas High-SiO₂ samples align closer to CIA reference values, typical of tonalite to granite trends.

The CIW_[mol] index, or "potassium-free CIA," defined as $CIW_{[mol]} = \frac{Al_2O_3}{Al_2O_3 + CaO^* + Na_2O} \times 100$, measures the transformation of feldspar to clay during weathering ([12, 13, 23]). It ranges from 50 in unaltered rocks to 100 in fully weathered ones, accounting for the immobility of Al₂O₃ and the mobility of Na₂O and CaO during weathering ([12]). For Alutaguse, the High-SiO₂ group shows an average CIW of 69.17 (55.05–94.43), approaching UCC values (58.87), while the Low-SiO₂ group averages 78.86 (60.99–99.76), aligning closer with PAAS values (82.72). This variation in CIW and CIA values reflects the differential loss of mobile cations and mineral structural changes ([19]).

The ICV_[wt.%] index, defined as $ICV_{[wt.\%]} = \frac{Fe_2O_3 + K_2O + Na_2O + CaO + MgO + TiO_2}{Al_2O_3}$, assesses compositional maturity and origins of sediment recycling in metasediments ([24, 25]). An ICV less than 1 signifies mature sediments, indicative of recycling or intense weathering, whereas an ICV greater than 1 suggests immaturity, common in first-cycle deposits from tectonically active regions. Non-clay minerals typically exhibit higher ICV values, which decrease with weathering or as sediments mature ([22]). In Alutaguse (Fig. S4b), High-SiO₂ metasediments average an ICV of 1.50 (0.45–2.32), while Low-SiO₂ metasediments average 2.06 (1.06–4.85) (Fig. S4b).

The Th/U ratio indicates weathering conditions in rocks, typically ranging from 3.5 to 4.0 in upper crustal rocks ([10]). This ratio increases with the oxidation of U⁴⁺ to U⁶⁺ during weathering and its subsequent loss in sedimentation ([26–28]), whereas sediments from active tectonic zones often show ratios below 3.5 ([29]). The ratio varies significantly among different rock types: 21 in metapelites and metagreywackes, 12 in garnet-alkali-feldspar gneisses, and 3.6 in calc-silicate rocks ([26]). Mantle-derived volcanic rocks typically exhibit lower ratios due to geochemical depletion ([30]). Notable increases to 8–10 in amphibolites and up to 25 in granulites suggest metamorphic shifts or uranium enrichment ([31]). In metapelites, graphite indicates low redox conditions and reduced uranium solubility ([32]), while titanite stability in calc-silicate rocks prevents uranium leaching during intense metamorphism ([33]). For Alutaguse (Fig. S4c), High-SiO₂ metasediments show an average Th/U ratio of 8.68 (ranging from 1.07 to 37.37), suggesting more extensive weathering, while Low-SiO₂ samples average 4.33 (ranging from 0.12 to 29.56). Graphite-rich mica gneisses in the region display broad thorium ranges with low Th/U ratios, reflecting influences from mantle sources or metamorphic processes ([31]).

The CIA, PIA, ICV, and CIW indices are validated metrics for assessing alteration in volcanic rocks, showing elevated values in Alutaguse metavolcanic samples [16, 17, 19, 34]. These indices help discern alteration-induced changes. The alteration index $AI_{[wt.\%]} = 100 \times \frac{MgO + K_2O}{MgO + K_2O + CaO + Na_2O}$ highlights the shift from sodic plagioclase to sericite and chlorite [15]. The $CCPI_{[wt.\%]} = 100 \times \frac{FeO + MgO}{FeO + MgO + Na_2O + K_2O}$ tracks the transition from albite and K-feldspar to Mg-Fe chlorite [16]. The $WIP_{[wt.\%]} = 100 \times \left(\frac{2Na_2O}{0.35} + \frac{MgO}{0.9} + \frac{2K_2O}{0.25} + \frac{CaO}{0.7} \right)$ emphasizes mobile alkali and earth metal ratios [17]. Additionally, the $SI_{[wt.\%]} = \frac{K_2O}{Na_2O + K_2O}$ signifies the formation of sericite from altered plagioclases [29].

K₂O+Na₂O versus SI analysis (Fig. S4d) indicates that Alutaguse metavolcanic samples undergo metasomatic hydrothermal alteration, placing them outside the typical igneous spectrum. The CaO versus SI plot shows a decline in CaO as SI increases (Fig. S4d inset), suggestive

of carbonation alteration ([19]). According to the alteration box plot by [16], these samples predominantly fall within the basalt/andesite zone, indicating significant hydrothermal influence, possibly linked to high SO_3 concentrations in Low- SiO_2 Alutaguse metasediments (Fig. ??). This alteration has led to enhanced carbonate-sericite alteration and elevated CCPI values, with a trend towards the calcite-epidote node, which shows similarities with Fe-Ca amphiboles. However, it is advised against concluding alkalinity or petrographic affinity based solely on the analysed mineralogy of (meta)volcanic compositions ([16, 19, 34]).

6.2 Metasedimentary rocks

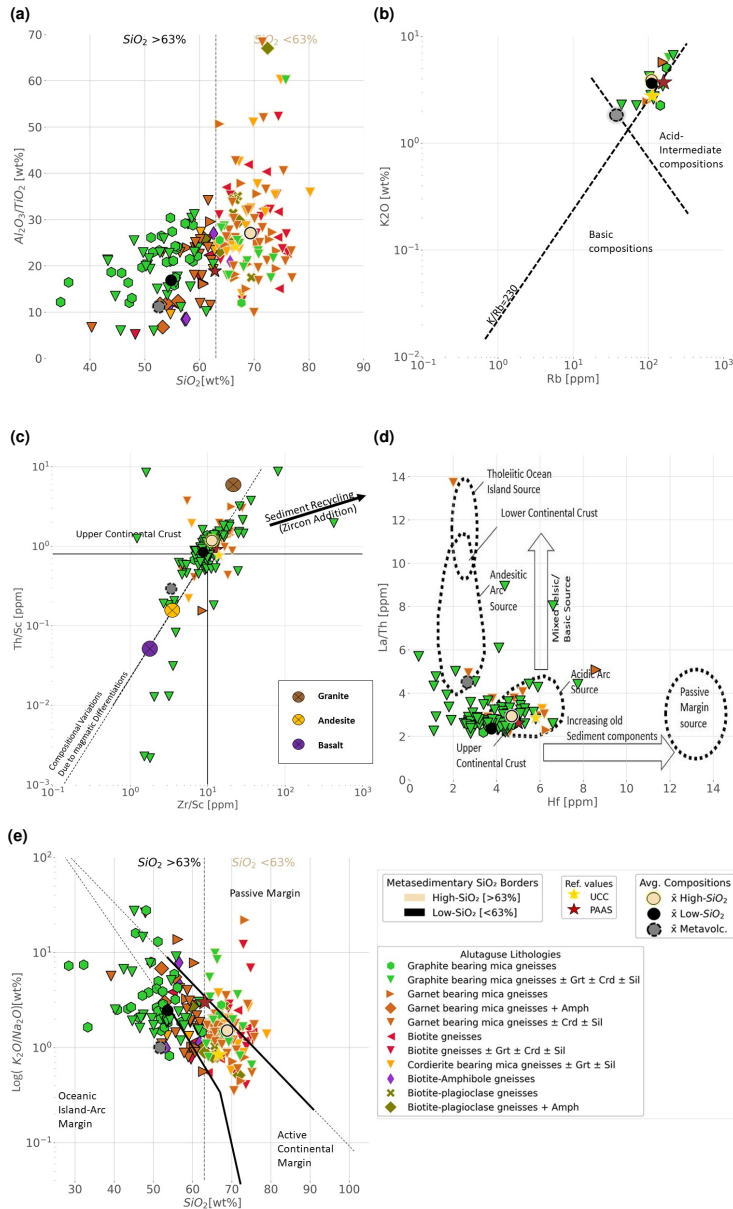


Fig. S5. Bivariate provenance and tectonic plots of Alutaguse metasedimentary samples. (a) Al_2O_3/TiO_2 vs. SiO_2 (After Hayashi (1997) [35]); (b) K_2O and Rb correlation, where a K/Rb value of 230 indicates a magmatic trend, adapted after Shaw (1968) [36]; (c) Th/Sc vs. Zr/Sc plots for identifying sedimentary provenance follow McLennan (1993) [37]; (d) La/Th vs. Hf to discriminate tectonic settings, after Floyd (1987) [38], and (e) K_2O-Na_2O vs. SiO_2 (After Roser (1986) [39]).

6.3 Metavolcanic rocks

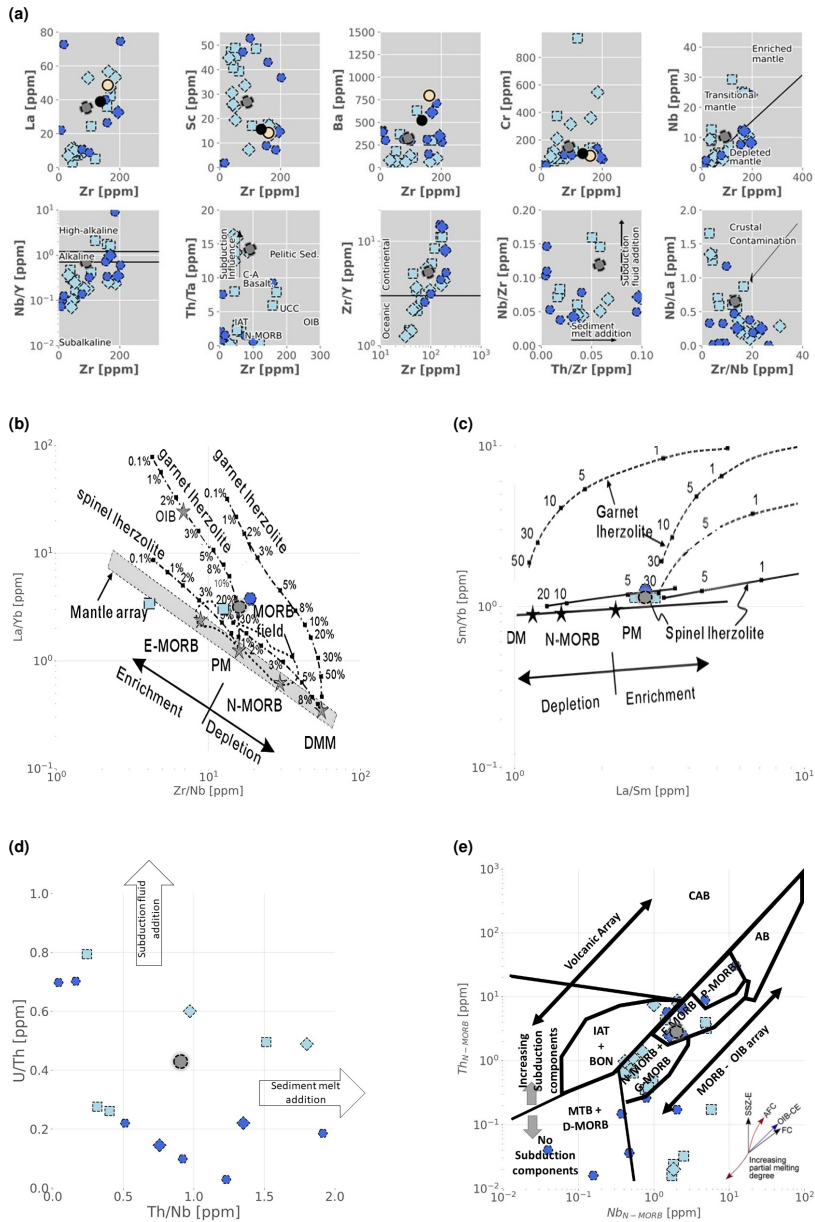


Fig. S6. Bivariant metavolcanic source plots. (a) Zr binary plots versus trace elements for Alutaguse metavolcanic samples.; (b) La/Yb vs. Zr/Nb and (c) Sm/Yb vs. La/Sm (After Yang et al. (2014) [40]); (d) U/Th vs. Th/Nb (After Ye et al. (2018) [41]); (e) N-MORB-normalised Th vs. Nb for volcanic and metavolcanic rocks (After Saccani (2015) [42]). Legend as in Figure S7

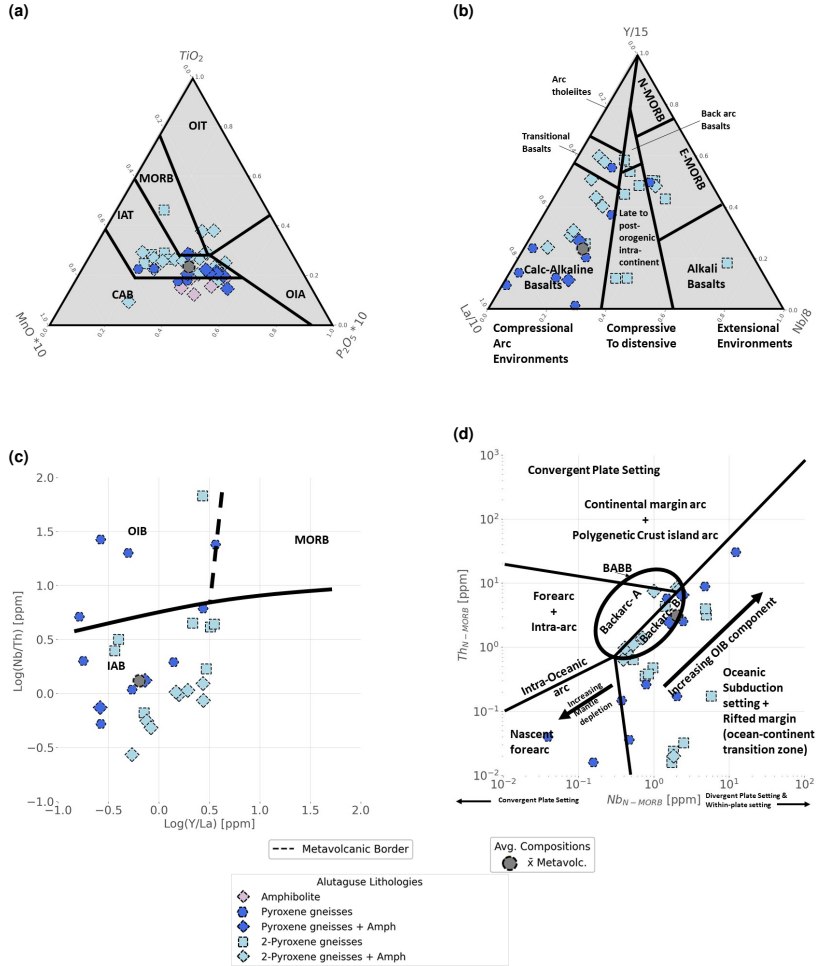


Fig. S7. Metavolcanic tectonic affinities of the Alutaguse samples: (a) TiO_2 -10(MnO)-10(P_2O_5) triangular plots (After Mullen (1983) [43]); (b) Y/15-La/10-Nb/8 triangular plot (After Cabanis and Lecolle (1989) [44]); (c) $\log(Nb/Th)$ vs. $\log(Y/La)$ plot (After Liu et al. (2018) [45]); (d) N-MORB-normalised Th vs. Nb tectonic setting plot for volcanic and metavolcanic rocks (After Saccani (2015) [42]).

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Appendix III

J D Solano-Acosta, A Soesoo, and R Hints. Integrated Geophysical and Emplacement Modelling of the Märjamaa and Kloostri rapakivi Granitoids, Estonia: Insights into Intrusion Geometry and Tectonic Controls. *Precambrian Research*, 430:107938, 2025. doi:[10.1016/j.precamres.2025.107938](https://doi.org/10.1016/j.precamres.2025.107938)



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Research paper

Integrated geophysical and emplacement modelling of the Märjamaa and Kloostri rapakivi granitoids, Estonia: Insights into intrusion geometry and tectonic controls

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ABSTRACT

The ~1.62 Ga buried Märjamaa and Kloostri rapakivi granitoids in western Estonia consist of three magmatic phases. This study utilizes processed Bouguer gravity and magnetic anomaly data to model the internal architecture and emplacement geometry of the granitic bodies, testing how inherited crustal structures influenced these processes within the context of the Mesoproterozoic Anorthosite–Mangerite–Charnockite–Granite (AMCG) magmatism in Fennoscandia. Potential-field data were processed through spectral separation and derivative filters (TDR, TDX, AS), complemented by lineament extraction and density mapping. A cross-gradient joint inversion using the SimPEG Python module, conducted down to 10 km, delineates three-dimensional density and susceptibility contrasts. Potential lineaments align with Riedel shear patterns along the NW-trending, dextral-oblique Åland–Paldiski–Pskov Deformation Zone (PPDZ), indicating transtensional pull-apart reactivation that focused magma ascent and suggesting that 1.6–1.4 Ga AMCG rapakivi granites and coeval basins formed through reactivation of Svecofennian fabrics. Phase I rooted deeply, formed as a steep-sided tabular granodiorite body that evolved into a piston–cauldron structure through roof collapse and block assimilation, producing positive magnetic and Bouguer anomalies. Phase II intruded as a surrounding granite ring during continued subsidence, associated with negative gravity anomalies. Phase III represents a late-stage trachytic Kloostri leucogranite emplaced by asymmetric subsidence, characterized by positive magnetic and negative Bouguer signatures. The phases were placed during the final stages of the first Fennoscandian rapakivi AMCG event (i.e. Wiborg suite), driven by asthenospheric upwelling and mafic underplating from superswell activity, within Nuna's breakup configuration. Overall, the results support a crustal structure-controlled emplacement in which shear-zone reactivation affect the geometry of the Fennoscandian rapakivi intrusions.

1. Introduction

Understanding how inherited crustal structures govern the geometry and emplacement of anorogenic (A-type) granites remains a key open question in Precambrian geology (Groccott et al., 1999; Puura and Flodén, 1999, 1996, 2000; Rämö and Haapala, 2005; Vigneresse, 2005; All et al., 2006; Bonin, 2007; Larin, 2009; Sharkov, 2010; Dall'Agnol et al., 2012; Frost and Frost, 2013; Siachoque et al., 2017; Salminen et al., 2021; Johansson, 2023). Rapakivi intrusions - a distinctive A-type subtype - provide a critical natural laboratory, as their emplacement often coincides with shearing zones and strike-slip corridors (Branigan, 1987; Elo and Korja, 1993; Eklund et al., 1996; Groccott et al., 1999; Alviola et al., 1999; Kosunen, 1999; Vigneresse,

2005; Karell et al., 2009, 2014; Nordbäck et al., 2024). Within a craton, episodic reactivation under transpressional–transtensional stress fields generates shearing and pull-apart structures that focus magma ascent and create space for pluton emplacement (Sylvester, 1988; Xu et al., 2013; Tomek et al., 2014; Pirajno and Santosh, 2015; Mondal and Mamtani, 2016; Siachoque et al., 2017; Burchard, 2018; Nordbäck et al., 2024). The Märjamaa rapakivi granitoid and its Kloostri satellite exemplify this setting in western Estonia, emplaced near the NW-trending Åland–Paldiski–Pskov Deformation Zone (PPDZ), where integrated potential-field analysis and lineament mapping allow testing of inherited structural controls on emplacement mechanisms (Puura

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and Flodén, 2000; Kirs et al., 2004; All et al., 2006; Solano-Acosta et al., 2023).

A-type granites are low-H₂O to anhydrous felsic rocks emplaced in post- to anorogenic settings (Loiselle and Wones, 1979; Eby, 1992; Frost et al., 2001; Vigneresse, 2005; Dall'Agnol and de Oliveira, 2007). Rapakivi granites, characterized by large and rounded crystals of orthoclase each with a rim of oligoclase, are widespread in Precambrian cratons, reported throughout most Precambrian cratons, including Baltica (or East European Craton), Laurentia, Amazonia, and the Ukrainian Shield (Haapala and Rämö, 1992; Rämö and Haapala, 1995, 2005; Bogdanova et al., 2006, 2015; Sharkov, 2010; Salminen et al., 2021; Grabarczyk et al., 2023), but rare in Phanerozoic orogens (Rajesh, 2000; Christiansen et al., 2007; Luo et al., 2024). They are compositionally evolved (high Fe/Mg, K/Na, Ga/Al, HFSE, and most REEs except Eu) (Haapala, 1995; Rämö and Haapala, 2005; Heinonen et al., 2010a; Valkama et al., 2016) with isotopic signatures indicating strong crustal input ($\epsilon_{\text{Nd}} < 0$, unradiogenic Hf, elevated Th and Zn) (Larin, 2009; Heinonen et al., 2010b). In Fennoscandia, rapakivi magmatism is closely tied to four-age AMCG (Anorthosite–Mangerite–Charnockite–Granite) suites (Emslie et al., 1994; Rämö and Haapala, 1995; Ahl, 1997; Alviola et al., 1999; Larin, 2009), emplaced after the Gothian orogeny and distinguished from Archean analogues by stronger assimilation and ferroan character (Frost and Frost, 2013; Ashwal and Bybee, 2017).

Fennoscandian AMCG rapakivi intrusions range from small plutons to batholiths up to 18,000 km² (Laitakari et al., 1996; Salminen et al., 2021), with most emplaced between 1.8–1.0 Ga and peaking at 1.6–1.4 Ga during the breakup of Nuna (Columbia) (Fig. 1a) (Emslie, 1978; Sharkov, 2010; Pirajno and Santosh, 2015; Wang et al., 2020; Johansson et al., 2022). Seismic and potential-field data show that magmatic pathways were focused along horst-graben systems (e.g., Landsort, Satakunta; Fig. 1a) and listric faults that shaped intrusion geometry and subsurface architecture (Korja et al., 1993; Elo and Korja, 1993; BABEL Working Group, 1993a,b; Korja and Heikkinen, 1995; All et al., 2006). These structures coincide with crustal thinning and mafic underplating during Subjotnian diabase emplacement (~1.65–1.55 Ga) (Puura and Flodén, 1996, 2000; All et al., 2006), and in the absence of subduction signatures (Puura and Flodén, 1999, 2000), emplacement has been explained by cauldron subsidence (Selonen et al., 2005; Karell et al., 2009, 2014), strike-slip faulting (Branigan, 1987; Persson, 1999; Nordbäck et al., 2024), and/or post-orogenic thermal relaxation (Kukkonen and Lauri, 2009).

The classical AMCG magmatism model invokes continental rifting – marked by crustal thinning, listric faulting, and a shallow, undulating Moho – as the main trigger for AMCG magmatism (Emslie, 1978; Korja et al., 1993, 2001; Sharkov, 2010). Korja et al. (2001) attribute ~42 km Moho depths to aborted intracontinental listric faults. Yet Salminen et al. (2021) show that present-day crust is thickest beneath Baltica's Late Paleo- to Mesoproterozoic massif-type anorthosites (~55 km under 1.64–1.62 Ga Finnish rapakivi; ~42 km beneath 1.47–1.44 Ga Swedish batholiths), whereas coeval complexes in Laurentia and Amazonia overlie thinner crust (~38–41 km). Locally, Moho depths beneath Märijamaa and Kloostri are 56–66 km (Solano-Acosta et al., 2023) (Fig. 1f). Overall, there is no correlation between present-day Moho thickness and the global distribution of massif-type anorthosites and related magmatic rocks; this neither supports nor negates rift-based interpretations (Salminen et al., 2021).

However, a purely continental-rift origin is further challenged because: (1) extended and coeval sedimentary successions of major stretching are absent (Hoffman, 1989; Salminen et al., 2021), (2) magmatism persisted for ~500 Myr (Ashwal and Bybee, 2017), and (3) present-day thermal data do not support whole-lithosphere thinning (Kukkonen and Lauri, 2009). Hotspot models predicting age-progressive belts (e.g., Persson 1999) likewise lack support, as the spatial distribution of massif-type anorthosites shows no clear age progression among coeval batholiths, either between continents or

within a single craton (Ashwal and Bybee, 2017; Salminen et al., 2021; Mitchell et al., 2021). Instead, Paleomagnetic evidence places Baltica and Laurentia at low latitudes between ~1.65–1.45 Ga, consistent with plume trapping, thermal insulation, and superswell-driven asthenospheric upwelling (Hoffman, 1989; Trubitsyn et al., 2003; Pirajno and Santosh, 2015; Salminen et al., 2021). In Fennoscandia, these conditions promoted mafic underplating and crustal extension that reactivated Svecofennian orogeny (SO) structural fabrics, focusing magma ascent and AMCG emplacement (Branigan, 1987; Elo and Korja, 1993; Puura and Flodén, 1999, 2000; Bonin, 2007; Larin, 2009; Sharkov, 2010; Salminen et al., 2021; Johansson, 2023; Nordbäck et al., 2024), while favouring hybrid sub-continental lithospheric mantle (SCLM)-crustal melts (Bonin, 2007; Larin, 2009; Heinonen et al., 2010b; Pirajno and Santosh, 2015).

In the Baltic countries, rapakivi intrusions are largely buried beneath Phanerozoic cover; their distribution and architecture are inferred from seismic and potential-field data (Korja et al., 1993; Puura and Flodén, 1996, 1999, 2000; All et al., 2006; Bogdanova et al., 2006, 2015; Solano-Acosta et al., 2023). In Estonia, the Rapakivi suite includes the large Riga Batholith and smaller-older intrusions such as Märijamaa, Kloostri, Naissaare, Taebala, Neeme, and Ereda, with A-type stocks (e.g. Abja, Muhu, Virstu; Fig. 1b) (Klein et al., 1994; Kirs et al., 2004). These plutons are mainly pink, coarse monzogranite–syenogranite, locally cut by aplitic and microsyenitic dykes; although classic rapakivi texture is rare, geochemical and isotopic signatures match the broader Fennoscandian rapakivi suite (Soesoo and Niin, 1992; Klein et al., 1994; Rämö et al., 1996; Ahl, 1997; Rämö and Haapala, 2005).

Integrated geophysical approaches are well suited to constrain internal architecture and emplacement (Puura and Flodén, 2000; Karell et al., 2014; Geshi et al., 2021a), and potential-field datasets resolve subsurface geometry, density contrasts, and structural fabrics (Elo and Korja, 1993; Korja and Heikkinen, 1995; Ardestani et al., 2021; Solano-Acosta et al., 2023; Mitchell et al., 2023; Senger et al., 2025). Like the Märijamaa and Kloostri granites, many Fennoscandian rapakivi bodies are located over shear corridors, show circular outlines and/or inward-dipping listric faults consistent with cauldron–piston subsidence under low differential stress (Branigan, 1987; Elo and Korja, 1993; Ahl, 1997; Kosunen, 1999; Alviola et al., 1999; Korja et al., 2001; Selonen et al., 2005; All et al., 2006; Karell et al., 2009, 2014). Although the geochemistry of Estonian rapakivi intrusions is well documented (Soesoo, 1993; Klein et al., 1994; Puura and Flodén, 1996; Rämö et al., 1996; Raukas and Teedumäe, 1997; Kivisilla et al., 1999; Kirs et al., 2004), few studies integrate potential-field modelling for these bodies (e.g., All et al. 2004, Solano-Acosta et al. 2023) and those remain regional in scope; none provides a detailed morphological-structural analysis to reconstruct phase-by-phase emplacement.

This study investigates whether inherited structures, particularly the PPDZ, influenced the emplacement of the Märijamaa pluton and its Kloostri satellite. We utilize Bouguer and magnetic datasets over the granitic bodies, along with spectral decomposition, residual derivative filters, automated lineament mapping, and 3-D cross-gradient joint residual-potential inversion (SimPEG). These methods help to reconstruct the geometry and emplacement style of the Märijamaa–Kloostri intrusions and assess the influence of SO structures. Placed within the Mesoproterozoic AMCG framework, the results support a structure-controlled model in which crustal inheritance governs rapakivi granite emplacement in Fennoscandia.

2. Geological setting

2.1. Regional context

Baltica, located northeast of the Trans-European Suture Zone, consists of three Archean–Proterozoic proto-cratons: Fennoscandia, Volgo-Uralia, and Sarmatia (Fig. 1a). The amalgamation of Sarmatia and

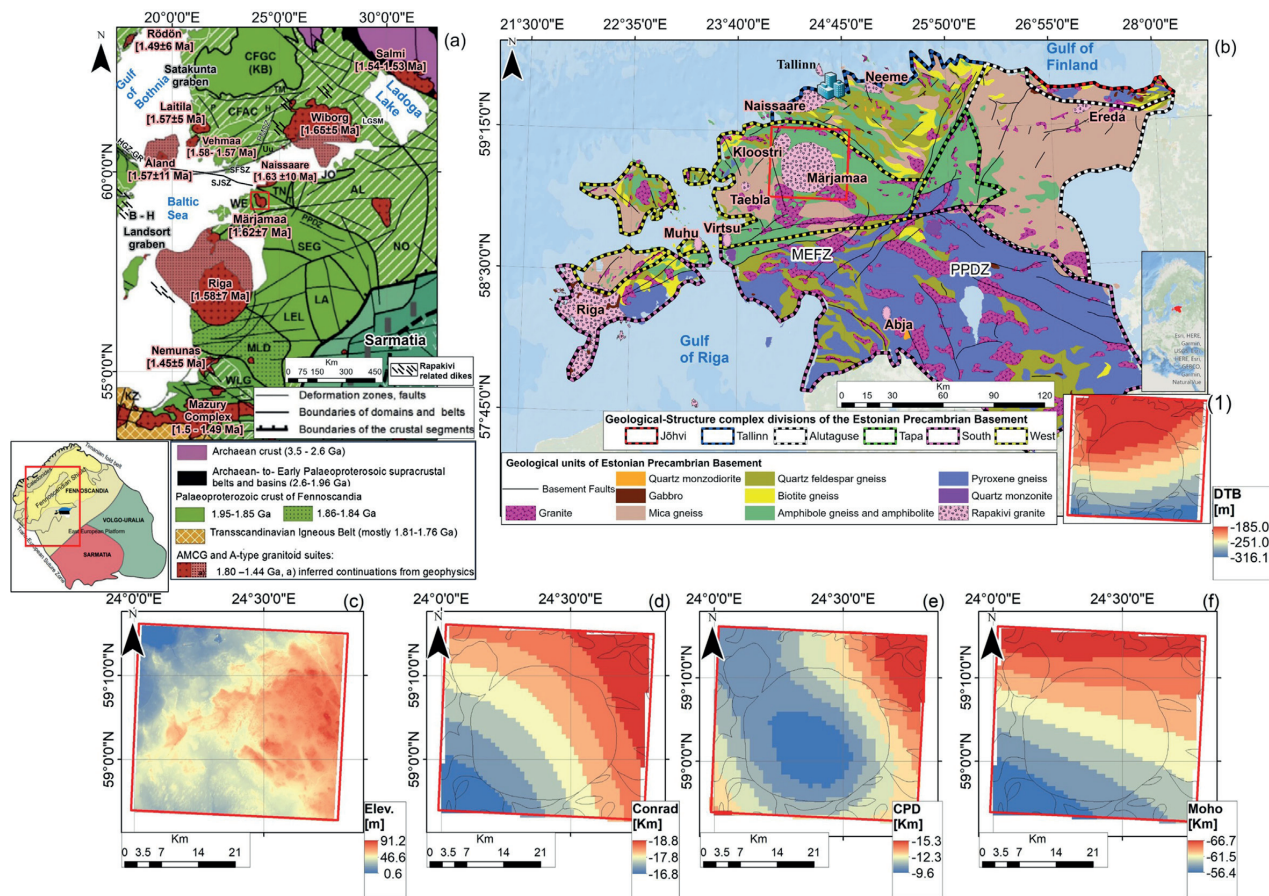


Fig. 1. Geological setting of the study zone. (a) Major Paleoproterozoic tectonic domains of Fennoscandia, modified after (Bogdanova et al., 2015), highlighting the most prominent rapakivi bodies, with the Marjamaa and Kloostri granitoids study area marked by a red rectangle. The white diagonal ruling over the greenish area indicates Svecofennian sedimentary basins. The abbreviations are AL — Alutaguse, B-H — Breven-Hällefors dike swarms, CFAS — Central Finland Arc Complex, CFGC — Central Finland Granitoid Complex, H — Häme belt, HGZ-GR — Hagsta-Gävrle-Rättvik Zone, JO — Jöhvi, KB — Keitele microcontinent, LA — Latgale, LEL — Latvian-East Lithuanian, MLD — Mid-Lithuanian domain, P — Pirkanmaa belt, SEG — South Estonian granulite domain, T — Tapa, TN — Tallinn, TM — Tampere belt, Uu — Uusimaa belt, WE — West Estonian domain, WL — West Lithuanian granulite domain. Shear and fault zones are abbreviated as follows: MEFZ — Middle Estonian Fault Zone, PMSZ — Porkkala-Mäntsälä Shear Zone, PPDZ — Paldiski-Pskov Deformation Zone, SFSZ — South Finland Shear Zone, and SJSZ — Sottunga-Jurmo Shear Zone. Age rapakivi granitoids of Fennoscandia are constrained from (Laitakari et al., 1996; Ahl, 1997; Rämö et al., 1996; Skridlaite et al., 2007; Larin, 2009; Sharkov, 2010; Johansson et al., 2016; Grabarczyk et al., 2023). (b) Precambrian geological map highlighting the Estonia basement, showing metalliferous anomalies based on (Soesoo et al., 2020). The upper left-corner inset shows the distribution of granulite- and amphibolite facies metamorphic rocks after (Bogdanova et al., 2015). Geological data are available from the Land Board Geoportal of the Republic of Estonia (<https://xgis.maaamet.ee/xgis2/page/app/geoloogia400k>), including (1) Depth to Basement (DTB) measurements from the same source. (c) Elevation data is derived from the 10-meter resolution GeoTIFF obtained from: <https://geoportaal.maaamet.ee/eng/Maps-and-Data/Elevation-data/Download-Elevation-Data-p664.html>. (d–f) Geophysical measurements from the studied zone updated from (Solano-Acosta et al., 2023): (d) Conrad discontinuity, (e) Curie Point Depth (CPD), and (f) Moho discontinuity.

Volgo-Uralia at 2.1 Ga (Terentiev and Santosh, 2020; Wang et al., 2020) initiated a prolonged phase of crustal accretion, arc magmatism, and basin closure in Fennoscandia between 2.1 and 1.79 Ga, during the SO. This orogeny stabilized the crust through arc accretion, deformation, and granitoid intrusions across southern Finland, central Sweden, and northern Estonia, involving the Bergslagen microcontinent, and extended to the southern Baltic countries with the Livonia microcontinent, culminating in continent–continent collision with Sarmatia during the Sveco Baltic phase (1.84–1.79 Ga) (Gorbatschev and Bogdanova, 1993; Nironen, 1997; Nironen et al., 2000; Nironen, 2017; Bogdanova et al., 2006, 2015; Lahtinen et al., 2005; Johansson et al., 2022; Solano-Acosta et al., 2025). By 1.77 Ga, crustal stabilization formed the core of Baltica. Final amalgamation of Sarmatia and Fennoscandia occurred by ~1.70 Ga, followed by the Gothian orogeny (1.73–1.48 Ga), which added juvenile terranes along Baltica's western margin (Bogdanova et al., 2008, 2015; Nironen, 2017). The Fennoscandian Shield comprises the Archean Kola and Karelia cratons, bounded by Paleoproterozoic mobile belts (Korsman et al., 1999; Korhonen et al., 2002; Lahtinen et al., 2005).

Fennoscandian rapakivi granites are divided into four AMCG age groups: 1.67–1.62, 1.59–1.56, 1.55–1.53, and 1.53–1.44 Ga (Rämö et al., 1996; Ahl, 1997; Puura and Flodén, 2000; Salminen et al., 2021; Grabarczyk et al., 2023). The Wiborg Suite (1.67–1.62 Ga) includes Wiborg, Bodom and Obbnäs, Naissaare, Märjamaa, and Kloostri bodies (Klein et al., 1994; Rämö et al., 1996; Kosunen, 1999; Puura and Flodén, 2000). The Åland Suite (1.59–1.56 Ga) includes Riga, Åland, Laitila, and Vehmaa (Suominen, 1991; Selonen et al., 2005). The Salmi Suite (1.55–1.53 Ga) is located at the Archean–Proterozoic boundary (Larin, 2009; Sharkov, 2010). The Ragunda Suite (1.53–1.47 Ga) includes intrusions in Sweden (Rödon), Lithuania (Nemunas), and Poland (Mazury complex) (Persson, 1999; Dörr et al., 2002; Duchesne et al., 2010; Grabarczyk et al., 2023). These phases coincide with fault-bounded basins (Puura and Flodén, 1996; Korja et al., 2001; Bogdanova et al., 2015; Salminen et al., 2021; Grabarczyk et al., 2023).

2.2. Local context

The geological framework of Estonia is defined by a Paleoproterozoic to Mesoproterozoic crystalline basement, unconformably overlain by Ediacaran–Palaeozoic sedimentary strata and a Quaternary cover (Fig. 1b) (Raukas and Teedumäe, 1997). Following Mesoproterozoic peneplanation and late Proterozoic denudation, the sedimentary succession dips gently southward (~0.10–0.20°, or 2–3.5 m/km), thickening from approximately 100 m in the north to over 900 m in the south (Kirs et al., 2004, 2009; Soesoo et al., 2004, 2020; Soosalu et al., 2022) (Fig. 1b1). The Precambrian basement is known from drill cores, potential field anomalies, and correlations with better-exposed terranes in Finland and Sweden (Raukas and Teedumäe, 1997; Puura et al., 2004; Bogdanova et al., 2015).

Six structural–petrological zones are recognized in northern Estonia and Latvia: Tallinn, Alutaguse, Jõhvi, West-Estonian, Tapa, and South-Estonian (Fig. 1b). These zones are delineated by lithological, geochemical, magnetic, and gravity contrasts (All et al., 2004; Bogdanova et al., 2015; Soesoo et al., 2020; Solano-Acosta et al., 2023). The PPDZ, a NW–SE-trending crustal-scale dextral shear zone approximately 30 km wide, separates the northern and eastern blocks from the western and southern domains. The Middle Estonian Fault Zone (MEFZ), which trends E–W and was reactivated during Mesoproterozoic extension, juxtaposes basement blocks of granulite- and amphibolite-facies metamorphism (Soesoo et al., 2004; Puura et al., 2004).

The Tallinn and Alutaguse domains in northeastern Estonia are characterized by amphibolite-facies metamorphism (3–5 kbar) and correlate with the Uusimaa belt in southern Finland and the Swedish Bergslagen region (Bogdanova et al., 2015; Soesoo et al., 2020). The Tallinn domain is interpreted as a continuation of the Uusimaa volcanic

arc, whereas the Alutaguse domain represents the corresponding back-arc basin (All et al., 2004), Soesoo et al. (2004), Kukkonen and Lauri (2009), Solano-Acosta et al. (2025). These domains later accreted against northern belts such as Häme (~1.89 Ga), Pirkkanmaa (~1.90 Ga), and Tampere (~1.92 Ga) within a double subduction system and arc amalgamation along proto-Fennoscandia and the Bergslagen microcontinent, culminating in Paleo-Svecofennian ocean closure and block collision at ~1.87 Ga (Kukkonen and Lauri, 2009; Nironen, 2017; Bogdanova et al., 2015; Solano-Acosta et al., 2025).

The West Estonian domain, bounded by the PPDZ and MEFZ, consists of high-grade metasedimentary and metaigneous rocks with granulite-facies overprint (Puura et al., 2004; Soesoo et al., 2004, 2006). The South Estonian domain, delineated by NW–SE-trending potential field anomalies, contains high-grade orthogneisses and mafic-felsic granulites (Soesoo et al., 2020; Solano-Acosta et al., 2023).

Regional geophysical patterns across the Baltic Sea – especially NW-aligned gravity and magnetic anomalies – suggest that Estonian basement domains are continuations of the oblique Svecofennian crustal blocks in southern Finland, Sweden, and Latvia (Korhonen et al., 2002; Bogdanova et al., 2006, 2015; Korja and Kosonen, 2015). The NW–SE lithological grain reflects inherited shear zones such as the PPDZ and its Finnish continuation, the Sottunga–Jummo Shear Zone (SJSZ), which further links to the Hagsta-Gävle-Rättvik Zone (HGZ–GRZ) zone in Bergslagen (Bogdanova et al., 2006, 2015; Högdahl et al., 2009; Soosalu et al., 2022; Nordbäck et al., 2024) (Fig. 1a). These structures, active during the Mesoproterozoic, control present-day seismicity, dominated by strike-slip mechanisms along reactivated basement faults (Wüestefeld, 2007; Soosalu et al., 2022).

U–Pb zircon dating divides the Estonian basement into three main age groups. The oldest (1.92–1.80 Ga) corresponds to the Fennoscandian basement, with ages of 1.92 ± 10 Ga for North Estonian metavolcanics (Petersell and Levchenkov, 1994; Soesoo et al., 2004; Kirs et al., 2009; Bogdanova et al., 2015), 1.83 ± 22 and 1.82 ± 7 Ga for southern granulites (Petersell and Levchenkov, 1994; Soesoo et al., 2004), and 1.82 ± 26 Ga for Tapa tonalites (Soesoo et al., 2006). Jõhvi belt magnetite gneisses yield 1.87 ± 18, 1.83 ± 10, and 1.79 ± 19 Ga (Soesoo et al., 2020). Tonalites in southern Estonia date to 1.79 ± 16 Ga; charnockites and garnet gneisses to 1.76 ± 11 and 1.78 ± 2 Ga (Soesoo et al., 2004, 2006); and a gabbro-norite dyke to 1.77 ± 20 Ga (Soesoo et al., 2006). The third group (1.64–1.58 Ga) includes A-type granitoids: Abja (1.64 ± 7 Ga) (Kirs and Petersell, 1994), Virtsu (1.61 ± 17 Ga) (Soesoo and Hade, 2012), and rapakivi intrusions at Märjamaa (1.63 ± 7 Ga), Naissaare (1.62 ± 7 Ga), and Riga (1.58 ± 2 Ga) (Rämö et al., 1996).

These Paleo to Mesoproterozoic rapakivi intrusions in Estonia include the Riga batholith and smaller granites such as Märjamaa–Kloostri, Naissaare, Taebala, Neeme, and Ereda (Fig. 1b) (Soesoo and Niin, 1992; Kirs et al., 2004). Their emplacement has been related to crustal-scale faults such as the PPDZ and MEFZ, reactivated under transtensional regimes during AMCG magmatism (Raukas and Teedumäe, 1997; Puura and Flodén, 1999, 2000; All et al., 2006; Bogdanova et al., 2015). These plutons are completely buried, with their geometry and zonation constrained by Bouguer and magnetic data (Solano-Acosta et al., 2023).

The Märjamaa pluton is a 40 × 25 km composite intrusion with three internal phases: Phase I comprises xenolith-rich melanocratic granodiorites; Phase II consists of biotite–hornblende granites; and Phase III, corresponding to the Kloostri satellite, is a leucocratic granite with trachytic textures (Soesoo and Niin, 1992; Klein et al., 1994; Laitakari et al., 1996). Petrographically (Soesoo and Niin, 1992; Raukas and Teedumäe, 1997), two generations of anhedral quartz (20%–30%) are present: one in potassium feldspar and microcline, the other in the groundmass. Microcline (20%–40%) forms large perthitic phenocrysts (2–3 cm) with quartz, biotite, and rare titanite inclusions. Plagioclase (20%–40%) is euhedral oligoclase-andesine. Biotite (2%–10%) forms anhedral clusters, while hornblende appears in Phase I and II, and muscovite is in Phase III. Magnetite and ilmenite were reported (3%–5%), with accessories like apatite, fluorite, zircon, titanite, and epidote.

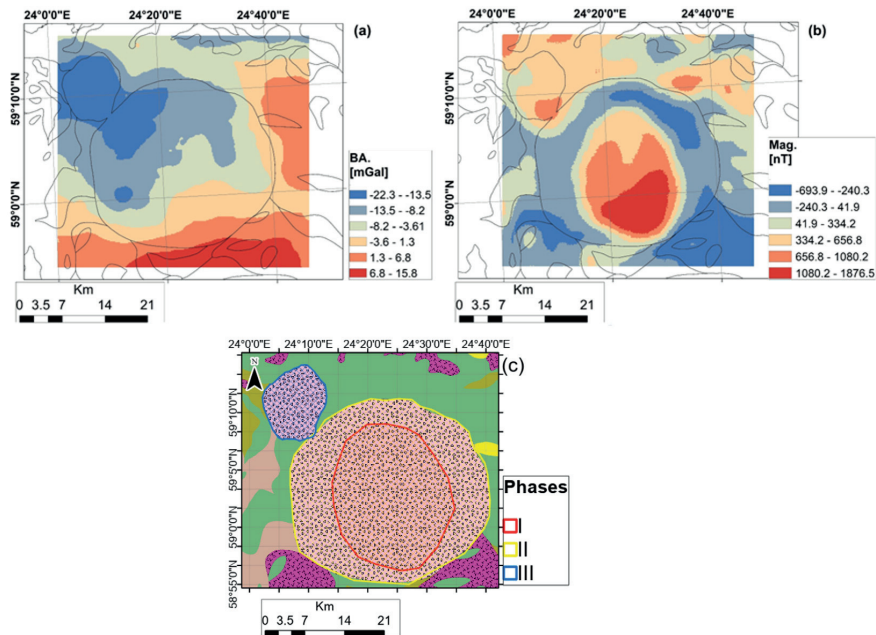


Fig. 2. Geophysical datasets of the Märjamaa pluton and its Kloostri satellite. (a) Bouguer anomaly (BA), and (b) magnetic anomaly, both Estonian Geological Service (EGT), and (c) Granitoid phase distribution map of the Märjamaa pluton and its Kloostri satellite (after Klein et al. (1994)).

Solano-Acosta et al. (2023) present a potential field model for the Estonian basement, showing across Märjamaa–Kloostri a deepening of the Conrad discontinuity from 16 to 18 km from SW to NE (Fig. 1e), Curie Point Depth (CPD) of approximately 10 km (Fig. 1f), and an increase in Moho depth from 56 to 66 km (Fig. 1g).

3. Geophysical data

To examine the geophysical features of the Märjamaa pluton and Kloostri satellite, potential datasets from the Estonian Geological Service (EGT) were used, including Bouguer (Fig. 2a) and magnetic anomaly data (Fig. 2b). The data were clipped with a 45×45 km² grid for accurate anomaly values. The raw potential-field datasets are accessible via the EGT web platform (<https://gis.egt.ee/portal/apps/webappviewer/index.html?id=e808917692cd4ba89a85ff8ffe17816f>).

The Bouguer anomaly reveals subsurface density variations, helping identify geological structures, crustal thickness, and rock types based on density (Oja et al., 2021; Osagie et al., 2021; Solano-Acosta et al., 2023; Senger et al., 2025). Gravimetric data, compiled by various institutions since the mid-20th-century, were consolidated by the Estonian Land Board and Geological Survey, covering 74,480 km² with 10,882 measurement points at an average density of 0.15 points/km² (Maasik, 1959; Jürgenson, 2001, 2003; All et al., 2004; All and Gromov, 2005; Oja et al., 2019, 2021; Geological Survey of Estonia (EGT), 2025) (Fig. 2a).

Magnetometric analyses are crucial for detecting magnetic mineral concentration variations, offering insights into crustal structure and composition (Omietimi et al., 2021; Saada et al., 2021; Solano-Acosta et al., 2023). In Estonia, aeromagnetic surveys were carried out during the 20th century, with measurements at 1 : 25,000 scale on land and 1 : 50,000 offshore (1987–1992), using a 250 m flight-line spacing and a 100 m flight altitude (All et al., 2004; Soesoo et al., 2020; Geological Survey of Estonia (EGT), 2025). Data were processed to the *DGRF* – 65 anomaly level (Korhonen et al., 2002), with a resolution of 1×1 km² (All et al., 2004) (Fig. 2b). For more information about

EGT potential datasets direct to: <https://www.egt.ee/maapouealane-teave/geoloogilised-andmed/ruumiandmed-ja-kaardid#andmed>.

4. Methodology

4.1. Gravity and magnetic preprocessing

The EGT potential-field datasets were processed in the Universal Transverse Mercator (UTM) coordinate system, Zone 34N of the Northern Hemisphere (UTM 34°N). The data were subsequently separated into regional and residual components using MAGMAP in Geosoft Oasis Montaj, allowing the isolation of long- and short-wavelength anomalies (Mickus et al., 1991; Lowrie and Fichtner, 2020; Solano-Acosta et al., 2023). Fourier transformation, wavelength filtering, and phase matching generated grids highlighting crustal density and magnetic susceptibility (Geosoft, 2004; Osagie et al., 2021; Saada et al., 2021).

A prior reduction-to-pole (RTP) transformation was performed on the complete magnetic data (Fig. 2b) to vertically align the peaks above the sources (Baranov and Naudy, 1964; Ganiyu et al., 2013). This transformation simulates the distribution of anomalies as if they were at the geomagnetic pole, using IGRF-derived values for inclination (72.992°) and declination (9.679°) for the centric 59.05°N , 24.36°E reference point (Finlay et al., 2010; Omietimi et al., 2021; Solano-Acosta et al., 2023). RTP result is shown in Fig. 3c1.

Wavelength filtering with high- and low-pass filters segmented complete Bouguer and RTP spectral components. The radial average power spectrum (Fig.S1) identified a critical wavenumber of 0.05 cycles/km, for both potential datasets, where lower wavenumbers relate to deeper sources (Figs. 3a1,3c2) and higher ones capture shallower features (Figs. 3a2,3c3) (Phillips, 2001).

A pseudo-gravity (PSG) transformation was applied to the RTP field data set (Fig. 3a3). This method uses fast Fourier transform (FFT) and Poisson's relation to convert magnetic to gravity anomalies (Pratt and Shi, 2004; Geosoft, 2004; Solano-Acosta et al., 2023). All potential procedures described were done within Geosoft Oasis Montaj software.

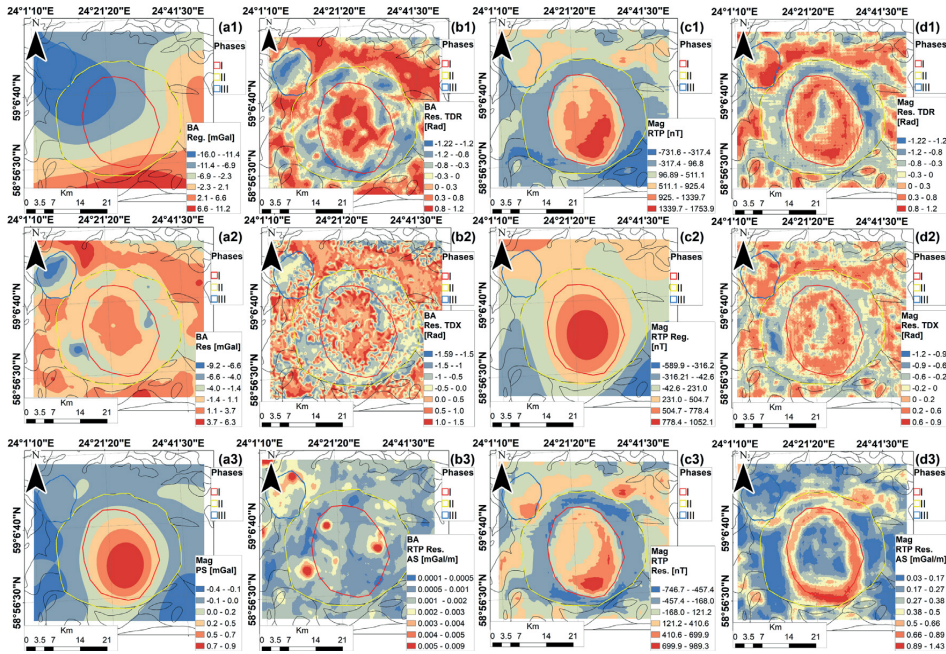


Fig. 3. Geophysical regional and residual wave-cut, and residual derivative results. (a) Bouguer maps: (a1) regional component, (a2) residual anomalies (cut at 0.05 cycles/km), (a3) pseudo-gravity transformation derived from the RTP magnetic map. (b) Bouguer residual derivatives: (b1) Tilt Derivative Ratio (TDR), (b2) Tilt Angle Derivative (TDX), (b3) Analytic Signal (AS). (c) Magnetic Reduce-To-Pole (RTP) maps: (c1) RTP magnetic anomaly, (c2) regional field, (c3) residual anomaly. (d) RTP residual derivatives: (d1) TDR, (d2) TDX, (d3) AS.

4.2. Derivative operators (TDR, TDX, AS)

To enhance structural interpretation, derivative filters were applied to the residual Bouguer and RTP datasets (Saada et al., 2021; Solano-Acosta et al., 2023). These filters are commonly used to highlight subsurface boundaries and potential structural discontinuities (Geosoft, 2004; Osagie et al., 2021; Ibraheem et al., 2018; Ndikum and Tabod, 2024). The derivative filters used here are expressed as:

Tilt Derivative Ratio (TDR):

$$\tan^{-1} \left(\frac{\partial f / \partial z}{\sqrt{(\partial f / \partial x)^2 + (\partial f / \partial y)^2}} \right)$$

Tilt Angle Derivative (TDX):

$$\tan^{-1} \left(\frac{\sqrt{(\partial f / \partial x)^2 + (\partial f / \partial y)^2}}{\partial f / \partial z} \right)$$

Analytic Signal (AS):

$$\sqrt{\left(\frac{\partial f}{\partial x} \right)^2 + \left(\frac{\partial f}{\partial y} \right)^2 + \left(\frac{\partial f}{\partial z} \right)^2}$$

Where, $\partial f / \partial x$, $\partial f / \partial y$, and $\partial f / \partial z$ represent the first derivatives of the potential field in the x , y , and z directions, respectively. TDR and TDX enhance density and magnetization contrasts, while AS emphasizes source geometry independently of any potential inclination (Geosoft, 2004; Solano-Acosta et al., 2023). Residual derivatives are shown in Figs. 3b1–b3, 3d1–d3.

4.3. Lineament extraction, length–density and orientation

Geophysical lineaments were extracted from residual TDR, TDX, and AS grids using hillshade illumination. For each derivative raster, eight Hillshade-azimuth rasters (0° , 45° , 90° , 135° , 180° , 225° , 270° ,

315° ; solar elevation 45°) were generated and further merged into two-composite rasters: Combination 1 (0° – 135°) and Combination 2 (180° – 315°) (e.g., Radaideh et al. (2016), Solano-Acosta et al. (2023)). Vectors from the hillshade composite rasters were generated with LINE module (PCI-Geomatica 2012) and analysed in ArcGIS 10.8 (COGO) module to derive azimuth, length and linear-density maps, and Microsoft Excel to plot the rose diagrams (Figs. 4a–c).

For kinematic context, basement faults and derived lineaments were categorized by orientation relative to the dextral PPDZ maximum horizontal stress ($\sigma_H = 115^\circ$ – 135° , 295° – 315°) in central Fennoscandia, over Estonia (Korja and Kosonen, 2015; Soosalu et al., 2022; Solano-Acosta et al., 2023). Lineaments were grouped into expected Riedel families for a dextral regime: R (125° – 145° , 305° – 325°), R' (20° – 40° , 200° – 220°), P (100° – 120° , 280° – 300°), Tension (155° – 175° , 335° – 355°), and Extension (σ_3) (65° – 85° , 245° – 265°) (Xu et al., 2013; Radaideh et al., 2016; Duan et al., 2024) (Figs. 4d–g). Many mapped lineaments coincide with Riedel structures, whereas non-Riedel lineaments – classified here as “other” – likely mark intrusive contacts, magmatic layering, alteration fronts, or rheological contrasts (Nabighian, 1972; Salem et al., 2007; Radaideh et al., 2016; Saada et al., 2021).

Linear and joint-density distributions (Figs. 4c–e) were computed for each residual derivative-based potential-field dataset. A composite joint high-density map was then compared with the Estonian tectonic framework, and regional heat-flow and lineament-density patterns from (Solano-Acosta et al., 2023), together with earthquake epicentres and focal mechanisms from Soosalu et al. (2022), were overlain for reference (Fig. 4g).

4.4. Upward continuation (qualitative depth filtering)

The upward continuation (UC) method was applied to residual Bouguer and RTP datasets to estimate theoretical anomaly depths,

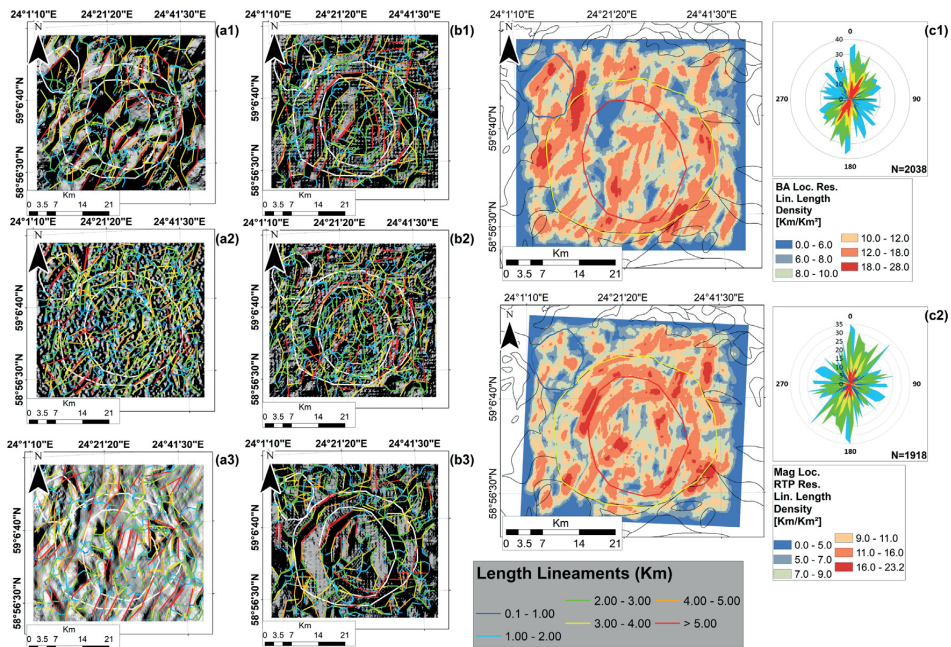


Fig. 4. Lineament analysis of the potential residual data. (a) Bouguer derivative maps: 1 TDR, 2 TDX, 3 AS. (b) RTP derivative maps: 1 TDR, 2 TDX, 3 AS. (c) Length density maps and rose diagrams for Bouguer (1) and RTP (2). (d–e) Residual lineaments organized by derivative type: Bouguer (d1–d3) and RTP (e1–e3), with their respective joint-density map in the right-bottom part. (f) Merged lineaments classified by Riedel shear types with rose diagrams for Bouguer (1) and RTP (2). (g) Estonian rapakivi bodies overlaid on zones with the highest High Flow values (in red) and the highest density of tectonic features derived from regional potential lineaments and basement faults (in grey), as updated from (Solano-Acosta et al., 2023). The Märrjamaa and Kloostri granitoid phases were plotted together with the merged high-density lineament-joint field compiled from all residual-derivative datasets. Basement faults and lineaments were categorized based on their orientation relative to the dextral Åland-Paldiski-Pskov Deformation Zone (PPDZ) maximum horizontal stress (σ_H) of 115–135°, 295–315° in central Fennoscandia (Korja and Heikkinen, 1995; Soosalu et al., 2022). The angle relative to σ_H and the expected mode of movement were represented as follows: blue indicates transfer (P) and normal (R), purple indicates reverse (R'), green indicates tensional sinistral strike-slip (T), and red indicates extensional dextral strike-slip (σ_s). The epicentres of earthquakes analysed by Soosalu et al. (2022) are shown as black dots labelled with a number (code), each with its respective focal mechanism, highlighting the oblique shearing arrangement of the Åland-PPDZ corridor.

attenuating near-surface noise and enhancing deeper features (Blakely, 1995; Saibi et al., 2006; Ganiyu et al., 2013; Okpoli and Akingboye, 2019). The UC formula is:

$$F_0(p, q) = e^{-\zeta \sqrt{p^2 + q^2}} F_\zeta(p, q)$$

Where $F_0(p, q)$ is the surface potential field, ζ is the continuation height, and $F_\zeta(p, q)$ is the upward-continued transform.

Following Jacobsen (1987), the continuation height at which the anomaly fades provides an estimate of the depth extent of the measurable signal. In this study the method was applied specifically to the residual Bouguer and RTP datasets, so the inferred depths represent the vertical range over which the residual density and magnetic susceptibility contrasts remain detectable, rather than the absolute maximum depth of the causative bodies. Heights from 500 to 9500 m were analysed (Figs. 5a1–a10, b1–b10), and anomalies in both residual datasets become negligible around 10000 m (Fig.S1). This attenuation trend indicates that spectral energy decreases with depth, reflecting reduced sensitivity of residual anomalies to density and magnetization contrasts. Nonetheless, the depth at which these anomalies vanish can provide a useful estimate of the causative body's extent.

4.5. SimPEG inversion

To enhance the geometry of the analysed granitoids, a cross-gradient joint inversion of residual gravity and RTP magnetic data using SimPEG python module was done (<https://simpeg.xyz/user-tutorials/>; Cockett et al. (2015)). This method is effective in modelling igneous and hydrothermal systems (Miller et al., 2017; Ardestani et al., 2021; Ardestani, 2022; Mitchell et al., 2023).

The inversion was carried out on the residual potential datasets (Figs. 3a2,3c3), within a 3-D tensor mesh with 70,200 cells, extending to –10,000 m depth, with 54 cells in the x-direction (Extension: 46,240 m; Cell-width: 850 to 1020 m), 50 cells in the y-direction (Extension: 42,840 m; Cell-width: 850 to 1020 m), and 26 cells in the z-direction (400 m thickness). A total of 67,500 active cells (96.1%) were defined below the topography (Fig. 1c), covering the 10 km anomaly depth akin to the UC results (Fig. 5).

Thirty vertical cross-sections – 15 along each axis (X_{1-15} and Y_{1-15}) – were generated to map the distributions of density and magnetic susceptibility, respectively (Fig. Figs. 6a–b). Residual datasets were inverted using cross-gradient joint inversion (SimPEG; <https://simpeg.xyz/user-tutorials/plot-inv-3-cross-gradient-pf>). The resulting profiles (Figs. 6[c–f]) delineate the three magmatic phases and their contrasting internal structures. Background values were set to 1×10^{-3} g/cc for

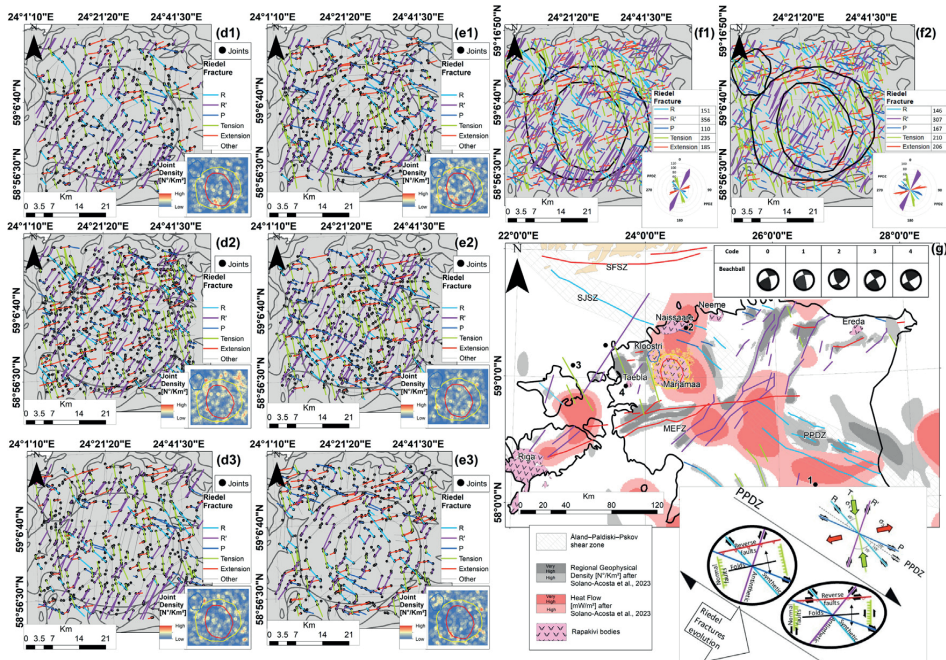


Fig. 4. (continued).

density and 1×10^{-3} SI for magnetic susceptibility. Figure S1 shows the mesh grid with observed, predicted, and misfit responses.

5. Results

5.1. Gravity and magnetic processing

The RTP magnetic map (Fig. 3c1) shows a narrow composite field, defining the regional fabric, whereas anomalies correlate with the mapped granitoid tripartite phases. Spectral separation isolates the long-wavelength regional component (Figs. 3a1,c2) and enhances short-wavelength residuals that sharpen the plutonic phases' signatures (Figs. 3a2,c3). The pseudo-gravity RTP transform highlights the lateral density contrasts (Fig. 3a3), illustrating the boundaries of Märijämsa Phase I, which is the most ferrous and dense (Soesoo and Niin, 1992; Soesoo, 1993; Klein et al., 1994). Residual Bouguer and RTP maps (Figs. 3a2,c3) define three concentric domains matching Phases I–III: Phase I shows coincident gravity and magnetic highs; Phase II yields paired lows; and Phase III produces a negative Bouguer anomaly with a local magnetic high on the NW margin.

5.2. Derivative and tectonic-lineament analysis

Residual derivatives of Bouguer and RTP data (Figs. 3b1–b3,3d1–d3) delineate the internal geometry of the Märijämsa-Kloostri granitoids. TDR and TDX show concentric-radial patterns that define tripartite magmatic zoning and ring-like structures (Figs. 3b1–b2,3d1–d2). AS sharpens gradients: in the residual Bouguer field it locally resolves narrow, high-amplitude lineations consistent with possible dykes (Fig. 3b3), and the residual RTP it accentuates rims around Phase I and the northern Phase II–host contact (Fig. 3d3). AS responses are depth dependent (Nabighian, 1972), where shallow sources yield edge-aligned maxima, whereas greater burial shifts maxima towards anomaly centres — explaining rim-focused signals at Phase I.

Automated lineament extraction from TDR, TDX, and AS grids yields dense clusters around the Märijämsa phases and the Kloostri body; length-density maxima occur along Phase II margins and in the NW sector (Figs. 4a–c). Rose diagrams define two dominant sets: short lineaments (0–2 km) trending NW–SE to NNE–SSW, and longer lineaments (> 2 km) trending NE–SW (Figs. 4c1,c2). In a dextral Riedel framework, distinct R, R', P, Tensional, and extensional (σ_3) sets are recognized, with joint-density peaks near Phase II boundaries and within the PPDZ σ_H -parallel corridor (Figs. 4d–g), identify as the principal fault structure (Korja and Kosonen, 2015; Soosalu et al., 2022; Solano-Acosta et al., 2023), where the dominance of antithetic R' fractures followed by tensile and extensional elements consistent with pull-apart shearing (Figs. 4f–g) (Sylvester, 1988; van Wyk de Vries and Merle, 1998; Xu et al., 2013; Radaideh et al., 2016; Siachoque et al., 2017). Within Phase II, the RTP residual map shows a fabric akin to a porphyroclastic dextral flow shape, pointing to a ductile-to-brittle overprint (Fig. 3c3).

5.3. Upward continuation (UC) behaviour

UC of the residual Bouguer and RTP fields from 0.5–9.5 km (Fig. 5a–b,S1) shows progressive attenuation of short-wavelength content and persistence of long-wavelength anomalies. At 0.5–1.5 km, Phase I produces pronounced residual highs in both gravity and magnetics, whereas Phases II–III are lower amplitude; by 2.5–5.5 km, the Phase I signal broadens and weakens and Phases II–III merge towards background; by 6.5–9.5 km, Phase I contrasts only weakly as more of the area approaches background.

Consistent with the RTP pseudo-gravity transform (Fig. 3a3) and the RTP AS (Fig. 3d3), UC anomaly maxima shift towards anomaly centres, yielding more centre-weighted responses at larger continuation heights. Importantly, the 10 km continuation at which residual contrasts remain detectable reflects UC persistence (signal detectability) and does not equate to the top or base depth of the sources; disappearance with

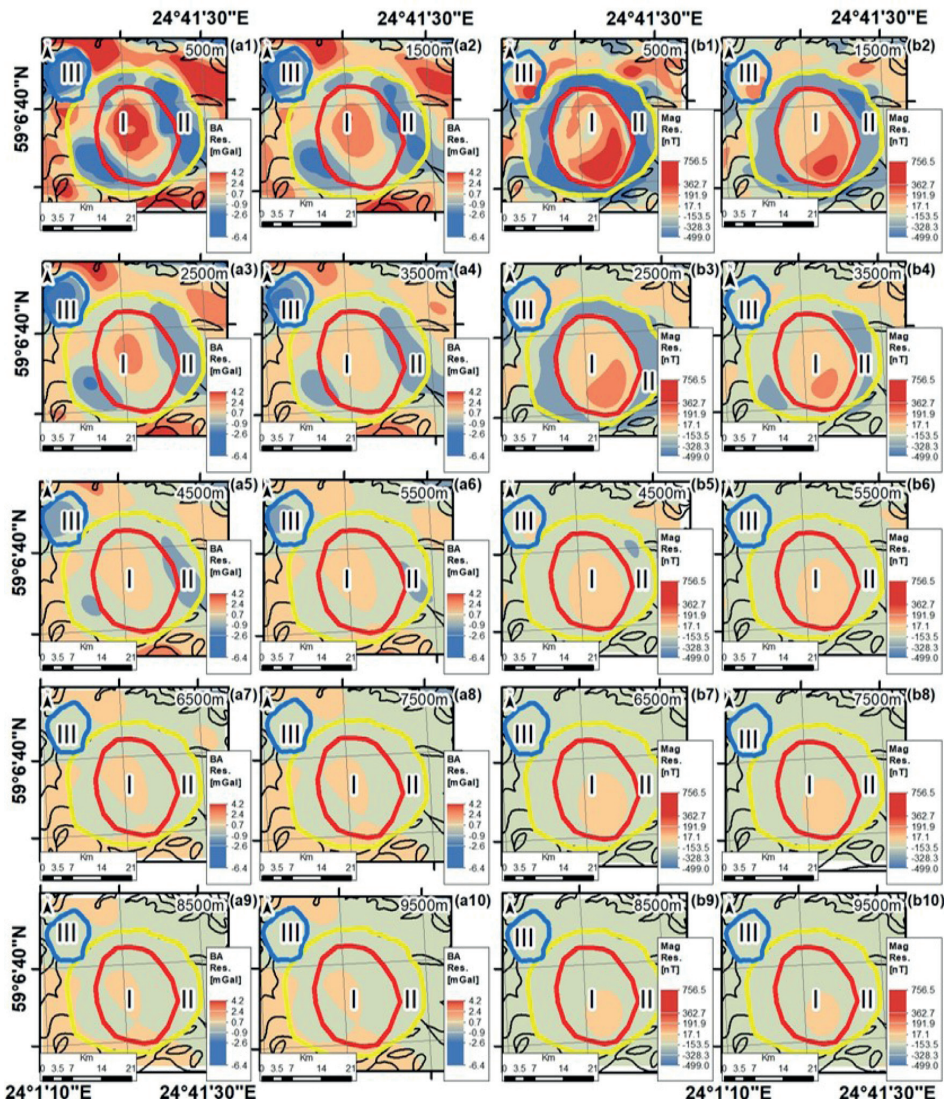


Fig. 5. Residual potential theoretical depths obtained from the Upper continuation (UC) method for the residual potential (a) Bouguer and (b) RTP data from (1–10) 500 m to 9500 m.

upward continuation can also arise from reduced magnetization/density contrast, interference between adjacent anomalies, finite lateral extent, or remanent directions that partially cancel the field (Jacobsen, 1987; Ganiyu et al., 2013). CPD estimates – where magnetite loses remanence near $\sim 580^\circ\text{C}$ – from regional work (Fig. 1e; Solano-Acosta et al. 2023) reach up to ~ 9.6 km beneath the study area, providing a thermal ceiling for magnetic sources that is consistent in order of magnitude with the UC persistence heights of the residual magnetic anomalies. This analysis therefore uses UC only to bracket the effective depth of analyzable signal in the residual component, not to infer absolute source depth.

5.4. 3-D joint gravity–magnetic inversion

The 10 km-depth SimPEG cross-gradient inversion of residual Bouguer (Figs. 6c,e) and RTP magnetic data (Figs. 6d,f) delineates three geophysically distinct zones corresponding to the identified magmatic phases. Phase I is characterized by a negative density anomaly and high magnetic susceptibility, while Phase II exhibits a positive density contrast and lower susceptibility. Phase III shows positive anomalies in both datasets. The cross sections X_{6-8} and Y_{5-8} define a central Phase I body flanked by Phase II on either side, with sharper contrasts at the top of Phase I and increased susceptibility at depth. Sections X_8 and Y_8 display a focused density high and a conical magnetic structure,

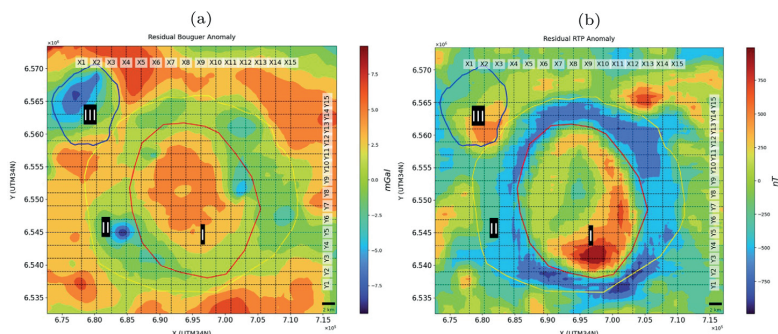


Fig. 6. Residual Bouguer and RTP SimPEG cross-gradient joint inversion results for the study area, displayed in 3×5 grids. (a) Bouguer data and (b) RTP data. The inversion models include fifteen slices per XY direction, presented as follows: (c) Bouguer data (X direction), (d) RTP data (X direction), (e) Bouguer data (Y direction), and (f) RTP data (Y direction), providing a detailed representation of spatial density and magnetic susceptibility variations, highlighting the Märijamaa pluton and its Kloostris satellite tripartite phases with interpretative dashed grey lines.

suggestive of localized downward collapse. At Phase I–II boundaries ($X_{4-5,11-12}$; $Y_{2-3,11-12}$), linear and cylindrical features may represent intrusive conduits, dike-like bodies, or localized heterogeneities. Phase III appears in X_{1-3} and Y_{13-15} as a smaller body with a similar vertical susceptibility profile to Phase I. The 10 km to-depth inversion reproduces the principal features of the residual fields and indicates that the dominant density-susceptibility contrasts, in agreement with the attenuation trends from UC and the CPD constraints. To obtain deeper insights into the base depth of the body and its interaction with host rocks, it is necessary to analyse the system in a larger scale using potential-field datasets and/or seismic profiles.

6. Discussion

The Märijamaa–Kloostris granitic complex provides insights into A-type granite emplacement within the Fennoscandian Shield. Although compositionally diverse, A-type granitoids share petrogenetic features controlled by local P – T conditions and tectonic setting (Bonin, 2007; Sharkov, 2010; Salminen et al., 2021; Johansson, 2023). In Fennoscandia, rapakivi intrusions formed at varying depths, recording a multistage evolution from melt generation to emplacement (Vigneresse, 2005; Haapala and Lukkari, 2005; Heinonen et al., 2010b; Karell et al., 2014). At Märijamaa–Kloostris system, structural and geophysical evidence suggests a transtensional-related emplacement (Fig. 4), with no preserved volcanic sequences reported (Rämö et al., 1996; Puura and Flodén, 1996; Kirs et al., 2004). Here we synthesize competing A-type granite emplacement models and test them against integrated gravity-magnetic and lineament data from the Märijamaa–Kloostris system.

Three schematic frameworks have been proposed to interpret A-type magmatism and emplacement. Bonin (2007) proposed a vertically zoned magmatic system in which ultradeep mantle-derived melts (30–40 km, > 10 kbar) generate mafic cumulates ($ol + cpx + opx \pm Ca-Ti-amph$), intermediate reservoirs (15–20 km, 5–7 kbar) host rapakivi batholiths, and shallow levels (1–4 km) crystallize felsic ring complexes ($af + pl + cpx + ap + zr \pm Ca-amph$). A-type granite compositions thus reflect contributions from both the SCLM and mafic lower crust, with ascending melts evolving from mafic to felsic through assimilation, emplacement guided by crustal structures, and mid-crustal formation driven by repeated injections linked to mantle upwelling. In contrast, Sharkov (2010) emphasized a plume-driven transcrustal system beneath thick sialic crust, where basaltic melts stall at multiple levels, forming anorthosite and gabbro-norite cumulates overlain by buoyant silicic magmas. Density contrasts, thermal input, and chemical exchange promote K-enrichment, isotopic heterogeneity, and multiphase AMCG batholiths that evolve asynchronously, ascending in successive pulses through gravitational instability. A third framework, proposed

by Johansson (2023) for the Bornholm rapakivi complex, envisions mantle-derived basalts underplating continental crust, generating heat that induces partial melting of overlying rocks. These basalts evolve into silicic compositions through fractional crystallization and assimilation, accumulate in lower crustal reservoirs with associated mafic cumulates, and gradually shift upward as a single, compositionally evolving melt column. Unlike bimodal regimes with discrete basaltic and granitic pulses, this model highlights a long-lived, internally sustained magmatic system beneath a stable supercontinent, independent of external tectonic triggers.

Taken together, the models of Bonin (2007), Sharkov (2010), and Johansson (2023) offer distinct yet complementary frameworks for A-type granite and rapakivi magmatism. While all invoke mantle-derived sources, they differ in thermal regime, crustal configuration, and the role of plume activity. Bonin (2007) emphasizes melt generation in the SCLM and ascent through crustal discontinuities in intraplate or mildly extensional settings, without requiring plume input. In contrast, Sharkov (2010) proposes a plume-driven scenario beneath thick sialic crust, where gravitational instability leads to successive intrusions—consistent with superswell dynamics. Johansson (2023) highlights long-term thermal insulation beneath stable Nuna, where basaltic underplating triggers progressive crustal melting and melt evolution. Though not explicitly plume-related, this model aligns with superswell conditions. All three frameworks converge on key processes such as fractional crystallization, crustal assimilation, and deep magma storage. The superswell hypothesis of Salminen et al. (2021) – invoking widespread mantle upwelling beneath equatorial Nuna – supports both (Sharkov, 2010) and Johansson (2023), while remaining compatible with the intraplate scenario of Bonin (2007).

These mantle-driven models are consistent with the broader planetary framework proposed by Pirajno and Santosh (2015), which links mantle plumes to supercontinent breakup. In this view, plume-induced thermal uplift, lithospheric doming, and rifting serve as primary mechanisms for continental fragmentation. Geophysical and tomographic data support this interpretation, with mantle plumes viewed as thermal-chemical upwellings originating either from the core–mantle boundary (CMB; one-layer model) or the mantle transition zone (MTZ; 410–660 km; two-layer model). In the supercontinent self-destruction model, subducting slabs stagnate in the MTZ, triggering the rise of superplumes that thermally weaken the overlying lithosphere. P-wave tomography from the western Pacific reveals stagnant slabs and high-velocity anomalies above the CMB – interpreted as a “slab graveyard” – potentially driving plume ascent. Pirajno and Santosh (2015) distinguishes primary plumes (from the D” layer), secondary plumes (from plume domes), and tertiary plumes (from near the 660 km discontinuity). Modern superswells beneath Africa and the Pacific exemplify

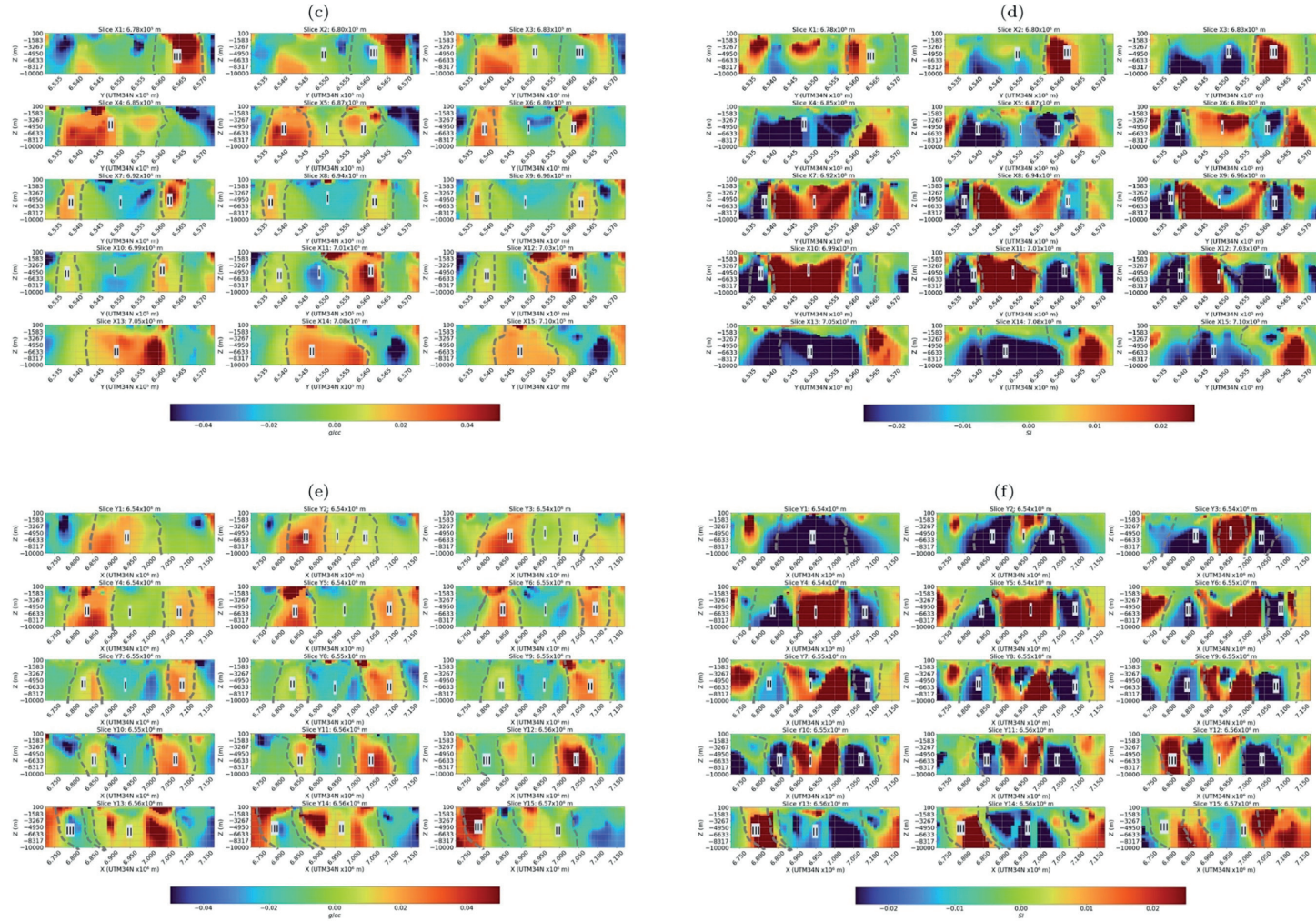


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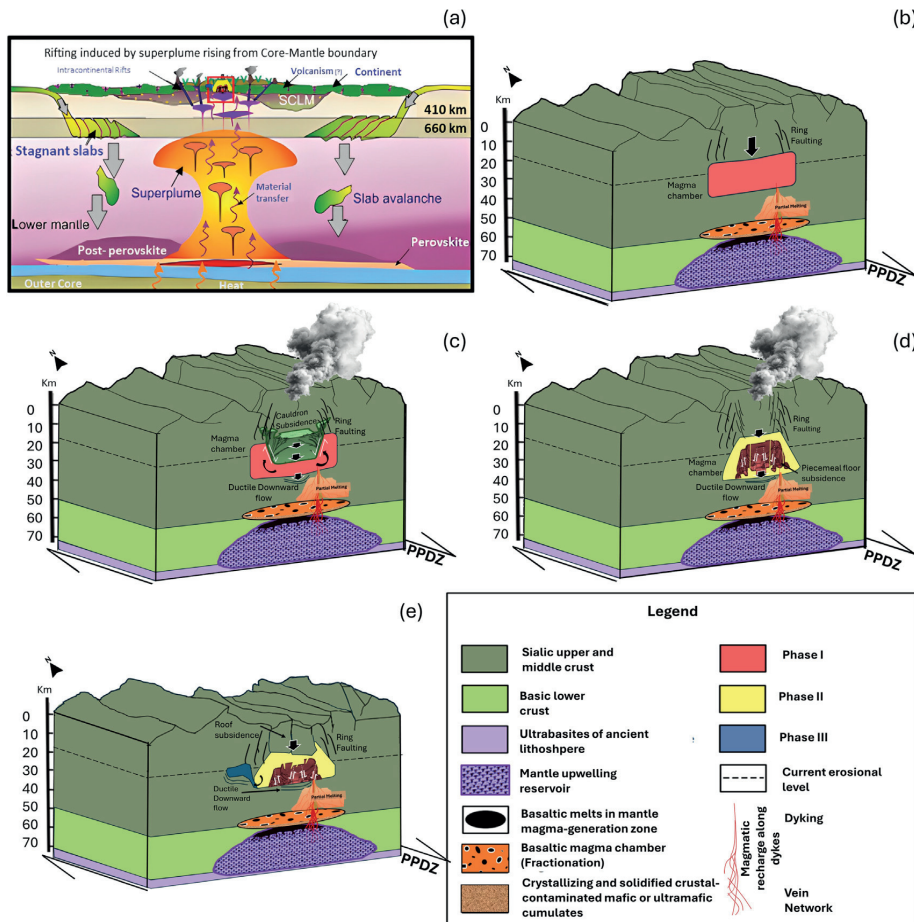


Fig. 7. The geodynamic model for the emplacement system of the Märjamaa–Kloostri rapakivi system is presented as follows: (a) Rifting induced by superplume rising from the core–mantle boundary, fuelled by ‘slab graveyards,’ exemplifying the superswell plume model for Fennoscandian rapakivi bodies as suggested by Salminen et al. (2021). A block model, adapted from (Pirajno and Santosh, 2015), references the structural crustal position of the Märjamaa and Kloostri bodies within a red square. Interpretive block diagrams (note: the shapes and dimensions of the magma chamber and collapsing blocks are approximations and not to scale) illustrate the emplacement mechanism, inferred magma flow paths, and associated strain patterns in the Märjamaa and Kloostri rapakivi granitoids. The tripartite emplacement is described as follows: (b) Proto-Phase I chamber: a steep-sided, tabular intrusion formed by magma ascent through a southeast conduit during PPDZ-related extension; (c) Piston–cauldron subsidence: roof collapse along ring faults due to magma decompression, inducing vertical and convergent subhorizontal flow; (d) Phase I–II development: Phase I consolidates through block assimilation, while Phase II intrudes peripherally and cross-cuts earlier structures during piecemeal collapse; (e) Phase III emplacement, a satellite wedge-shaped body formed by asymmetric (trapdoor) subsidence, reflecting late-stage intrusion during crustal relaxation. The legend for the Märjamaa and Kloostri figures b to e is approximate in scale, with the principal purpose being to enhance the geodynamic understanding of the emplacement system.

this model and are analogous to the superswell proposed beneath low-latitude Nuna by Salminen et al. (2021) (Fig. 7a).

Deep-Earth processes tied to supercontinent breakup promoted mantle upwelling, transient lithospheric thinning, mafic underplating, and anorogenic low- H_2O magmatism (Pirajno and Santosh, 2015) and references therein). In such settings, A-type granitoids are commonly metallogenic (e.g., Olympic Dam IOCG), and during 1.8–1.1 Ga – coincident with Nuna’s fragmentation – SEDEX, IOCG, Sn, and stratiform Cu–Au–U systems were favoured (Pirajno and Santosh, 2015). Lithospheric-scale shear zones can sustain long-lived fluid circulation that modifies melts and focuses metals (Sn–W–Mo, Au, U–Th, REE); topaz and fluorite indicate F-rich fluids enhancing HFSE transport; and brittle structures act as high-permeability pathways (Bonin, 2007;

Vignerresse, 2005; Radaideh et al., 2016; Pirajno and Santosh, 2015; Sialchoque et al., 2017). Although the balance of magmatic versus crustal contributions remains debated (Bonin, 2007; Dall’Agnol et al., 2012), rapakivi plutons can host Sn–(W–Be–Zn–Cu–Pb) skarns and Fe-oxide Cu–Au–U–Ag systems (e.g., Salmi; Haapala 1995), and Li–As–W–Zn–Mn or Cu–As–In enrichments in the Wiborg Batholith (Valkama et al., 2016). Although mineral exploration is not our focus, a distinct NE-trending RTP residual anomaly in northern Märjamaa (Fig. 3d3) and coincident SimPEG density/susceptibility contrasts in X_{12-15} and Y_{15} profiles (Figs. 6c–f) suggest a structurally controlled subsurface feature warranting targeted mineralogical and geochemical follow-up.

Within this planetary context, Fennoscandian rapakivi intrusions – such as Märjamaa–Kloostri – represent mid- to shallow-crustal expressions of AMCG magmatism linked to superswell-driven mantle

processes. Their Mesoproterozoic emplacement beneath a thermally insulated, low-latitude Nuna supercontinent (Elming et al., 2010; Mitchell et al., 2021; Salminen et al., 2021) places them in a global suite of plume-related AMCG granitoids (Fig. 7a). Regionally, AMCG magmatism – including rapakivi granites and massif-type anorthosites – occurred during Nuna breakup assembly and was triggered by superswell-induced perturbation of the SCLM, partial melting, and lithospheric extension (Bonin, 2007; Pirajno and Santosh, 2015; Vigneresse, 2005; Sharkov, 2010; Johansson, 2023).

In Fennoscandia, NE-SW Svecofennian (SO) compression established NW-SE to NNW-SSE shearing corridors and oblique strike-slip later reactivated during Mesoproterozoic extension, with structural deflections near rapakivi massifs, and present shallow seismicity Rg-wave propagation confirm ongoing oblique strike-slip (Branigan, 1987; Korja and Heikkinen, 1995; Korja et al., 2001; Bogdanova et al., 2006, 2015; Korja and Kosonen, 2015; Soosalu et al., 2022). Within this oblique-extensional, corridor-focused regime, late-stage Wiborg Suite rapakivi intrusions (including Märjamaa-Kloostri) formed sheet-like, tabular plutons along dilatational ramps (i.e., listric faults) imaged on Gulf of Bothnia and SW Finland seismic profiles (Figs. 4g, 1a) (BABEL Working Group, 1993a,b; Korja and Heikkinen, 1995; Puura and Flodén, 1999; Korsman et al., 1999; Puura and Flodén, 2000; Korja et al., 2001; Selonen et al., 2005; All et al., 2006; Karell et al., 2009, 2014). During N-S extension at 1.65–1.60 Ga, reactivation of the SJJSZ and Porkkala-Mäntsälä Shear Zone (PMSZ) (Fig. 1a, Fig. 4g) overprinted older E-W fabrics and enabled emplacement of the ~1.64 Ga Bodom and Obbnäs stocks (Kosunen, 1999; Korsman et al., 1999; Torvela et al., 2008; Nironen, 2017; Nordbäck et al., 2024). Along the Åland-PPDZ, elevated heat flow – with joint-density peaks near Phase II boundaries and within the σ_H -parallel corridor (Figs. 4d–e, g) – coincides with the regional NW-SE, high-density potential-field structural lineaments (Solano-Acosta et al., 2023) and shear-wave-splitting fast polarizations parallel to the corridor (Wüestefeld, 2007). Mapped Riedel extensional and tensional sets along the Åland-PPDZ (Fig. 4f–g) indicate reactivated Paleoproterozoic (SO) shear fabrics during Nuna's breakup, consistent with pull-apart kinematics and coeval AMCG magmatism and basin development (Bogdanova et al., 2013, 2015; Pirajno and Santosh, 2015; Nironen, 2017; Nordbäck et al., 2024; Johnston et al., 2024).

Together, these observations suggest deep structural links among the Märjamaa, Kloostri, Naissaare, and Neeme intrusions—potentially via shared conduits (Fig. 4g) (Puura and Flodén, 1999, 2000)—whereas the Riga pluton follows the MEZF, also in a zone of high structural density and heat flow (Kirs et al., 2004; All et al., 2006; Solano-Acosta et al., 2023). Estonian mainland rapakivi bodies share Obbnäs-like compositional traits (accessory titanite, low F, high Ba, low Rb/Sr) (Soesoo, 1993; Klein et al., 1994; Rämö et al., 1996; Kosunen, 1999), where NW-SE (σ_h) shearing at ~1.62 Ga reactivated SO fabrics and focused ascent along the PPDZ during Märjamaa-Kloostri emplacement (Puura and Flodén, 2000; All et al., 2006; Bogdanova et al., 2015).

Understanding the emplacement depth of the Märjamaa-Kloostri and related rapakivi complexes requires evaluating post-magmatic exhumation and crustal restructuring across Fennoscandia. Uplift and erosion after rapakivi magmatism likely removed several kilometres of cover, exposing shallow-level plutonic suites (Haapala and Rämö, 1992; Puura and Flodén, 1999), but the magnitude and cause of crustal thinning remain debated. Puura and Flodén (1999) argued that erosion alone cannot account for the estimated 10–15 km of thinning, citing limited preservation of volcanic cover and a lack of unequivocal denudation markers, and instead proposed mantle- and crustal-scale uplift – possibly related to a thermal swell or plume – that fractured the crust and promoted ascent of mantle and lower-crustal magmas beneath gravitationally unstable orogenic lithosphere. Expanding this view, Puura and Flodén (2000) suggested a hybrid scenario of differential erosion plus tectonic tilting: early rapakivi intrusions postdate Svecofennian thickening and are followed by removal of 5–10 km of

crust, consistent with variable exposure of volcanic remnants across the Wiborg, Riga, and Åland massifs. In contrast, Sharkov (2010) advocated a predominantly erosional model, inferring 16–20 km of crustal stripping from 4.5–5 kbar metamorphic pressures in Svecofennian rocks and reconstructing a Paleoproterozoic crust 75–78 km thick, with 50–60 km of sialic upper crust over ~20 km of mafic lower crust. Despite differences, all models imply substantial post-orogenic crustal reworking.

A-type granites are often emplaced at shallow levels (~4 km) with magma chambers extending to ~20 km (Bonin, 2007), along ascent paths from ~15 km (~6 kbar) at 450–650°C to final crystallization at 720–780°C (Vigneresse, 2005). Emplacement depths of AMCG-rapakivi massifs range between 7–20 km (2–6 kbar). The Salmi massif crystallized at shallow levels (2–3 kbar; Sharkov and Bogina 2006). The Korosten pluton formed at transitional depths (3–4 kbar; Sharkov 2010), with zoned pegmatites yielding 2.6–3.0 kbar (Sharkov, 2010). The Wiborg batholith intruded at 2.5–5.4 kbar based on Al-in-hornblende geobarometry (Elliott, 2001). The Mazury Complex reached zircon saturation above 840°C at ~4 kbar (Grabarczyk et al., 2023). The Riga pluton represents the deepest case (5–7 kbar; 14–17 km; Rämö et al. 1996, Kirs et al. 2004, Sharkov 2010), supported by subsolidus coronas in gabbroids that formed at 5–6 kbar and 900–1100°C (Sharkov et al., 2004; Sharkov, 2010). For Märjamaa-Kloostri, Buddington–Barth geothermometry indicates 550–650°C and Al content of amphiboles 5.2–6.5 kbar (Klein et al., 1994), suggesting ~17–22 km emplacement depth.

The Estonian basement shows a south-to-north tilt of ~100 m at the Precambrian surface across the Märjamaa-Kloostri bodies (Fig. 1b1), implying asymmetric erosional denudation and block tilting (Soesoo et al., 2004, 2020). Peak metamorphic conditions in amphibolitic host rocks (~700 °C, 4.5–5.5 kbar; Soesoo et al. 2020) agree with pressures inferred by Sharkov (2010). The Conrad discontinuity lies at ~17–19 km beneath Märjamaa-Kloostri (Fig. 1d; Solano-Acosta et al. 2023), separating felsic upper crust (SIAL) from mafic lower crust (SIMA). Allowing for ~20 km of erosion places the original Conrad near ~40 km depth, consistent with the SIAL-SIMA boundary reconstructed at ~40–50 km by Sharkov (2010) (Fig. 7). Regional potential-field profiles (All et al., 2004) show both intrusions extending to comparable structural levels, with the present base of Märjamaa at ~10 km, akin with CPD and UC estimates (Figs. 1e,5). Accounting for ~20 km of post-emplacement erosion (Sharkov, 2010) implies an original emplacement depth of ~20–30 km, closely matching hornblende geothermometry results from Klein et al. (1994).

Some Fennoscandian rapakivi plutons – such as Vehmaa, Ahvenisto, Märjamaa-Kloostri, Mårdsjö, and the Wiborg-Kymi stock – commonly display circular planforms with inward-dipping listric faults indicative of cauldron–piston subsidence, and their limited brecciation points to emplacement under low differential stress (Klein et al., 1994; Laitakari et al., 1996; Alviola et al., 1999; Haapala and Lukkari, 2005; Selonen et al., 2005; Karell et al., 2009, 2014). Many systems also record piston–cauldron features and protracted chamber evolution, although the boundary with caldera collapse remains blurred: both involve ring faults, but calderas typically form at depths ≤ 7 km and erupt, whereas piston-cauldrons occur deeper (7–30 km) and need not produce surface volcanism (Smith and Bailey, 1968; Myers, 1975; Elming et al., 2021; Grocott et al., 1999; Burchardt, 2009, 2018; Karell et al., 2009, 2014; Currier et al., 2019; Geshi et al., 2021a). Surface ignimbrites may have existed (Puura and Flodén, 2000) but are mostly eroded (Haapala and Rämö, 1992; Sharkov, 2010). Taken together, erosion estimates and structural geometry place the Märjamaa and Kloostri intrusions firmly in the piston–cauldron category.

The proto-Phase I magma chamber likely ranged from ~15 to 30 km in depth, forming a steep-sided, ovoid body with a vertical extent of ~15 km and a lateral footprint of 40 × 25 km (Fig. 7b) (Cruden, 1998; Grocott et al., 1999; Tomek et al., 2014). Transtensional pull-apart shearing along the PPDZ triggered piston-type floor collapse,

generating ring fractures and a proto-piston block that accommodated magma and expanded the chamber (Smith and Bailey, 1968; Myers, 1975; Selonen et al., 2005; Stevenson et al., 2008; Burchardt et al., 2012; Burchardt, 2018). Additional evidence from UC modelling (Fig. 5) and SimPEG inversion X_{9-10} and Y_{13-14} profiles (Figs. 6) supports a southeastern feeder conduit to Phase I, implying vertical magma ascent and a central recharge zone. The Märjamaa chamber thus functioned as the principal reservoir feeding a broadly symmetrical intrusion.

The transition from the proto-Phase I chamber (Fig. 7b) to piston-cauldron subsidence (Fig. 7c) was likely driven by shearing and pull-apart extension along the PPDZ (Figs. 4g), which generated concentric and radial fractures leading to roof collapse. Magma initially ascended vertically and along fractures, then transitioned into subhorizontal, inward-directed flow (Clough et al., 1909; Smith and Bailey, 1968; McNulty et al., 2000; Burchardt et al., 2012; Tomek et al., 2014; Siachoque et al., 2017). SimPEG inversion cross-sections X_{7-9} and Y_{7-9} (Fig. 6) reveal a cylindrical structure at the centre of Märjamaa, showing a positive gradient in density and magnetic susceptibility at 3 km depth, extending 4 km in thickness. This anomaly, possibly a paleo-conduit or syenitic residual core, supports piston-style collapse and potential ignimbrite-forming pyroclastic activity. As the extension of PPDZ continued, the release of magma and the tectonic discharge reduced the pressure of the chamber, initiating the sagging of the roof and the subsidence of the crust along the faults of the rings, inducing radial stress, decompression, magma ascent and possible ignimbrite eruptions (Smith and Bailey, 1968; Eklund et al., 1996; Grocott et al., 1999; Karell et al., 2009; Tomek et al., 2014; Geshi et al., 2021b).

Piecemeal roof collapse enabled melt exchange between upper and lower zones of the chamber (Burchardt, 2009, 2018; Burchardt et al., 2012; Tomek et al., 2014), allowing accumulation and eventual coalescence of discrete pulses into a consolidated Phase I body. Phase II magma intruded peripherally into Phase I-piston via PPDZ-related fractures (Fig. 7d). It is associated with negative magnetic and Bouguer anomalies, may be influenced due to signs of advanced fractional crystallization and crustal assimilation (Soesoo and Niin, 1992; Soesoo, 1993; Klein et al., 1994).

Leucogranitic Phase III marks the final subsidence stage (Soesoo, 1993; Klein et al., 1994), forming an asymmetric trapdoor-style emplacement at the NW of the principal body (Fig. 7e) (Grocott et al., 1999; Tomek et al., 2014). The intrusion likely resulted from chamber depressurization and late-stage piecemeal floor collapse, where intensified piston-cauldron dynamics at the chamber base remobilised residual melts. Magma ascended along NW-oriented conduits during final emplacement.

The gravity and magnetic patterns reflect systematic petrophysical contrasts among the three phases and fit a piston-cauldron emplacement along the PPDZ. Phase I (ferroan, melanocratic granodiorite) is magnetite-bearing and denser, yielding coincident gravity and magnetic highs; measured bulk density is $\rho \approx 2.71 \text{ g/cm}^3$ (Soesoo and Niin, 1992) and opaque Fe-Ti oxides (magnetite-ilmenite) reach $\sim 3\%$ (Soesoo, 1993; Klein et al., 1994). Phase II (quartz-K-feldspar-rich ring) is lighter and magnetite-poor (Soesoo, 1993; Klein et al., 1994), giving paired gravity and magnetic lows; its bulk density is $\rho \approx 2.66 \text{ g/cm}^3$ (Soesoo and Niin, 1992). Phase III (late Na-rich trachytic leucogranite) is low-density— $\rho \approx 2.59\text{--}2.63 \text{ g/cm}^3$ (Soesoo and Niin, 1992; Soesoo, 1993; Klein et al., 1994) producing a gravity low; its local magnetic high at the NW margin is best explained by superposition due to the central Phase I magnetite-rich collapse and/or localized concentration of magnetic carriers along NW-trending late conduits active during trapdoor subsidence.

Future studies that integrate host-rock interaction, mineral textures, and seismic imaging are essential to refine the timing of emplacement, depth, and structural relationships of the Märjamaa and Kloostri bodies. Similar analysis is pertinent for the Naissaare granite, to improve our understanding of the influence of the PPDZ on the emplacement of Estonian rapakivi bodies.

7. Conclusions

The Märjamaa-Kloostri complex records structure-controlled emplacement along the NW-trending Åland-PPDZ, where reactivated Paleoproterozoic (SO) shear fabrics under transpressional–transensional stress focused Mesoproterozoic AMCG magma ascent, driving pull-apart basin development and cauldron-style subsidence. Emplaced at $\sim 1.62 \text{ Ga}$, the system preserves a multiphase evolution: Phase I formed a steep-sided tabular intrusion that evolved into a piston-cauldron; Phase II intruded as a peripheral ring during continued subsidence and lateral expansion; Phase III was emplaced asymmetrically (trapdoor-style) during late crustal relaxation. An integrated workflow – potential-field filtering, spectral derivatives, automated lineament analysis, and 3-D cross-gradient joint residual-potential inversion (SimPEG) – delineates sharp phase contacts and structural features consistent with this kinematic framework. Overall, the complex shows how structural inheritance and shear-zone kinematics govern magma pathways, chamber geometry, and the staged evolution of AMCG rapakivi granitoids in the Fennoscandian Shield.

CRedit authorship contribution statement

Juan David Solano-Acosta: Writing – review & editing, Writing – original draft, Visualization, Methodology, Investigation, Formal analysis, Data curation, Conceptualization. **Alvar Soesoo:** Writing – review & editing, Validation, Supervision, Project administration, Investigation, Conceptualization. **Rutt Hints:** Writing – review & editing, Validation, Supervision, Project administration, Investigation.

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Declaration of competing interest

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this paper.

Appendix A. Supplementary data

Supplementary material related to this article can be found online at <https://doi.org/10.1016/j.precamres.2025.107938>. Geophysical datasets and the SimPEG code are available upon request. Supplementary Figure S1 is attached to the repository.

Data availability

Data will be made available on request.

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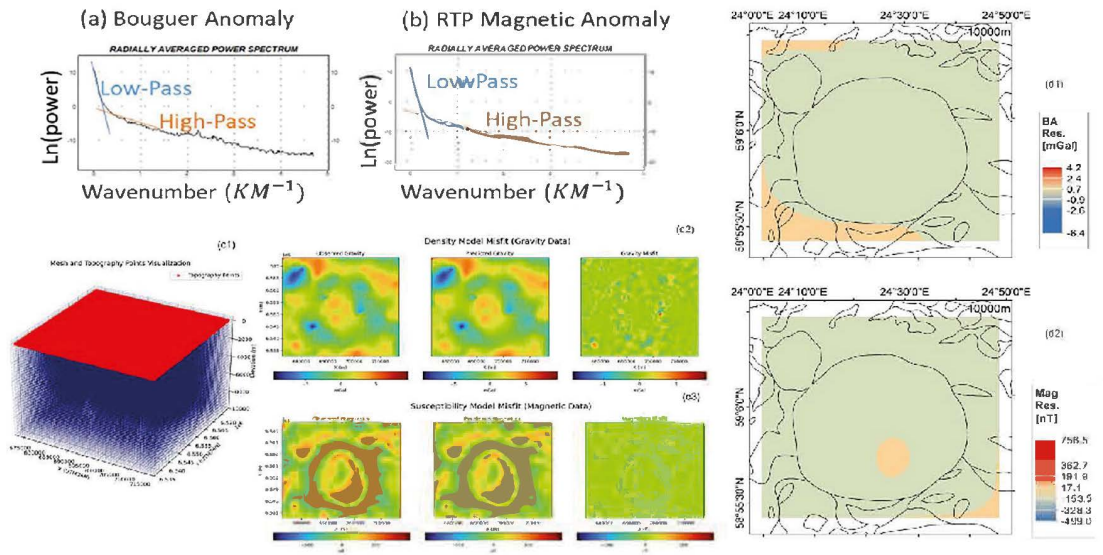


Fig.S1. Radially averaged power spectra with high-pass and low-pass filters based on wavenumber analysis for **(a)** Bouguer and **(b)** RTP magnetic data. **(c)** SimPEG inversion setup and model-data misfit. **(c1)** 3D tensor mesh and topography points used in the inversion; the mesh is refined near the surface and fully active below the red topographic surface. **(c2)** Gravity data: observed residual Bouguer anomaly, predicted response from the recovered density model, and misfit. **(c3)** Magnetic data: observed RTP anomaly, predicted response from the recovered susceptibility model, and misfit. **(d)** UC depth analysis up to 10,000 m showing signal attenuation and enhanced signal sources. **(e1)** Bouguer anomaly, **(e2)** RTP magnetic residual anomaly.

Appendix IV

J D Solano-Acosta, A Soesoo, and R Hints. An updated review of major element geochemistry and petrogenesis of the oxidised Märjamaa and Kloostri Rapakivi Granites, Estonia. 2025. Manuscript Under review

An Updated Review of Major Element Geochemistry and Petrogenesis of the oxidised Märjamaa and Kloostri Rapakivi Granites, Estonia

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An Updated Review of Major Element Geochemistry and Petrogenesis of the oxidised Märjamaa and Kloostri Rapakivi Granites, Estonia

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Abstract

The ~ 1.62 Ga Märjamaa and Kloostri rapakivi granitoids in western Estonia, consist of three distinct magmatic phases that reflect progressive differentiation throughout their emplacement. Phase I represents ferrous, melanocratic granodiorite; Phase II forms a quartz- and K-feldspar-enriched granite ring; and Phase III corresponds to a late-stage, Na-rich trachytic leucogranite. Updated whole-rock major-element data reveal a systematic compositional evolution from silica-poor to silica-rich phases, driven by progressive feldspar accumulation and concomitant depletion in Fe–Ti oxides and phosphates. Moderate $FeO^*/(FeO^* + MgO)$ ratios indicate consistently oxidising crystallisation conditions across all phases. These results refine the geochemical characterisation of the Märjamaa–Kloostri intrusions, establishing them as late-stage, oxidised members of the Wiborg rapakivi suite and providing new insights into Mesoproterozoic A-type magmatism in the Estonian Precambrian basement.

Keywords:

Märjamaa–Kloostri rapakivi granitoids, Tripartite magmatic phases, fractional crystallisation, Oxidised magmatism.

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1. Introduction

A-type granites are anhydrous felsic rocks enriched in alkaline metals and High Field Strength Elements (HFSE), typically emplaced in post- to anorogenic settings such as continental interiors or rift zones (Loiselle and Wones, 1979; Eby, 1992; Frost et al., 2001; Vigneresse, 2005; Dall’Agnol and de Oliveira, 2007). A-type granites typically have high incompatible element contents, including high rare earth elements (*HREE*), *Zr*, *Nb*, and *Ta*, but low *Co*, *Sc*, *Cr*, *Ni*, *Ba*, *Sr*, and *Eu* contents (Frost and Frost, 1997).

Rapakivi granites, a distinctive A-type subtype, are characterised by ovoid alkali-feldspar megacrysts mantled by oligoclase (Haapala and Rämö, 1992; Rämö and Haapala, 1995, 2005) and range from quartz syenites to peralkaline granites (Bonin, 2007; Dall’Agnol et al., 2012). These granites are widely reported from Precambrian cratons, including Baltica (East European Craton), Laurentia, Amazonia, and the Ukrainian Shield (Puura and Flodén, 1996, 1999, 2000; Bogdanova et al., 2006, 2015; Sharkov, 2010; Salminen et al., 2021; Grabarczyk et al., 2023; Hinchey et al., 2024), as well as in Phanerozoic belts (Rajesh, 2000; Christiansen et al., 2007; Luo et al., 2024).

In Fennoscandia (Fig.1a), rapakivi granites are a hallmark of Palaeo to Mesoproterozoic AMCG (Anorthosite–Mangerite–Charnockite–Granite) suites, which formed between 1.8–1.0 Ga, peaking at 1.6–1.4 Ga during the assembly of the Nuna–Columbia supercontinent (Emslie, 1978; Vigneresse, 2005; Salminen et al., 2021; Johansson et al., 2022). These bimodal associations commonly include rapakivi granites, massif-type anorthosites, gabbros, ferrodioritic dykes, and mangerites (Haapala and Rämö, 1992; Frost et al., 1999; Dall’Agnol and de Oliveira, 2007). Petrogenetic models attribute their formation to partial melting of crustal sources triggered by mantle upwelling and mafic underplating, with tectonic settings ranging from post-orogenic extension to mantle superswell-related and thermal perturbations (Bonin, 2007; Kukkonen and Lauri, 2009; Sharkov, 2010; Salminen et al., 2021; Johansson, 2023).

In Estonia, the AMCG suite comprises by smaller ~ 1.6 Ga granitoids, in-

cluding Märjamaa (with its Kloostri satellite), Naissaare, Taebla, Neeme, and Ereda, and the massif ~ 1.5 Ga Riga batholith (Fig.1b) (Soesoo and Niin, 1992; Klein et al., 1994; Rämö et al., 1996; Kirs et al., 2004). These granitoids are pink, medium- to coarse-grained syeno- to monzogranites with porphyroceous textures, in which the classic plagioclase-mantled alkali-feldspar ovoids are scarce (Soesoo and Niin, 1992; Klein et al., 1994). Despite their small size and the absence of associated anorthosites, they are geochemically and isotopically consistent with the Fennoscandian rapakivi suites (Klein et al., 1994; Rämö et al., 1996; Ahl, 1997; Kirs et al., 2004).

The Märjamaa-Kloostri intrusions were emplaced within a Svecofennian-inherited crustal framework, where deep-seated shear zones provided pathways for Palaeoproterozoic magmatism (e.g., Åland-PPDZ (Paldiski-Pskov Deformation Zone); Fig.1c) (Raukas and Teedumäe, 1997; Puura and Flodén, 2000; Kirs et al., 2004; All et al., 2006; Bogdanova et al., 2015; Soosalu et al., 2022; Solano-Acosta et al., 2023). Previous studies have outlined their petrography and isotopic signatures (Soesoo and Niin, 1992; Klein et al., 1994; Kirs et al., 2004), yet a systematic synthesis of their major, trace and REE elemental geochemistry and magmatic evolution remains lacking.

This study provides an updated analysis of compiled whole-rock major element data for the Märjamaa rapakivi pluton and its Kloostri satellite (Figs.1,2). We examine the geochemical evolution of its tripartite magmatic phases, focusing on their elemental trends and redox conditions, and compare them with Fennoscandian rapakivi granites within the broader AMCG context.

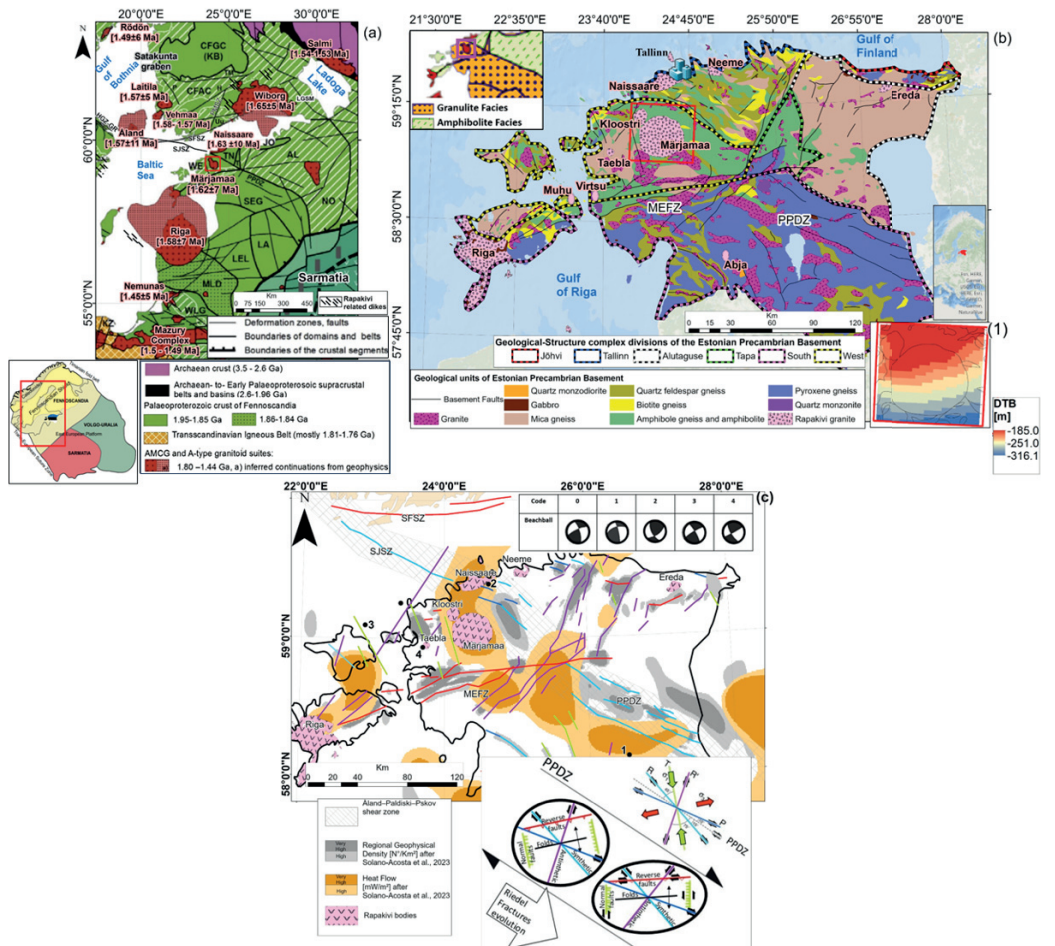


Figure 1: Geological setting of the study zone. (a) Major Paleoproterozoic tectonic domains of Fennoscandia, modified after Bogdanova et al. (2015), highlighting the most prominent rapakivi bodies, with the Märjamaa and Kloostri granitoids study area marked by a red rectangle. The white diagonal ruling over the greenish area indicates Svecofennian sedimentary basins.

Figure 1: (*continued*) The abbreviations are AL – Alutaguse, B-H – Breven-Hällefors dike swarms, CFAS – Central Finland Arc Complex, CFGC – Central Finland Granitoid Complex, H – Häme belt, HGZ-GR – Hagsta-Gävle-Rättvik Zone, JO – Jöhvi, KB – Keitele microcontinent, LA – Latgalia, LEL – Latvian-East Lithuanian, MLD – Mid-Lithuanian domain, P – Pirkanmaa belt, SEG – South Estonian granulite domain, T – Tapa, TN – Tallinn, TM – Tampere belt, Uu – Uusimaa belt, WE – West Estonian domain, WLJ – West Lithuanian granulite domain. Shear and fault zones are abbreviated as follows: MEFZ – Middle Estonian Fault Zone, PMSZ – Porkkala-Mäntsälä Shear Zone, PPDZ – Paldiski-Pskov Deformation Zone, SFSZ – South Finland Shear Zone, and SJSZ – Sottunga-Jurmo Shear Zone. Age rapakivi granitoids of Fennoscandia are constrained from [Laitakari et al. \(1996\)](#); [Ahl \(1997\)](#); [Rämö et al. \(1996\)](#); [Skridlaite et al. \(2007\)](#); [Larin \(2009\)](#); [Sharkov \(2010\)](#); [Johansson et al. \(2016\)](#); [Grabarczyk et al. \(2023\)](#). (b) Precambrian geological map highlighting the Estonia basement, showing metalliferous anomalies based on [Soesoo et al. \(2020\)](#). The upper left-corner inset shows the distribution of granulite- and amphibolite facies metamorphic rocks after [Bogdanova et al. \(2015\)](#). Geological data are available from the Land Board Geoportal of the Republic of Estonia (<https://xgis.maaamet.ee/xgis2/page/app/geoloogia400k>), including (1) Depth to Basement (DTB) measurements from the same source. (c) Estonian rapakivi bodies were overlaid on zones with the highest High Flow values and the highest density of tectonic features derived from regional potential lineaments and basement faults, derived from [Solano-Acosta et al. \(2023\)](#). Basement faults were categorized based on their orientation relative to the dextral PPDZ maximum horizontal stress (σ_H of 115° – 135° , 295° – 315°) in central Fennoscandia ([Korja and Kosonen, 2015](#); [Soosalu et al., 2022](#); [Solano-Acosta et al., 2023](#)). The angle relative to σ_H and the expected mode of movement were represented as follows: blue indicates (transfer (P) and normal (R)), purple indicates 90° (reverse - R'), green indicates 30° – 40° (sinistral strike-slip (T)), and red indicates 30° – 40° (dextral strike-slip (σ_3)). The epicentres of earthquakes analysed by [Soosalu et al. \(2022\)](#) are shown as black dots labeled with a number (code), each with its respective focal mechanism.

2. Geological setting

2.1. Regional context

Baltica, northeast of the Trans-European Suture Zone, consists of three Archean–Proterozoic proto-cratons—Fennoscandia, Volgo-Uralia, and Sarmatia (Fig. 1a). Their amalgamation between 2.1 and 1.79 Ga during the Svecofennian orogeny (SO) produced a stable continental core through arc accretion, deformation, and granitoid intrusions, with the Svecobaltic phase (1.84–1.79 Ga) marking the final collision of the Livonia microcontinent with Sarmatia ([Gorbatschev and Bogdanova, 1993](#); [Nironen, 1997](#); [Bogdanova et al., 2006, 2015](#); [Johansson et al., 2022](#)). By ~ 1.70 Ga, Baltica’s crust was stabilized, followed by the Gothian orogeny (~ 1.73 Ga) along its western margin ([Bogdanova et al., 2008](#); [Nironen, 2017](#)).

66 AMCG suites are emplaced after the Gothian orogeny within the Svecofen-
 67 nian Province and Transscandinavian Igneous Belt to the south, marked by bi-
 68 modal magmatism in Baltica (Ahl, 1997; Vigneresse, 2005; Larin, 2009; Salminen
 69 et al., 2021; Grabarczyk et al., 2023). Fennoscandian rapakivi granites are com-
 70 monly grouped into four main age intervals: 1.67–1.62, 1.59–1.56, 1.55–1.53,
 71 and 1.53–1.44 Ga (Rämö et al., 1996; Ahl, 1997; Puura and Flodén, 2000;
 72 Salminen et al., 2021; Grabarczyk et al., 2023). The oldest, the Wiborg Suite
 73 (1.67–1.62 Ga), comprises the Wiborg, Naissaare, Märjamaa, and Kloostri in-
 74 trusions (Klein et al., 1994; Rämö et al., 1996; Puura and Flodén, 2000). The
 75 Åland Suite (1.59–1.56 Ga) encompasses bodies such as Riga, Åland, Laitila,
 76 and Vehmaa (Suominen, 1991; Selonen et al., 2005). The Salmi Suite (1.55–1.53
 77 Ga) occurs along the Archean–Proterozoic transition zone (Larin, 2009; Sharkov,
 78 2010), while the Ragunda Suite (1.53–1.47 Ga) includes intrusions in Sweden
 79 (Rödon), Lithuania (Nemunas), and Poland (Mazury complex) (Persson, 1999;
 80 Dörr et al., 2002; Duchesne et al., 2010). These magmatic episodes are spatially
 81 associated with fault-controlled blocks (Puura and Flodén, 1996; Bogdanova
 82 et al., 2015; Salminen et al., 2021; Nordbäck et al., 2024).

83 *2.2. Estonian context*

84 The Estonian geological framework comprises a Paleo to Mesoproterozoic
 85 crystalline basement overlain by gently south-dipping (~ 0.10 – 0.20°) Ediacaran–Palaeozoic
 86 strata and a Quaternary cover (Fig.1b1) (Raukas and Teedumäe, 1997; Kirs
 87 et al., 2004; Soesoo et al., 2020). The basement, known from drilling and geo-
 88 physical data, correlates with better-exposed terranes in southern Finland and
 89 Sweden (Puura et al., 2004; Bogdanova et al., 2015; Solano-Acosta et al., 2025).

90 Six structural–petrological zones — Tallinn, Alutaguse, Jõhvi, West-Estonian,
 91 Tapa, and South-Estonian — are distinguished by lithological, geochemical,
 92 and geophysical contrasts, bounded by major crustal-scale features such as the
 93 NW–SE Paldiski–Pskov Deformation Zone (PPDZ) and E–W Middle Estonian
 94 Fault Zone (MEFZ) (All et al., 2004; Bogdanova et al., 2015). The Tallinn and
 95 Alutaguse domains, correlating with the Uusimaa arc and its back-arc basin, ac-

creted with other Svecofennian terranes during 1.92–1.87 Ga double subduction and arc–microcontinent collisions (Nironen, 2017; Solano-Acosta et al., 2025). West and South Estonia consist of high-grade metasedimentary and orthogneissic complexes overprinted by granulite-facies metamorphism (Fig.1b) (Soesoo et al., 2004, 2020). Regional gravity and magnetic patterns show these domains as continuations of oblique Svecofennian blocks in Finland, Sweden, and Latvia, shaped by inherited NW–SE shear zones, including the PPDZ and its Sottunga–Jürmo extension, which remain seismically active (Fig.1c) (Korhonen et al., 2002; Högdahl et al., 2009; Soosalu et al., 2022; Nordbäck et al., 2024).

Estonia’s Precambrian basement consists of three main age groups. The oldest (1.92–1.80 Ga) corresponds to the Fennoscandian crust and includes North Estonian metavolcanics (1.92 ± 10 Ga), granulitic metavolcanics in the south (1.83 ± 22 and 1.82 ± 7 Ga), and tonalites in Tapa (1.82 ± 26 Ga) (Petersell and Levchenkov, 1994; Soesoo et al., 2004, 2006; Kirs et al., 2009; Bogdanova et al., 2015). Magnetite-bearing gneisses from the Jõhvi belt yield U – Pb zircon ages of 1.87 ± 18 , 1.83 ± 10 , and 1.79 ± 19 Ga (Soesoo et al., 2020). Southern Estonian tonalites date to 1.79 ± 16 Ga, while charnockites and orthopyroxene–garnet gneisses are 1.76 ± 11 and 1.78 ± 2 Ga, respectively; a gabbro–norite dyke is 1.77 ± 20 Ga (Soesoo et al., 2004, 2006). The younger group (1.64–1.58 Ga) comprises A-type granites such as the Abja pluton (1.64 ± 7 Ga), Virtsu quartz monzonite (1.61 ± 17 Ga), and rapakivi granitoids including Märjamaa (1.63 ± 7 Ga), Naissaare (1.62 ± 7 Ga), and the Riga batholith (1.58 ± 2 Ga) (Kirs and Petersell, 1994; Rämö et al., 1996; Soesoo and Hade, 2012).

Estonian rapakivi bodies, and the High-K intrusions such as Virtsu, Muhu, and Abja share geochemical similarities with Finnish rapakivi granites, with some showing shoshonitic signatures (Figs.1b,c) (Soesoo, 1993; Klein et al., 1994; Kirs and Petersell, 1994; Kirs et al., 2004). Their emplacement was structurally controlled by the Paldiski–Pskov Deformation Zone (PPDZ) and Middle Estonian Fault Zone (MEFZ), with reactivation of these and related shear zones, including the Sottunga–Jürmo extension, likely guiding magma ascent (Fig.1c) (Puura and Flodén, 2000; Bogdanova et al., 2015; Nordbäck et al., 2024).

127 These plutons are medium- to coarse-grained pink syeno- to monzogranites
 128 with prominent microcline megacrysts, locally cut by aplitic and microsyenitic
 129 dykes (Soesoo and Niin, 1992). Biotite (annitic to siderophyllitic) dominates
 130 the mafic fraction, with hornblende present in Märjamaa, Naissaare, and Neeme
 131 (Soesoo, 1993).

132 The Märjamaa intrusion is a 40×25 km composite pluton emplaced into
 133 West-Estonian Proterozoic gneisses (Fig.1b)(Kirs et al., 2004; All et al., 2004;
 134 Solano-Acosta et al., 2023). It comprises three phases (Fig.2a): Phase I, melano-
 135 cratic granodiorite with abundant gneiss xenoliths (up to 20 cm); Phase II,
 136 biotite–hornblende granite forming a K-feldspar-rich ring; and Phase III, the
 137 Kloostri leucogranite, a Na-rich trachytic phase (Soesoo and Niin, 1992; Klein
 138 et al., 1994; Laitakari et al., 1996). Quartz (20–30%) occurs in two genera-
 139 tions: as inclusions in microcline (20–40%) and in the groundmass. Plagioclase
 140 (20–40%), biotite (2–10%) forms anhedral clusters, hornblende occurs in Phases
 141 I–II, and muscovite is confined to Phase III. Magnetite and ilmenite (3–5%)
 142 are common, with zircon, apatite, titanite, fluorite, and epidote as accessories
 143 (Soesoo and Niin, 1992; Raukas and Teedumäe, 1997).

144 3. Geochemical data

145 In the 20th-century, geochemical studies of Estonia’s Precambrian base-
 146 ment lithologies used wet-chemistry methods to assess major element concen-
 147 trations at various drill-core depths (Kivisilla et al., 1999). The dataset cov-
 148 ers elements like SiO_2 , TiO_2 , Al_2O_3 , Fe_2O_{3tot} , MnO , MgO , CaO , Na_2O ,
 149 K_2O , P_2O_5 , SO_3 , and LOI in weight percentages [wt.%], available on the Es-
 150 tonian Research Infrastructure portal "GEOkirjandus" [https://kirjandus.](https://kirjandus.geoloogia.info/reference/21247)
 151 [geoloogia.info/reference/21247](https://kirjandus.geoloogia.info/reference/21247). For this study, data from Kivisilla et al.
 152 (1999) were used, focusing on the Märjamaa pluton and Kloostri satellite,
 153 with 45 samples organised by drill-core positions (Fig.2), based on Estonian
 154 Geological Survey reports ([https://gis.egt.ee/portal/apps/dashboards/](https://gis.egt.ee/portal/apps/dashboards/99f758ac4ef548f686b831adb3199378)
 155 [99f758ac4ef548f686b831adb3199378](https://gis.egt.ee/portal/apps/dashboards/99f758ac4ef548f686b831adb3199378)). Elemental concentrations were nor-

Id	Kivisilla et al. (1999) Id	Long	Lat	Core Id	Depth (m)	Phase	SiO ₂	TiO ₂	Al ₂ O ₃	Fe ₂ O ₃ _{tot}	MnO	MgO	CaO	NiO	K ₂ O	P ₂ O ₅	SO ₃	LOI	Fe ₂ O ₃	FeO	FeO*
3023300	374	24.43	58.980	302	330	I	64.30	0.10	12.56	4.18	0.01	0.40	1.77	3.19	3.97	0.01	0.07	0.41	0.54	0.28	1.07
3023320	375	24.43	58.980	302	332	I	63.45	1.08	15.66	7.68	0.03	1.25	3.78	2.53	3.58	0.00	0.22	0.96	2.52	3.50	4.33
3023344	377	24.43	58.980	302	334.4	I	67.26	0.75	13.06	5.12	0.11	1.14	3.27	2.50	4.31	0.04	0.40	0.73	2.11	2.71	4.01
3023344	376	24.43	58.980	302	334.42	I	67.14	0.80	13.14	5.18	0.11	1.13	3.42	2.50	4.31	0.07	0.40	0.68	2.11	2.76	4.06
3024031	380	24.43	58.980	302	403.1	I	73.00	0.11	13.34	2.05	0.01	0.43	1.35	2.80	6.31	0.03	0.08	0.49	0.94	1.00	1.85
3024051	381	24.43	58.980	302	405.1	I	65.95	0.78	13.79	5.25	0.01	1.40	1.04	2.84	3.59	0.04	0.22	0.80	2.24	2.70	4.72
3024384	385	24.43	58.980	302	438.4	I	65.56	0.76	13.75	5.88	0.01	1.41	1.03	2.70	5.47	0.02	0.20	0.93	2.31	3.01	5.25
3024600	388	24.43	58.980	302	460	I	64.85	0.82	13.34	5.73	0.10	1.04	4.86	2.48	5.23	0.05	0.22	0.96	2.43	3.06	5.16
3024785	390	24.43	58.980	302	478.5	I	66.01	0.93	13.50	5.74	0.02	1.60	2.93	2.88	3.00	0.00	0.20	0.77	2.47	3.01	5.16
3024805	391	24.43	58.980	302	805.5	I	65.85	0.90	14.21	6.17	0.14	1.23	2.86	2.75	4.77	0.03	0.25	0.54	2.88	3.31	5.55
3024850	393	24.43	58.980	302	485	I	66.32	0.80	13.78	5.48	0.14	1.39	2.94	3.00	5.10	0.30	0.22	0.54	2.29	2.86	4.93
3024874	394	24.43	58.980	302	487.4	I	65.90	0.88	13.09	6.10	0.11	1.60	2.27	2.76	4.99	0.43	0.50	0.13	2.47	3.28	5.48
9-3250	125	24.53	58.980	9c	125	I	64.53	0.98	13.02	6.94	0.14	1.59	3.66	2.47	4.43	0.06	0.55	1.22	2.07	3.54	6.24
9-3270	126	24.53	58.980	9c	127	I	63.13	0.88	13.78	7.13	0.28	1.79	3.35	2.46	5.07	0.47	0.44	1.27	2.54	3.86	6.42
P3053760	1374	24.51	59.180	P305	376	I	69.62	0.37	13.74	3.30	0.00	0.64	2.18	2.57	6.36	0.17	0.22	0.35	1.90	2.13	8.85
P3063064	1376	24.47	59.080	P306	306.4	I	67.40	1.11	13.56	4.88	0.10	1.35	3.11	2.57	4.80	0.32	0.22	0.55	1.96	2.03	4.30
P3063270	1377	24.47	59.080	P306	327	I	63.71	1.08	14.07	6.78	0.13	1.36	3.70	3.12	4.73	0.49	0.22	0.60	2.77	3.61	6.10
P3063430	1378	24.47	59.080	P306	343	I	67.48	0.65	13.59	5.89	0.09	1.25	3.12	2.80	5.17	0.30	0.22	0.43	2.01	2.50	4.41
P3152875	1420	24.30	59.120	P315	287.5	I	67.91	0.88	13.29	5.06	0.10	1.34	3.17	2.60	4.42	0.27	0.30	0.66	1.99	2.76	4.55
P3152813	1421	24.30	59.120	P315	281.3	I	66.33	0.90	13.99	5.64	0.12	1.35	3.29	2.80	4.32	0.34	0.27	0.76	2.14	3.14	5.07
P3153087	1422	24.30	59.120	P315	308.7	I	64.97	0.94	13.56	5.48	0.13	1.51	3.40	3.00	5.53	0.52	0.30	0.68	2.51	2.94	4.93
P3032833	1368	24.61	59.120	P303	283.3	II	69.32	0.53	13.29	4.47	0.06	0.69	2.07	2.44	6.28	0.14	0.22	0.32	1.97	2.25	4.02
P3033190	1369	24.61	59.120	P303	319	II	67.54	0.66	13.80	5.46	0.08	0.84	2.65	2.50	5.27	0.31	0.22	0.56	2.20	2.80	4.92
P3033338	1370	24.61	59.120	P303	333.8	II	69.72	0.17	13.78	4.49	0.03	0.38	1.62	2.07	7.44	0.03	0.23	0.44	0.71	0.70	1.34
P3042790	1371	24.56	59.140	P304	279	II	69.91	0.26	13.11	3.64	0.06	0.50	1.62	2.65	7.40	0.04	0.22	0.56	1.74	1.71	3.28
P3043035	1372	24.56	59.140	P304	303.5	II	71.03	0.19	13.62	3.23	0.07	0.51	1.67	2.65	6.07	0.08	0.22	0.32	1.37	1.30	3.91
P3142442	1411	24.30	59.180	P314	244.2	II	72.50	0.34	11.81	3.31	0.05	0.50	1.50	2.37	6.47	0.03	0.22	0.90	1.49	1.64	2.98
P3142558	1412	24.30	59.180	P314	255.8	II	73.66	0.17	13.69	2.24	0.03	0.42	1.91	2.53	5.99	0.03	0.22	0.51	0.99	1.12	2.91
P3142976	1413	24.30	59.180	P314	297.6	II	72.09	0.34	13.10	3.91	0.09	0.46	1.50	2.38	6.48	0.02	0.22	0.90	1.67	1.85	3.36
P3142906	1415	24.30	59.180	P314	290.6	II	73.24	0.22	12.92	2.16	0.03	0.46	1.91	2.56	6.31	0.03	0.22	0.84	0.97	1.07	1.94
P3192660	1438	24.294	59.134	P319	266	II	66.02	0.60	14.06	4.47	0.07	1.26	2.31	2.76	6.47	0.13	0.22	1.43	1.99	2.23	4.03
P3193090	1439	24.294	59.134	P319	309	II	67.61	0.43	12.65	4.35	0.07	0.63	2.13	2.77	6.03	0.06	0.22	0.46	1.47	1.51	2.82
P3242760	1456	24.2	59.090	P324	276	II	66.63	0.73	13.22	6.55	0.13	1.11	3.24	2.61	4.02	0.37	0.25	0.53	2.60	3.56	5.96
P3243450	1457	24.2	59.090	P324	345	II	68.68	0.32	12.71	6.01	0.11	0.74	2.65	2.86	6.25	0.31	0.22	0.38	2.09	3.02	5.22
P3243510	1458	24.2	59.090	P324	351.2	II	68.33	0.56	12.75	6.10	0.11	0.89	3.11	2.51	4.76	0.30	0.22	0.36	2.14	3.29	5.45
P3222408	1440	24.14	59.220	P322	240.8	III	76.12	0.63	12.19	3.58	0.13	0.80	2.63	3.00	4.93	0.08	0.23	0.68	1.93	2.80	4.15
P3222707	1450	24.14	59.220	P322	270.7	III	76.14	0.37	12.89	2.60	0.07	0.79	1.99	3.13	6.19	0.03	0.22	0.46	1.20	1.26	2.44
P322760	1451	24.14	59.220	P322	276	III	76.34	0.08	12.89	1.05	0.01	0.22	1.34	4.67	3.88	0.01	0.22	0.28	0.47	0.52	0.95
P3232892	1452	24.13	59.170	P323	289.2	III	72.92	0.14	13.04	2.47	0.03	0.45	1.72	3.66	4.62	0.02	0.23	0.69	1.07	1.26	2.23
P3233042	1453	24.13	59.170	P323	304.2	III	72.49	0.25	13.27	2.11	0.04	0.53	1.71	2.80	6.00	0.03	0.22	0.54	0.94	1.09	1.90
P3233110	1454	24.13	59.170	P323	311	III	72.43	0.25	13.83	2.28	0.07	0.36	1.48	3.04	5.66	0.04	0.23	0.31	1.02	1.14	2.06
P3233300	1455	24.13	59.170	P323	333	III	71.54	0.33	13.24	2.36	0.07	0.62	1.89	3.19	3.88	0.03	0.22	0.53	1.08	1.18	2.13
P3282580	1467	24.17	59.130	P328	258	III	69.33	0.40	13.28	2.90	0.09	0.62	2.06	3.47	6.21	0.06	0.22	0.34	1.28	1.52	2.66
P3282810	1468	24.17	59.130	P328	281	III	69.84	0.44	13.33	3.06	0.07	0.58	2.31	3.30	6.25	0.06	0.22	0.46	1.43	1.47	2.70
P3283190	1469	24.17	59.130	P328	319	III	70.74	0.42	12.64	3.34	0.10	1.07	2.26	2.94	5.79	0.04	0.23	0.43	1.48	1.50	2.80

Table 1: The major elemental data [wt.%] sourced from the Kivisilla et al. (1999) dataset are organised by location, drill core, and the tripartite phases, as illustrated in Figure 2a. The data is normalised to 100%. The values for Fe_2O_3 and FeO were derived from Fe_2O_{3tot} following the method outlined by Le Maitre (1976). The calculation used for plutonic rocks is as follows: $FeO^* = Fe_2O_3(0.8998) + FeO$.

malised to 100%. Summarized data is provided in Table 1.

4. Methodology and Results

4.1. Geochemical Classification

The geochemical classification follows the phase boundaries in Figure 2a. Average values and full compositional ranges are reported to capture intra-phase variability. Elemental relations are presented in Table 2. Figure 3 highlights major elemental trends from Phase I to III

Id	Phase	Geochemical Analyses										CIPW after Buckle et al. (2023)									
		<i>Mg</i> #	<i>Na</i> ₂ <i>O</i> + <i>K</i> ₂ <i>O</i>	<i>K</i> ₂ <i>O</i> / <i>Na</i> ₂ <i>O</i>	<i>FeO</i> / <i>MgO</i>	<i>FeO</i> ²⁺ / <i>(FeO</i> ²⁺ + <i>MgO</i>)	<i>MAFI</i>	<i>ASI</i>	<i>A</i> / <i>NK</i>	<i>A</i> / <i>CNK</i>	<i>DF1</i>	<i>DF2</i>	<i>T</i> _{SiO₂} [°C]	<i>T</i> _{MgO} [°C]	<i>Qtz</i>	<i>An</i>	<i>Ab</i>	<i>Or</i>	<i>Ilm</i>	<i>Mag</i>	<i>Ap</i>
3023300	I	25.16	9.16	1.88	2.68	0.73	7.39	0.84	1.07	0.84	-3.16	-1.82	670.98	556.79	29.89	2.36	27.11	35.50	0.19	0.80	0.02
3023320	I	20.52	7.50	1.56	3.49	0.78	3.73	0.85	1.40	0.82	-1.25	-1.44	824.50	645.53	19.44	10.78	25.16	27.46	2.08	4.16	1.08
3023344	I	18.24	7.81	1.70	4.03	0.80	4.54	0.84	1.30	0.82	-1.73	-0.68	770.57	713.70	24.49	8.25	24.87	29.45	1.45	3.11	0.80
30233442	I	17.86	7.62	1.63	4.14	0.81	4.20	0.84	1.33	0.82	-1.60	-0.67	772.35	713.90	24.82	9.03	24.85	28.32	1.55	3.10	0.87
3024031	I	17.34	9.11	2.25	4.29	0.81	7.76	0.96	1.17	0.96	-3.17	-2.85	689.33	557.12	29.16	5.21	23.89	37.54	0.21	1.37	0.07
3024051	I	21.02	8.43	1.96	3.38	0.77	5.39	0.87	1.29	0.85	-1.90	-2.85	789.20	556.90	20.80	8.49	24.38	33.45	1.50	3.29	0.80
3024384	I	19.37	8.46	1.83	3.75	0.79	5.44	0.86	1.27	0.84	-1.83	-2.84	794.68	556.79	19.83	8.03	25.56	32.74	1.46	3.69	0.75
3024600	I	15.40	7.71	2.11	4.94	0.83	2.85	0.73	1.37	0.72	-1.82	-0.74	804.72	706.33	21.17	9.97	21.33	31.40	1.59	3.43	0.82
3024785	I	21.80	7.92	1.75	3.23	0.76	5.00	0.90	1.32	0.87	-1.62	-2.04	788.34	602.77	22.37	9.13	24.68	30.21	1.79	3.49	0.92
3024805	I	16.67	7.53	1.73	4.50	0.82	4.66	0.98	1.47	0.95	-1.65	-0.41	790.59	730.35	23.50	12.21	23.58	28.54	1.72	3.65	0.77
3024850	I	20.29	8.10	1.70	3.54	0.78	5.16	0.89	1.32	0.87	-1.83	-0.43	783.93	730.47	21.78	9.17	25.65	30.47	1.53	3.36	0.70
3024874	I	21.35	7.65	1.77	3.32	0.77	4.40	0.86	1.33	0.83	-1.61	-0.71	794.04	713.71	22.72	9.05	23.71	29.38	1.70	3.65	1.02
9v3250	I	18.66	6.91	1.79	3.92	0.80	3.25	0.87	1.47	0.84	-1.34	-0.41	809.23	730.83	24.03	11.59	21.40	26.80	1.91	3.95	1.09
9v3270	I	19.71	7.53	2.06	3.67	0.79	4.18	0.92	1.45	0.88	-1.73	-0.16	829.06	781.35	20.10	11.84	21.27	30.60	1.70	4.21	1.12
F3052780	I	13.45	8.84	2.43	5.79	0.85	6.65	0.89	1.20	0.88	-2.82	-0.88	737.24	699.32	25.53	6.14	21.95	37.32	0.71	2.79	0.28
F3063064	I	22.01	7.37	1.86	3.19	0.76	4.26	0.92	1.44	0.90	-1.66	-0.31	768.68	706.68	26.25	11.39	22.02	28.67	2.13	2.87	0.75
F3063270	I	16.73	7.85	1.51	4.48	0.82	4.14	0.86	1.37	0.83	-1.44	-0.60	820.82	725.50	18.86	10.53	26.76	28.27	2.08	4.06	1.15
F3063430	I	20.79	8.06	1.79	3.43	0.77	4.94	0.87	1.31	0.85	-1.88	-0.98	767.45	699.43	23.48	8.94	24.66	30.86	1.24	2.94	0.70
F3152675	I	20.44	7.02	1.70	3.40	0.77	3.85	0.92	1.47	0.90	-1.47	-0.35	761.35	706.81	27.74	11.71	22.27	26.44	1.69	2.92	0.63
F3152813	I	21.56	6.62	1.88	3.27	0.77	3.33	1.00	1.65	0.97	-1.36	-0.44	780.93	719.64	27.38	14.29	19.72	25.87	1.73	3.15	0.80
F3153087	I	21.60	8.32	1.78	3.27	0.77	4.87	0.84	1.31	0.82	-1.87	-0.40	803.90	725.36	19.00	9.02	25.70	31.89	1.81	3.40	0.75
F3032833	II	13.04	8.71	2.57	6.00	0.86	6.65	0.92	1.23	0.91	-2.78	-0.96	741.50	671.97	26.10	6.85	20.84	37.46	1.02	2.88	0.33
F3033190	II	13.31	7.77	2.11	5.86	0.85	5.12	0.96	1.41	0.94	-2.11	-0.72	766.60	691.33	25.87	11.03	21.41	31.59	1.27	3.31	0.45
F3033338	II	20.32	10.11	2.79	3.53	0.78	8.48	1.03	1.27	1.03	-3.68	-1.72	735.79	627.65	21.72	7.92	22.74	44.29	0.33	1.04	0.07
F3042790	II	12.05	10.05	2.79	6.57	0.87	8.44	0.86	1.06	0.86	-3.60	-0.77	733.10	671.83	22.87	2.12	22.67	44.16	0.50	2.54	0.09
F3043035	II	14.32	9.15	2.54	5.38	0.84	7.48	0.95	1.20	0.95	-3.21	-1.12	717.25	682.40	26.49	6.23	21.99	39.05	0.36	2.14	0.12
F3142442	II	13.09	8.84	2.72	5.97	0.86	7.34	0.87	1.08	0.87	-3.29	-0.08	696.44	659.86	30.90	2.51	20.36	38.71	0.65	2.18	0.07
F3142558	II	15.79	8.52	2.37	4.80	0.83	7.51	1.05	1.23	1.05	-3.17	-1.33	680.01	627.43	33.16	4.86	21.60	35.77	0.33	1.45	0.07
F3142676	II	10.98	8.87	2.72	7.30	0.88	7.37	0.89	1.11	0.89	-3.36	0.35	702.14	672.06	30.07	3.20	20.37	38.72	0.65	2.45	0.05
F3142906	II	17.56	8.87	2.47	4.23	0.81	7.86	1.01	1.17	1.00	-3.30	-1.06	685.90	627.40	31.44	4.87	21.90	37.74	0.42	1.43	0.07
F3192560	II	21.95	9.13	2.31	3.20	0.76	6.82	0.93	1.26	0.92	-2.65	-0.77	788.09	682.24	19.37	8.15	23.77	38.38	1.16	2.95	0.31
F3193090	II	16.67	9.69	2.74	4.50	0.82	7.47	0.81	1.07	0.81	-3.33	-0.14	723.12	682.27	24.90	2.23	21.92	41.92	0.82	2.14	0.12
F3242760	II	14.53	7.23	1.77	5.29	0.84	3.98	0.90	1.42	0.87	-1.57	-0.65	779.47	725.67	25.61	10.84	22.34	27.61	1.41	3.81	0.87
F3243450	II	8.21	9.05	2.23	10.06	0.91	6.41	0.86	1.21	0.85	-2.65	-0.86	787.62	713.40	20.14	6.56	24.08	37.47	1.00	3.94	0.45
F3243510	II	12.73	7.27	1.90	6.17	0.86	4.16	0.88	1.37	0.85	-1.76	-0.86	755.45	713.50	28.00	9.58	21.43	28.39	1.07	3.57	0.70
F3222408	III	14.93	7.92	1.64	5.13	0.84	5.29	0.81	1.19	0.81	-2.10	0.61	730.17	725.80	28.22	5.31	25.65	29.46	1.22	2.83	0.19
F3222707	III	23.32	9.42	1.92	2.96	0.75	7.43	0.89	1.16	0.89	-2.91	-0.18	729.55	681.61	23.14	5.17	27.51	36.90	0.70	1.75	0.09
F3222760	III	17.32	8.55	0.83	4.29	0.81	7.21	0.90	1.08	0.90	-2.17	-1.90	656.18	557.06	30.94	2.77	39.75	23.07	0.15	0.68	0.02
F3232892	III	15.46	8.28	1.26	4.92	0.83	6.56	0.92	1.18	0.92	-2.32	-1.10	690.41	627.81	30.02	5.55	31.29	27.62	0.27	1.57	0.05
F3233042	III	20.08	8.80	2.14	3.58	0.78	7.09	0.94	1.20	0.93	-2.99	-0.68	696.60	645.62	29.25	5.98	23.91	35.79	0.48	1.38	0.07
F3233310	III	13.64	8.70	1.86	5.70	0.85	7.21	1.00	1.24	1.00	-2.87	-0.53	697.15	682.45	29.28	7.14	25.85	33.67	0.48	1.48	0.09
F3233330	III	20.74	9.18	1.88	3.44	0.77	7.29	0.88	1.13	0.87	-2.93	0.05	710.02	682.16	26.14	4.15	27.25	35.68	0.63	1.58	0.07
F3282580	III	17.22	9.69	1.79	4.32	0.81	7.63	0.82	1.07	0.82	-2.90	-0.38	727.12	671.53	22.53	2.31	29.62	36.90	0.76	2.92	0.12
F3282810	III	15.89	9.64	1.84	4.76	0.83	7.32	0.81	1.08	0.81	-2.89	-0.15	734.12	682.18	22.11	2.72	28.95	37.22	0.84	2.09	0.12
F3283100	III	24.32	8.73	1.97	2.80	0.74	6.46	0.83	1.14	0.83	-2.68	0.44	721.26	706.96	26.24	4.26	25.05	34.51	0.81	2.16	0.09

Table 2: Analysis of major elemental relationships in Tripartite rapakivi samples, including temperature, pressure and *CIPW*. $Mg\#_{[wt.\%]} = 100 \cdot \frac{MgO}{MgO+FeO^*}$; $MALI_{[wt.\%]} = (Na_2O + K_2O) - CaO$; $ASI_{[mol]} = \frac{Al_2O_3}{CaO+Na_2O+K_2O}$; $A/NK_{[mol]} = \frac{Al_2O_3}{Na_2O+K_2O}$. Multi-dimensional discriminant functions DF1 and DF2, based on tectonic analysis, after [Verma et al. \(2013\)](#). $DF1 = 0.051 \ln \frac{TiO_2}{SiO_2} + 0.226 \ln \frac{Al_2O_3}{SiO_2} - 1.77 \ln \frac{Fe_2O_3}{SiO_2} + 1.83 \ln \frac{FeO}{SiO_2} - 0.065 \ln \frac{MnO}{SiO_2} + 0.134 \ln \frac{MgO}{SiO_2} + 0.225 \ln \frac{CaO}{SiO_2} + 0.742 \ln \frac{Na_2O}{SiO_2} - 1.78 \ln \frac{K_2O}{SiO_2} + 0.146 \ln \frac{P_2O_5}{SiO_2} - 2.12$. & $DF2 = 1.09 \ln \frac{TiO_2}{SiO_2} - 1.65 \ln \frac{Al_2O_3}{SiO_2} - 1.19 \ln \frac{Fe_2O_3}{SiO_2} + 1.03 \ln \frac{FeO}{SiO_2} + 0.82 \ln \frac{MnO}{SiO_2} + 0.026 \ln \frac{MgO}{SiO_2} + 0.023 \ln \frac{CaO}{SiO_2} + 0.212 \ln \frac{Na_2O}{SiO_2} + 0.085 \ln \frac{K_2O}{SiO_2} - 0.85 \ln \frac{P_2O_5}{SiO_2} + 2.54$. *CIPW* Abbreviations: *Qtz*=Quartz, *An*=Anorthite, *Ab*=Albite, *Or*=Orthoclase, *Ilm*=Ilmenite, *Mag*=Magnetite, *Ap*=Apatite

163 SiO_2 content increases with magmatic evolution: Phase I averages 66.69%
164 (63.13–74.30%), Phase II 69.76% (66.02–73.66%), and Phase III 71.59% (69.84–75.34%).
165 Al_2O_3 remains stable but slightly declines: 13.51% (12.56–14.21%) in Phase I,
166 13.31% (11.81–15.78%) in Phase II, and 13.16% (12.19–13.89%) in Phase III.
167 These values fall within the typical 11–14 *wt.*% for A-type granites ([Dall’Agnol](#)
168 [and de Oliveira, 2007](#)), though up to 17 *wt.*% has been observed ([Duchesne](#)
169 [et al., 2010](#); [Grabarczyk et al., 2023](#)). The alignment of samples along the
170 $CaO/(FeO + MgO + TiO_2)$ vs. Al_2O_3 trend confirms their A-type affinity
171 ([Dall’Agnol and de Oliveira, 2007](#)) (Figs.S1).
172 TiO_2 and FeO^* contents decrease on average from Phase I to III, with TiO_2
173 dropping from an average of 0.79% to 0.33% and FeO^* from 4.80% to 2.41%.
174 MnO also shows a slight average reduction from 0.09% in Phase I to 0.06% in
175 Phase III. The $Mg\#$ displays a non-linear pattern, decreasing from an average
176 of 19.55 in Phase I to 14.61 in Phase II, then increasing to 18.29 in Phase III,
177 with a wide range across phases (8.21–25.16).
178 Alkaline oxides reflect progressive fractionation: CaO averages decrease from
179 3.1% in Phase I to 1.94% in Phase III, while Na_2O averages increase to 3.34%
180 in Phase III. K_2O peaks in Phase II at an average of 6.23%. Incompatible
181 components show clear range contractions and average declines: P_2O_5 falls from
182 0.32% (range 0.01–0.49%) in Phase I to 0.04% (range 0.01–0.08%) in Phase III,
183 while SO_3 content also decreases from Phase I to III, but with average values
184 of $\sim 0.22\%$ for all phases. LOI values reflect a declining trend from Phase I to
185 III, consistent with progressive differentiation and volatile loss.

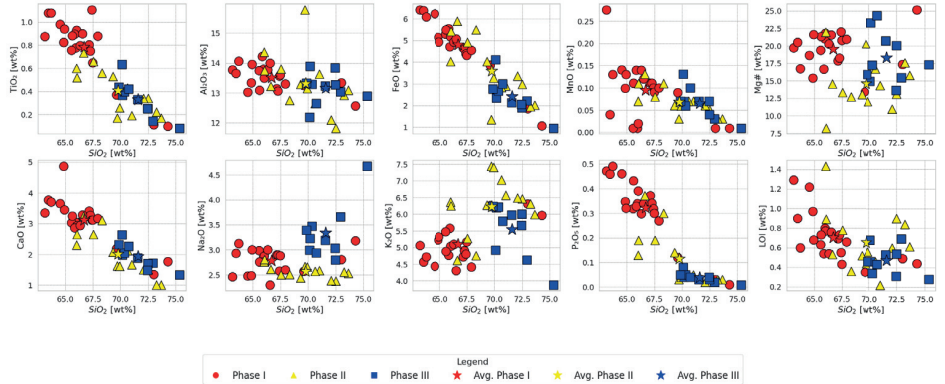


Figure 3: Major element distribution of samples from Kivisilla et al. (1999), categorized into the three phases of the analysed rapakivi granites, shown as Harker bi-plots comparing SiO_2 with other major elements.

186 The granitic samples were analysed using the updated CIPW normative cal-
 187 culation via the Python-based *webNORM* application (Buckle et al., 2023),
 188 available at <https://webnorm.streamlit.app/>. This tool builds on the algo-
 189 rithm of Verma et al. (2003) and applies the *Fe*-correction method for plutonic
 190 rocks recommended by Le Maitre (1976), which is widely used for igneous rock
 191 classification. However, CIPW norms represent an anhydrous, idealised mineral
 192 assemblage and do not account for the hydrous phases (e.g., biotite, hornblende),
 193 as it is present in the Märjamaa–Kloostri rapakivi granites (Klein et al., 1994;
 194 Soesoo, 1993). Consequently, while normative mineral proportions (e.g., quartz,
 195 feldspar, Fe–Ti oxides) provide a useful framework for comparing bulk composi-
 196 tional trends and classification, CIPW-phases like pyroxenes and corundum are
 197 excluded, as they are not represented in the typical rapakivi mineral assemblage
 198 (Andersson and Eklund, 1994; Eklund and Shebanov, 1999); they cannot substi-
 199 tute for measured mineral chemistry when interpreting crystallisation conditions
 200 or magma evolution. For this reason, the normative data presented here are used
 201 exclusively to describe bulk chemical trends and feldspar–quartz–oxide propor-
 202 tions, without attempting to derive thermobarometric parameters (P – T – H_2O)

or detailed fractionation pathways, which require mineralogical analyses (Klein et al., 1994; Eklund et al., 1994; Andersson and Eklund, 1994).

Normative quartz increases with magmatic evolution, from 23.44% (18.86–29.89%) in Phase I to 26.19% (19.37–33.16%) in Phase II and 26.79% (22.11–30.94%) in Phase III.

Feldspars: Normative albite increases progressively from 23.83% (19.72–27.11%) in Phase I and 21.96% (20.36–24.08%) in Phase II to 28.48% (23.91–39.75%) in Phase III, while anorthite declines from 9.39% (2.36–14.29%) in Phase I to 4.54% (2.31–7.14%) in Phase III. Orthoclase peaks in Phase II at 37.23% (27.61–44.29%), compared to 30.53% (25.87–37.54%) in Phase I and 33.09% (23.07–37.22%) in Phase III.

Oxides: Normative magnetite decreases from 3.21% (0.80–4.21%) in Phase I to 2.56% (1.04–3.94%) in Phase II and 1.75% (0.68–2.83%) in Phase III. Ilmenite shows a similar trend, decreasing from 1.51% (0.19–2.13%) in Phase I to 0.79% (0.33–1.41%) in Phase II and 0.63% (0.15–1.22%) in Phase III.

Phosphates: Normative apatite also declines markedly from early to late phases—0.76% (0.02–1.15%) in Phase I, 0.27% (0.05–0.87%) in Phase II, and 0.09% (0.02–0.19%) in Phase III.

A correlation matrix of CIPW normative mineral compositions for the Märrjamaa and Kloostri granitoids (Fig. 4) highlights systematic mineralogical normative associations across Phases I–III

Phase I (Fig. 4a) shows a strong oxide–phosphate cluster (magnetite, ilmenite, apatite) with strong positive intercorrelations, reflecting their co-crystallisation from a relatively oxidised melt. Quartz exhibits negative correlations with these phases, suggesting silica enrichment occurs at the expense of Fe–Ti oxides and apatite. Feldspar correlations are decoupled: orthoclase negatively correlates with anorthite and oxides, while albite shows mostly weak negative correlations across phases, indicating limited compatibility during early crystallisation.

Phase II (Fig. 4b) displays weaker overall correlations, but the oxide–phosphate cluster (magnetite, ilmenite, apatite) persists, albeit less coherently. Quartz maintains its negative association with oxides and apatite. Anorthite shows

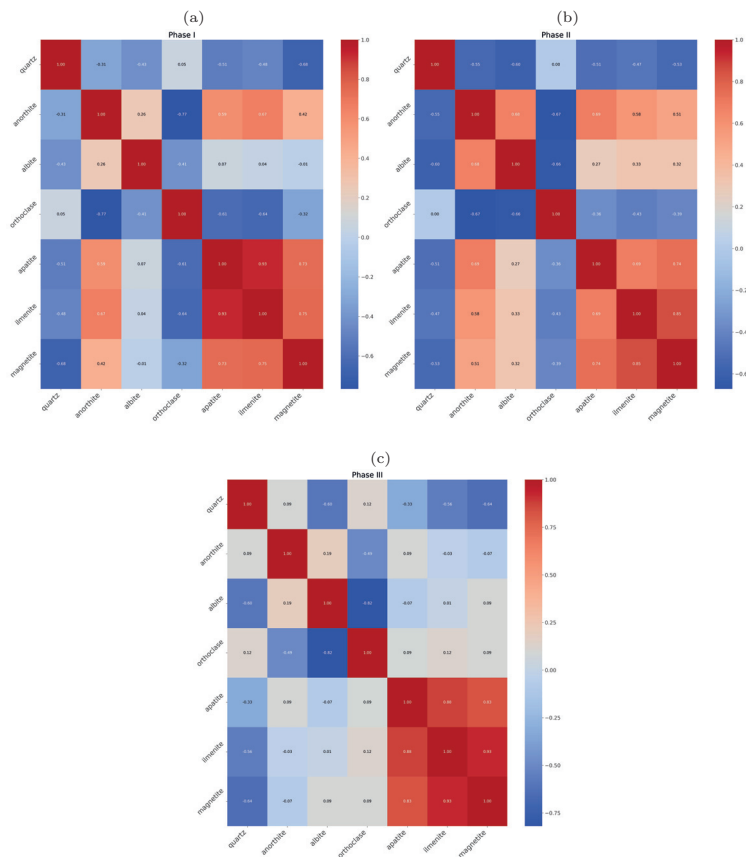


Figure 4: Comparative analysis of mineral correlation matrices across three phases: (a) Phase I matrix, (b) Phase II matrix, and (c) Phase III matrix.

234 moderate positive correlations with apatite and ilmenite, while orthoclase ex-
 235 hibits negative correlations with these phases, reflecting geochemical partition-
 236 ing between Ca-rich and alkali-rich components. Albite shows more variable
 237 correlations, with some positive links to anorthite and apatite, indicating pro-
 238 gressive plagioclase–alkali feldspar interaction during crystallisation.

239 Phase III (Fig.4c) reveals a felsic, evolved assemblage. Quartz shows posi-
 240 tive correlations with feldspars (albite, anorthite) and weak to negative corre-
 241 lations with oxides, reflecting its strong association with late-stage silica–alkali
 242 enrichment. Albite strongly correlates with quartz and anorthite, indicating a
 243 consolidated felsic framework, but negatively with oxides and apatite. Ortho-
 244 clase shows a shift to positive correlations with oxides and apatite, suggesting
 245 late-stage coupling of alkalis with Fe–Ti–P-bearing phases. Magnetite and il-
 246 menite remain tightly intercorrelated, maintaining their role as indicators of an
 247 oxidised residual melt.

248 $An - Ab - Or$ (Barker, 1979; Luo et al., 2024) (Fig.5a) and *TAS* diagrams
 249 (Le Maitre, 2002) (Fig.5b), plot Märijamaa and Kloostri samples evolve from
 250 quartz-monzonites to granodiorites (Phase I) to granite fields (Phases II and III
 251), consistent with Fennoscandian rapakivi suites trends (Kosunen, 1999; Rämö
 252 and Haapala, 1995; Haapala et al., 2005). Alkali content ($Na_2O + K_2O$) in-
 253 creases progressively: 7.88% (6.62–9.16%) in Phase I, 8.80% (7.23–10.11%) in
 254 Phase II, and 8.89% (7.92–9.69%) in Phase III, matching Finnish rapakivi trends
 255 (Fig.5b). In the K_2O vs. SiO_2 diagram (Rickwood, 1989), all samples fall within
 256 the high-*K* alkaline series, similar to Finnish rapakivi (Fig.S2; Kosunen (1999)).

257 Samples range from alkali-calcic to alkali ($MALI > 3.8$), with *ASI* mostly <
 258 1.0, indicating a dominantly metaluminous character. Phase I shows *MALI* 4.76
 259 (2.85–7.76), *ASI* 0.88 (0.73–1.00); Phase II, 6.79 (3.98–8.48), 0.92 (0.81–1.05);
 260 and Phase III, 6.95 (5.29–7.63), 0.88 (0.81–1.00), reflecting increasing alkalinity
 261 and metaluminous to marginally peraluminous compositions (Table2; Figs.5c,
 262 5d), unlike Finnish rapakivi granites, which range from metaluminous to pera-
 263 luminous (Rämö and Haapala, 1995, 2005).

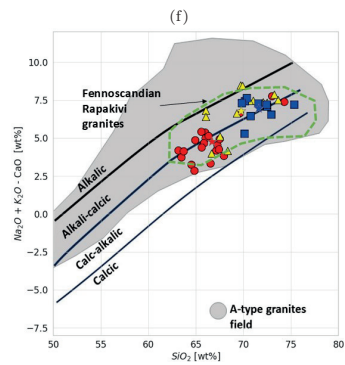
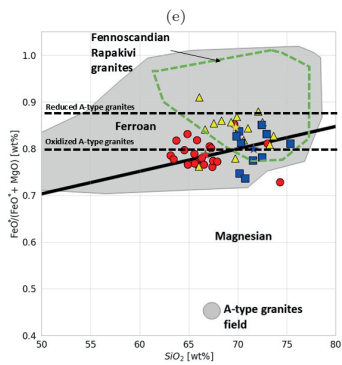
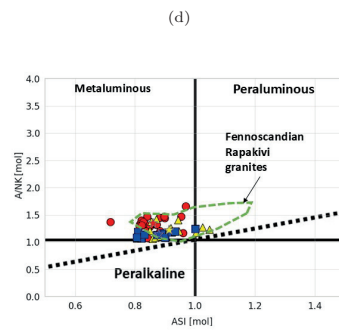
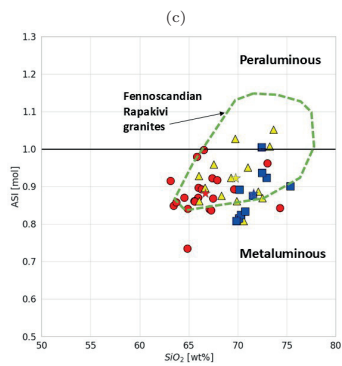
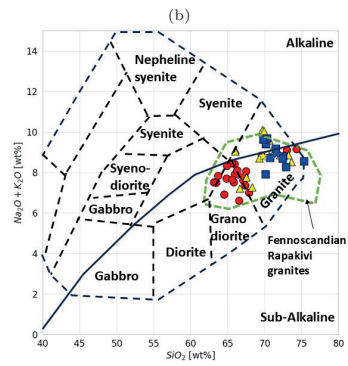
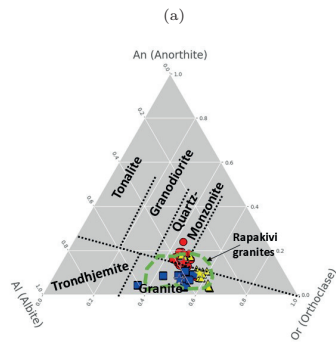


Figure 5: Geochemical classification of the Märjamaa and Kloostri tripartite phases of rapakivi granitoids: (a) $An - Ab - Or$ diagram from Barker (1979) with rapakivi classification after Luo et al. (2024); (b) $K_2O + Na_2O$ vs. SiO_2 diagram from Cox et al. (1979) after Kosunen (1999); (c) ASI vs. SiO_2 after Christiansen et al. (2007); (d) A/CNK vs. A/NK diagram from Maniar and Piccoli (1989); (e) $FeO^*/(FeO^* + MgO)$ vs. SiO_2 diagram after Frost et al. (2001), modified from Hinchey et al. (2024); and (f) $(Na_2O + K_2O - CaO)$ vs. SiO_2 diagram after Frost et al. (2001). Sources of Fennoscandian rapakivi plutons are based on the work of Dall’Agnol and de Oliveira (2007) and Grabarczyk et al. (2023), utilizing results from Rämö and Haapala (1995, 2005) for the Wiborg area, Kosunen (1999) for the Bodom and Obbnäs rapakivi plutons, Heinonen et al. (2010b) for the Ahvenisto body, and Sharkov (2010) for the Salmi granitoids.

Most samples fall within the ferroan A-type granite field (Figs.5e, 5f), overlapping with typical Finnish rapakivi suites (Frost et al., 2001; Soesoo and Hade, 2012; Grabarczyk et al., 2023; Hinchey et al., 2024).

4.2. Petrogenesis

A-type granites in the Laurentia–Baltica region are classified into ilmenite-, magnetite-, and two-mica (peraluminous) types (Ishihara, 1977; Anderson and Morrison, 2005; Dall’Agnol and de Oliveira, 2007). In Laurentia, magnetite-series granites typically have $FeO^*/(FeO^* + MgO)$ ratios of 0.80–0.88, while ilmenite-series granites exceed 0.88, indicating reduced conditions (Frost and Frost, 1997; Anderson and Morrison, 2005). Finnish rapakivi granites span a broader range (0.79–1.00), often associated with reduced A-type compositions (Rämö and Haapala, 1995, 2005; Grabarczyk et al., 2023), though their relationship with oxidised A-types is debated (Bonin, 2007; Dall’Agnol and de Oliveira, 2007; Dall’Agnol et al., 2012).

The Märjamaa and Kloostri intrusions predominantly reflect oxidised conditions, with $FeO^*/(FeO^* + MgO)$ ratios of 0.73–0.85 (*avg.*0.79) in Phase I, 0.76–0.91 (*avg.*0.84) in Phase II, and 0.74–0.85 (*avg.*0.80) in Phase III. These ratios fall within the oxidised field in Fig.5e (Frost et al., 2001; Hinchey et al., 2024), matching same classification fields in Figure 6a (Dall’Agnol and de Oliveira, 2007).

FeO^*/MgO ratios distinguish melt evolution, with values < 4 indicating unfractionated and 4–16 marking fractionated felsic granites, typical of A-type systems (Whalen et al., 1987; Bonin, 2007). In the analysed samples: 3.79

(2.68–5.79) in Phase I suggests less fractionated melts; 5.63 (3.20–10.06) in Phase II marks peak fractionation; and 4.19 (2.80–5.70) in Phase III indicates moderate to borderline fractionation.

From experimental data (Patino Douce, 1997; Patiño Douce, 1999), A-type granites form through low-pressure (≤ 5 kbar) hybridization of calc-alkaline granitoids with high-Al olivine tholeiites or metagraywackes, producing melts with plagioclase and orthopyroxene as residues (Dall’Agnol and de Oliveira, 2007). Our samples plot along this low-pressure trend in the CaO/Al_2O_3 vs. $CaO + Al_2O_3$ diagram (Fig.6b), matching Finnish rapakivi granites. Higher-pressure (12–15 kbar) melts yield clinopyroxene-rich residues and plot with calc-alkaline granitoids (Dall’Agnol and de Oliveira, 2007).

Duan et al. (2022) critique zircon-saturation thermometry for estimating magma temperatures due to limited Zr influence in rock compositions, despite its use in rapakivi studies (e.g., Grabarczyk et al. (2023)). They propose two formulas for granitoid studies: $T_{SiO_2}[^{\circ}C] = -14.16 \times SiO_2[wt.\%] + 1723$ and $T_{MgO}[^{\circ}C] = 887.6 \times (MgO[wt.\%])^{0.0989}$. These models effectively represent granitoid temperatures in mixtures of melt and crystals prior to solidification (Duan et al. (2022)). Comparative analyses reveal a decreasing temperature trend from Phase I to Phase III (Table 2). Notably, the estimates for T_{SiO_2} shows a decreasing trend from Phase I at $778.66^{\circ}C$ (670.98 – $829.06^{\circ}C$), to Phase II at $735.18^{\circ}C$ (680.01 – $788.09^{\circ}C$), and Phase III at $709.26^{\circ}C$ (656.18 – $734.12^{\circ}C$). Meanwhile, T_{MgO} estimates are approximately $100^{\circ}C$ lower for each phase.

The multidimensional discrimination tectonic diagram for acid rocks, based on the analysis by Verma et al. (2013), has been well-established in A-type complexes (Verma et al., 2022) (Fig.6c). This setting is defined by two functions, $DF1$ and $DF2$, as presented in Table 2. Analysed samples plot mostly within the continental rift domain.

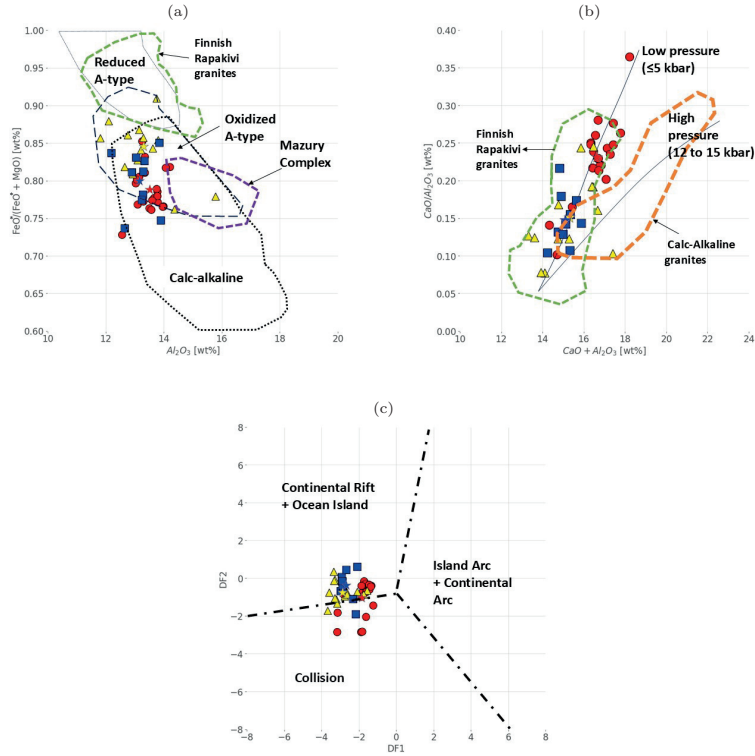


Figure 6: Petrogenetic classification diagrams for the tripartite Märjamaa and Kloostri rapakivi granite phases. (a) $FeO^*/(FeO^* + MgO)$ vs. Al_2O_3 classification plot of calc-alkaline and A-type granites, including fields for reduced and oxidised subtypes, based on [Dall'Agnol and de Oliveira \(2007\)](#). (b) Whole-rock CaO/Al_2O_3 vs. $CaO + Al_2O_3$ for pressure ranges in A-type granites, after [\(Patino Douce, 1997; Patiño Douce, 1999\)](#) modified from [Dall'Agnol and de Oliveira \(2007\)](#). Sources of the Finnish rapakivi plutons are derived from the work of [Dall'Agnol and de Oliveira \(2007\)](#) and references therein, while the domain corresponding to the Mazury complex domain follows [Grabarczyk et al. \(2023\)](#). (c) Discriminant-function (DF1 and DF2; Table 2) multidimensional diagrams based on log-transformed major and trace element ratios for tectonic discrimination of acid granitic rocks, after [Verma et al. \(2013\)](#).

314 5. Discussion

315 Indices based on major elements are fundamental for classifying A-type gran-
 316 ites, as they provide insights into crystallization history and parental magma
 317 composition (Frost et al., 2001; Bonin, 2007; Dall’Agnol et al., 2012). These
 318 granites are subdivided into reduced and oxidised suites, with crystallisation
 319 conditions largely governed by oxygen fugacity (fO_2) and water content (Ishi-
 320 hara, 1977; Dall’Agnol and de Oliveira, 2007; Frost and Frost, 1997). Experi-
 321 mental studies demonstrate that their formation depends on fO_2 , H_2O content,
 322 and pressure at the magma source (Rajesh, 2000; Dall’Agnol and de Oliveira,
 323 2007; Grabarczyk et al., 2023). Oxidised A-type magmas typically derive from
 324 hydrous (≥ 4 wt.%) quartz–feldspathic lower crustal melts under oxidising con-
 325 ditions, leaving clinopyroxene as a key residual phase. In contrast, reduced A-
 326 type granites may originate from quartz–feldspathic sources with metasedimen-
 327 tary components or from evolved tholeiitic magmas (Dall’Agnol and de Oliveira,
 328 2007). These contrasting sources generate distinct geochemical and petrological
 329 signatures (Clemens et al., 1986; Frost and Frost, 1997; Dall’Agnol et al., 1999).

330 Proterozoic increases in O_2 levels likely influenced the redox state of fluids
 331 and magmatism (Vigneresse, 2005). Oxidised granites crystallising near the
 332 Ni – NiO (NNO) buffer form magnetite-bearing assemblages at oxygen fugac-
 333 ities slightly above the Fayalite-Magnetite-Quartz (FMQ) buffer (Dall’Agnol
 334 et al., 2005; Dall’Agnol and de Oliveira, 2007), as documented in Fennoscan-
 335 dia (Grabarczyk et al., 2023) and other Phanerozoic shear-controlled oxidised
 336 A-type granites (Rajesh, 2000). The Märjamaa and Kloostri suites crystallized
 337 under similarly oxidised conditions (Figs. 5e, 6a), reflecting a petrogenetic evo-
 338 lution influenced by magma source, fO_2 , and H_2O content. These conditions
 339 favoured the crystallisation of H_2O -bearing silicates such as biotite and horn-
 340 blende (Soesoo and Niin, 1992; Soesoo, 1993; Klein et al., 1994), underscoring
 341 the combined roles of crustal sources and water content in shaping oxidised A-
 342 type signatures (Frost and Frost, 1997; Bonin, 2007; Dall’Agnol and de Oliveira,
 343 2007; Dall’Agnol et al., 2012).

344 The Polish Mazury Complex granitoids (Fig.1a) exhibit a distinctly oxidised
 345 signature (Fig.6a) (Grabarczyk et al., 2023), with elevated alumina contrasting
 346 with the more ferric Märjamaa–Kloostri granites. They formed through partial
 347 melting of dark-grey granodiorite at mid- to lower-crustal depths, producing
 348 F-rich quartz monzonitic magmas that ascended along lithospheric weaknesses,
 349 such as the Białystok fault (Duchesne et al., 2010; Grabarczyk et al., 2023).
 350 A similar evolution is inferred for the Märjamaa–Kloostri intrusions, emplaced
 351 within the PPDZ corridor (Fig.1c). In both complexes, fluorine, titanite, and
 352 apatite—elements and minerals that enhance oxidised signatures—are abun-
 353 dant, highlighting their role in magmatic differentiation and redox evolution
 354 (Soesoo and Niin, 1992; Klein et al., 1994; Kirs et al., 2004; Grabarczyk et al.,
 355 2023).

356 Although redox states in A-type granites are influenced by host rocks, melt
 357 composition, and crystallisation conditions, the imprint of magma sources re-
 358 mains the primary control on their geochemical and petrological contrasts (Patiño Douce,
 359 1999; Frost and Frost, 1997; Dall’Agnol and de Oliveira, 2007; Dall’Agnol et al.,
 360 2012). In Finnish rapakivi intrusions, significant high-level contamination is
 361 unlikely (e.g., Heinonen et al. 2010a), supporting a source-dominated redox
 362 control. Thus, these conditions do not define strictly oxidised granites in the
 363 narrow sense, as no absolute correspondence exists between oxidised and reduced
 364 A-type suites (Dall’Agnol and de Oliveira, 2007); however, normative magnetite
 365 contents in Märjamaa–Kloostri are notably higher than ilmenite. This oxidised
 366 trend contrasts with the reduced Wiborg, Bodom, Obbnäs, and Ahvenisto gran-
 367 itoids (Figs.5e, 6a) (Rämö and Haapala, 1995, 2005; Kosunen, 1999).

368 Crystallization of Märjamaa and Kloostri rapakivi granites occurred in three
 369 distinct stages, marking a progression of differentiation from Phase I to Phase
 370 III. Phase I of Märjamaa is enriched in magnetite, ilmenite, and apatite, reflect-
 371 ing a $Fe-Ti-P$ -rich CIPW normative composition. These components progres-
 372 sively decline in Phases II and III, indicating a felsic evolutionary trend driven
 373 by feldspar-dominated differentiation. This transition marks a shift from ox-
 374 ide–phosphate–rich assemblages to quartz–albite–orthoclase enrichment (Fig.4).

Phase I exhibits strong clustering of oxides and apatite, which weakens in Phase II and becomes decoupled in Phase III, where quartz and alkali feldspar dominate. This progression reflects fractional crystallisation under evolving redox conditions and changing mineral compatibilities, corroborated by petrographic evidence of iron-rich accessory phases across all stages (Soesoo and Niin, 1992; Klein et al., 1994; Raukas and Teedumäe, 1997).

Thermobarometric estimates for the Märjamaa and Kloostri intrusions are consistent with other Fennoscandian rapakivi complexes. Klein et al. (1994) applied Buddington’s and Barth’s geothermometers to amphiboles, yielding crystallization temperatures of 550–650°C and pressures of 5.2–6.5 kbar. Our T_{SiO_2} results suggest higher crystallization temperatures, whereas T_{MgO} values align with these estimates (Table 2). Regional potential field profiles indicate that the Märjamaa and Kloostri granitoids extend to depths of up to 10 km (All et al., 2004; Solano-Acosta et al., 2023). Comparable conditions are reported for other rapakivi systems, including magma chambers at ~ 20 km depth with emplacement at ~ 4 km (Bonin, 2007), isothermal ascent from ~ 15 km at ~ 6 kbar (Vigneresse, 2005), and crystallisation temperatures between 450–780°C. Rapakivi batholiths across Fennoscandia typically formed at 2–8 kbar (7–20 km) (Sharkov and Bogina, 2006), such as the Salmi massif (2–3 kbar) and the Korosten pluton (3–4 kbar), consistent with zoned pegmatites at 2.6–3.0 kbar (Sharkov, 2010). The Wiborg pluton crystallized at 2.5–5.4 kbar based on hornblende Al content (Elliott, 2001), while zircon saturation thermometry indicates $> 840^\circ\text{C}$ for the Mazury Complex granitoids, crystallized at up to 4 kbar (Grabarczyk et al., 2023). The Riga pluton, Baltica’s largest anorthosite–rapakivi massif, formed at 14–17 km depth and 5–7 kbar (Rämö et al., 1996; Kirs et al., 2004; Sharkov, 2010). Subsolidus double-layer coronas in associated gabbroids further suggest that some anorthosites were entrained upward by granitic magmas from deeper crustal levels (Sharkov et al., 2004; Sharkov, 2010). Large composite rapakivi bodies dominate central subprovinces, whereas smaller intrusions and diabase dikes prevail at the margins (Puura and Flodén, 1999; All et al., 2006). The Märjamaa–Kloostri granitoids, linked to the early Wiborg suite, likely repre-

sent late-stage differentiates with enhanced oxidation (Puura and Flodén, 1996, 1999, 2000). Gneiss xenoliths in Phase I of Märjamaa highlight such interactions (Soesoo and Niin, 1992), suggesting significant assimilation and H_2O input from host rocks in the studied bodies.

Future research should prioritise comparative analyses of fluorine-bearing assemblages, xenoliths, and host rocks to better constrain oxidation states, which appear more influenced by local emplacement conditions, melt evolution, and source lithology than by regional trends. This requires integrated studies of fluid inclusions in quartz and fluorite, whole-rock REE geochemistry, and detailed mineral chemistry (biotite, magnetite, feldspars). Core F314 (Fig.2), showing a contact between Phase II and biotite–amphibole gneiss, underscores the need for further investigation of host-rock assimilation. In parallel, high-precision $U\text{--}Pb$ zircon and monazite geochronology should refine the crystallisation sequence of the three phases, complemented by advanced thermobarometry, isotopic analyses, and assessments of magmatic evolution in relation to post-thermal tectonic relaxation. These approaches will significantly advance our understanding of the Märjamaa–Kloostri system and, more broadly, rapakivi granite emplacement within the Estonian Precambrian basement and the Fennoscandian Shield.

6. Conclusions

The Märjamaa and Kloostri rapakivi granites formed through three distinct phases, reflecting the progression of magmatic differentiation from Phase I to Phase III. Phase I represents the least evolved melts, characterised by lower SiO_2 , higher TiO_2 , FeO^* , P_2O_5 , and abundant normative magnetite, ilmenite, and apatite, reflecting a Fe–Ti–P-rich, moderately fractionated oxidised composition. Phase II shows peak K_2O , elevated orthoclase, and reduced oxide–phosphate contents, marking intensified alkali enrichment and advanced feldspar-dominated fractionation. Phase III records the most evolved stage, with the highest SiO_2 and albite, lowest oxides and phosphates, and declining theoretical temperatures, indicating a silica- and alkali-rich residual melt dominated

by quartz–feldspar assemblages. Together, these trends define a multi-stage, oxidised A-type magmatic system evolving through progressive fractional crystallisation.

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8. Declaration of competing interest

The authors state that they have no known competing financial interests or personal ties that may have influenced the work described in this study.

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Appendix V

J D Solano-Acosta, S Graul, A Soesoo, R Hints, T All, J Vind, L Maala, and S Liivamägi. Comparative geochemistry, provenance, and tectonic evolution of Paleoproterozoic metavolcanic and metasedimentary units in the Alutaguse Zone (North Estonia) and Southern Svecofennian domains (Russia–Finland): Implications for Eastern Fennoscandian crustal development. 2025. Manuscript Under review

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Fennoscandian Crustal Development

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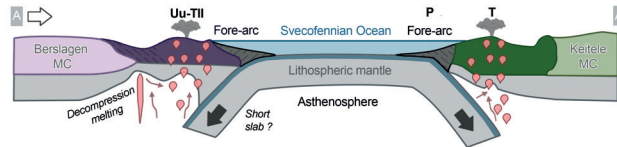
Graphical Abstract

Comparative Geochemistry, Provenance, and Tectonic Evolution of Paleoproterozoic Metavolcanic and Metasedimentary Units in the Alutaguse Zone (North Estonia) and Southern Svecofennian Domains (Russia–Finland): Implications for Eastern Fennoscandian Crustal Development

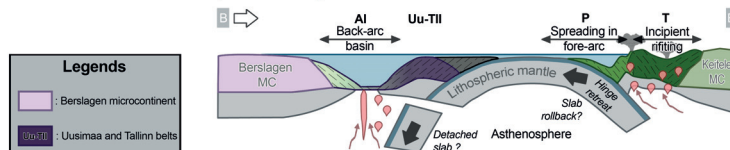
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Graphical Abstract: Schematic geodynamic model cross-sections illustrating the evolution of Fennoscandia within the North Estonia and South Finland Precambrian domains for the time intervals: 1) 1.92–1.90 Ga; 2) 1.90–1.89 Ga; 3) 1.89–1.87 Ga; 4) 1.87–1.86 Ga; and 5) 1.84–1.83 Ga.

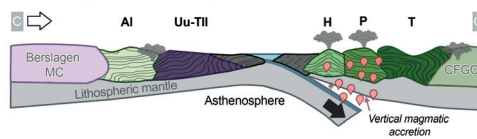
1. 1.91-1.90 Ga: Divergent double subduction system



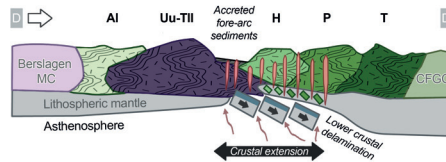
2. 1.90-1.89 Ga : Multiple arc-back systems



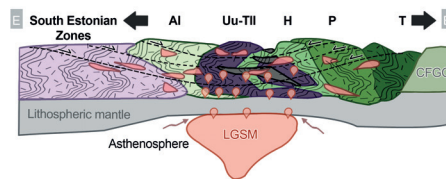
3. 1.89-1.87 Ga: Single subduction system



4. 1.87-1.86 Ga: Collision



5. 1.84-1.83 Ga: Post collisional rifting



Legends	
	: Berslagen microcontinent
	: Uusimaa and Tallinn belts
	: Alutaguse
	: Hääme belt
	: Pirkkanmaa belt
	: Tampere belt
	: CFGC
	: For-arc sediments
	: Oceanic crust
	: Relative plate motion
	: Tectonic transport
	: Magmatic bodies
	: Mantle flow

Highlights

Comparative Geochemistry, Provenance, and Tectonic Evolution of Paleoproterozoic Metavolcanic and Metasedimentary Units in the Alutaguse Zone (North Estonia) and Southern Svecofennian Domains (Russia–Finland): Implications for Eastern Fennoscandian Crustal Development

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- Geochemistry shows Alutaguse formed as a Paleoproterozoic back-arc basin.
- High- and low-silica sediments record mixed arc and rift sources in Fennoscandia.
- Tectonic proxies link Alutaguse to the Uusimaa–Bergslagen margin at 1.90–1.87 Ga.
- Data show Svecofennian double subduction toward the Bergslagen–Fennoscandian margin.

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Abstract

This study investigates the geochemistry of Paleoproterozoic metasedimentary and metavolcanic units from the North Estonian Alutaguse zone and South Svecofennian (SS) domains (Ladoga, Saimaa, Häme, Uusimaa), to elucidate the tectonic evolution of the Svecofennian Orogeny in Eastern Fennoscandia. Representative metasediments include micaceous gneisses ($\pm Grt \pm Crd \pm Sil$), subdivided into High- SiO_2 ($> 63wt\%$) and Low- SiO_2 ($\leq 63wt\%$), while metavolcanics comprise amphibolites and pyroxenic gneisses. Weathering and alteration indices confirm dataset reliability for provenance and tectonic analysis. Major elemental data in High- SiO_2 metasediments indicate felsic higher affinities with

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arc-related settings, whereas Low- SiO_2 types reflect mafic–intermediate sources akin to rift-related environments. Metasedimentary trace elemental data in general enhance a continental island arc setting. Metavolcanics are sub-alkaline, with $La/Yb-Zr/Nb$ relations indicating enrichment from Ladoga southwards and mixed-depleted sources in Uusimaa and Alutaguse. Th/Nb and Nb/Zr ratios suggest asthenospheric mantle sources for Alutaguse and subducted oceanic crust for SS. Tectonic proxies show dominant oceanic arc affinities. Elevated $CaO-MnO$ in Alutaguse and Uusimaa High- and Low- SiO_2 metasediments, and $CaO-MgO$ in metavolcanics, suggest a genetic link between the two domains, positioning Alutaguse as a 1.90-1.89 Ga back-arc to the (Tallinn)-Uusimaa Belt, situated in the northern part of the Bergslagen microcontinent. Subsequent accretion of the Tallinn–Uusimaa belt to proto-Fennoscandia (e.g., Keitele microcontinent) at 1.89–1.87 Ga closed the Alutaguse basin and culminated in the final closure of the Svecofennian ocean.

Keywords:

Svecofennian Orogeny (SO), Estonian Alutaguse domain, South Svecofennian (SS) zone, Bergslagen microcontinent, Whole-rock geochemical analysis, Metasediment & Metavolcanic units

0 1. Introduction

1 The Estonian Paleo- to Mesoproterozoic basement, entirely buried beneath
2 Lower Palaeozoic cover and lacking surface outcrops, remains only partly stud-
3 ied, with its geology reconstructed solely from drill-core material supported
4 by geophysical and geochemical datasets (Puura and Huhma, 1993; Kivisilla
5 et al., 1999; Soesoo et al., 2004, 2020; All et al., 2004; Puura et al., 2004; Bog-
6 danova et al., 2015; Nirgi and Soesoo, 2021; Solano-Acosta et al., 2023, 2025a,b).
7 Despite limited exposure, Paleoproterozoic metavolcanic and metasedimentary
8 rocks have been identified in northern Estonia (i.e. Tallinn, Alutaguse, and
9 Jõhvi domains; Fig.1). These rocks are comparable to similar units found in
10 southwestern Finland and the Bergslagen region of Sweden (Raukas and Tee-
11 dumäe, 1997; Puura et al., 2004; Kirs et al., 2009; Beunk and Kuipers, 2012;

12 Bogdanova et al., 2015; Soesoo et al., 2020; Solano-Acosta et al., 2025a). How-
13 ever, more data are needed to establish reliable correlations among the various
14 Estonian domains.

15 The present Baltica (or East European Craton) comprises the amalgamated
16 Archean–Paleoproterozoic cratons of Fennoscandia, Volgo Uralia, and Sarma-
17 tia (Fig.1a), forming the stable core of northeastern Europe (Bogdanova et al.,
18 2008, 2015; Pesonen et al., 2012; Nironen, 2017). Positioned at the centre of the
19 Nuna (Columbia) supercontinent, Baltica underwent Paleoproterozoic crustal
20 growth through the Sarmatia–Volgo Uralia collision at 2.20–2.05 Ga followed
21 by major arc accretion in Fennoscandia during the Svecofennian Orogeny (SO)
22 at 1.92–1.80 Ga (Bogdanova et al., 2006, 2015; Pesonen et al., 2021; Teren-
23 tiev and Santosh, 2020; Wang et al., 2020). This orogeny produced nearly
24 1×10^6 km² of new crust through episodic arc–continent amalgamation (Lahti-
25 nen et al., 2009). The central and southern SO consists of NW–SE megado-
26 mains (100–300 km) bounded by dextral shear zones that formed at 1.86–1.75
27 Ga during southwest-younging accretion, with igneous ages decreasing in this
28 direction (Beunk and Kuipers, 2012; Bogdanova et al., 2015). These domains
29 record episodic rollback of a long-lived south–southwest-retreating subduction
30 system and crustal growth via multiple subduction collisions (Lahtinen, 1997;
31 Lahtinen et al., 2011, 2022; Bogdanova et al., 2006, 2013, 2015; Kukkonen
32 and Lauri, 2009). A major tectonic reorganisation at 1.82–1.80 Ga followed
33 the oblique Volgo–Sarmatia–Fennoscandia collision, disrupting crustal growth
34 (Bogdanova et al., 2015; Nironen, 2017; Salminen et al., 2021a,b; Soesoo et al.,
35 2020; Pesonen et al., 2021), halting northeastward accretion of the 1.87–1.75 Ga
36 Bergslagen–Livonia microcontinent, and establishing the transpressional frame-
37 work of southern Fennoscandia (Beunk and Page, 2001; Beunk and Kuipers,
38 2012; Bogdanova et al., 2015). In the southeast, this collision drove a $\sim 45^\circ$ an-
39 ticlockwise rotation of Sarmatia and subsequent extension at 1.80–1.75 Ga (Bog-
40 danova et al., 2008, 2013, 2015), followed by widespread Nuna-related AMCG
41 (Anorthosite–Mangerite–Charnockite–Granite) magmatism between 1.64 and
42 1.44 Ga, with rapakivi granites representing a major expression extending across

43 Fennoscandia (Fig.1b) (Pesonen et al., 2021; Nironen, 2017; Salminen et al.,
44 2021a; Solano-Acosta et al., 2025b).

45 In the Fennoscandian region, microcontinents such as Keitele and Both-
46 nia along the passive margins of the Archean Karelia and Norrbotten cratons
47 have motivated models of Svecofennian orocline development at around 1.87
48 Ga (Lahtinen et al., 2014; Nironen, 1997, 2017; Salminen et al., 2021b). Com-
49 parable accretionary deformation at 1.83–1.82 Ga has been proposed for the
50 Bergslagen area in south-central Sweden (Beunk and Kuipers, 2012; Bogdanova
51 et al., 2015). According to the Trans-Baltic model of Bogdanova et al. (2015)
52 (Fig.1a), the southern SO evolved through the northeastward accretion of the
53 Bergslagen–Livonia microcontinent, possibly including fragments of Amberland.
54 In southwestern Fennoscandia (Fig.1b), three major Paleoproterozoic arc sys-
55 tems are recognised: the Savo Belt (1.93–1.80 Ga), the Central–Western Finland
56 Arc and its granitoid complexes (1.90–1.87 Ga), and the Southern Svecofennian
57 domain (1.90–1.82 Ga), which comprises the Häme and Uusimaa belts together
58 with the Saimaa and Ladoga zones (Kähkönen, 2005; Bogdanova et al., 2015;
59 Nironen, 2017). The Häme and Uusimaa belts (1.90–1.88 Ga) extend eastward
60 into the Saimaa and Ladoga zones, which collided with the Karelian craton (Ko-
61 rja et al., 2006; Kotova and Podkovyrov, 2014; Nironen, 2017). Their curved
62 geometries reflect deformation associated with the Bothnian and Saimaa oro-
63 clines (Lahtinen et al., 2014; Nironen, 2017; Lahtinen et al., 2022). The Uusimaa
64 belt may continue beneath the Gulf of Finland toward the Tallinn zone (Figs.1b-
65 c) (Kirs et al., 2009; Kukkonen and Lauri, 2009; Bogdanova et al., 2015; Soesoo
66 et al., 2020; Kara et al., 2021), adjacent to the Bergslagen microcontinent (Bog-
67 danova et al., 2015), while the Ladoga zone defines its southeastern continuation
68 along the Archean boundary (Kotova et al., 2009; Myskova et al., 2012; Balty-
69 baev, 2013; Kotova and Podkovyrov, 2014).

70 The Precambrian basement of Estonia (Fig.1c), buried beneath 100–900 m
71 of Neoproterozoic–Devonian shallow-marine sediments that thicken southward
72 (e.g. Fig.1d2) (Puura and Huhma, 1993; Soesoo et al., 2004, 2020). Northern
73 Estonia consists of 1.92–1.88 Ga metavolcanic and metasedimentary sequences

74 intruded by orogenic granites that correlate with units in southern Finland and
 75 Swedish Bergslagen area (Beunk and Page, 2001; Puura et al., 2004; Bogdanova
 76 et al., 2015; Nironen, 2017; Soesoo et al., 2020; Kara et al., 2021), whereas
 77 the southern Estonian–Latvian granulite belt contains younger 1.84–1.77 Ga
 78 high-grade rocks (Bogdanova et al., 2015; Soesoo et al., 2004, 2020). These do-
 79 mains are separated by the Paldiski–Pskov Deformation Zone (PPDZ) (Raukas
 80 and Teedumäe, 1997; Puura et al., 2004; Kirs et al., 2009; Bogdanova et al.,
 81 2015; Soesoo et al., 2020) and bounded by NW–SE basement-scale shear zones
 82 (Soosalu et al., 2022; Solano-Acosta et al., 2023, 2025b) (Figs.1a-c).

83 Within this trans-Baltic configuration, the 1.92–1.88 Ga Tallinn, Alutaguse
 84 and Jõhvi domains reported *Fe–S–Si* metalliferous signatures (Fig.1a) (Bog-
 85 danova et al., 2015; Soesoo et al., 2020; Nirgi and Soesoo, 2021; Nirgi et al.,
 86 2024), suggesting to constitute the eastern continuation of a Bergslagen-type
 87 metallogenic province extending into the Uusimaa region of southern Finland
 88 and associated with 1.91–1.89 Ga felsic volcanic, subvolcanic, turbiditic, car-
 89 bonate and skarn systems (Beunk and Kuipers, 2012; Bogdanova et al., 2015;
 90 Kara et al., 2021). Northern Estonia subsequently underwent migmatisation
 91 and granitoid intrusion at approximately 1.79 Ga (Raukas and Teedumäe, 1997;
 92 All et al., 2004; Soesoo et al., 2004, 2020), synchronous with the late Svecofen-
 93 nian granite–migmatite zone of Finland (Bogdanova et al., 2015; Nironen, 2017).
 94 The Alutaguse domain, consisting of aluminous and graphitic gneisses, metavol-
 95 canics, quartzites and carbonates intruded by small-volume Svecofennian gran-
 96 ites, interpreted as a ca. 1.88 Ga rifted back-arc segment of the Uusimaa–Tallinn
 97 arc (Solano-Acosta et al., 2025a). Subsequent post-SO compression sealed and
 98 deformed this basin, producing favourable conditions for polymetallic VMS-
 99 style *Cu–Pb–Zn* mineral systems (Nirgi and Soesoo, 2021; Nirgi et al., 2024;
 100 Solano-Acosta et al., 2025a). Drill-core and geochemical evidence supports the
 101 development of a short-lived back-arc basin dominated by aluminous gneisses
 102 (Fig.1a) (Kivisilla et al., 1999; Soesoo et al., 2004, 2020; All et al., 2004; Kirs
 103 et al., 2009; Bogdanova et al., 2015; Solano-Acosta et al., 2025a). However, re-
 104 gional correlations and geochronological constraints remain incomplete, and an

105 alternative model proposes that Alutaguse represents the continuation of the
106 1.95–1.88 Ga Kalevian marginal basin with metapelitic input from the Kare-
107 lian Craton in southeast Finland and west-northwest Russia (Kirs et al., 2009;
108 Bogdanova et al., 2015).

109 This study evaluates the competing interpretations of the Alutaguse suc-
110 cession by integrating its metavolcanic and metasedimentary units with coeval
111 formations in the SS domains. Major, trace and *REE* geochemistry in Pre-
112 cambrian metasedimentary and metavolcanic rocks provides key constraints on
113 source characteristics and tectonic settings, offering a robust basis for recon-
114 structing crustal evolution and provenance (Bhatia and Crook, 1986; McLennan
115 et al., 1995; Lahtinen et al., 2002; El-Bialy, 2013; de Carvalho Mendes et al.,
116 2021; Solano-Acosta et al., 2025a,c; Gebremicale et al., 2025). Here, whole-rock
117 geochemistry is used to determine the genetic relationships among the Alu-
118 taguse, Ladoga, Saimaa, Häme and Uusimaa units (Fig.1b) and to refine the
119 geodynamic role of Alutaguse within the Fennoscandian SO.

Figure 1: Geological setting of the study zone, highlighting key features: Geological setting of the study zone, highlighting key features: (a) Central and Southern Svecofennian crustal structure crosses the Baltic Sea, encompassing Bergslagen, Livonia, Amberland, Keitele and Bothnia zones, that may be considered microcontinents. Pinkish box enhances the study zone. Notice the box at the bottom left with Baltica, enhancing the position of the bigger figure. The ages shown in the frames correspond to the main accretionary events within the specified zones. The dashed triangle lines represent the upper surfaces of dipping mantle reflectors, as identified by Lahtinen et al. (2009). The notation $Fe - S - Si$ signals the existence of diverse iron and sulphide ore deposits.; (b) Major Palaeoproterozoic tectonic zones over the Fennoscandian area. The white diagonal ruling over the greenish area indicates Svecofennian sedimentary basins. This map depicts the South Svecofennian (SS) and Estonian Alutaguse zones. Figures a and b are modified from Bogdanova et al. (2015), whose abbreviations are: AL - Alutaguse, BB - Bothnian, BS - Bergslagen, BPG - Belarus-Podlasie granulite belt, CE - Ciechanow, CFAC - Central Finland Arc Complex; CFGC-Central Finland Granitoid Complex (Keitele microcontinent), OKSZ-Orijärvi and Kemiö shear zones. KB - Keitele microcontinent, JO - Jöhvi, LEL-Latvian-East Lithuanian, MLD-Mid-Lithuanian zone, NO - Novgorod, SEG - South Estonian granulite zone, Ta - Tapa, Tll - Tallinn, WE - West Estonian zone, WLK - West Lithuanian granulite zone. Deformation zones: HGZ-GR - Hagsta-Gävle-Rättvik Zone, HSZ - Hassela Shear Zone, MEFZ - Middle Estonian Fault Zone, PPDZ - Paldiski-Pskov Deformation Zone, SFSZ - South Finland Shear Zone, and for volcanic belts and sedimentary basins: Uu-Uusimaa, H - Häme, P-Pirkanmaa, T - Tampere, PoB-Pohjanmaa, O-J - Oskarshamn-Jönköping. Please refer to Bogdanova et al. (2015) for further information and a comprehensive understanding. (c) Geological map scheme of the Precambrian basement of Estonia, showing geochemical anomalies according to Soesoo et al. (2020), and the upper right-corner inset showing the distribution of granulite- and amphibolite facies metamorphic rocks. Redish symbols correspond to Rapakivi lithologies. modified from Bogdanova et al. (2015). The Anomaly symbols represent the most prominent metal of the mineralisation.; (c1) Zoom of the geological basement map of the Alutaguse zone, depicting the location of the cores and the prominent metallogenic Alutaguse zones.; (d) Geophysical maps of the Alutaguse zone depicting the (1) 10x10m topography, (2) Depth to basement, (3) Bouguer gravitational anomaly, (4) residual Bouguer anomaly, 5) Regional magnetic anomaly, (6) Residual magnetic anomaly. Residual maps present the metalliferous anomalies' locations. figures d3 to d6 are modified after Solano-Acosta et al. (2023). The (c) geological data, (d1) topographical data, and (d2) depth-to-basement data can be downloaded from the Geoportal of the Land Board of the Republic of Estonia (<https://geoportaal.maaamet.ee/>).

2. Geological Setting

2.1. Fennoscandian context

Between $\sim 2.10\text{--}1.90$ Ga, a pre-Svecofennian ocean separated proto-Fennoscandia from the Bergslagen block, with the Tallinn–Uusimaa arc system accreting to the Häme and CFAC (Central Finland Arc Complex) belts by ~ 1.90 Ga, triggering crustal thickening and widespread magmatism across southern Finland, central Sweden, and northern Estonia, (Nironen, 1997; Puura et al., 2004; Soesoo et al., 2004; Kukkonen and Lauri, 2009; Lahtinen et al., 2009; Hanski, 2015; Bogdanova et al., 2015; Nironen, 2017; Kara, 2021; Solano-Acosta et al., 2025a)

(Figs.1a-b). Crustal growth continued through alternating transtensional to transpressional cycles along a southward-retreating active margin (Hermansson et al., 2008; Saalman et al., 2009; Bogdanova et al., 2015; Stephens and Weihed, 2020; Kara, 2021; Soosalu et al., 2022), consistent with tectonic switching (Gaál and Gorbatshev, 1987; Collins, 2002; Kukkonen and Lauri, 2009; Kara, 2021; Lahtinen et al., 2022). The Skellefte, Tampere, Häme, and Uusimaa arcs are interpreted as either separate belts from subduction retreat or double-plunging zones (Kukkonen and Lauri, 2009; Nironen, 2017; Kara et al., 2021; Solano-Acosta et al., 2025a,c), or as segments of a single curvilinear arc (Lahtinen et al., 2022), defining Finland’s tectonic provinces (Korsman et al., 1999; Nironen, 1997, 2017) under a N–NE stress field (Hermansson et al., 2008; Saalman et al., 2009; Bogdanova et al., 2015; Soosalu et al., 2022). Further crustal assembly occurred via oblique accretion of southern terranes, including the Belarus–Baltic, Vitebsk, and Osnitsk–Mikashевичi belts (1.92–1.79 Ga) (Claesson et al., 2001; Bogdanova et al., 2015), culminating in the amalgamation of Fennoscandia with Volgo-Sarmatia by ~ 1.77 Ga (Bogdanova et al., 2006, 2015; Johansson et al., 2022). The Svecobaltic Orogeny (1.84–1.79 Ga) marked the last major crustal thickening, followed by the Gothian Orogeny (1.73–1.48 Ga), which added westward-younging belts to Baltica’s margin (Bogdanova et al., 2008, 2015; Nironen, 2017). Together, these events contributed to the assembly of the Nuna supercontinent (1.90–1.60 Ga), with Baltica at its core, drifting from high to equatorial latitudes between 1.90–1.30 Ga at rates up to 14 cm/yr during the SO, 6.5 cm/yr during the Gothian, and 9 cm/yr during the Danopolonian orogeny (Rogers and Santosh, 2002; Condie et al., 2021; de Oliveira Chaves, 2021; Pesonen et al., 2021; Salminen et al., 2021b).

The SS domain comprises several tectonic belts with complex histories. The Häme belt (1.88–1.87 Ga) contains amphibolite-facies metavolcanic and metasedimentary rocks intruded by granites, migmatites, and pegmatites, with older arc-type units and younger rift-affinity basalts (Kähkönen, 2005; Kukkonen and Lauri, 2009; Hanski, 2015; Kara, 2021; Kara et al., 2021). The Uusimaa belt (2.10–1.91 Ga) shows a dual rifting–collision signature, hosting juvenile

arc rocks at 1.85–1.87 Ga and older isotopic signals from Paleoproterozoic crust (Lahtinen and Huhma, 1997; Rämö et al., 2001; Väisänen et al., 2002; Kukkonen and Lauri, 2009; Kara et al., 2021). Both belts contain 1.90–1.88 Ga metavolcanics and calcareous metasediments, especially in western Uusimaa (Kähkönen, 2005; Kara, 2021), though their linkage remains debated—interpreted as either distinct terranes or a shared arc system (Kukkonen and Lauri, 2009; Bogdanova et al., 2015; Kara et al., 2021; Kara, 2021).

The eastern SS domain comprises metapelites, metagreywackes, and metavolcanics from andalusite-bearing schists to diatexites. The South Finland Shear Zone (SFSZ; Fig.1b) forms a major tectonic boundary with sillimanite- and andalusite-grade metamorphism. Orijärvi records intense premetamorphic alteration, while West Uusimaa preserves granulite-facies rocks formed under $\sim 800^\circ\text{C}$ and ~ 6 kbar, with three metamorphic events between 1.91–1.80 Ga (Hölttä and Heilimo, 2017; Nironen, 2017).

The Saimaa zone, along the Raahe–Ladoga Suture Zone (RLSZ; Fig.1b), contains metasedimentary and metavolcanic units including mica schists, black shales, and diatexitic granulites. A 2.4 – 2.2 Ga depositional gap is recorded by zircon ages (Lahtinen et al., 2022). After a 1.92 Ga collision, at least five deformation events occurred between 1.9 – 1.8 Ga, and a 1.91 – 1.92 Ga marine signal with metabasalt and shale is linked to the Bothnian oroclinal and conductivity anomalies (Korja et al., 2002, 2006).

In the NW Ladoga region (SE Finland and Russian Karelia), the contact between the Archaean Karelian Craton and the Svecofennian domain occurs along the RLSZ (Fig.1b). Here, Archaean granite gneiss domes are juxtaposed with Palaeoproterozoic Kalevian sediments within a 50 – 150 km-wide zone. Ladoga granitoids exhibit low ϵNd , indicating Archaean crustal input (Kotova et al., 2009; Baltybaev, 2013; Kotova and Podkovyrov, 2014). The Ladoga complex includes metamorphosed Sortavala Group volcanics (2.06 – 1.98 Ga) and Kalevian siliciclastics (1.97 – 1.89 Ga), linked to subduction processes. A NW-trending shear zone divides the region into two blocks, with $Sm - Nd$ data indicating 2.7 – 2.5 Ga sources and zircon provenance from both Proterozoic (2.0 – 1.9 Ga)

191 and Archean (2.74–2.54, 3.01–2.90 Ga) crust (Myskova et al., 2012). NE–SW
 192 shears and SE–NW folds mark the transition south of the RLSZ, where gran-
 193 itoids resemble those of the Central Finland complex. Ladoga metasediments
 194 reflect erosion from Archean gneisses and Sortavala volcanics, with Svecofen-
 195 nian assimilation (Kotova et al., 2009; Kotova and Podkovyrov, 2014). Western
 196 Ladoga volcanogenic units align with southern Savo and Tampere belts, indi-
 197 cating shared tectonic origins (Kotova and Podkovyrov, 2014).

198 2.2. Estonian Basement

199 The Estonian basement comprises Paleoproterozoic–Mesoproterozoic crys-
 200 talline crust overlain by gently south-dipping (~ 0.10 – 0.20°) Ediacaran–Palaeozoic
 201 strata and Quaternary cover (Fig.1d2) (Raukas and Teedumäe, 1997; Soesoo
 202 et al., 2004, 2020). Geological and geophysical datasets further distinguish
 203 six structural and petrological zones within the Estonian basement (Fig.1c-d):
 204 Tallinn, Alutaguse, Jõhvi, West Estonian, Tapa, and South Estonian, each de-
 205 fined by specific lithological assemblages, metamorphic grades, and potential-
 206 field signatures (Soesoo et al., 2004, 2020; Solano-Acosta et al., 2023), and
 207 bounded by major crustal features such as the NW–SE PPDZ and the E–W
 208 Middle-Estonian Fault Zone (MEFZ) (All et al., 2004; Soesoo et al., 2004, 2020;
 209 Bogdanova et al., 2015; Solano-Acosta et al., 2023).

210 The Tallinn domain—within the Finnish Uusimaa arc, with Alutaguse as its
 211 back-arc—accreted at 1.92–1.87Ga in a double-subduction, arc–microcontinent
 212 collision framework (Puura et al., 2004; Kukkonen and Lauri, 2009; Bogdanova
 213 et al., 2015; Nironen, 2017; Kara, 2021; Solano-Acosta et al., 2025a). Aluta-
 214 guse, a folded metasedimentary basin that postdates arc–back-arc closure, re-
 215 mains poorly dated, yet geochemical–geophysical links to SO and ca.1.88 Ga
 216 aluminogneiss ages in adjacent areas support its continuation of this back-arc
 217 system (Kivisilla et al., 1999; Soesoo et al., 2020; Solano-Acosta et al., 2023,
 218 2025a,c; Bogdanova et al., 2015; Lahtinen et al., 2017; Kara et al., 2021).
 219 Western and southern Estonia comprise high-grade metasedimentary and or-
 220 thogneiss complexes metamorphosed to granulite facies (Fig.1b), later intruded

221 by ~ 1.6 Ga rapakivi suites (Soesoo et al., 2004, 2020; Rämö et al., 1996; All
222 et al., 2004; Solano-Acosta et al., 2025b).

223 Regional gravity and magnetic data indicate continuation of Estonian do-
224 mains northward into oblique Svecofennian blocks in Finland and Sweden and
225 southward into Latvian terranes forming Livonia (Fig.1a); these are shaped
226 by inherited NW–SE shear zones (the PPDZ and its Sottunga–Jurmo exten-
227 sion) Beunk and Kuipers (2012); Bogdanova et al. (2015); Solano-Acosta et al.
228 (2025b), which remains seismically active (Fig.1c) (Bogdanova et al., 2006, 2008,
229 2015; Soosalu et al., 2022; Solano-Acosta et al., 2023).

230 Three main age groups occur in the Precambrian basement. (1) 1.92–1.80 Ga
231 Fennoscandian crust: North Estonian metavolcanics (1.92 ± 10 Ga), southern
232 granulitic metavolcanics (1.83 ± 22 , 1.82 ± 7 Ga), Tapa tonalites (1.82 ± 26 Ga),
233 and Jõhvi magnetite-bearing gneisses (1.87 ± 18 , 1.83 ± 10 , 1.79 ± 19 Ga)
234 (Petersell and Levchenkov, 1994; Soesoo et al., 2004, 2006; Kirs et al., 2009;
235 Bogdanova et al., 2015; Soesoo et al., 2020). (2) 1.79–1.77 Ga southern Esto-
236 nian intrusives and high-grade gneisses: tonalites (1.79 ± 16 Ga), charnockites
237 (1.76 ± 11 Ga), orthopyroxene–garnet gneisses (1.78 ± 2 Ga), and a gabbro-
238 norite dyke (1.77 ± 20 Ga) (Soesoo et al., 2004, 2006). (3) 1.64–1.58 Ga A-type
239 and rapakivi magmatism: Abja granite (1.64 ± 7 Ga), Virtsu quartz monzonite
240 (1.61 ± 17 Ga), and rapakivi bodies at Märjamaa (1.63 ± 7 Ga), Naissaare
241 (1.62 ± 7 Ga), and the Riga batholith (1.58 ± 2 Ga) (Kirs and Petersell, 1994;
242 Rämö et al., 1996; Soesoo and Hade, 2012; Solano-Acosta et al., 2025b).

243 Alutaguse records high-temperature amphibolite-facies conditions (3–5 kbar;
244 Puura et al. 1983) and comprises Al-rich mica gneisses, metapelites, graphite
245 gneisses, and minor pyroxene-rich metavolcanics and intrusions (Bogdanova
246 et al., 2015; Soesoo et al., 2020). It has been linked to the Kalevian-age
247 marginal basin extending east to Lake Ladoga (All et al., 2004), though the tim-
248 ing of sedimentation and metamorphism remains uncertain (Bogdanova et al.,
249 2015; Soesoo et al., 2020). Core mineralogy (Klein, 1986) shows Al-rich garnet-
250 bearing gneisses dominate in Alutaguse (90.45%) versus Tallinn (25.4%); bi-
251 otite–plagioclase gneisses are rare in Alutaguse (1%) but common in Tallinn

(24.4%), while $Bi - Pl$ and $Bi - Hbl - Pl$ types represent 50.2% and 6.1%, respectively. Southwestward, the Estonian–Latvian Granulite Belt contains charnockitized amphibolites and feldspar gneisses formed at 5–6 kbar (Soesoo et al., 2006; Bogdanova et al., 2015).

Geophysically, Alutaguse underlies subdued relief (max. 166 m; Fig.1d1) with basement depths of 130 – 450 m (Fig.1d2). Regional gravity and magnetic fields are generally low (Figs.1d3,d5), but local positive residual anomalies coincide with $Cu-Pb-Zn$ enrichments near Uljaste, Assamalla, and Haljala mineralised zones (Figs.1c1,d4,d6, (Kivisilla et al., 1999; Soesoo et al., 2004, 2020; Solano-Acosta et al., 2023, 2025a).

3. Materials and Datasets

This study integrates three Alutaguse geochemical datasets to assess compositional variability, provenance, and tectonic setting (cf. Solano-Acosta et al. (2025a,c)). The compiled dataset is available at https://kirj.ee/wp-content/plugins/kirj/pub/Geochemical_Data_Alutaguse_EESTI_20250205113208.xlsx?v=a57b8491d1d8. (1) 216 whole-rock major-element analyses (wet chemistry) compiled by Kivisilla et al. (1999)— SiO_2 , TiO_2 , Al_2O_3 , Fe_2O_{3tot} , MnO , MgO , CaO , Na_2O , K_2O , P_2O_5 , SO_3 , LOI (wt.%)—archived in SARV (<https://kirjandus.geoloogia.info/en/reference/21247>) and based on Estonian Geological Survey (EGT) core logs (<https://gis.egt.ee/portal/apps/dashboards/99f758ac4ef548f686b831adb3199378>); unlike Solano-Acosta et al. (2025a), only samples with complete major-element data are retained here. (2) Sixteen samples (13 metasedimentary, 3 metavolcanic) analysed by Solano-Acosta et al. (2025a) for major, trace, and *REE* chemistry, classified from EGT core descriptions and consistent with Kivisilla et al. (1999). (3) 149 samples from newly released EGT trace-element datasets for Uljaste drill cores (Fig.1c1), reporting trace elements with *REE* limited to *La* and *Ce*, classified as above and accessible via eMaapõu (<https://geoloogia.info/en/analytical-data>). Comparative data from the SS domain — $n = 206$ for the Finnish Uusimaa,

281 Häme, and Saimaa domains — were extracted from Rasilainen et al. (2007),
 282 based on a subset of 6544 samples in the GTK geochemical database ([https://](https://tupa.gtk.fi/paikkatieto/meta/rock_geochemical_data_of_finland.html)
 283 tupa.gtk.fi/paikkatieto/meta/rock_geochemical_data_of_finland.html).
 284 These samples can be freely downloaded in Esri File Geodatabase and MS Excel
 285 formats (<https://hakku.gtk.fi/en/locations/?id=4>). They are properly
 286 georeferenced, and only the samples classified as “schist, gneiss, amphibolite,
 287 and skarn rock” were used. These samples were filtered out from the “sedi-
 288 mentary” and “volcanic” categories and further segmented by SS domain, as
 289 shown in the Figure 1b. Additional Ladoga domain data include 26 siliciclas-
 290 tic metasediments from Kotova et al. (2009) and 8 metavolcanic samples from
 291 Kotova and Podkovyrov (2014).

292 Complete elemental data from Alutaguse and SS can be observed in the
 293 Supplementary data. Table 1 presents the average and minimum-maximum
 294 concentrations of the major elements (in *wt.%*), while Table 2 summarizes the
 295 trace and *REE* element concentrations (*ppm*) for all samples.

296 3.1. Data Integration and Classification Procedures

297 The Alutaguse dataset was treated as in Solano-Acosta et al. (2025a,c), in-
 298 tegrating (1) legacy major element data from Kivisilla et al. (1999), (2) thirteen
 299 samples with full major, trace, and *REE* data (Solano-Acosta et al., 2025a),
 300 and (3) trace-elemental data from Uljaste drill cores published by the EGT.
 301 Lithological classifications follow EGT core descriptions, standardised based on
 302 Kivisilla et al. (1999).

303 Metasedimentary samples were categorized by SiO_2 content following Verma
 304 and Armstrong-Altrin (2013), with High- SiO_2 defined as $> 63 \text{ wt.}\%$ and Low-
 305 SiO_2 as $\leq 63 \text{ wt.}\%$. Although a few lithologies span both groups, most show
 306 consistent silica-based distributions (Table 1; Fig.2a1). Lithologies predomi-
 307 nantly in the High- SiO_2 group include: Biotite gneisses ($\pm Grt \pm Crd \pm Sil$),
 308 Garnet-bearing mica gneisses $\pm Crd \pm Sil$, Cordierite-bearing mica gneisses
 309 $\pm Grt \pm Sil$, Graphite-bearing mica gneisses $\pm Grt \pm Crd \pm Sil$, Biotite-plagioclase
 310 gneisses $\pm Amph$, and Biotite-Amphibole gneisses. The Low- SiO_2 group in-

311 cludes similar lithologies but with lower average silica contents, including Graphite-
312 bearing mica gneisses ($\pm Grt \pm Crd \pm Sil$), Garnet-bearing mica gneisses $\pm Amph$,
313 and Biotite–Amphibole gneisses.

314 Since the third Alutaguse dataset from EGT includes only trace elements and
315 lacks major oxides, samples were assigned to hypothetical High- or Low- SiO_2
316 groups based on the average $\bar{\chi}_{SiO_2}$ values of each lithology. This classification
317 follows the methodology of Solano-Acosta et al. (2025a,c) (Fig. 2a; Table 2).

318 The SS dataset was processed consistently to allow direct comparison (Table
319 1; Fig. 2a3). Major oxides from Rasilainen et al. (2007) do not include LOI and
320 were therefore normalized to 100%. Non-normalized LOI values from other
321 datasets (Kivisilla et al., 1999; Kotova et al., 2009; Kotova and Podkovyrov,
322 2014) were retained where available (Table 1). SS metasediments were grouped
323 by High or Low SiO_2 content and per SS domain. Lithologies include micaceous
324 gneisses $\pm Grt \pm Crd \pm Sil$ (Rasilainen et al., 2007), and quartz micaceous gneisses
325 in the Ladoga area (Kotova et al., 2009). SS samples are of Svecofennian age
326 (Rasilainen et al., 2007; Kotova et al., 2009; Kotova and Podkovyrov, 2014), and
327 none reported graphite.

328 No further subdivision was applied to metavolcanic samples (Figs. 2a2, 2a4).
329 Alutaguse types include Amphibolites, Pyroxene gneisses ($\pm Amph$), and 2-
330 pyroxene gneisses ($\pm Amph$). For SS, metavolcanic samples correspond to am-
331 phibolites, hornblende gneisses, and mafic volcanites (Rasilainen et al., 2007),
332 as well as biotite-clinopyroxene and amphibole schists for the Ladoga domain
333 (Kotova and Podkovyrov, 2014).

Group	Classification	Metasedimentary units																				ΣREE	LREE	HREE	HREE/LREE	(100 + Ta_2O_5/Zr)
		Tb/U	Tb/Ta	Zr/Sm	Tb/Zr	Nb/Zr	Nb/Ta	Nb/Y	Zr/Y	Ba/Bi	La/Nb	Tb/La	Sm/Yb	Eu*	(Gd/Yb) _{cn}	(La/Sm) _{cn}	(La/Yb) _{cn}	ΣREE	LREE	HREE	HREE/LREE					
Alutaguse High- SiO_2 Metasedimentary samples	Garnet-bearing mica gneissos ± Crd ± Sil	8.20	23.68	11.11	0.31	0.07	10.32	0.78	2.31	9.36	4.38	0.31														
	Garnet-bearing mica gneissos ± Crd ± Sil	1.07-0.37	11.50-32.99	4.66-27.53	0.02-0.09	0.02-0.04	10.00-33.04	0.15-1.00	3.71-19.41	3.79-17.49	2.44-15.79	0.07-0.45														
	Garnet-bearing mica gneissos ± Crd ± Sil	2.48-15.37	17.30-41.37	7.60-20.30	0.07-0.30	0.07-0.30	13.17-37.42	0.57-1.57	7.50-33.39	4.75-30.30	3.14-13.19	0.26-0.41	8.00	0.27	4.85	4.8	30.09	31.158	307.1	4.47	0.05			0.32		
	Corundum-bearing mica gneissos ± Crd ± Sil	8.40	25.38	10.08	0.12	0.07	10.27	0.71	2.31	9.36	4.38	0.31														
	Graphitic bearing mica gneissos ± Crd ± Sil	1.48-12.53	10.16-45.57	1.67-11.09	0.04-0.31	0.05-0.31	11.32-30.00	0.30-0.55	5.13-35.35	2.48-8.57	2.34-10.1	0.05-0.41														
	Graphitic bearing mica gneissos ± Crd ± Sil	0.12-0.29	0.14-22.21	1.22-42.33	0.00-0.28	0.00-0.27	7.17-47.33	0.00-0.03	0.65-110.68	1.35-23.42	0.03-0.03	0.25-0.41	0.22-0.77	0.36-1.18	1.07-7.41	0.31-0.77	56.42-245.56	39.39-237.66	2.00-39.67	0.02-0.41	0.23-1.85					
	Graphitic bearing mica gneissos ± Crd ± Sil	4.20	52.41	17.25	0.15	0.08	11.30	0.50	13.31	0.31	2.66	0.45	1.60	1.00	12.06	17.50	160.35	14.54	0.12	0.77						
	Graphitic bearing mica gneissos ± Crd ± Sil	1.02-0.56	0.16-29.25	1.22-47.33	0.00-0.28	0.00-0.27	7.17-47.33	0.00-0.03	0.65-110.68	1.35-23.42	0.03-0.03	0.25-0.41	0.22-0.77	0.36-1.18	1.07-7.41	0.31-0.77	56.42-245.56	39.39-237.66	2.00-39.67	0.02-0.41	0.23-1.85					
	Graphitic bearing mica gneissos ± Crd ± Sil	4.76	11.41	8.24	0.12	0.12	11.97	0.96	2.66	2.66	0.41	1.41	0.36	1.82	1.56	3.22	185.56	176.21	0.34	0.06	0.34					
	Total	7.28	9.33	8.25	0.02	0.04	20.39	0.16	0.30	0.30	2.13	0.2	0.29	0.25	1.39	31.31	30.09	106.22	89.37	0.38	0.38			0.23		
SS Zones High- SiO_2 Metasedimentary samples	Ussimaa	0.12-0.29	0.14-22.21	1.22-42.33	0.00-0.28	0.00-0.27	7.17-47.33	0.00-0.03	0.65-110.68	1.35-23.42	0.03-0.03	0.25-0.41	0.22-0.77	0.36-1.18	1.07-7.41	0.31-0.77	56.42-245.56	39.39-237.66	2.00-39.67	0.02-0.41	0.23-1.85					
	Hine	1.89-7.11	4.79-33.87	6.05-41.53	0.02-0.09	0.03-0.14	8.88-32.81	0.16-1.37	4.01-22.80	0.61-37.54	1.72-3.38	0.14-0.40	1.03-5.0	0.25-1.10	0.35-0.63	1.09-4.51	0.86-40.94	64.07-375.79	43.48-304.43	1.49-71.36	0.02-0.23	0.07-0.54				
	Silman	2.82	7.32	11.07	0.07	0.07	12.98	0.46	2.66	4.41	8.21	0.31	2.56	0.72	2.00	5.48	16.88	176.62	160.66	12.36	0.07	0.30				
	Ladoga	1.60-10.61	6.72-27.31	5.98-13.96	0.20-0.12	0.03-0.14	7.85-18.72	0.33-0.62	3.88-11.30	0.76-2.29	1.09-0.48	0.18-0.32	2.05-3.11	0.44-1.04	0.63-1.29	1.12-4.43	1.98-19.27	70.75-226.80	65.68-222.05	5.08-15.13	0.06-0.09	0.25-0.58				
	Total	1.07-12.41	3.70-34.38	6.28-34.37	0.04-0.11	0.04-0.10	7.84-20.03	0.47-0.37-0.92	9.59	3.79-12.01	1.47-20.09	0.17-0.35	1.49-3.83	0.34-0.96	1.30-2.70	2.41-4.88	6.34-17.55	38.78-203.62	36.82-250.45	1.78-19.33	0.05-0.10	0.14-0.56				
	Ussimaa	2.50-7.92	7.00-15.71	5.98-13.96	0.20-0.12	0.02-0.10	11.14-14.52	0.36-1.36	3.33-20.29	2.24-17.77	1.54-3.00	1.14-3.55	0.65-1.06	1.14-3.95	2.19-4.64	2.53-44.54	43.95-340.89	30.17-229.40	4.78-11.49	0.05-0.12	0.15-1.36					
	Hine	4.57	14.47	12.54	0.07	0.07	14.96	0.46	2.66	4.41	8.21	0.31	2.56	0.70	2.00	5.48	16.88	176.62	160.66	12.36	0.07	0.30				
	Ladoga	1.07-12.41	3.70-34.38	5.98-41.53	0.02-0.12	0.02-0.10	7.83-32.81	0.16-1.37	3.33-22.80	0.61-37.54	1.54-3.41	0.14-0.56	1.03-5.00	0.25-1.10	0.35-0.63	1.09-4.64	0.86-44.54	38.58-375.79	36.82-204.43	1.49-71.36	0.02-0.23	0.07-0.54				
	Total	1.07-12.41	1.70-34.38	6.28-34.37	0.04-0.11	0.04-0.10	7.84-20.03	0.47-0.37-0.92	9.59	3.79-12.01	1.47-20.09	0.17-0.35	1.49-3.83	0.34-0.96	1.30-2.70	2.41-4.88	6.34-17.55	38.78-203.62	36.82-250.45	1.78-19.33	0.05-0.10	0.14-0.56				
	Alutaguse Metavolcanic samples	Ussimaa	1.80-7.92	1.09-24.35	1.39-11.33	0.05-0.14	0.05-0.14	10.00-33.04	0.24-0.03	3.66-1.37	3.66-1.37	0.41	0.36	0.25	0.36	0.25	0.36	15.87	15.87	74.43-385.45	74.43-385.45	0.06-0.14				
Hine		5.20	12.02	9.80	0.08	0.10	11.31	0.82	2.90	2.68	3.38	0.42	1.02	0.72	1.78	1.16	0.00	30.38	151.95	12.03	0.08	0.41				
Silman		1.46-8.28	3.14-19.29	2.45-10.12	0.06-0.10	0.10-0.19	11.32-16.19	0.18-0.22	3.26-20.03	0.16-0.45	0.16-0.37	0.16-0.45	2.08-12.58	0.37-1.10	1.08-2.41	1.78-4.45	2.08-12.58	59.25-201.74	47.45-177.09	4.12-17.45	0.02-0.25	0.08-0.40				
Ladoga		1.60-10.61	0.17-0.43	1.84-10.40	0.04-0.12	0.04-0.12	10.28-20.07	0.31-0.69	4.33-17.82	0.08-0.60	1.06-1.13	0.55-0.36	2.14-0.79	0.36-1.10	0.57-0.81	0.94-4.73	1.47-17.53	148.30-366.69	139.05-165.77	4.88-20.99	0.02-0.05	0.08-0.65				
Total		2.96	8.16	7.41	0.10	0.11	11.96	1.06	4.06	4.01	1.96	0.36	2.01	0.36	2.02	3.86	12.39	113.14	107.00	7.74	0.09	1.14				
Ussimaa		1.27-17.90	4.73-11.45	2.08-17.57	0.04-0.17	0.04-0.17	12.02-16.10	0.27-0.30	0.55-0.86	0.92-0.60	0.92-0.60	1.62-0.83	0.36-0.94	1.14-0.89	1.14-0.89	2.42-2.22	0.93-17.70	64.68-10.10	61.01-153.42	1.65-12.09	0.06-0.11	0.66-2.00				
Hine		0.67-11.30	8.22	0.09	0.10	0.10	14.67	0.60	6.73	8.54	2.84	0.62	2.83	0.67	2.08	0.57	11.61	194.01	182.10	12.42	0.07	0.63				
Ladoga		0.60-10.81	5.13-25.93	2.50-78.89	0.01-0.17	0.04-0.12	10.28-20.07	0.19-0.42	0.65-0.66	0.65-0.66	0.65-0.66	0.65-0.66	0.65-0.66	0.65-0.66	0.65-0.66	0.65-0.66	0.65-0.66	0.65-0.66	0.65-0.66	0.65-0.66	0.65-0.66	0.65-0.66	0.65-0.66	0.65-0.66		
PAAS		4.51	11.4	10.12	0.07	0.09	11.84	0.73	2.57	4.06	2.01	0.35	0.64	1.33	1.20	0.2	64.58	171.82	12.96	0.07	0.47					
UCC		3.89	10.5	13.97	0.06	0.08	12	0.50	5.84	4.91	2.5	0.35	2.05	0.64	1.20	0.16	3.20	186.37	156.28	10.09	0.07	0.33				
Alutaguse Metavolcanic samples	2-Pyroxene gneissos ± Amph	1.53	24.91	10.98	0.06	0.06	10.36	0.31	3.79	36.95	44.4	0.36														
	2-Pyroxene gneissos	0.50-3.92	0.88-41.74	0.71-13.25	0.00-0.19	0.03-0.29	13.50-22.17	0.07-0.36	1.23-5.58	13.83-30.00	0.74-0.35	0.01-0.41														
	Pyroxene gneissos ± Amph	1.86	22.94	12.96	0.06	0.06	2.96	0.96	3.94	3.94	1.94	0.35														
	Pyroxene gneissos	4.54-6.90	21.41-24.04	12.52-13.20	0.04-0.10	0.04-0.07	17.79-29.28	0.35-0.98	8.49-13.74	3.70-6.64	3.88-3.98	0.19-0.35														
	Total	0.08-34.77	0.28-45.00	0.16-25.56	0.00	0.13	2.11	5.4	34.17	60.88	1.14		1.25	0.69	1.96	4.39	13.63	75.26	61.84	13.43	0.22	-0.63-1.44				
	Ussimaa	3.87	14.21	9.90	0.06	0.12	18.02	0.76	3.96	27.34	22.07	0.20	1.17	0.96	1.36	3.45	6.88	44.88	38.38	0.28	0.25					
	Hine	0.08-14.77	0.07-1.74	0.71-25.56	0.00-0.12	0.04-0.34	4.08-40.00	0.01-0.30	0.18-0.30	0.18-0.30	0.18-0.30	0.18-0.30	1.13-1.25	0.45-0.69	0.47-1.09	1.68-4.39	2.08-13.63	55.35-70.79	47.07-61.84	6.22-13.43	0.21-0.29	-0.63-1.44				
	Silman	2.31	5.61	3.97	0.02	0.06	11.00	0.20	1.55	10.28	2.33	0.16	0.24	0.60	2.38	1.67	13.77	60.01	79.01	10.48	0.13					
	Ladoga	0.90-1.20	0.17-11.82	0.13-14.07	0.00-0.05	0.00-0.05	2.67-22.21	0.04-0.55	1.15-11.41	2.06-5.52	0.07-0.49	0.04-0.47	1.01-2.81	0.36-0.89	0.36-0.89	0.36-0.89	9.31-29.04	1.55-44.02	1.55-29.04	2.27-24.27	0.04-0.27					
	Total	1.97	5.61	3.97	0.02	0.06	11.00	0.20	1.55	10.28	2.33	0.16	0.24	0.60	2.38	1.67	13.77	60.01	79.01	10.48	0.13					
SS Metavolcanic Samples	Hine	1.73-1.05	0.15-11.86	0.17-13.05	0.00-0.05	0.00-0.05	1.40-10.55	0.00-0.62	1.15-11.76	0.02-0.40	0.08-0.43	0.01-0.10	0.02-0.43	0.01-0.10	0.02-0.43	0.01-0.10	0.02-0.43	30.14-80.15	13.95-45.59	43.94-20.46	0.04-0.55	0.14-0.20				
	Silman	2.92	8.36	7.44	0.01	0.06	10.16	1.02	4.06	3.79	1.96	0.36	1.96	0.68	1.72	3.86	18.17	98.48	13.4	0.17	0.07					
	Ladoga	1.20-2.96	0.06-0.41	1.17-0.87	0.00-0.03	0.00-0.03	0.17-0.56	0.07-0.56	0.42-0.57	1.18-2.38	0.04-0.10	0.02-0.43	1.24-3.30	0.70-0												

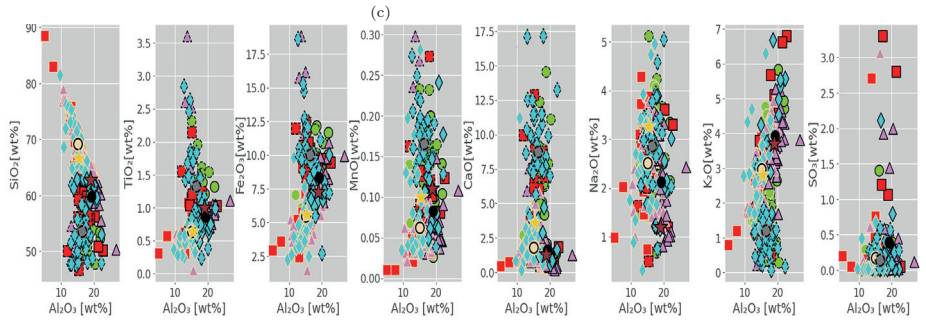
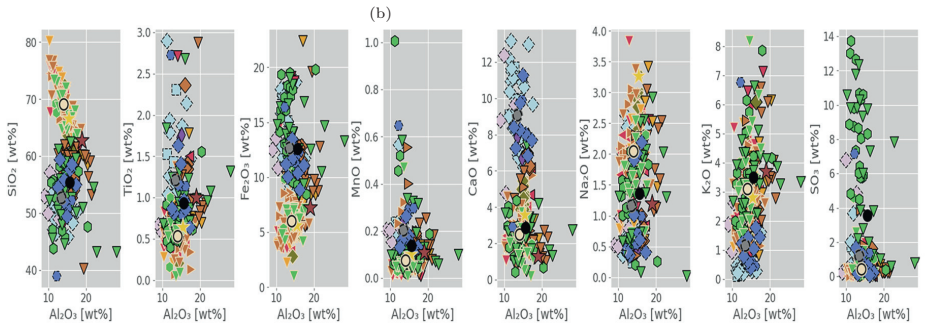
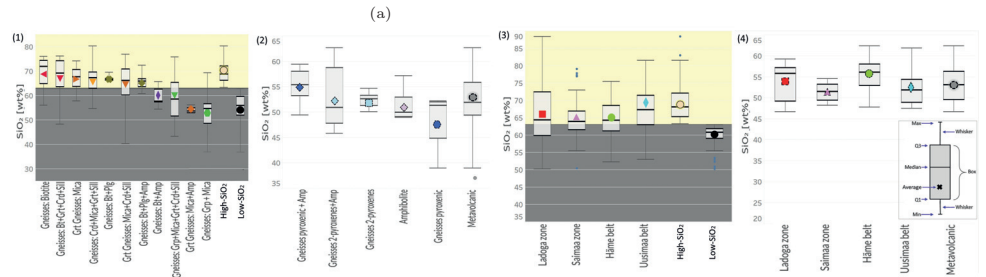
334 4. Results

335 Geochemical comparisons were conducted between metasedimentary and
336 metavolcanic samples from the Alutaguse and SS zones. Metasedimentary com-
337 positions were evaluated against established crustal reference values, including
338 the Upper Continental Crust (UCC; Rudnick and Gao (2003)) and Post-Archean
339 Australian Shale (PAAS; Taylor and McLennan (1985); McLennan (2001)).

340 4.1. Geochemistry of Major Elements

341 Al_2O_3 –Harker diagrams show distinct major element distributions for Alu-
342 taguse (Fig.2b) and SS samples (Fig.2c). High- SiO_2 metasediments resemble
343 UCC compositions, while Low- SiO_2 samples align more with PAAS.

344 In the High- SiO_2 metasediments, Alutaguse has a lower average SiO_2 con-
345 tent than Uusimaa but higher than Häme and Saimaa. Ladoga has the highest
346 average and widest range of SiO_2 among the zones. TiO_2 is the lowest in Aluta-
347 guse, with a wide range partly overlapping Uusimaa, whereas Häme and Saimaa
348 show distinctly higher averages and Ladoga remains low. Al_2O_3 is intermedi-
349 ate, overlapping all SS zones, averaging lower than Häme and Saimaa but higher
350 than Ladoga, which has the lowest mean and widest spread. Fe_2O_3 is clearly en-
351 riched in Alutaguse, with a higher mean and wider range than Uusimaa, Häme,
352 and Saimaa, while Ladoga overlaps Uusimaa at the lower end; within the SS do-
353 mains, Häme and Saimaa exceed Uusimaa in average values. MnO contents are
354 generally similar, though the broader Alutaguse range includes higher maxima
355 absent in Häme and Saimaa. MgO is higher than Uusimaa on average but lower
356 than Häme, with Saimaa intermediate; Ladoga exhibits the widest variability,
357 spanning both high and low extremes. CaO is most enriched in Alutaguse, with
358 higher averages than all SS zones, though Uusimaa shows a wider range that ex-
359 tends above it, while Häme, Saimaa, and Ladoga decrease progressively. Na_2O
360 is lowest in Alutaguse and Saimaa, higher in Uusimaa and Ladoga, and peaks
361 in Häme, which defines the SS maximum. K_2O is higher than Uusimaa, Häme,
362 and Ladoga but broadly comparable to Saimaa, whose range is wider; among



Alutaguse lithologies

- Gneisses: Grp + Mica
- Gneisses: Grp + Mica + Grt + Crd + Sill
- Gneisses: Grt + Mica
- Gneisses: Grt + Mica + Amp
- Gneisses: Grt + Mica + Crd + Sill
- Gneisses: Biotite
- Gneisses: Bt + Grt + Crd + Sill
- Gneisses: Crd + Mica + Grt + Sill
- Gneisses: Bt + Amp
- Gneisses: Bt + Plg
- Gneisses: Bt + Plg + Amp
- Amphibolite
- Gneisses: pyroxenic
- Gneisses: pyroxenic + Amp
- Gneisses: 2-pyroxenes
- Gneisses: 2-pyroxenes + Amp
- Metavolcanic origin
- Meta sedimentary borders
- High- SiO_2 > 63%
- Low SiO_2 < 63%

Avg. compositions

- High- SiO_2
- Low- SiO_2
- Metavolcanic

SS zones

- Ladoga zone
- Saimaa zone
- Häme belt
- Uusima belt

References

- UCC
- PAAS

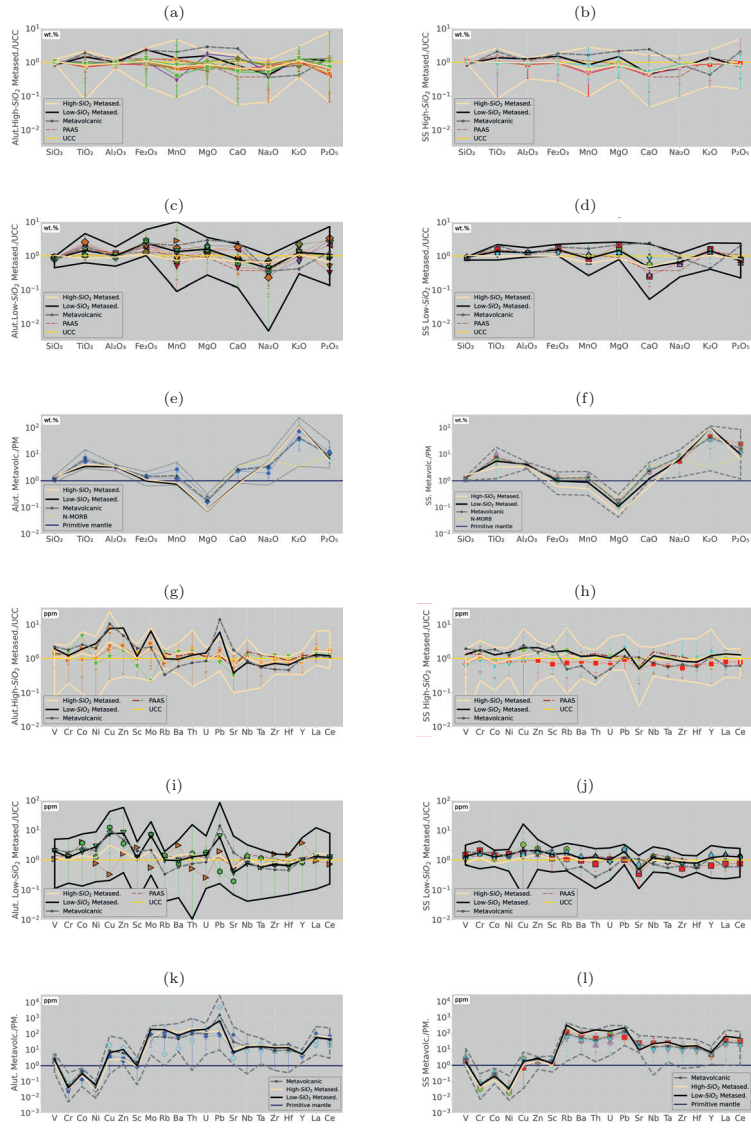
Figure 2: Boxplot and Harker diagram summary of metasedimentary and metavolcanic samples from the Alutaguse and South Svecofennian (SS) domains. (a) Boxplot of SiO_2 concentrations by lithological group for Alutaguse: (1) metasedimentary and (2) metavolcanic rocks; and for SS: (1) metasedimentary and (2) metavolcanic units. (b) Bivariate Harker diagrams of Alutaguse samples and (c) SS samples, plotting major oxides vs. Al_2O_3 . Sample marker borders distinguish each lithological group as follows: High- SiO_2 (light brown), Low- SiO_2 (black), and metavolcanics (dashed grey). This distinction is consistent throughout the text.

the SS zones, Saimaa has the highest averages, Uusimaa and Häme are intermediate, and Ladoga the lowest. P_2O_5 is lowest in Alutaguse and Uusimaa, while Häme, Saimaa, and Ladoga show higher averages. SO_3 is elevated in Alutaguse relative to Uusimaa, Häme, and Saimaa, and overlaps Ladoga, which is the most enriched SS zone, although Alutaguse displays the widest range extending to the highest values. Overall, the Alutaguse High- SiO_2 metasediments are geochemically similar to their SS counterparts but show slightly higher Fe_2O_3 , MgO , SO_3 , and K_2O , with moderate depletion in Al_2O_3 and Na_2O (Figs.3a,3b).

In the Low- SiO_2 metasediments, Alutaguse has the lowest average SiO_2 and the broadest range, extending to values far below the SS domains. Uusimaa and Saimaa are the most siliceous, while Häme and Ladoga are intermediate. TiO_2 in Alutaguse overlaps the SS range but is lower than Häme and Ladoga, with Uusimaa at the minimum. Al_2O_3 is depleted in Alutaguse compared to all SS domains, which peak in Saimaa and Häme, remain intermediate in Uusimaa, and are slightly lower in Ladoga. Fe_2O_3 is distinctly enriched in Alutaguse, exceeding Uusimaa, Häme, and Saimaa, while Ladoga is also ferruginous but below Alutaguse. MnO contents are generally low and similar across zones, though Alutaguse spans the widest range. MgO in Alutaguse is higher than Uusimaa, Häme, and Saimaa but lower than Ladoga, which is the most Mg-rich; Uusimaa is the least. CaO is more enriched in Uusimaa and Alutaguse, while Häme is lower, Saimaa strongly depleted, and Ladoga the lowest. Na_2O is lowest in Alutaguse and Saimaa, higher in Häme and Ladoga, and peaks in Uusimaa. K_2O in Alutaguse is comparable to Uusimaa, Häme, and Saimaa but lower than Ladoga. P_2O_5 is slightly higher in Alutaguse than in SS domains, which remain uniformly low. Finally, SO_3 is strongly enriched in Alutaguse, far

388 above Uusimaa, Häme, and Saimaa; among the SS domains, Ladoga is the most
 389 enriched but remains well below Alutaguse. Compared with the High- SiO_2
 390 metasediments, the Low- SiO_2 group is characterized by systematically lower
 391 SiO_2 and Al_2O_3 , but stronger enrichment in Fe_2O_3 , MgO , and SO_3 , along
 392 with broader variability across most oxides (Figs.3c,3d).

393 In the metavolcanic group, Alutaguse has SiO_2 values overlapping all SS
 394 zones, with averages close to Uusimaa and Ladoga, higher than Saimaa, but
 395 slightly lower than Häme. TiO_2 in Alutaguse is comparable to Uusimaa and
 396 Häme, lower than Saimaa, and higher than Ladoga. Al_2O_3 is depleted in Aluta-
 397 guse relative to all SS zones, which peak in Häme, followed by Uusimaa, Ladoga,
 398 and Saimaa. Fe_2O_3 is enriched in Alutaguse, higher than Uusimaa, Häme, and
 399 Ladoga, though Saimaa shows the highest average values. MnO is slightly
 400 higher in Alutaguse and Saimaa compared to the other belts. MgO is strongly
 401 enriched in Alutaguse, exceeding all SS zones, with Ladoga and Uusimaa inter-
 402 mediate and Häme the lowest. CaO in Alutaguse is comparable to Uusimaa and
 403 higher than Häme and Saimaa, while Ladoga is intermediate. Na_2O is lowest
 404 in Alutaguse, distinctly below Uusimaa, Häme, and Saimaa, with Ladoga in-
 405 termediate. K_2O is also lowest in Alutaguse, overlapping Uusimaa and Saimaa
 406 but clearly below Häme and Ladoga. P_2O_5 in Alutaguse is lower than Häme,
 407 Saimaa, and Ladoga, and slightly below Uusimaa. Finally, SO_3 is highly ele-
 408 vated in Alutaguse, well above all SS domains; within the SS zones, Uusimaa
 409 and Häme show moderate values, Saimaa is lower, and Ladoga lacks data. Over-
 410 all, the Alutaguse metavolcanics differ from their SS counterparts by distinct
 411 enrichment in Fe_2O_3 , MgO , and SO_3 , and depletion in Al_2O_3 , Na_2O , K_2O ,
 412 and P_2O_5 (Figs.3e,3f). Compared with the High- and Low- SiO_2 metasediments,
 413 the metavolcanics show consistently higher MgO and CaO , stronger SO_3 en-
 414 richment in Alutaguse units, and lower alkali contents, reflecting a more mafic
 415 and volatile-rich character.



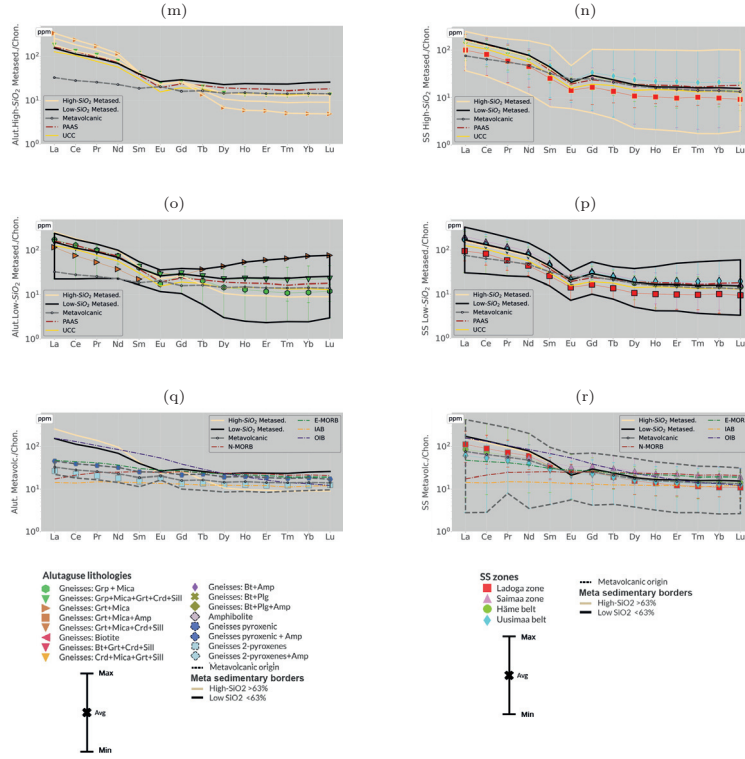


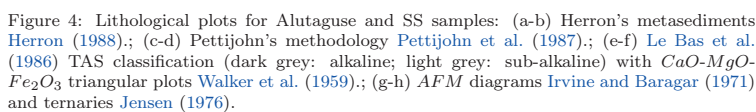
Figure 3: Normalised spider plots comparing geochemical trends between Alutaguse (left; after [Solano-Acosta et al. \(2025a\)](#)) and South Svecofennian (right) samples. (a-b) High- SiO_2 metasediments (major oxides) normalized to Upper Continental Crust (UCC; [Rudnick and Gao \(2003\)](#)); (c-d) Low- SiO_2 metasediments (major oxides) normalized to UCC; (e-f) Metavolcanic rocks (major oxides) normalized to Primitive Mantle (PM; [Sun and McDonough \(1989\)](#)); (g-h) High- SiO_2 metasediments (trace elements) normalized to UCC; (i-j) Low- SiO_2 metasediments (trace elements) normalized to UCC; (k-l) Metavolcanic rocks (trace elements) normalized to PM; (m-n) High- SiO_2 metasediments (REE patterns) normalized to chondrite ([Sun and McDonough \(1989\)](#)); (o-p) Low- SiO_2 metasediments (REE patterns) normalized to chondrite; (q-r) Metavolcanic rocks (REE patterns) normalized to chondrite.

416 Alutaguse (Fig.4a) and SS (Fig.4b) High- SiO_2 metasediments plot predom-

inantly within the litharenite field of Herron (1988), while Low- SiO_2 samples align with the graywacke and shale fields. In the Pettijohn et al. (1987) classification (Figs.4c,4d), High- SiO_2 samples again cluster within litharenite, while Low- SiO_2 samples fall into the graywacke field. Alutaguse and SS metavolcanics have similar $Mg\#$ (Table 1) and are subalkaline, tracking tholeiitic basalt fractionation trends, with most samples plot within the ortho-amphibolite domain (Figs.4e,4f), although few SS metavolcanics of Häme and Uusimaa extend to para-amphibolite affinities. In TAS plots (Le Bas et al., 1986), both groups fall within the subalkaline basalt-andesite series, though SS samples display higher $Na_2O + K_2O$ concentrations. Alutaguse metavolcanics follow a tholeiitic trend in AFM and Jensen diagrams (Fig.4g), whereas SS samples trend more calc-alkaline in AFM but remain dominantly tholeiitic in the Jensen plots (Fig.4h).

4.2. Geochemistry of Trace and Rare Earth Elements (REE)

In the High- SiO_2 metasediments, transition metals (Fig.3g) show that Alutaguse is enriched in V , Cr , Co , Ni , Cu , and Zn relative to Uusimaa and Ladoga, with values closer to Häme and Saimaa but showing wider ranges that extend to higher maxima. Sc is moderate in Alutaguse, overlapping Uusimaa, while Häme and Saimaa record higher averages and narrower spreads. Among the LILE (Fig.3h), Alutaguse is distinguished by very high Ba and Sr , higher than most SS zones except for the upper range of Saimaa, while Rb , Th , and Pb fall within the intermediate field. U contents are relatively low and overlap Ladoga, contrasting with slightly higher averages in Uusimaa and Häme. HFSE (Fig.3h) shows moderate Nb , Ta , and Hf in Alutaguse, overlapping all SS domains, but Zr and Y are consistently lower than in Uusimaa and Saimaa, and closer to Ladoga. REE patterns (Figs.3m, 3n) reveal that Alutaguse has the highest average $\sum REE$ and is strongly dominated by LREE, with narrow ranges and systematic depletion in HREE. In contrast, Uusimaa, Häme, and Saimaa display higher HREE contents and wider spreads, while Ladoga records the lowest overall REE abundances. The HREE/LREE ratios are lowest in Alutaguse, confirming strong LREE enrichment, whereas Uusimaa



447 shows the broadest variability. Negative *Eu* anomalies are present across all
 448 zones, but they are most pronounced in Uusimaa and Häme, while Alutaguse
 449 and Ladoga retain weaker anomalies. Overall, Alutaguse metasediments are
 450 marked by transition-metal enrichment, elevated *Ba–Sr*, moderate *HFSE*, and
 451 strong *LREE* dominance, setting them apart from the more *HREE*-balanced
 452 signatures of the Southern Svecofennian zones.

453 In the Low-*SiO*₂ metasediments, transition metals (Fig.3i) show systematic
 454 enrichments relative to the High-*SiO*₂ group. Alutaguse stands out with the
 455 highest averages and the broadest ranges in *V*, *Cr*, *Ni*, *Cu*, and *Zn*, reaching
 456 extreme maxima that exceed all SS domains. Häme and Saimaa also record
 457 elevated *Co* and *Ni*, but with narrower spreads, whereas Uusimaa and Ladoga
 458 display lower averages and more restricted variability. *Sc* contents are highest
 459 in Uusimaa, Häme, and Saimaa but consistently lower in Ladoga. *Mo* is only
 460 reported in Alutaguse samples, and its content is higher in the Low silica sam-
 461 ples. Among the *LILE* (Fig.3j), *Ba* and *Sr* are most enriched in Alutaguse and
 462 Saimaa, with wide ranges that exceed those in Häme, Uusimaa, and Ladoga. *Rb*
 463 and *Pb* show intermediate levels across all zones, though Uusimaa and Häme
 464 yield slightly higher averages than Ladoga. *Th* is lowest in Ladoga, while *U*
 465 overlaps across all belts but tends to be higher in Uusimaa and Häme. The
 466 *HFSE* (Fig.3j) display moderate *Nb* and *Ta* in all zones, with Alutaguse and
 467 Ladoga overlapping the lower Svecofennian ranges, while *Zr* and *Y* are clearly
 468 elevated in Uusimaa and Saimaa, extending to maxima well above those of Häme
 469 and Ladoga. *Hf* is moderate across all belts but slightly higher in Uusimaa and
 470 Saimaa. *REE* patterns (Figs.3o,3p) reveal that Alutaguse has relatively high
 471 $\sum\text{REE}$ but is dominated by *LREE*, with *HREE* restricted to narrow ranges.
 472 Uusimaa and Saimaa yield the highest $\sum\text{REE}$ and broader variability, with
 473 Häme intermediate and Ladoga consistently the lowest. The *HREE/LREE*
 474 ratios are lowest in Alutaguse, confirming strong *LREE* enrichment, whereas
 475 Uusimaa and Saimaa exhibit broader spreads. *Eu* anomalies are present across
 476 all belts, weak in Ladoga but more evident in Uusimaa and Saimaa, while Aluta-
 477 guse shows intermediate values. Overall, Low-*SiO*₂ metasediments are charac-

478 terised by strong transition-metal enrichment, high $Ba - Sr$, moderate $HFSE$,
 479 and elevated $\sum REE$ with pronounced $LREE$ dominance, setting them apart
 480 from the more $HREE$ -balanced patterns of the Southern Svecofennian belts.

481 The Low- SiO_2 group is distinguished by higher and more variable transition
 482 metals, stronger $Ba - Sr$ enrichment, and broader REE ranges, whereas the
 483 High- SiO_2 group shows lower transition metals, extreme but narrower $Ba - Sr$
 484 enrichments, and consistently higher $LREE$ dominance with more subdued
 485 $HREE$. Negative Eu anomalies occur in both groups but are weaker in Alu-
 486 taguse and Ladoga than in Uusimaa, Häme, and Saimaa. Overall, Low- SiO_2
 487 metasediments record stronger metal enrichment and wider variability, while
 488 High- SiO_2 counterparts display more homogeneous $LREE$ -dominated signa-
 489 tures.

490 In the metavolcanics (Fig.3k), transition metals are overall enriched relative
 491 to the metasedimentary groups. Alutaguse records the highest V , Cr , Ni , Cu ,
 492 and Zn , with maxima that greatly exceed the SS domains, while Uusimaa also
 493 shows elevated V and Cr but with narrower spreads. Häme and Saimaa yield
 494 intermediate values, and Ladoga consistently the lowest. Sc is notably high
 495 in Saimaa and Uusimaa. Mo content is higher than the High- SiO_2 metased-
 496 iments but lower than the Low- SiO_2 samples (Table 2). Among the $LILE$
 497 (Fig.3l), Ba and Sr are markedly enriched in Alutaguse and Häme, with ranges
 498 extending above Uusimaa and Ladoga. Rb and Pb are highly variable in Alu-
 499 taguse, while Th and U are generally low but slightly higher in Häme and
 500 Uusimaa. The $HFSE$ (Fig.3l) are moderate overall, with Nb and Ta show-
 501 ing overlapping fields across all belts, while Zr , Hf , and Y are most elevated
 502 in Saimaa and Häme and lowest in Ladoga. REE distributions (Figs.3q,3r)
 503 highlight Alutaguse volcanics with the highest $\sum REE$, dominated by $LREE$,
 504 whereas Uusimaa and Häme display broader spreads and greater $HREE$ con-
 505 tributions. Saimaa plots intermediate, and Ladoga remains the lowest. The
 506 $HREE/LREE$ ratios are lowest in Alutaguse, consistent with strong $LREE$
 507 enrichment, while Uusimaa and Häme exhibit wider variability. Eu anomalies
 508 are weak in Alutaguse and Ladoga but more pronounced in Uusimaa, Häme,

509 and Saimaa. Overall, metavolcanics are characterised by strong transition-metal
 510 enrichment, very high $Ba - Sr$, moderate $HFSE$, and elevated $\sum REE$ with
 511 marked $LREE$ dominance, distinguishing Alutaguse from the more $HREE$ -
 512 balanced SS samples.

513 Together, the three lithological groups reveal systematic trends in metal and
 514 trace element distributions across the Alutaguse and SS domains. Transition
 515 metals show a progressive enrichment from High- to Low- SiO_2 metasediments
 516 and reach their maximums in metavolcanics, with Alutaguse consistently record-
 517 ing the highest averages and the widest ranges, in contrast to the more restricted
 518 values of Uusimaa, Häme, Saimaa and Ladoga. $LILE$ display a similar pro-
 519 gression, with Ba and Sr dominating Alutaguse in all lithologies and extending
 520 beyond the ranges of the SS, while Rb , Th , U , and Pb remain more variable and
 521 often overlap. $HFSE$ remain moderate across groups, but Uusimaa, Häme, and
 522 Saimaa consistently yield higher Zr , Hf , and Y than Alutaguse and Ladoga.
 523 REE patterns highlight the strong $LREE$ enrichment of Alutaguse across all
 524 lithologies, with consistently low $HREE/LREE$ ratios and subdued Eu anoma-
 525 lies, whereas Uusimaa, Häme, and Saimaa exhibit broader ranges and stronger
 526 Eu depletions. Overall, the Alutaguse succession is marked by elevated transi-
 527 tion metals, extreme $Ba - Sr$ enrichment, moderate $HFSE$, and strong $LREE$
 528 dominance, distinguishing it from the more balanced REE and $HFSE$ signa-
 529 tures of the SS zones.

530 5. Discussion

531 Geochemical signatures in metasedimentary and metavolcanic rocks reflect
 532 weathering, alteration, and depositional processes, with key controls in metased-
 533 iments including sediment transport and sorting (McLennan et al., 1993; Cullers,
 534 1994), diagenesis or metamorphism during burial (Fedó et al., 1995, 1996), and
 535 source composition and depositional setting (Bhatia and Crook, 1986; Roser
 536 and Korsch, 1986), making it essential to evaluate these factors in Alutaguse
 537 and SS rocks before interpreting petrogenetic trends (Taylor and McLennan,

1985; Cullers et al., 1997; Gao and Wedepohl, 1995; Gao et al., 1999; Large
et al., 2001; Sifeta et al., 2005; Karakacs and Güçtekin, 2021; Jian et al., 2013;
Faisal et al., 2020; Gebremicale et al., 2025).

5.1. Weathering & Alteration indices

Weathering depletes Na , K , and $LILE$, while enriching Al_2O_3 , TiO_2 , $HFSE$
and REE . However, $HFSE$ and REE show minimal variation due to their low
mobility (Cullers et al., 1997; McLennan, 1989, 1993). To assess alteration
in metasediments, chemical indices are applied: the Chemical Index of Alter-
ation (CIA ; Nesbitt and Young (1982)), Plagioclase Index of Alteration (PIA ;
Fedo et al. (1995)), Chemical Index of Weathering (CIW ; Harnois (1988)),
and Index of Compositional Variation (ICV ; Cox et al. (1995)). For metavol-
canics, indices include the Alteration Index (AI ; Ishikawa et al. (1976)), Chlo-
rite–Carbonate–Pyrite Index ($CCPI$; Large et al. (2001)), Parker Index (WIP ;
Parker (1970)), and the Sericitisation Index (SI ; MacLean and Hoy (1991);
Karakacs and Güçtekin (2021)).

The CIA quantifies feldspar alteration into aluminous phases by tracking
the loss of mobile cations (Na , Ca , K), reflecting chemical weathering intensity
and paleoclimate conditions, with values < 60 indicating initial weathering,
60–80 intermediate stages, and > 80 extreme alteration (Nesbitt and Young,
1982; McLennan, 1989; Fedo et al., 1995; Han et al., 2019). Representative
 CIA values include basalts (0–45), granites and granodiorites (45–55), illite
and smectite (75–85), and kaolinite or gibbsite (approaching 100). It is defined
as $CIA_{[mol]} = (Al_2O_3 / (Al_2O_3 + CaO^* + Na_2O + K_2O)) \times 100$, where CaO^* is
the silicate-bound calcium fraction (Fedo et al., 1995). Following the correction
procedure of McLennan (1993), CaO is adjusted in two steps: (1) the apatite
fraction is subtracted using $CaO' = CaO - (10/3 \times P_2O_5)$; (2) if $CaO' > Na_2O$,
then $CaO = Na_2O$, otherwise $CaO^* = CaO'$. The PIA , defined as $PIA_{[mol]} =$
 $((Al_2O_3 - K_2O) / (Al_2O_3 + CaO^* + Na_2O - K_2O)) \times 100$, tracks plagioclase
weathering, ranging from ~ 50 in unaltered rocks to 100 in clays (Fedo et al.,
1995). The $CIW_{[mol]} = Al_2O_3 / (Al_2O_3 + CaO^* + Na_2O) \times 100$, or potassium-free

568 *CIA*, isolates feldspar-to-clay alteration, spanning ~ 50 (unaltered feldspar) to
569 100 (kaolinite) (Harnois, 1988; Fedo et al., 1995; Saleh et al., 2022). Increasing
570 *CIA* and *CIW* values reflect cation loss and structural change during alteration
571 (Karakacs and Güçtekin, 2021). The $ICV_{[wt.\%]} = (Fe_2O_3 + K_2O + Na_2O +$
572 $CaO + MgO + TiO_2)/Al_2O_3$ measures compositional maturity and sediment
573 recycling, with values < 1 reflecting mature or recycled sediments, and values
574 > 1 denoting immature, first-cycle deposits (Cullers and Podkovyrov, 2000;
575 El-Bialy, 2013; Han et al., 2019).

576 The *A-CN-K* ternary diagram (Nesbitt and Young, 1984) also provides
577 insight into weathering intensity and source composition. Ideal weathering
578 trends align with the *A-CN* line, while their intersection indicates the origi-
579 nal source composition (Nesbitt and Young, 1984; Fedo et al., 1995). In both
580 Alutaguse (Fig.S1a) and the SS metasediments (Fig.S1b), this approach comple-
581 ments chemical indices: Low-*SiO*₂ samples project toward the *Al*₂*O*₃ apex, close
582 to the PAAS, implying stronger weathering, while High-*SiO*₂ samples cluster
583 near UCC reference fields from tonalitic to granitic sources.

584 In the High-*SiO*₂ metasediments (Table 1), *CIA*, *PIA*, and *CIW* values
585 cluster in intermediate fields (60–70) with narrow spreads. Alutaguse overlaps
586 Uusimaa and Häme but extends to higher *PIA* and *CIW* maxima, while Saimaa
587 consistently yields the highest averages, reflecting more advanced feldspar al-
588 teration. Ladoga shows the lowest intervals, suggesting subdued weathering.
589 *ICV* values hover near unity: Alutaguse, Uusimaa, and Häme slightly above 1,
590 pointing to immature signatures, whereas Saimaa falls below 1, consistent with
591 greater recycling and compositional maturity (Figs.5a,5b).

592 In the Low-*SiO*₂ metasediments (Table 1), weathering indices rise system-
593 atically, with higher averages and broader ranges extending into the extreme
594 weathering field. Alutaguse shows the widest spreads and highest maxima,
595 highlighting intense feldspar and plagioclase alteration compared to the Sve-
596 cofennian belts. Saimaa again yields elevated averages, while Häme and Uusi-
597 maa plot at intermediate levels with narrower ranges. Ladoga remains the least
598 weathered, with consistently restricted variability. Unlike the High-*SiO*₂ group,

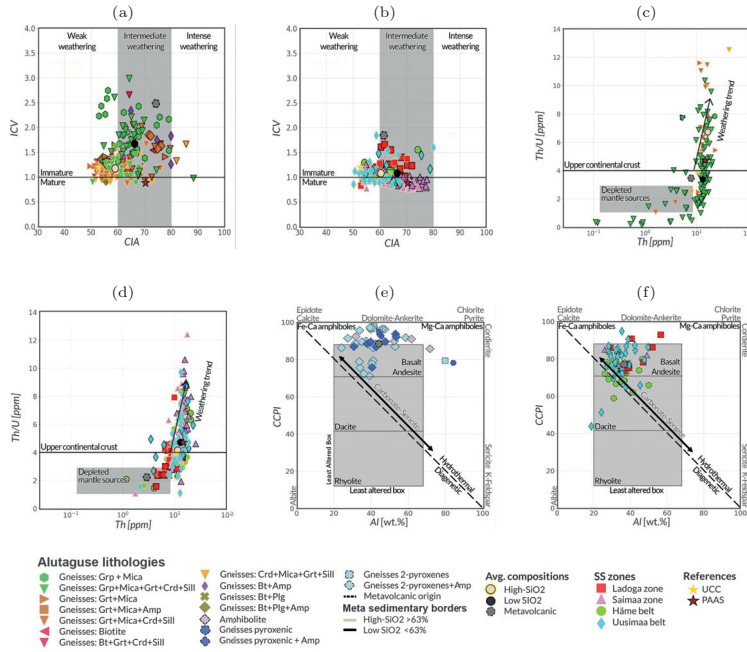


Figure 5: Weathering and alteration discrimination plots for Alutaguse and South Svecofennian (SS) samples: (a-b) ICV vs. CIA (after Nesbitt and Young (1984); Cox et al. (1995)).; (c-d) Th/U vs. Th (after McLennan (1993)).; (e-f) $CCPI$ vs. AI alteration box plots for metavolcanics (after Large et al. (2001)).

599 ICV values are distinctly higher in Alutaguse, marking immature (Fig.5a), first-
600 cycle compositions, whereas Saimaa remains below unity, showing a recycled,
601 compositionally mature character (Fig.5b).

602 The Th/U ratio in rocks reveals weathering conditions, with upper crustal
603 values typically in the 3.5 – 4.0 range (McLennan, 1993). Increased weathering
604 raises this ratio due to the oxidation of U^{4+} into more soluble U^{6+} , which
605 is readily lost during sedimentation (Taylor and McLennan, 1985; McLennan
606 et al., 1990, 1995). Sediments from tectonically active settings often display
607 ratios below 3.5 (McLennan and Taylor, 1991). Very high values are reported
608 in metapelites and metagreywackes (up to 21), garnet–alkali feldspar gneisses

609 and quartz-rich garnet gneisses (~ 12), while calc-silicate rocks average near
610 the upper crustal reference (3.6) (Taylor and McLennan, 1985). Mantle-derived
611 volcanic rocks typically have lower ratios due to their depleted signatures (Singh
612 and Khan, 2017). Variations in Th/U can also record metamorphic overprint or
613 U enrichment: metamorphism may elevate ratios to 8–10 in upper amphibolite
614 facies and up to 25 in granulites, whereas low ratios may indicate U addition
615 and enhanced mobility (Wedepohl, 1991). Most samples exceed both UCC
616 (0.62 ppm) and PAAS (4.71 ppm), indicating moderate–strong U depletion from
617 oxidative weathering or metamorphism.

618 In the High- SiO_2 metasediments (Table 2), Th/U ratios are highly vari-
619 able, with Alutaguse showing the highest average and widest range, consistent
620 with its elevated CIA , PIA , and CIW maxima. This suggests that stronger
621 feldspar alteration was accompanied by selective U loss, enhancing the Th/U
622 signal. Saimaa also records relatively high Th/U , in line with its high weathering
623 indices, whereas Uusimaa and Häme plot in intermediate fields with narrower
624 spreads. Ladoga yields the lowest Th/U , consistent with its subdued weath-
625 ering and restricted index variability. ICV values close to unity across most
626 zones highlight immaturity in Alutaguse, Uusimaa, and Häme, contrasting with
627 Saimaa, where low ICV aligns with both higher CIA – PIA – CIW and moder-
628 ately elevated Th/U , indicating a more mature, recycled character.

629 In the Low- SiO_2 metasediments (Table 2), Th/U ratios shift systematically
630 with weathering intensity. Saimaa shows the highest averages, consistent with
631 its elevated CIA , PIA , and CIW values, confirming strong feldspar and pla-
632 gioclase alteration coupled with U depletion. Uusimaa and Häme yield interme-
633 diate Th/U , matching their moderate weathering indices, while Ladoga records
634 the lowest ratios, in agreement with its limited chemical alteration and relatively
635 high ICV . Alutaguse stands out for its wide Th/U range, reflecting variable
636 U mobility under strong weathering conditions, which also corresponds to its
637 highest CIA – PIA – CIW values and distinctly immature ICV .

638 CIA , PIA , ICV , and CIW indices have proven significant in evaluating
639 volcanic rock alteration (Karakacs and Güçtekin, 2021). In the metavolcanics

(Table 1), *CIA*, *PIA*, and *CIW* values are generally higher than in the metasedimentary groups, pointing to stronger feldspar and plagioclase alteration. Alutaguse samples span the widest ranges and extend to the highest maxima, while the SS domains are more tightly constrained. Within the SS, Saimaa consistently yields elevated averages, whereas Häme and Uusimaa occupy the lower end with the narrowest spreads. *ICV* values in Alutaguse metavolcanics are also broader and higher, reflecting stronger compositional variability and alteration effects, while the SS domains cluster within narrower, more homogeneous fields. The *Th/U* ratios reinforce this contrast: Alutaguse records the highest averages and extreme ranges, spanning signatures from upper continental crust to depleted mantle, whereas the SS domains concentrate closer to depleted mantle values (Table 2). Häme and Uusimaa display the most restricted spreads, while Ladoga and Saimaa extend slightly higher but remain less variable than Alutaguse. Overall, Alutaguse metavolcanics preserve greater variability in both alteration indices and *Th/U* (Fig.5c), while the SS metavolcanics are more constrained and depleted-like (Fig.5d).

The *AI*, *CCPI*, *WIP*, and *SI* indices provide complementary constraints on volcanic alteration (Parker, 1970; Large et al., 2001; Bailie et al., 2011; Karakacs and Güçtekin, 2021). $AI_{[wt.\%]} = 100 \times (MgO + K_2O)/(MgO + K_2O + CaO + Na_2O)$ reflects sodic plagioclase replacement by sericite and chlorite (Ishikawa et al., 1976); $CCPI_{[wt.\%]} = 100 \times (FeO + MgO)/(FeO + MgO + Na_2O + K_2O)$ tracks albite–K-feldspar alteration to Mg–Fe chlorite (Large et al., 2001); $WIP_{[wt.\%]} = 100 \times (2 \times Na_2O/0.35 + MgO/0.9 + 2 \times K_2O/0.25 + CaO/0.7)$ emphasises alkali and alkaline-earth mobility (Parker, 1970); and $SI_{[wt.\%]} = K_2O/(Na_2O + K_2O)$ records sericite formation from altered plagioclase (McLennan and Taylor, 1991). Moderately altered rocks typically fall within the least altered fields or higher *AI*, whereas strongly altered samples yield very high *AI* (> 80) and/or *CCPI* (> 95).

In the metavolcanics (Table 1), Alutaguse extends into these more altered fields, with consistently high *CCPI* and variable *AI*, reflecting enhanced carbonate–sericite alteration. *WIP* values are lower and more variable than in the

671 SS domains, marking stronger alkali mobility, while SI spans the widest ranges,
672 consistent with heterogeneous sericitisation. In contrast, the SS metavolcanics
673 remain confined to narrower $AI-CCPI$ ranges: Häme and Uusimaa at the lower
674 end, and Saimaa and Ladoga clustering in intermediate fields. When compared
675 with the $CIA-PIA-CIW-ICV$ system, these indices confirm the same trend:
676 Alutaguse metavolcanics exhibit the widest spreads and strongest alteration,
677 while SS domains preserve more restricted, depleted-like alteration fields.

678 This contrast is also evident in alteration plots. The Alutaguse $K_2O + Na_2O$
679 vs. SI diagram (Fig.S1c) places most samples outside the igneous spectrum, in-
680 dicating metasomatic hydrothermal alteration, while the CaO vs. SI plot shows
681 systematic CaO depletion with increasing SI , suggesting carbonation influx
682 (Karakacs and Güçtekin, 2021). SS metavolcanics plot largely within the igneous
683 spectrum, with concentrated $CaO-SI$ distributions and limited alteration, ex-
684 cept for Ladoga samples extending into the hydrothermal field (Fig.S1d). On
685 the alteration box plot of Large et al. (2001), most Alutaguse (Fig.5e) and SS
686 samples (Fig.5f) fall in the basalt/andesite field, though some Häme and Uusi-
687 maa extend into the dacite zone above the diagenetic field. Alutaguse is more
688 affected by hydrothermal alteration, with stronger carbonate-sericite overprint,
689 elevated $CCPI$, and SO_3 (Fig.2). Eventhought, LOI values tend to be lower
690 than 2wt% 1. SS samples mostly remain in the least altered fields. Both groups
691 trend toward the calcite-epidote node, suggesting $Fe-Ca$ amphibole affinities.

692 5.2. Metasedimentary rocks

693 Most Alutaguse and SS metasediments show average values consistent with
694 moderate weathering ($CIA < 80$) and chemical immaturity ($ICV > 1$, $SiO_2/Al_2O_3 <$
695 6; Table 1), making them suitable for provenance and tectonic analysis (El-Bialy,
696 2013; Han et al., 2019). Generally low LOI values (avg. $< 4\%$) indicate lim-
697 ited post-depositional alteration, supporting the reliability of the dataset (Han
698 et al., 2019), although the Alutaguse Low- SiO_2 ; principally the graphitiferous
699 gneisses subset, records the highest LOI (Table 1), consistent with its elevated
700 weathering indices.

701 5.2.1. Provenance

702 Al_2O_3/TiO_2 ratios, utilized in rock origin studies (El-Bialy (2013); Ofili et al.
703 (2022)), often mirror those of the source rocks in sandstones and mudstones
704 (Hayashi et al. (1997)). Differences in Al and Ti between silts or shales and their
705 origins are minimal because Ti mainly exists within clay minerals or as ilmenite
706 inclusions (Hayashi et al. (1997)). This ratio is a reliable indicator for igneous
707 source rocks (Han et al. (2019)). 3–16, 8–21, and 21–70 ranges indicate mafic,
708 intermediate, and felsic sources, respectively (Hayashi et al. (1997); Kelemen
709 et al. (2003)). In the High- SiO_2 metasediments, averages align closely with the
710 UCC value (24.06), pointing to felsic sources. Alutaguse records the highest
711 averages and the widest ranges, extending well beyond the felsic field, while the
712 SS domains remain more constrained around UCC (Table 1). Among the SS,
713 Saimaa extends to the highest ratios, Uusimaa and Häme cluster nearer to UCC,
714 and Ladoga yields the lowest and most restricted spreads. In the Low- SiO_2
715 group, values plot closer to PAAS (18.9), consistent with intermediate–felsic
716 inputs. Alutaguse again shows the broadest range, spanning mafic to felsic
717 fields, while Uusimaa and Saimaa register higher averages in the felsic range,
718 Häme remains intermediate, and Ladoga consistently plots at the lower end with
719 narrow variability.

720 The $A - CN - K$ ternary diagram for Alutaguse (Fig.S1a) and SS metased-
721 iments (Fig.S1b) reveals source compositions by projecting weathering trends
722 toward the feldspar join (Nesbitt and Young, 1984, 1989; Fedo et al., 1995).
723 Both datasets accommodate preferably within tonalite–granite trends, suggest-
724 ing intermediate to felsic igneous sources—consistent with the Al_2O_3/TiO_2 ratio
725 interpretations.

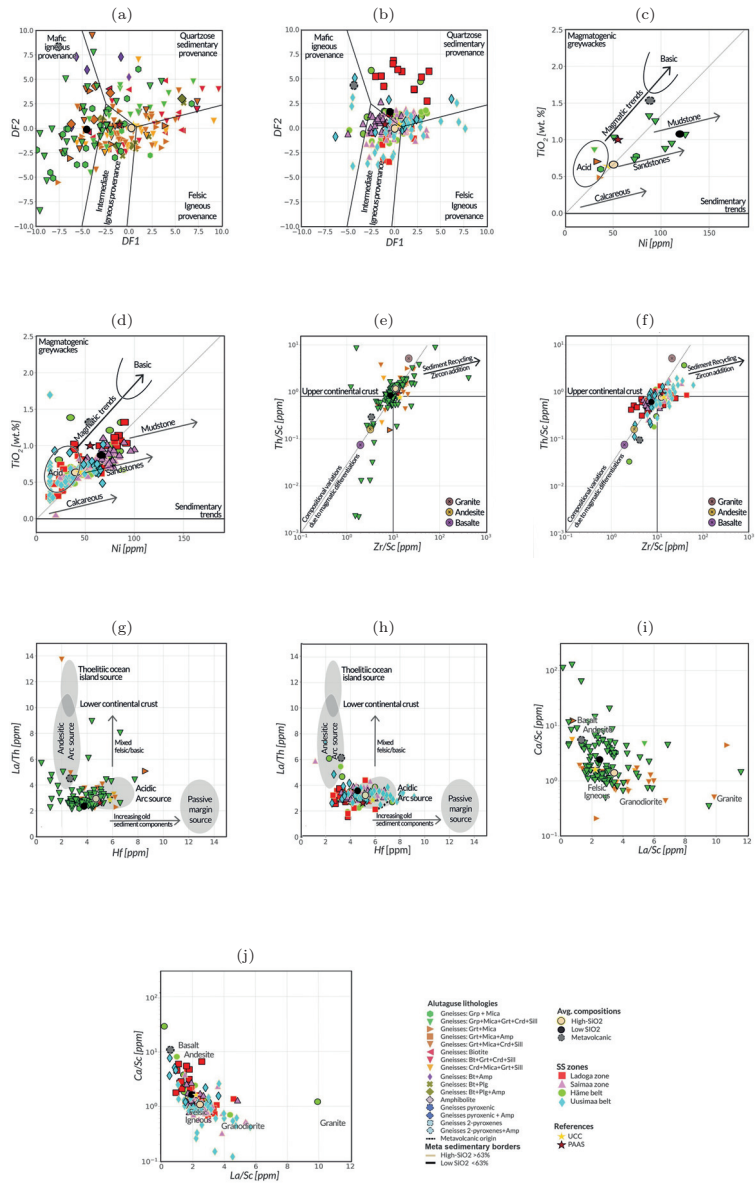


Figure 6: Major-element provenance plots of the Alutaguse and South Svecofennian (SS) metasediments. Discriminant function plots (a–b) show provenance signatures using major elements with equations after [Roser and Korsch \(1988\)](#), where $DF_1 = 30.638 \frac{TiO_2}{Al_2O_3} - 12.541 \frac{Fe_2O_3}{Al_2O_3} + 7.329 \frac{MgO}{Al_2O_3} + 12.031 \frac{Na_2O}{Al_2O_3} + 35.402 \frac{K_2O}{Al_2O_3} - 6.382$ & $DF_2 = 56.500 \frac{TiO_2}{Al_2O_3} - 10.879 \frac{Fe_2O_3}{Al_2O_3} + 30.875 \frac{MgO}{Al_2O_3} - 5.404 \frac{Na_2O}{Al_2O_3} + 11.112 \frac{K_2O}{Al_2O_3} - 3.890$; (c–d) TiO_2 vs. Ni plots after [Floyd et al. \(1989\)](#); Th/Sc vs. Zr/Sc (e–f) per [McLennan et al. \(1993\)](#); [Condie \(1993\)](#); La/Th vs. Hf (g–h) after [Floyd and Leveridge \(1987\)](#) and La/Sc vs. Co/Th (i–j) after [Gu et al. \(2002\)](#).

726 [Roser and Korsch \(1988\)](#) proposed an enhanced chemical index for determin-
 727 ing clastic sediment sources. In both Alutaguse (Fig.6a) and SS (Fig.6b), most
 728 High- SiO_2 metasediments plot near the intermediate–felsic boundary. SS Low-
 729 SiO_2 samples generally indicate intermediate sources, with Ladoga trending to-
 730 ward the quartzose sedimentary field. Alutaguse Low- SiO_2 samples span mafic
 731 to intermediate fields, with a stronger mafic signal, especially among graphite-
 732 bearing mica gneisses.

733 K and Rb , as mobile lithophile elements, may be altered during diagenesis
 734 and low-grade metamorphism but retain value for identifying igneous sources
 735 ([Floyd and Leveridge, 1987](#)). In both Alutaguse (Fig.S1g) and SS (Fig.S1h)
 736 metasediments, high K_2O and Rb values follow a trend close to the magmatic
 737 differentiation line at a 230 ratio ([Shaw, 1968](#)), implying felsic to intermediate
 738 source rocks.

739 $TiO_2 - Ni$ plots offer additional provenance insight. In both Alutaguse
 740 (Fig.6c) and SS (Fig.6d), High- SiO_2 metasediments cluster within the acidic
 741 source field, indicating minimal mafic input. Alutaguse Low- SiO_2 samples,
 742 mainly graphite-bearing mica gneisses, align with the sandstone sedimentary
 743 trend. SS Low- SiO_2 samples follow a magmatic trend from basic toward in-
 744 termediate compositions, though some Ladoga and Saimaa samples also show
 745 sedimentary affinity, though not as clearly as Alutaguse.

746 Transition metals concentrate in mafic minerals such as pyroxene and olivine
 747 ([Cullers et al., 1997](#)), whereas felsic rocks are enriched in $HFSE$ through melt
 748 differentiation ([Feng and Kerrich, 1990](#)). Both groups are relatively immo-
 749 bile during alteration, making them reliable provenance indicators ([Armstrong-](#)

Altrin et al., 2004). In the High- SiO_2 metasediments, Alutaguse is enriched in *LILE* and transition metals compared to the SS domains and UCC (Figs.3g, 3m), while Uusimaa and Saimaa consistently yield higher *HFSE*. In the Low- SiO_2 group, Alutaguse shows even stronger enrichment in *LILE*, *Pb*, and transition metals—particularly *Co* (Table 2)—with maxima often surpassing those of the SS domains (Fig.3i), though *HFSE* again remain lower than in Uusimaa or Saimaa. These patterns point to dominantly felsic sources for the High- SiO_2 rocks, whereas the Low- SiO_2 Alutaguse samples reflect a wider provenance range, with also mafic insights.

Trace element ratios such as La/Sc , La/Co , Th/Sc , Th/Co , and Cr/Th distinguish mafic from felsic sources and provide key insights into sedimentary provenance (Armstrong-Altrin et al., 2004). These values for Alutaguse and SS are summarized in Table 3. High- SiO_2 metasediments align with felsic source signatures. Low- SiO_2 samples show elevated La/Sc , La/Co , and Th/Sc ratios above typical mafic values, suggesting a stronger intermediate rather than mafic source influence.

Th and *Zr* concentrate in felsic rocks due to their incompatibility, while *Sc* occurs in early-formed mafic minerals like olivine and pyroxene (McLennan and Taylor, 1991). Thus, Th/Sc and Zr/Sc ratios are effective provenance indicators (McLennan et al., 1990, 1993; Spalletti et al., 2008). In Alutaguse and SS metasediments, Th/Sc vs. Zr/Sc plots (Figs.6e,6f) mostly follow the igneous differentiation trend from andesite to granite. Low- SiO_2 samples plot at lower ratios than High- SiO_2 ones. Some Alutaguse samples trend toward basaltic compositions. This is supported by La/Th vs. Hf (Figs.6g,6h) and La/Sc vs. Co/Th (Figs.6i,6j) diagrams.

REE have exceptionally low partition coefficients between water and rock, enabling their effective transfer from source rocks to clasts (Gao and Wedepohl, 1995). Their stability during weathering, transport, diagenesis, and medium-grade metamorphism makes them reliable provenance indicators (Chaudhuri and Cullers, 1979; McLennan, 1989; Cullers et al., 1997). Mafic rocks typically show low $\sum REE$ and weak Eu^* anomalies, whereas felsic rocks display

		<i>La/Sc</i>	<i>La/Co</i>	<i>Th/Sc</i>	<i>Th/Co</i>	<i>Cr/Th</i>
Range of Sediments from Felsic Sources		2.50-16.3	1.80-13.8	0.84-20.05	0.67-19.4	4-15
Range of Sediments from Mafic Sources		0.43-0.86	0.14-0.38	0.05-0.22	0.04-1.40	25-500
UCC		2.21	1.76	0.77	0.62	7.90
PAAS		2.39		0.91		7.53
Alutaguse High- <i>SiO</i> ₂ Metasedimentary samples	Garnet bearing mica gneisses $\pm$ Crd $\pm$ Sil	4.28, 1.20-9.87	4.11, 1.55-13.81	1.33, 0.41-3.74	1.10, 0.53-2.29	6.86, 1.59-16.99
	Garnet bearing mica gneisses	4.86, 2.03-10.71	4.48, 0.76-10.98	1.56, 0.69-3.15	1.75, 0.22-4.78	4.86, 2.58-8.35
	Cordierite bearing mica gneisses $\pm$ Grt $\pm$ Sil	3.12, 0.72-4.77	2.44, 0.58-3.54	1.09, 0.22-1.95	0.81, 0.18-1.09	5.28, 1.76-6.67
	Graphite bearing mica gneisses $\pm$ Grt $\pm$ Crd $\pm$ Sil	5.38, 4.12,	0.55, 3.64,	2.06, 1.33,	0.21, 1.09,	7.94, 6.35,
	Total	0.72-10.71	0.55-13.81	0.22-3.74	0.18-4.78	1.59-16.99
		4.36,	2.02,	1.43,	0.67,	32.18,
Alutaguse Low- <i>SiO</i> ₂ Metasedimentary samples	Graphite bearing mica gneisses $\pm$ Grt $\pm$ Crd $\pm$ Sil	0.07-77.88	0.03-21.75	0.00-33.25	0.00-9.64	0.30-945.45
	Graphite bearing mica gneisses	2.49, 1.46-3.52	0.62, 0.47-0.77	1.03, 0.67-1.39	0.26, 0.22-0.30	6.59, 5.48-7.70
	Garnet bearing mica gneisses	0.79	0.41	0.15	0.08	21.69
	Total	4.27, 0.07-77.88	1.98, 0.03-21.75	1.40, 0.00-33.25	0.65, 0.00-9.64	31.38, 0.30-945.45
SS Zones High- <i>SiO</i> ₂ Metasedimentary samples	Uusimaa	2.91, 1.07-5.65	5.60, 1.54-26.70	0.89, 0.28-1.95	1.70, 0.38-8.20	6.40, 1.22-14.30
	Häme	2.22, 0.99-4.37	2.59, 0.67-4.67	0.75, 0.18-1.33	0.87, 0.12-1.42	13.99, 5.38-82.95
	Saimaa	2.69, 1.51-5.44	3.70, 1.38-10.49	0.80, 0.26-1.83	1.09, 0.26-3.05	10.32, 3.00-25.06
	Ladoga	2.53, 1.55-4.59	2.24, 0.55-4.38	0.83, 0.52-1.34	0.74, 0.18-1.50	13.95, 5.23-33.81
	Total	2.64, 0.99-5.65	3.86, 0.55-26.70	0.82, 0.18-1.95	1.18, 0.12-8.20	10.34, 1.22-82.95
		2.02, 0.53-3.63	2.05, 0.63-3.35	0.63, 0.11-1.29	0.64, 0.13-1.01	17.02, 5.59-104.94
SS Zones Low- <i>SiO</i> ₂ Metasedimentary samples	Häme	2.51, 0.20-9.93	1.98, 0.21-3.63	0.83, 0.03-3.68	0.62, 0.03-0.94	12.80, 7.67-36.69
	Saimaa	2.21, 1.62-4.85	2.58, 1.21-4.06	0.69, 0.46-1.59	0.80, 0.36-1.25	10.54, 6.30-18.23
	Ladoga	1.53, 0.04-2.58	1.02, 0.48-2.68	0.52, 0.31-0.73	0.36, 0.15-0.93	24.99, 7.46-49.09
	Total	2.10, 0.20-9.93	2.08, 0.21-4.06	0.67, 0.03-3.68	0.65, 0.03-1.25	14.85, 5.59-104.94

Table 3: Range of elemental ratios of the Alutaguse and South Svecofennian (SS) metasedimentary units compared to the ratios of the upper continental crust (after Rudnick and Gao (2003)) and sediments derived from felsic rocks and mafic rocks (after Armstrong-Altrin et al. (2004)).

781 higher $\sum REE$ and stronger negative Eu^* signatures (Cullers et al., 1997). In
 782 the High- SiO_2 metasediments (Table 2), Alutaguse yields the highest $\sum REE$,
 783 dominated by $LREE$ and with very low $HREE/LREE$, while SS domains
 784 record lower totals and more balanced patterns. Within the SS, Uusimaa and
 785 Saimaa extend to the highest $HREE$ values, whereas Häme is intermediate and
 786 Ladoga consistently the lowest. In the Low- SiO_2 group, $\sum REE$ increases over-
 787 all, with Saimaa and Häme recording the highest averages, Uusimaa intermedi-
 788 ate, and Ladoga again the lowest. Alutaguse shows broad ranges but remains
 789 $LREE$ -dominated, with $HREE/LREE$ values consistently higher than in its
 790 High- SiO_2 counterpart. Negative Eu anomalies are ubiquitous, strongest in
 791 Uusimaa and Häme, weaker in Ladoga and Alutaguse, and moderate in Saimaa.
 792 Altogether, REE patterns highlight felsic to intermediate sources, with Aluta-
 793 guse marked by extreme $LREE$ enrichment and variability, while the SS do-
 794 mains retain more balanced $HREE$ contributions and stronger Eu anomalies
 795 (Figs. 3m–3p).

796 5.2.2. Sorting, recycling and maturation

797 Sorting during sedimentary transport and sedimentation determines sedi-
 798 ments' mineralogical and chemical characteristics. This process can be assessed
 799 by analysing the textural maturity, considering factors such as grain sizes,
 800 morphologies, and their corresponding mineralogical and geochemical profiles
 801 (McLennan et al. (1993)). Low SiO_2/Al_2O_3 ratios indicate insignificant sedi-
 802 mentary sorting (Taylor and McLennan (1985); Rudnick and Gao (2003); Chen
 803 et al. (2014)). In the High- SiO_2 metasediments, averages are close to the UCC
 804 benchmark (Table 1), suggesting moderate sorting. Alutaguse, Uusimaa, and
 805 Häme cluster near UCC, Saimaa is slightly lower, and Ladoga yields the highest
 806 averages and broadest ranges, indicating stronger variability. In the Low- SiO_2
 807 metasediments, ratios align closely with PAAS (Table 1), reflecting limited sort-
 808 ing. Alutaguse shows the widest spread, while Uusimaa and Saimaa yield higher
 809 averages, Häme remains the lowest, and Ladoga matches the PAAS reference.

810 The K_2O/Na_2O ratios of all metasediments exceed unity, confirming chemi-

cal immaturity (El-Bialy, 2013). Low- SiO_2 samples yield the highest values, with Alutaguse showing extreme variability beyond the SS domains, where Saimaa and Ladoga exceed Uusimaa and Häme. High- SiO_2 samples plot closer to the UCC, but with wider spreads in Alutaguse and Saimaa. Complementary Al_2O_3/TiO_2 ratios, where narrower ranges reflect stronger immaturity (El-Bialy, 2013), reinforce this contrast: Low- SiO_2 metasediments are more immature, while High- SiO_2 counterparts display broader felsic-extending ranges.

On the provenance discrimination diagrams of Roser and Korsch (1988) (Figs.6a,6b), most Alutaguse and SS metasediments—regardless of SiO_2 content—cluster within the intermediate igneous provenance field, with Alutaguse Low- SiO_2 samples showing a slightly more mafic trend. In contrast, the SS Low- SiO_2 Ladoga samples fall within the quartzose sedimentary field, indicating contributions from recycled, mature detritus and older sedimentary rocks (Kotova et al., 2009).

All metasediments yield $CIW/CIA > 1$, indicating derivation from broad provenance regions and long transport paths rather than local sources (Gao et al., 1999; El-Bialy, 2013). Alutaguse shows the widest ranges and highest ratios, reflecting stronger recycling and heterogeneous inputs, whereas the SS domains are more constrained. Within the SS, Saimaa and Uusimaa lie at intermediate levels, Häme is the most uniform and proximal, and Ladoga varies between higher ratios in the Low- SiO_2 group and lower values in the High- SiO_2 subset.

The $100 \times TiO_2/Zr$ ratio discriminates sediment maturity, with values < 0.33 denoting psammitic (immature, coarse-grained) and > 0.33 pelitic (mature, fine-grained) signatures (Garcia et al., 1994). In the High- SiO_2 group, Alutaguse and Uusimaa plots near the psammitic–pelitic threshold and the other SS records preferably pelitic signatures with the widest variability. In the Low- SiO_2 group, all zones yield pelitic averages, with Ladoga and Alutaguse showing the broadest ranges that extend to extreme values, indicating stronger recycling.

The immobile trace element ratios Th/Sc and Zr/Sc provide reliable indicators of sedimentary recycling (McLennan et al., 1990, 1993; Condie, 1993).

842 In the High- SiO_2 metasediments, Alutaguse shows wider Th/Sc variability but
843 more moderate Zr/Sc than Uusimaa and Saimaa, which are strongly influenced
844 by zircon recycling. Häme and Ladoga remain at the lower end, reflecting less
845 recycled signatures. In the Low- SiO_2 group, Alutaguse displays the broadest
846 Zr/Sc spread, far exceeding the SS domains, while Uusimaa, Häme, and Saimaa
847 show intermediate ranges and Ladoga stays most restricted. The combined
848 Th/Sc – Zr/Sc trends indicate that the SS domains largely preserve first-cycle
849 characteristics, whereas Alutaguse records significant recycling, particularly in
850 the Low- SiO_2 subset.

851 Heavy minerals such as zircon, monazite, and allanite may accumulate through
852 sedimentary reworking and sorting, generating elevated $(Gd/Yb)_n$ ratios, which
853 typically range between 1.0–2.0 in post-Archean sediments (McLennan and
854 Taylor, 1991; McLennan et al., 1993). High- SiO_2 Alutaguse samples show
855 anomalously high ratios (Table 2), indicating strong heavy-mineral enrichment,
856 whereas the SS domains remain mostly within or slightly above the upper-
857 crustal range. In the Low- SiO_2 group, Alutaguse overlaps the SS fields with
858 broad variability, but the lower average, while Uusimaa yields the highest values
859 and Häme, Saimaa, and Ladoga cluster near crustal levels. Additionally, the
860 negative Sr anomalies recorded in both Alutaguse (Figs.3g,3i) and SS metased-
861 iments (Figs.3h,3j) are characteristic of old, recycled sedimentary environments
862 (Mader and Neubauer, 2004).

863 In general, High- SiO_2 Alutaguse metasediments show mature, well-sorted
864 signatures with strong heavy-mineral enrichment, high SiO_2/Al_2O_3 , elevated
865 $(Gd/Yb)_n$, and broad Th/Sc – Zr/Sc variability, while Uusimaa is similar but
866 coarser, and Häme–Saimaa are more uniform and pelitic. Ladoga combines
867 good sorting with mature geochemistry, reflecting mixing. Low- SiO_2 samples
868 are less sorted but chemically more immature, with high K_2O/Na_2O , narrow
869 Al_2O_3/TiO_2 , and broader Th/Sc – Zr/Sc and $(Gd/Yb)_n$ ranges in Alutaguse,
870 while Uusimaa shows heavy-mineral input and Häme–Saimaa–Ladoga remain
871 consistently pelitic.

872 5.2.3. Tectonic affinities

873 Sediment geochemistry is widely used to infer tectonic settings (Bhatia, 1983;
874 Bhatia and Crook, 1986), although transport, sorting, and mafic input can
875 obscure provenance signals (McLennan et al., 1990; Floyd et al., 1989). Dis-
876 crimination plots based on major, trace, and *REE* data—originally developed
877 for Phanerozoic sediments— have been also applied to Precambrian sequences
878 (Roser and Korsch, 1986; Sifeta et al., 2005; El-Bialy, 2013; Han et al., 2019),
879 but should be interpreted with caution due to potential geochemical overprinting
880 (Floyd et al., 1989).

881 Major oxides such as TiO_2 , Al_2O_3 , Fe_2O_3 , MgO , and the Al_2O_3/SiO_2 ratio
882 vary with tectonic setting, distinguishing oceanic arcs from continental arcs
883 and margins (Bhatia, 1983). High concentrations of these oxides in Alutaguse
884 and SS samples suggest non-passive margin origins. Discrimination based on
885 K_2O/Na_2O and SiO_2 separates passive and active continental margins from
886 oceanic arcs (Roser and Korsch, 1986). Most samples from Alutaguse (Fig.7a),
887 SS (Fig.7b), and reference datasets (UCC, PAAS) plot in the active margin
888 field, with some overlap into the passive field. However, due to the mobility of
889 *Na* and *K*, such classifications must be interpreted cautiously (El-Bialy, 2013).

890 This study applies discriminant-function-based tectonic diagrams, which im-
891 prove upon traditional approaches by distinguishing island/continental arcs,
892 continental rifts, and collision settings for high- ($SiO_2 > 63wt.\%$) and low-silica
893 ($SiO_2 \leq 63wt.\%$) metasediments (Verma and Armstrong-Altrin, 2013; Chen
894 et al., 2014). High-silica samples from both Alutaguse and SS align mainly with
895 arc-related settings, with some inputs into the rift domain (Figs.7c,7d), whereas
896 low-silica samples plot within the rift and collision domains (Figs.7e,7f).

897 Trace elements such as *Th*, *Sc*, *La*, and *Zr* are stable in depositional settings
898 and effective for identifying source rocks (Bhatia and Crook, 1986; McLennan
899 et al., 1993; Roser and Korsch, 1986). Expanding on Bhatia (1983), Bhatia
900 and Crook (1986) used these immobile elements to discriminate four tectonic
901 settings. For Alutaguse and SS metasediments, the most appropriate diagrams

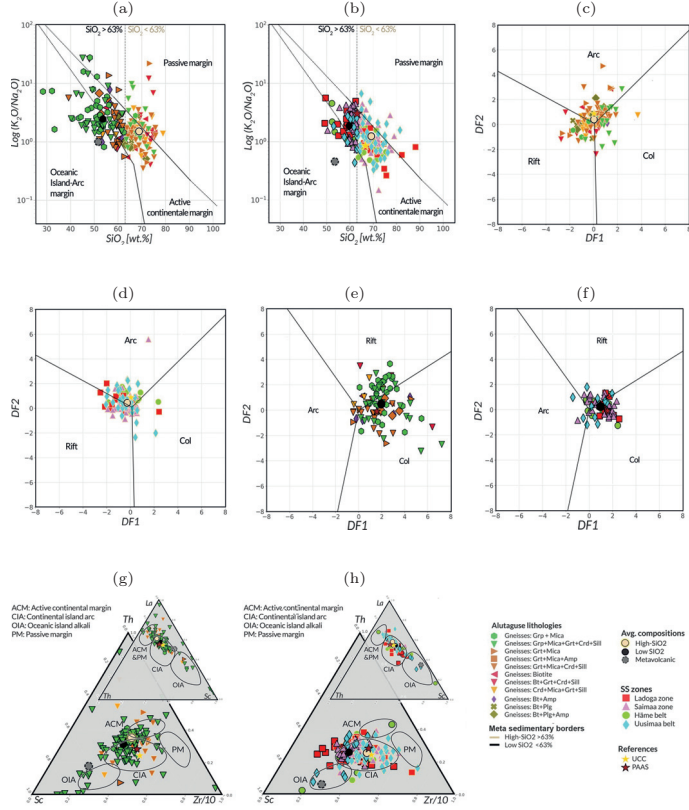


Figure 7: Tectonic discrimination plots for the Alutaguse and South Svecofennian (SS) metasediments are outlined in a series of graphics. Plots of K_2O-Na_2O vs. SiO_2 are shown in (a-b) after Roser and Korsch (1986). Discriminant functions, $DF1$ vs. $DF2$, are detailed for High- SiO_2 metasediments in (c-d) and Low- SiO_2 metasediments in (e-f) after Verma and Armstrong-Altrin (2013). For High- SiO_2 : $DF1 = [-0.263 * \ln(TiO_2/SiO_2)] + [0.604 * \ln(Al_2O_3/SiO_2)] + [-1.725 * \ln(Fe_2O_3/SiO_2)] + [0.660 * \ln(MnO/SiO_2)] + [2.191 * \ln(MgO/SiO_2)] + [0.144 * \ln(CaO/SiO_2)] + [-1.304 * \ln(Na_2O/SiO_2)] + [0.054 * \ln(K_2O/SiO_2)] + [-0.330 * \ln(P_2O_5/SiO_2)] + [1.588]$ & $DF2 = [-1.196 * \ln(TiO_2/SiO_2)] + [1.064 * \ln(Al_2O_3/SiO_2)] + [0.303 * \ln(Fe_2O_3/SiO_2)] + [0.436 * \ln(MnO/SiO_2)] + [0.838 * \ln(MgO/SiO_2)] + [-0.407 * \ln(CaO/SiO_2)] + [1.021 * \ln(Na_2O/SiO_2)] + [-1.706 * \ln(K_2O/SiO_2)] + [-0.126 * \ln(P_2O_5/SiO_2)] + [-1.068]$, and for Low- SiO_2 : $DF1 = [0.608 * \ln(TiO_2/SiO_2)] + [-1.854 * \ln(Al_2O_3/SiO_2)] + [0.299 * \ln(Fe_2O_3/SiO_2)] + [-0.550 * \ln(MnO/SiO_2)] + [0.120 * \ln(MgO/SiO_2)] + [0.194 * \ln(CaO/SiO_2)] + [-1.510 * \ln(Na_2O/SiO_2)] + [1.941 * \ln(K_2O/SiO_2)] + [0.003 * \ln(P_2O_5/SiO_2)] + [-0.294]$ & $DF2 = [-0.554 * \ln(TiO_2/SiO_2)] + [-0.995 * \ln(Al_2O_3/SiO_2)] + [1.765 * \ln(Fe_2O_3/SiO_2)] + [-1.391 * \ln(MnO/SiO_2)] + [-1.034 * \ln(MgO/SiO_2)] + [0.225 * \ln(CaO/SiO_2)] + [0.713 * \ln(Na_2O/SiO_2)] + [0.330 * \ln(K_2O/SiO_2)] + [0.637 * \ln(P_2O_5/SiO_2)] + [-3.631]$; (g-h) Ternary plots of $La-Th-Sc$ (inset) and $Th-Sc-Zr/10$ (back) after Bhatia and Crook (1986).

are the $Th - La - Sc$ and $Sc - Th - Zr/10$ triangular plots (Bhatia and Crook, 1986; El-Bialy, 2013; Han et al., 2019). Both plots place most samples within the continental island arc (CIA) field, consistent with UCC and PAAS references (Figs.7g,7h). Notably, CIA and active continental margins (ACM) share convergent tectonic regimes, marked by subduction, orogenic activity, and continental crustal substrates (El-Bialy, 2013).

The analysed metasediments display key geochemical features—negative Eu anomalies, depleted $Nb-Ta$, $LREE$ enrichment, and limited $HREE$ fractionation in both Alutaguse and SS samples (Fig.3)—suggesting a continental arc source (Sifeta et al., 2005; El-Bialy, 2013; Chen et al., 2014). However, as noted by McLennan et al. (1990), such signatures are not exclusive to particular tectonic settings, as continental crust often shows arc-like geochemistry by default. Overall, the data major and trace elemental data support deposition within a continental arc environment, with possible back-arc rift influence in the Low- SiO_2 metasediments, within a collisional environment.

5.3. Metavolcanic rocks

5.3.1. The effects of shallow-level open-system processes

Shallow processes such as crustal contamination, fractional crystallisation, and post-magmatic alteration can affect mafic magma compositions and must be considered when interpreting source signatures (Saccani et al., 2018; Ma et al., 2021; Gebremicale et al., 2025). $LILEs$ often reflect these shifts, especially in metavolcanic rocks. In Alutaguse, evidence of alteration by seawater or metamorphic fluids is suggested by carbonation and hydrothermal trends (Fig.S1c), though most samples plot near the least-altered domain (Fig.5e). Alutaguse LOI values show a low average, implying limited alteration. In contrast, SS metavolcanics show no LOI data (Rasilainen et al., 2007) and cluster in less-altered fields (Figs.S1d,5f). These patterns suggest a minimal open-system overprint, indicating that the observed geochemistry largely reflects reliable magmatic compositions (Rudnick and Gao, 2003; Karakacs and Güçtekin, 2021; Ma et al., 2021).

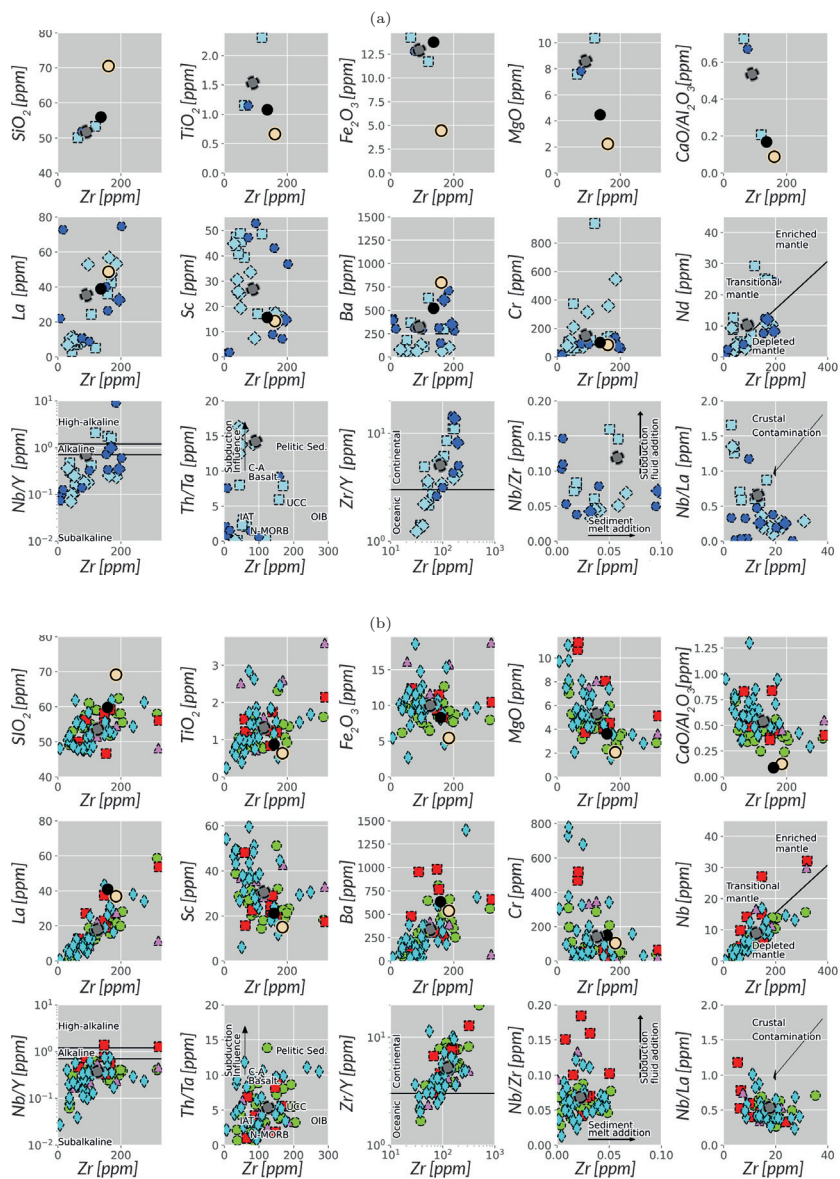


Figure 8: Zr binary plots versus major, trace and *REE* elements for Alutaguse and South Svecofennian (SS) metavolcanic samples. Legend as in Figure 2.

932 5.3.2. Crystallisation and partial melting

933 Alutaguse and SS metavolcanic samples generally display tholeiitic tenden-
 934 cies (Figs.4g,4h), though several SS samples also show calc-alkaline affinities
 935 (Figs.4h,S1d). Sub-alkaline *Nb/Y* ratios are observed in both Alutaguse (Fig.8a)
 936 and SS metavolcanics (Fig.8b). Mafic members of both groups are metalumi-
 937 nous (Figs.9a,9b), consistent with derivation from mafic precursors (Figs.4e,
 938 4f).

939 Major-element trends show increasing *MgO* with *SiO₂* and decreasing *TiO₂*
 940 and *K₂O* (Figs.3e,3f), coupled with low *Ni* and *Cr* contents (Figs.3k, 3l), in-
 941 dicating basaltic fractional crystallisation from a differentiated mantle source
 942 (Ma et al., 2021). Elevated *CaO/Al₂O₃* in Alutaguse, Uusimaa, and Ladoga
 943 (Figs.8a,8b), along with negative correlations between *MgO* and *Al₂O₃*, *CaO*,
 944 and *Na₂O* (Figs.3e, 3f), together with *Sr* enrichment and weak to positive Eu
 945 anomalies (Figs.3q,3r), point to limited plagioclase fractionation, as plagioclase
 946 preferentially incorporates Sr and Eu (Floyd et al., 1989; Faisal et al., 2020).
 947 The inverse *MgO-TiO₂* trend further indicates minimal crystallisation of *Fe-Ti*
 948 oxides (Ma et al., 2021), supporting a fractionation regime dominated by mafic
 949 silicates rather than feldspar or titanite.

950 Experimental constraints show that pelitic and psammitic melts are more
 951 aluminous than amphibolite-derived ones in the *Al-Fe-Mg-Ti-Ca* system (Patiño Douce,
 952 1999). Both Alutaguse and SS metavolcanics plot within the amphibolite-
 953 derived melt field (Figs.9c, 9d), consistent with partial melting of metabasaltic
 954 to metatonalitic sources rather than pelitic or graywacke protoliths, as corrob-
 955 orated by the *Al/(Mg + Fe)* vs. *Ca/(Mg + Fe)* discrimination of Altherr et al.
 956 (2000) (Figs.9c, 9d).

957 Trace-element systematics reinforce these contrasts. In *Nb-Zr* space (Ta-
 958 ble 2), Alutaguse samples span the widest field (Table 2), ranging from strongly
 959 depleted (*Nb* < 10, *Zr* < 200 ppm; (Pearce, 1996)) to enriched signatures,

960 indicating heterogeneous mantle inputs. Uusimaa and Häme plot mostly in
 961 intermediate fields, with Uusimaa closer to the depleted threshold and Häme
 962 extending to the highest Zr values (> 500 ppm), reflecting zircon-rich recycling
 963 or felsic contamination. Saimaa and Ladoga consistently yield higher Nb and
 964 Zr , plotting well beyond the depleted domain and forming the most enriched
 965 SS signatures.

966 This pattern is consistent with the $\delta Nb = [\log(Nb/Y) + 1.74 - (1.92 \times$
 967 $\log(Zr/Y))]$ relation of [Fitton et al. \(1997\)](#), where $\delta Nb > 0$ indicates enriched
 968 mantle and $\delta Nb < 0$ a depleted mantle. Alutaguse shows the widest spread,
 969 with both positive and negative values, again pointing to heterogeneous man-
 970 tle contributions (Table 2). By contrast, the SS domains are more restricted
 971 and dominantly depleted: Häme and Uusimaa yield the most negative values,
 972 Ladoga remains intermediate, and Saimaa trends toward slightly enriched sig-
 973 natures.

974 Similarly, La/Yb vs. Zr/Nb (Figs. 9e, 9f) and La/Sm vs. Sm/Yb (Figs. S1i–j)
 975 show that most Alutaguse and SS metavolcanics follow garnet-lherzolite trajec-
 976 tories ([Yang et al., 2014](#)). Alutaguse and Uusimaa cluster close to depleted
 977 fields with low La/Yb and La/Sm , whereas Ladoga and Saimaa show enriched,
 978 subduction-modified mantle signatures.

979 5.3.3. Nature of magma sources

980 Elemental composition in mafic magmas largely results from fractional crys-
 981 tallisation and rock assimilation, while trace element composition, especially for
 982 incompatible elements, is more influenced by the mantle source’s composition
 983 and degree of melting than by superficial crustal processes. Consequently, the
 984 trace element profiles of different magmas primarily reflect distinct source char-
 985 acteristics tied to specific tectono-magmatic environments ([Sifeta et al. \(2005\)](#);
 986 [Faisal et al. \(2020\)](#); [Ma et al. \(2021\)](#)). However, sample limitations blur the
 987 delineation of fractional crystallisation’s exact impact on these variations.

988 *HFSEs* and *REEs* remain stable during alteration and weathering and are
 989 instrumental in identifying magma sources and gauging partial melting extents

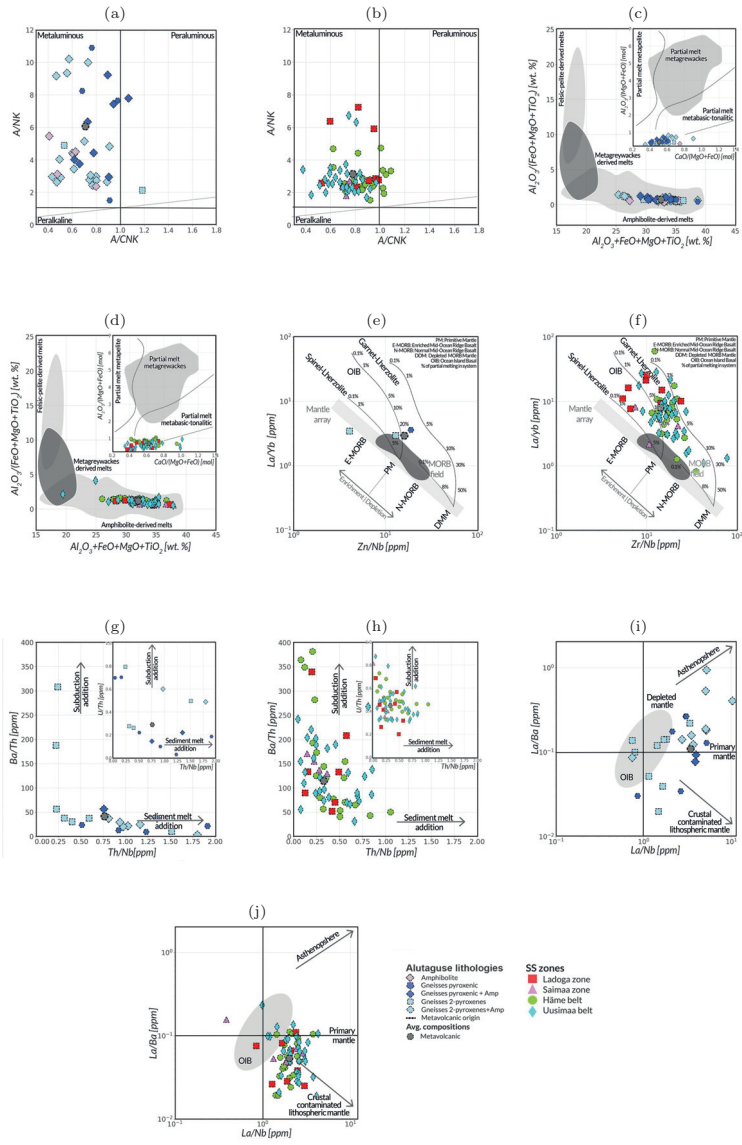


Figure 9: Metavolcanic source nature of Alutaguse and South Svecofennian (SS) zones: (a-b) A/CNK [molar ratio $Al_2O_3/(CaO + Na_2O + K_2O)$] vs. A/NK [molar ratio $Al_2O_3/(Na_2O + K_2O)$] diagram (modified from Faisal et al. (2020)); (c-d) $Al_2O_3/(FeO + MgO + TiO_2)$ vs. $(Al_2O_3 + FeO + MgO + TiO_2)$ after Patiño Douce (1999), with an insight of Altherr et al. (2000) discrimination $Al/(Mg + Fe)$ vs. $Ca/(Mg/Fe)_{[mol]}$; (e-f) La/Yb vs. Zr/Nb adapted from Yang et al. (2014); (g-h) Ba/Th and U/Th ratios vs. Th/Nb , after Saccani et al. (2018) and Ye et al. (2018) respectively; (i-j) La/Ba vs. La/Nb , after Hart (1988).

990 (Saccani et al. (2018); Wan et al. (2019); Han et al. (2019); Ma et al. (2021)).
 991 Certain trace element ratios, like Nb/Zr , Th/Zr , and others, remain relatively
 992 unchanged during mafic magma crystallisation, predominantly influenced by
 993 minerals like olivine, clinopyroxene, and plagioclase. Thus, in moderate frac-
 994 tionation, these ratios likely reflect the source’s inherent compositions (Saccani
 995 et al. (2018); Wan et al. (2019); Ma et al. (2021)).

996 SS samples exhibit elevated Nb/Zr (Fig.8b), Ba/Th , and U/Th (Fig.9d)
 997 ratios, indicating a dominant influence from subducted oceanic crust-derived
 998 solutions. In contrast, Alutaguse samples show increased Th/Nb (Fig.9c) and
 999 Th/Zr (Fig.8a) values. In the La/Ba vs. La/Nb plot, Alutaguse metavolcanics
 1000 lean towards an asthenospheric mantle origin (Fig.9g), while SS samples suggest
 1001 a crust-contaminated lithospheric mantle source (Fig.9h).

1002 5.3.4. Tectonic setting implications

1003 Averages from the Alutaguse and SS metavolcanics were juxtaposed with
 1004 metasedimentary samples, revealing oceanic tectonic mafic arc affinities (Figs.6,7).
 1005 The tectonic origins of the SS and Alutaguse metavolcanics have been a topic
 1006 of contention, with theories spanning from island arc collisions to rift mech-
 1007 anisms (Lahtinen (2000); Kirs et al. (2009); Bogdanova et al. (2015); Soesoo
 1008 et al. (2020)). Notably, the presence of *HFSE* and *HREE*, known for their
 1009 stable characteristics, is instrumental in determining tectonic settings of extru-
 1010 sive rocks (Pearce (1996); Sifeta et al. (2005); Saccani et al. (2018)). While the
 1011 conventional product of asthenospheric mantle melting is MORB, characterised
 1012 by lower *LREE* and *LILE*, the studied metavolcanics exhibit a contrasting
 1013 arc-like signature. This, particularly the notable *HFSE* depletion, points to-
 1014 wards the influence of subduction or associated phenomena (Faisal et al. (2020);

1015 [Ma et al. \(2021\)](#)). Delving deeper into *REE* analysis, the Alutaguse chondrite-
 1016 normalised metavolcanics (Fig.3q) showcase trends spanning from IAB (Island-
 1017 Arc Basalt) to E-MORB. In contrast, the SS samples depict broader *REE*
 1018 spectrums (Fig.3p), capturing values from IAB to OIB (Oceanic Island Basalt),
 1019 though their averages lean more towards the E-MORB trend. Different plots
 1020 further elucidate the tectonic orientations of these (meta)volcanic units ([Pearce](#)
 1021 [\(1996\)](#); [Kotova and Podkovyrov \(2014\)](#); [Saccani \(2015\)](#); [Wan et al. \(2019\)](#)).

1022 $TiO_2 - 10(MnO) - 10(P_2O_5)$ triangular plots ([Mullen \(1983\)](#)) indicate that
 1023 Alutaguse (Fig.10a) and SS (Fig.10b) metavolcanic samples mostly fall within
 1024 the IAT (island- arc tholeiite) domain. However, several SS samples intersect
 1025 with the OIA (Oceanic island alkali) region, notoriously from the Ladoga group.

1026 On the $Y/15 - La/10 - Nb/8$ plot ([Cabanis and Lecolle \(1989\)](#)), both Aluta-
 1027 guse (Fig.10a) and SS (Fig.10b) metavolcanic samples predominantly plot in the
 1028 calc-alkaline basalts zone, indicative of a compressional arc setting, notably, the
 1029 SS Ladoga samples edge towards the "Late to Post-Orogenic intra-continent"
 1030 zone, signalling a transition from a tectonic compressive to a distensive context.

1031 In the $Hf/3 - Th - Nb/16$ triangular diagram, most Alutaguse (Figs.10e),
 1032 and SS (Fig.10f) metavolcanic samples align with the calc-alkaline volcanic arc-
 1033 basalt (VAB) zone ([Wood \(1980\)](#)), pointing to subduction or crustal-magma
 1034 interactions. Specifically, Alutaguse samples indicate crustal magma interac-
 1035 tion, while SS samples trend towards subduction zone enrichment. Both exhibit
 1036 significant crustal contamination, as further supported by the Nb/La vs. Zr/Nb
 1037 plots (Figs. 8a,8b).

1038 On the $Log(Nb/Th)$ vs. $Log(Y/La)$ plot (Figs. 10g,10h) based on [Liu et al.](#)
 1039 [\(2018\)](#), both Alutaguse and SS metavolcanic samples largely align with the
 1040 island arc basalt (IAB) zone.

1041 Nb/Th vs. Zr/Nb diagram determines the mantle source (Figs.10i,10j) as
 1042 outlined by [Condie \(2005\)](#), both Alutaguse and SS metavolcanic samples pre-
 1043 dominantly fall within the arc domain, especially near the enriched component
 1044 (EN). This hints at influences from the Upper Continental crust and subcon-
 1045 tinental lithosphere, emphasizing a subduction component. The same is sup-



Figure 10: Metavolcanic tectonic affinities of the Alutaguse and South Svecofennian (SS) zones: (a-b) $TiO_2 - 10(MnO) - 10(P_2O_5)$ triangular plots after Mullen (1983); (c-d) $Y/15 - La/10 - Nb/8$ triangular plots after Cabanis and Lecolle (1989); (e-f) $Hf/3 - Th - Nb/16$ triangular plots after Wood (1980); (g-h) $Log(Nb/Th)$ vs. $Log(Y/La)$ plot after Liu et al. (2018); (i-j) Nb/Th vs. Zr/Nb plot after Condie (2005).

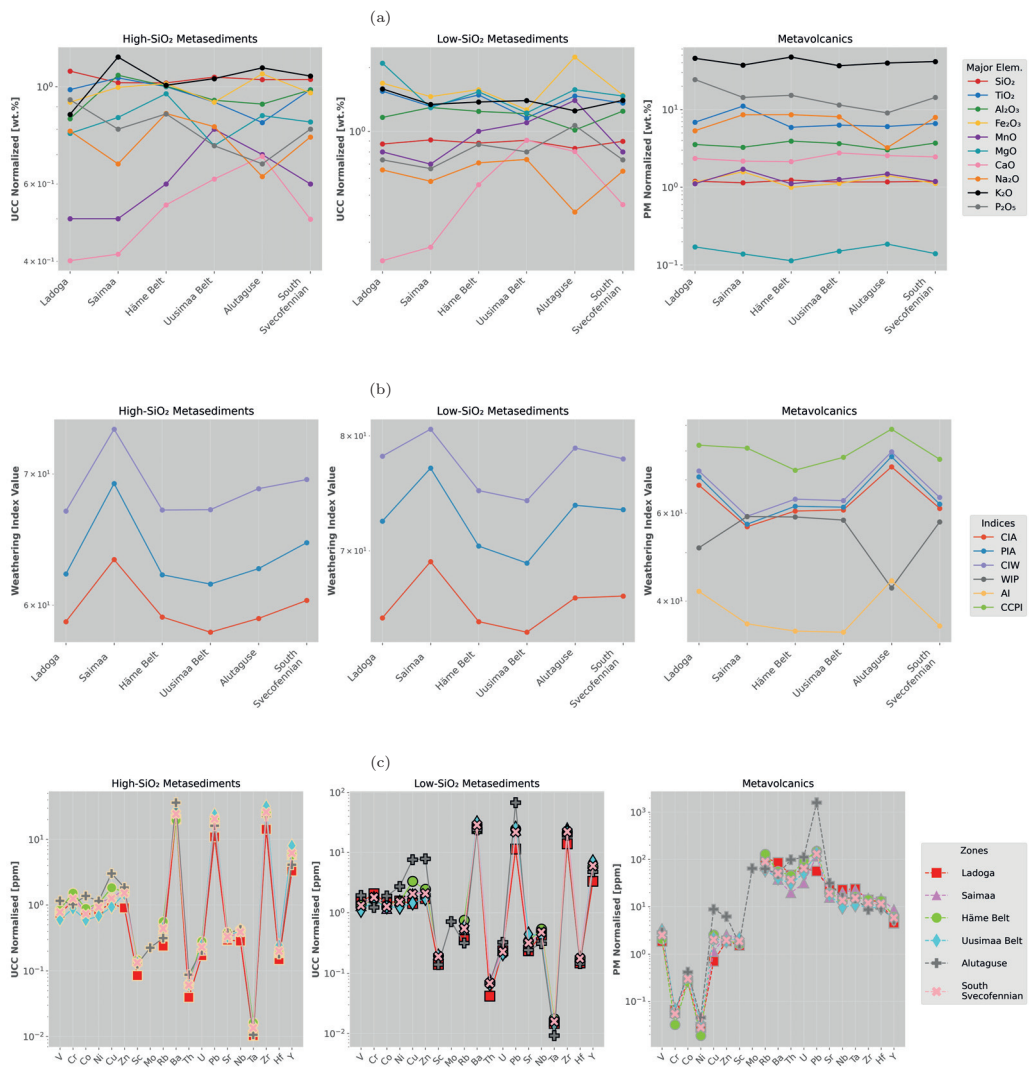
ported by the Th/Ta vs. Zr plots (Figs.8a,8b), with both samples distinctly away from the OIB zone.

The Zr/Y values, as proposed by Pearce and Norry (1979), differentiate between continental and oceanic basalts with a threshold value of 3. When charted on Zr/Y vs. Zr plots (Figs.8a,8b), both Alutaguse and SS samples predominantly occupy the continental domain. However, Alutaguse samples exhibit a broader Zr/Y range, notably extending into the oceanic domain, especially seen in the 2-Pyroxene gneisses samples with high Y values (Table 2). In contrast, the Ladoga SS samples consistently position within the continental domain.

Saccani (2015) introduces two discrimination plots based on the $Th-Nb_{N-MORB}$ relation. The first illustrates compositional variations (Figs.S1 k-l), while the second conveys tectonic interpretations (Figs.S1 m-n). Both Alutaguse (Fig.S1k) and SS (Fig.S1l) display back-arc basin basalt (BABB) affinities, aligning with the assimilation-fractional crystallisation (AFC) trend. Similarly, the Alutaguse (Fig.S1m) and SS (Fig.S1n) fall within the back-arc basin basalt field, spanning the "oceanic subduction setting domain + rifted margin" domain.

5.4. Geodynamic implications

A triad of models informs the evolution of the SO domain. The first envisions 1.90–1.88 Ga accretion of arc complexes to the Karelian craton, marking microcontinent–continent amalgamation (Lahtinen et al., 2005, 2009; Kukkonen and Lauri, 2009; Bogdanova et al., 2015; Nironen, 2017; Mikkola et al., 2018; Kara et al., 2021; Solano-Acosta et al., 2025a). The second, based on tectonic switching theory (Collins, 2002), invokes successive subduction and arc formation (Kukkonen and Lauri, 2009; Kara, 2021; Kara et al., 2021). Both imply formation of 1.92–1.87 Ga arc rocks from the Skellefte district (Sweden) to the Finnish Svecofennian belts (Fig.1b). The third model emphasises pre–1.91 Ga



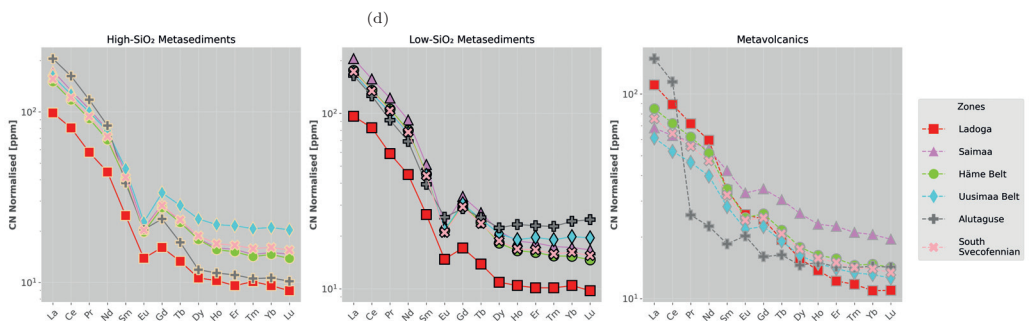


Figure 11: Bivariate average geochemical trends of the Alutaguse and South Svecofennian (SS) zones: (a) Major elemental data and (b) Weathering indices, both as shown in Table 1, (c) Trace elements, and (d) Rare Earth Elements (*REE*) are detailed in Table 2. Alutaguse *REE* metavolcanic *Pr* to *Lu* data corresponds to three samples (see supplementary dataset).

1072 events, interpreting the SO as a single evolving arc with continuous 1.90–1.88
 1073 Ga magmatism forming the CFAC and SS, culminating in oroclinal buckling at
 1074 1.87–1.86 Ga (Lahtinen et al., 2014, 2022). Amidst these models, two distinct
 1075 tales emerge. One perceives multiple arcs stemming from double subduction-
 1076 rifting events, while the other perceives the 1.90-1.88 Ga volcanic rocks as rem-
 1077 nants of a singular, later deformed, linear arc. The distinct geochemical con-
 1078 trasts between the SS and Alutaguse metasedimentary and metavolcanic units
 1079 (Tables 1 & 2; Fig.11), favour the first tale, distinguishing these belts as dis-
 1080 tinctive units.

1081 Early Paleoproterozoic inheritance set the stage. Intense plume activity at
 1082 2.74–2.69 Ga and protracted erosion of the Karelian craton (2.5–2.06 Ga) fed
 1083 rift-basin sedimentation prior to inversion and assembly of the SE Karelia do-
 1084 main (Weihed et al., 2005; Kukkonen and Lauri, 2009; Hanski, 2015; Baltybaev,
 1085 2013; Kotova and Podkovyrov, 2014; Lahtinen et al., 2022). Across the RLSZ,
 1086 ϵNd shifts from negative (Karelian) to positive (Svecofennian) values, reflecting
 1087 contrasted crustal reservoirs (Baltybaev, 2013). Around 2.1-2.0 Ga, the Keitele
 1088 microcontinent surfaced, as evidenced by detrital zircons ranging from 2.1 to
 1089 1.95 Ga and seismic data (Hanski, 2015; Nironen, 2017).

1090 At 2.05–1.92 Ga the Karelian margin in eastern Finland transitioned from
 1091 extension to compression, forming the Jormua–Outokumpu (J–O) ophiolites and
 1092 initiating arc–continent interactions (Fig.12a; [Nironen, 2017](#)). Collision of the
 1093 SW-derived Keitele microcontinent with Karelia triggered melting of arc-affinity
 1094 mafic rocks by mantle-derived magmas in the Bothnian sector, producing first
 1095 calc-alkaline then alkaline granitoids. Coevally, SW-derived sediments mixed
 1096 with passive-margin and foredeep deposits to form the allochthonous Pudasjärvi
 1097 and Iisalmi blocks (PB–LB), while the Northern Ostrobothnia nappe system
 1098 (ONS) was emplaced onto PB during early SO development (Fig.12a; [Nironen,](#)
 1099 [2017](#); [Mikkola et al., 2018](#)). These steps record breakup of the Karelian craton
 1100 and opening of an ocean basin west of Karelia by ≤ 2.0 Ga ([Weiheid et al., 2005](#);
 1101 [Hanski, 2015](#)).

1102 Convergence between the Karelian and Norrbotten blocks at ~ 1.92 Ga pro-
 1103 duced a V-shaped basin with remnants of oceanic crust and SW-directed sub-
 1104 duction ([Lahtinen et al., 2005](#); [Baltybaev, 2013](#); [Hanski, 2015](#); [Nironen, 2017](#)).
 1105 Slab break-off and polarity changes near the Knaften arc preceded deposition
 1106 of an oceanic metabasalt–metachert–metaturbidite package at 1.92–1.91 Ga
 1107 (Fig.12a1), extending toward the SE Bothnian microcontinent and matching a
 1108 regional conductance anomaly ([Korja et al., 2002, 2006](#)). Arc magmatism waned
 1109 along the Karelian margin by ~ 1.91 Ga (SB), even as SW-side activity persisted
 1110 in Tampere–Pirkanmaa–Uusimaa (Fig.12b1). In NW Ladoga, detrital zircon
 1111 spectra (dominant 2.0–1.9 Ga; youngest 1.91 ± 14 Ga) indicate mainly Protero-
 1112 zoic sources (Sortavala felsic volcanics; Lakhdenpokh’ya granitoids) with minor
 1113 Archaean input ([Kotova et al., 2009](#); [Myskova et al., 2012](#); [Kotova and Pod-](#)
 1114 [kovyrov, 2014](#); [Baltybaev, 2013](#)); the basal Kalevian andesites (1.92–1.91 Ga)
 1115 and tholeiitic tuffs tie Ladoga to arc processes also seen along the southern SB
 1116 (Fig.12a). In CFAC, the submarine Haveri Formation (1.92–1.90 Ga) reflects
 1117 rifting and E-MORB affinities in Tampere, transitioning to arc volcanism by
 1118 ~ 1.90 Ga and forearc–accretionary sedimentation in Pirkanmaa (1.92–1.89 Ga)
 1119 ([Kähkönen, 2005](#); [Kara, 2021](#); [Kara et al., 2021](#)) (Fig.12b1). Saimaa “Older
 1120 Svecofennian” supracrustals (≥ 1.91 Ga) show low Archaean contributions and

1121 juvenile peaks; younger units (post-1.89 Ga) record mixed sources (Lahtinen
1122 et al., 2022; Myskova et al., 2012).

1123 Uusimaa departs from the standard SS switching sequence by showing earlier
1124 magmatism (2.1–1.91 Ga) and juvenile signatures (Kähkönen, 2005; Weihed
1125 et al., 2005; Kukkonen and Lauri, 2009; Bogdanova et al., 2015; Kara, 2021).
1126 Together with Tallinn (North Estonia), 1.92–1.91 Ga felsic metavolcanics and
1127 *Fe–S–Si* anomalies, plus the Jöhvi magnetite quartzite (2.2–1.89 Ga deposition;
1128 1.91–1.90 Ga mineralisation), link Uusimaa–Tallinn to the northern Bergslagen
1129 margin (Kirs and Petersell, 1994; All et al., 2004; Kirs et al., 2009; Kukkonen
1130 and Lauri, 2009; Bogdanova et al., 2015; Soesoo et al., 2020; Nirgi and Soesoo,
1131 2021; Nirgi et al., 2024). Detrital zircon clusters in West Estonia at 1.97–1.90 Ga
1132 point to proximal Bergslagen–Svecofennian sources (Bogdanova et al., 2015). A
1133 double-subduction geometry with a shorter slab beneath Bergslagen plausibly
1134 explains the earlier onset of arc magmatism in Uusimaa–North Estonia, while
1135 arcs in Tampere evolved coevally by 1.92–1.90 Ga (Fig.12b1; Kukkonen and
1136 Lauri, 2009; Kara, 2021; Kara et al., 2021; Bogdanova et al., 2015).

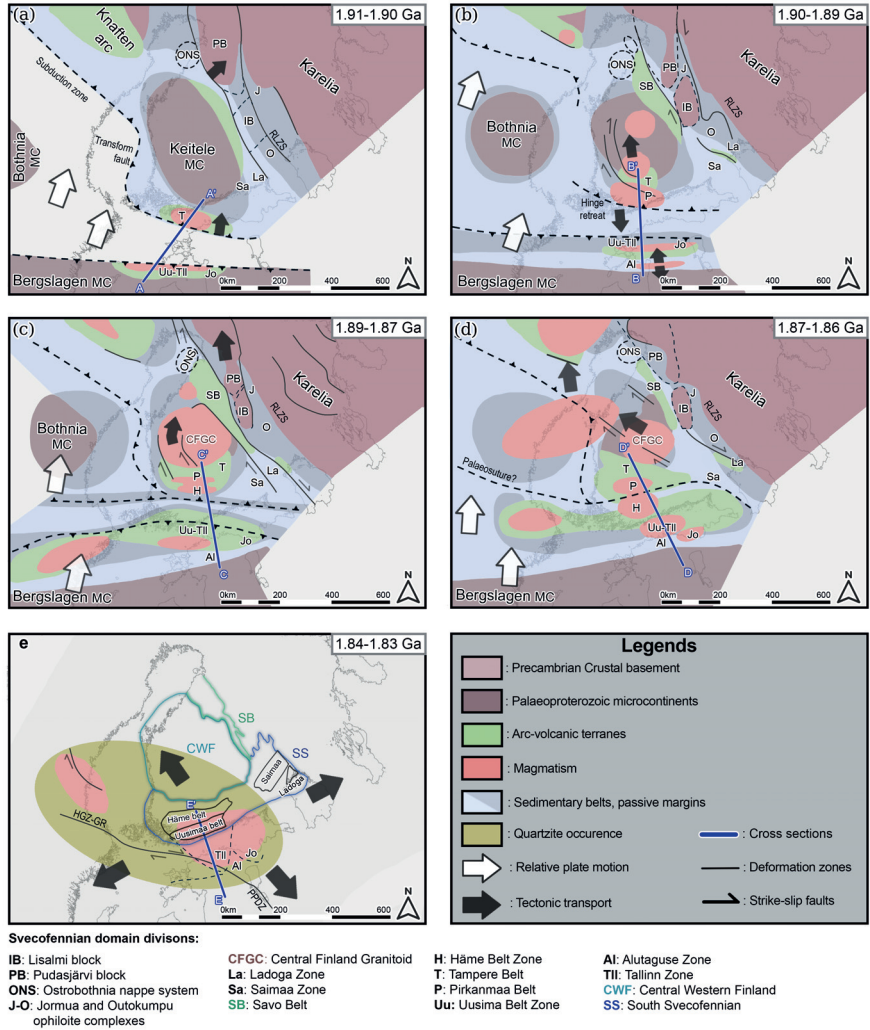
1137 Geochemically, High-*SiO*₂ metasediments show a consistent Alutaguse–Uusimaa
1138 affinity: both are relatively *CaO*-rich with low *TiO*₂ and slightly elevated
1139 *MnO*, contrasting with the more aluminous and *Ca*-poor Saimaa, the mod-
1140 erately aluminous Häme, and the highly quartzose Ladoga, which has the high-
1141 est *SiO*₂ and lowest *Al*₂*O*₃. Low-*SiO*₂ metasediments maintain this separa-
1142 tion but express it differently: Alutaguse is strongly *Fe*₂*O*₃–*TiO*₂–*SO*₃-rich
1143 and markedly *Na*₂*O*-poor, while Uusimaa again falls on a less aluminous and
1144 more *CaO*-rich trajectory than Saimaa and Häme, placing both domains to-
1145 ward the mafic–intermediate side of the SS spectrum. Saimaa remains the most
1146 aluminous and *Ca*-poor, Häme shows elevated *Al*₂*O*₃ and *K*₂*O*, and Ladoga
1147 trends toward higher *K*₂*O* and *MgO*. Metavolcanics strengthen the Aluta-
1148 guse–Uusimaa coupling through shared *CaO–MgO* enrichment and similar in-
1149 termediate *SiO*₂ ranges, whereas Häme represents the most aluminous and felsic
1150 pole, Ladoga a moderately mafic type with high *MgO* and *CaO*, and Saimaa the
1151 most *TiO*₂- and *Fe*₂*O*₃-rich mafic end-member within the SS domains. (Table 1;

Fig.11a). Weathering indices corroborate these contrasts (Saimaa most mature; Alutaguse wide ranges yet on average immature; Uusimaa–Häme least altered; Figs.5a,5b,11b). Systematic $CaO - MnO$ enrichment in metasediments and $CaO - MgO$ in metavolcanics ties Uusimaa–Alutaguse to the Bergslagen margin and distinguishes them from the other SS belts across a palaeo–Svecofennian ocean (Bogdanova et al., 2015; Solano-Acosta et al., 2025a).

Arc-type magmatism peaked across the system, while Bergslagen experienced back-arc rifting and granitoid magmatism, resulting in polymetallic deposits (e.g. VMS) (Allen et al., 1996; Beunk and Kuipers, 2012; Hanski, 2015; Stephens and Weihed, 2020; Bogdanova et al., 2015). In Uusimaa, 1.90–1.88 Ga volcanosedimentary successions include calcareous layers that mirror CaO -enriched metasediments in Uusimaa–Alutaguse (Fig.11a; Kähkönen, 2005; Kara et al., 2021). Alutaguse metavolcanics show a stronger asthenospheric imprint than SS equivalents (Figs.9i,9j), consistent with a back-arc established by ~1.89 Ga and followed by rift-related sedimentation (Fig.12b2; Kirs et al., 2009; Bogdanova et al., 2015; Solano-Acosta et al., 2025a). Tectonic discriminants confirm this pattern: High- SiO_2 metasediments are dominantly arc-derived with minor rift input, Low- SiO_2 subsets record stronger rift–collision affinity in both provinces (Figs.7c-f), and metavolcanics retain arc signatures but reflect mantle-enrichment sources in Alutaguse versus subduction-modified lithosphere in SS (Fig.10). The magnetite-rich pulse at ~1.89 Ga in the Jöhvi units likely marks the same upwelling (Bogdanova et al., 2015; Nirgi and Soesoo, 2021). Trace-element patterns show $Pb-Cu-Zn$ enrichment in Alutaguse relative to SS (Fig.11c), and Low- SiO_2 Alutaguse samples display enhanced $HREE$ (Fig.11d).

Collision of Bothnia with Keitele, pushed over Karelia thickening lithosphere, induced delamination, and drove asthenospheric upwelling, promoting decompression melting and renewed arc magmatism in Häme and Pirkanmaa (Nironen, 2017; Fig.12b3). The Saimaa domain records a metamorphic peak at 1.89–1.88 Ga and the onset of D-orocline shortening (NNW–SSE), followed by a ~1.87 Ga shift in compression that buckled the arc stack (Lahtinen et al., 2022). During this stage, the Central Finland Granitoid Complex (CFGC) de-

(a)



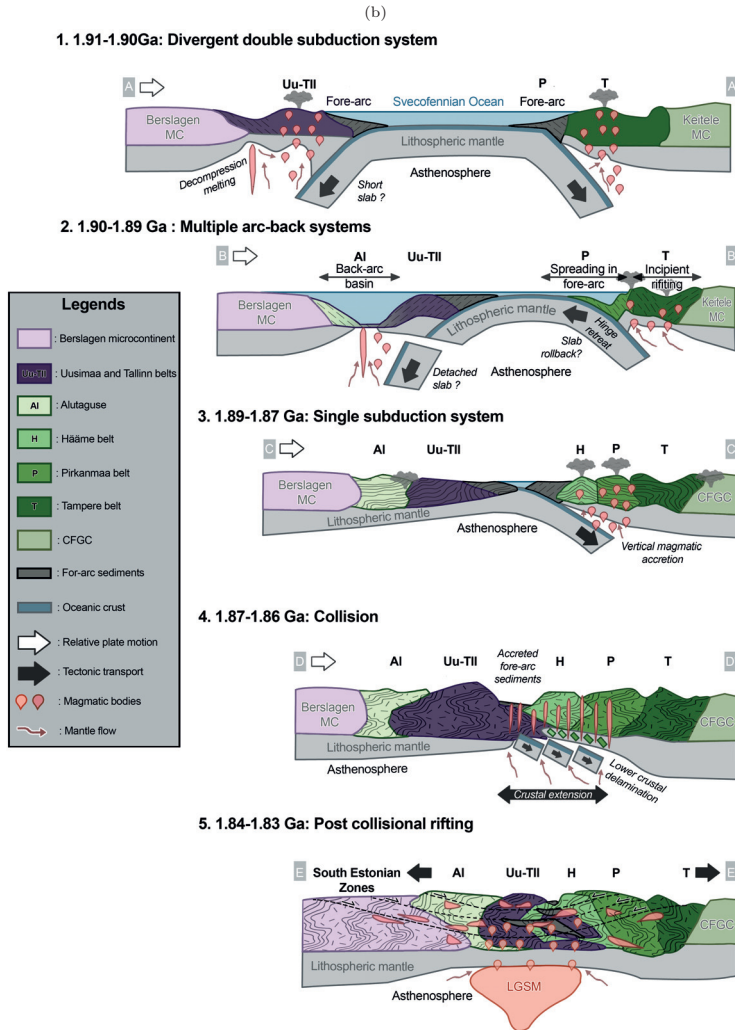


Figure 12: Schematic Geodynamic Model in a) XY (Modified from [Nironen \(2017\)](#), after [Kotova and Podkovyrov \(2014\)](#)) and b) YZ (After [Kukkonen and Lauri \(2009\)](#); [Kara et al. \(2021\)](#)) sections of the evolution of Fennoscandia from: 1) 1.92-1.90 Ga; 2) 1.90-1.89 Ga; 3) 1.89-1.87 Ga; 4) 1.87-1.86 Ga and 5) 1.84-1.83 Ga. The model is a not-to-scale conceptual representation designed to illustrate key dynamics rather than depict precise physical dimensions. XY cross-sections are indicated in Figure 12a with the blue line respectively.

veloped under transtension with post-kinematic granitoids, uplift, and erosion
(Mikkola et al., 2018; Nironen, 2017). Bimodal arc magmatism at ~1.90 Ga in
Tampere and younger rift-related magmas in Häme–Pirkanmaa are compatible
with switching and hinge retreat–advance cycles (Kähkönen, 2005; Kara, 2021;
Kara et al., 2021).

1.87–1.86 Ga collision of the Bergslagen microcontinent with the CFAC
closed the remaining oceanic tracts (Kukkonen and Lauri, 2009; Nironen, 2017;
Kara et al., 2021) (Fig.12b4). This collision could be the one reported in the
Low- SiO_2 samples (Figs.7e,7f). Almost immediately, a short-lived extensional
pulse is recorded by within-plate gabbroic dikes, lateritic paleosols, and high-
 Nb /adakite-like magmatism interpreted as slab detachment–delamination sig-
natures (Kara et al., 2021; Kara, 2021) (Fig.12b4).

A major tectonic shift occurred at 1.86 to 1.84 Ga when Volgo-Sarmatia
collided with southern Fennoscandia, presumably the southern Bergslagen mi-
crocontinent, causing inversion and imbrication of the pre-existing extensional
basin (Lahtinen et al., 2005; Kähkönen, 2005; Kukkonen and Lauri, 2009; Bog-
danova et al., 2015) (Fig.12b4). This event triggered widespread low-pressure
and high-temperature metamorphism at nearly 5 kbar and 750°C from 1.85 to
1.79 Ga, producing extensive migmatization and granite emplacement concen-
trated in the late Svecofennian granite and migmatite (LSGM) zone of southwest
Finland, a belt nearly 100 km wide and 500 km long that transects the southern
Svecofennides (Hölttä and Heilimo, 2017; Nironen, 2017; Mikkola et al., 2018).
Formed on thin volcanic, volcanoclastic, and carbonate units, the LSGM reflects
reactivation of an earlier extensional basin along inherited crustal weaknesses:
early recumbent folding was followed by crustal thickening and emplacement of
microcline granite sheets at about 1.84 to 1.83 Ga, and the zone is bounded
by major ductile shear systems (Ehlers et al., 1993; Hölttä and Heilimo, 2017;
Mikkola et al., 2018; Nironen, 2017) (Fig.12b5). Crustal stretching resumed
near 1.83 Ga with re-sedimentation of quartz sands in intracontinental rifts
(Lahtinen et al., 2010; Nironen, 2017), coincident with development of the Os-
karshamn Jönköping arc along the southern Fennoscandian margin, possibly

1214 extending into present-day Lithuania (Skridlaite and Motuza, 2001; Bogdanova
1215 et al., 2015).

1216 In sum, the study zone geochemistry delineate a Bergslagen-facing back-
1217 arc/rift in Alutaguse-(Tallinn)–Uusimaa system by ~ 1.90 – 1.89 Ga, progressive
1218 collision and oroclinal tightening at 1.89 – 1.86 Ga, brief post-collisional exten-
1219 sion at ~ 1.86 Ga, and a later LP–HT overprint (1.85 – 1.79 Ga). The spatial
1220 coherence of Ca – Mn – Mg systematics in metasediments and metavolcanics (Ta-
1221 ble 1; Figs. 11a, 11d, 10) underpins these linkages and explains the contrasting
1222 Karelian- versus Bergslagen-facing signals observed across Ladoga–Saimaa and
1223 Häme with Alutaguse–Uusimaa (Fig. 1a,b; Fig. 12a,b).

1224 6. Conclusions and recommendations

1225 Integrated major–trace– REE trends show High- SiO_2 metasediments cluster
1226 near UCC, whereas Low- SiO_2 approach PAAS. In the High- SiO_2 set, Alutaguse
1227 (with Uusimaa) is Ca -rich with episodic Mn peaks, relatively Fe – Mg – S rich
1228 and Al – Na poor; Saimaa is Al – K rich and Ca -poor; Ladoga is the most quart-
1229 zose. In the Low- SiO_2 set, Alutaguse is the least siliceous and most variable,
1230 remains Fe – Mg – S rich and Na -poor and, together with Uusimaa, is Ca -rich,
1231 whereas Saimaa is Ca -poor and Ladoga lower in Ca but more Mg rich. Weath-
1232 ering indices rank Saimaa most mature, Alutaguse spans the widest ranges yet
1233 is on average immature, and Häme–Uusimaa are least altered. Trace– REE pat-
1234 terns are consistent: in both silica groups Alutaguse is metal-rich with strong
1235 $LREE$ dominance and subdued $HREE$; SS belts are more $HREE$ -balanced
1236 with stronger negative Eu^* ; Ladoga shows $HREE$ depletion; Saimaa carries
1237 elevated $HREE$. Immobile-element discriminants place High- SiO_2 metasedi-
1238 ments in continental island-arc/active-margin fields, with added rift input in
1239 Low- SiO_2 subsets, strongest in Alutaguse.

1240 Metavolcanics are subalkaline, dominantly tholeiitic and metaluminous; Alu-
1241 taguse is Ca – Mg – Fe rich and Na – P poor, closest to Uusimaa but more de-
1242 pleted; within SS, Uusimaa is Ca – Na high and K low, Häme is Si – Al rich and

1243 *Mg* poor, Saimaa is *Fe–Ti–Mn* rich with high *Na*, and Ladoga is *P* rich and *Na*
1244 poor. Alteration is limited overall, so primary magmatic signals are preserved,
1245 yet still moderate in Alutaguse. Fractionation and Trace-*REE* ratios indicate
1246 partial melting of garnet-lherzolite trend in depleted to transitional mantle,
1247 with Alutaguse–Uusimaa toward the depleted end and Ladoga–Saimaa more
1248 enriched. Trace-element systematics point to asthenospheric back-arc input in
1249 Alutaguse, whereas the SS belts reflect slab-modified lithospheric sources with
1250 a sediment-derived component. Collectively, the data support Alutaguse as a
1251 ~ 1.90 – 1.89 Ga back-arc rift linked to Tallin–Uusimaa arc over north Bergslagen
1252 microcontinent, preceding accretion at 1.87 – 1.86 Ga.

1253 Still, future research must focus on: *U–Pb* zircon on magmatic and detrital
1254 grains for crystallisation and maximum depositional ages and provenance; zircon
1255 *Hf* with whole-rock *Nd–Sr* to separate juvenile vs. recycled sources and appor-
1256 tion arc vs. back-arc inputs; sulphide *S–Pb* chemistry to characterize sulphur
1257 and metal reservoirs; *O–H–C* in carbonates and quartzites to screen fluid ori-
1258 gin and water–rock interaction; fluid-inclusion microthermometry and Raman
1259 to estimate *P–T–X*; and integrated gravity–magnetics with *2D/3D* EM and
1260 IP methods along sulphide- and graphite-bearing corridors to rank VMS-style
1261 *Pb–Zn–Cu* targets.

1262 7. Supplementary data

1263 Complete geochemical datasets for the Alutaguse and South Svecofennian
1264 (SS) domains are provided in the supplementary data. Also, a document with
1265 supplementary plots is attached.

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1271 9. Declaration of competing interest

1272 The authors state that they have no known competing financial interests or
1273 personal ties that may have influenced the work described in this study.

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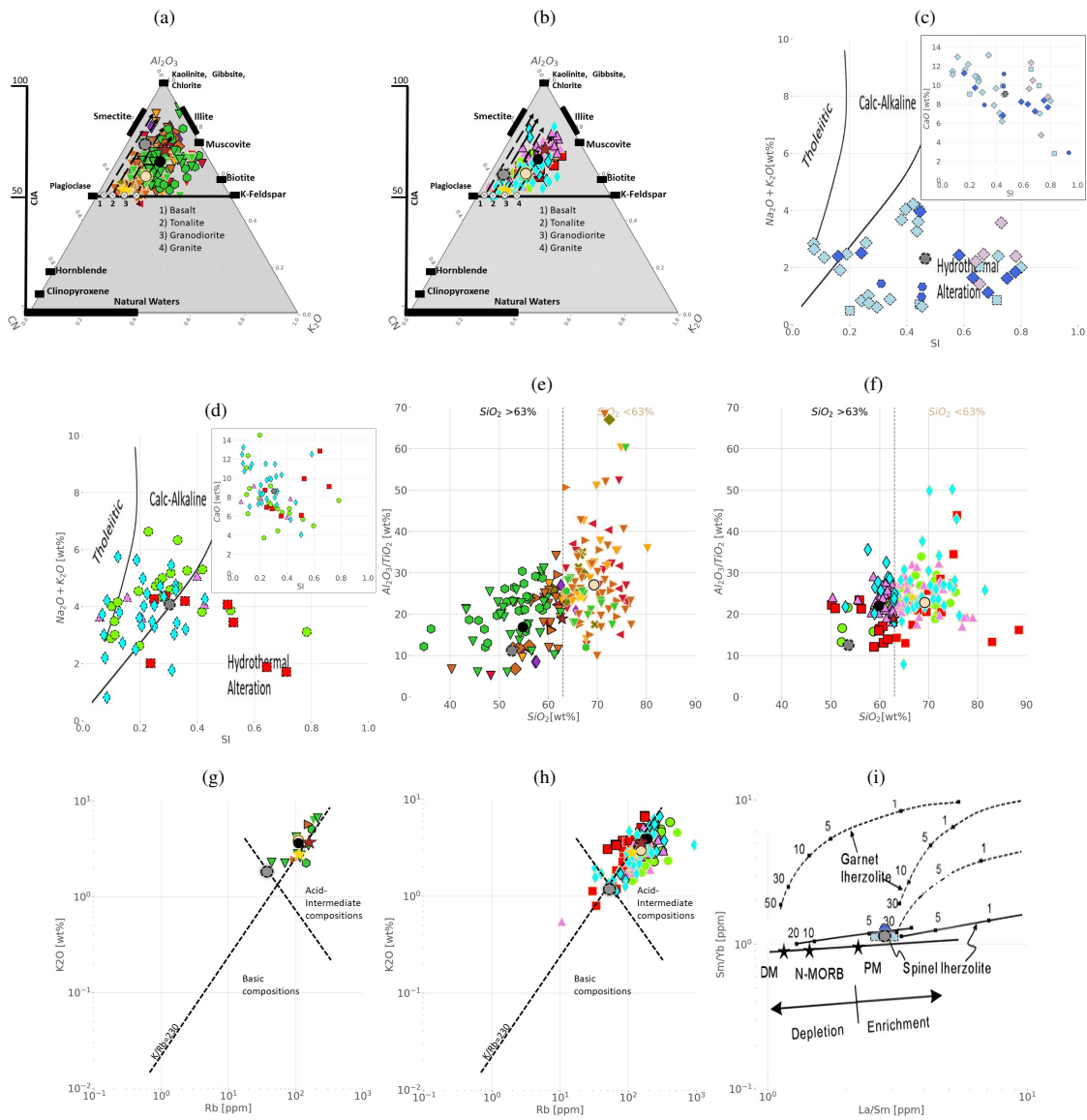
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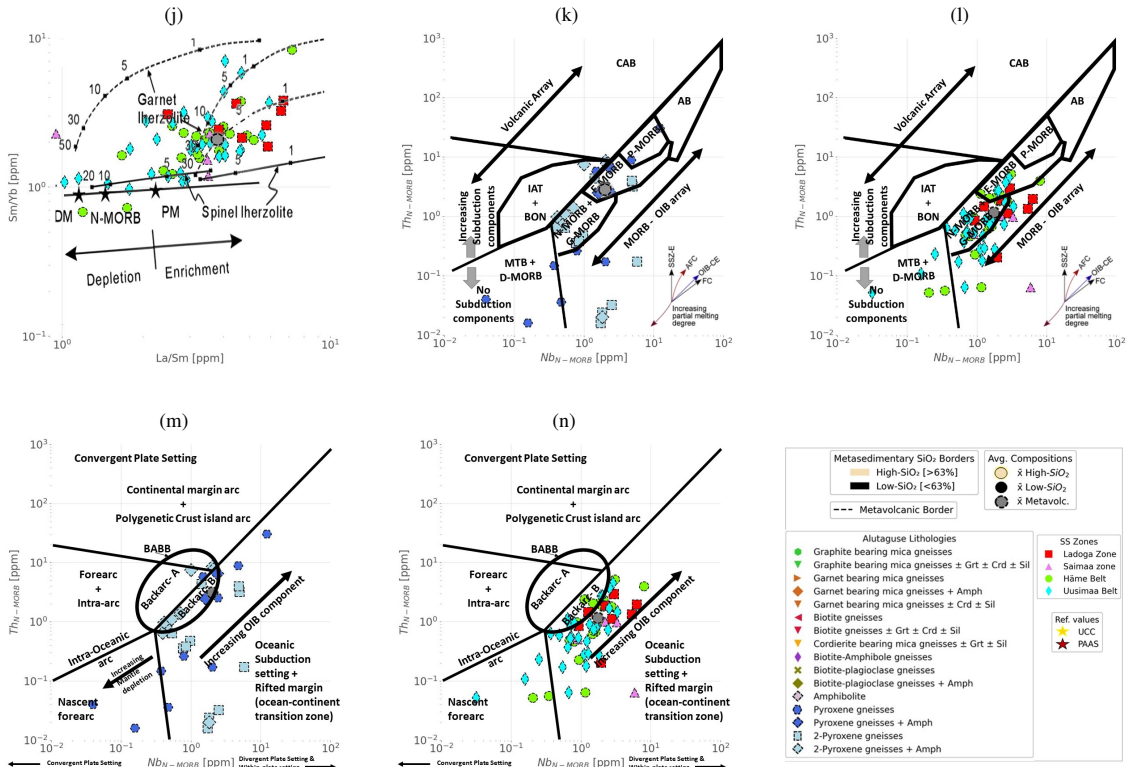
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Supplementary 1. Weathering and alteration discrimination plots for Alutaguse and South Svecofennian (SS) samples: (a-b) $A - CN - K$ [$Al_2O_3 - (CaO^* + Na_2O) - K_2O$] ternary (after Nesbitt and Young (1984, 1989), (c-d) sericitisation index SI vs. $K_2O + Na_2O$ for volcanic rocks with SI vs. CaO insets for metavolcanic rocks (after MacLean and Hoy (1991), (e-f) The Al_2O_3/TiO_2 vs. SiO_2 plots for both zones are showcased in (a-b) after Hayashi et al. (1997), (g-h) discriminant function plots K vs. Rb trends after Shaw (1968), (i-j) Sm/Yb vs. La/Sm after Yang et al. (2014), Saccani (2015) N-MORB-normalised Th vs. Nb for volcanic and metavolcanic rocks, for source (k-l) and tectonic setting (m-n).

Appendix VI

J D Solano-Acosta, S Graul, A Soesoo, R Hints, T All, J Vind, and S Liivamägi. Exploration of critical elements in the concealed North Estonian basement and its Eastern Fennoscandian context. 2025. Manuscript Under review

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Exploration of critical elements in the concealed North Estonian basement and its Eastern Fennoscandian context

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Abstract: This study evaluates the mineral potential of the NE Tallinn-Alutaguse-Jõhvi Estonian Paleoproterozoic corridor, composed of amphibolite- to granulite-facies gneisses and metal-rich volcanic-sedimentary units. Mesoproterozoic rapakivi bodies were also analysed. Borehole geochemistry reveals that the Alutaguse metavolcanic suite has high concentrations of Cu (2230 ppm), Zn (2650 ppm), and Pb (4030 ppm), hosted in magnetite-rich and sulphide-graphite gneisses, similar to those found in ore systems in Sweden and Finland. Data suggest that the metalliferous Alutaguse target zones are linked to intrusive bodies and/or hydrothermal-skarn processes akin to those at Jõhvi. Automated MSCL-XYZ scanning of drill cores reveals associations of critical metals, including Ni-Co-Cr, Mo-W-Bi, Sn-Zn-Cd, Cu-Ni, Nb-Y-P, and Au-Ag-As-Sb-Bi-W-Se-Sn, identifying new prospective intervals. Analyses of the Märjamaa rapakivi pluton reveal REE concentrations of up to 3600 ppm in the Si-poor Fe-Ti-P-rich Phase I, indicating phases with potential exploration value. Geophysical data also highlight zones of high magnetic susceptibility, possibly inferring altered or mineralised target zones.

Keywords: Critical Raw Minerals (CRM) exploration; Estonian Precambrian basement; Geophysical potential analysis; drill-core Geochemistry.

1. Introduction

The Baltica or East European Craton in northeastern Europe comprises Archean-Paleoproterozoic to Neoproterozoic cratons, including Fennoscandia, Volgo-Uralia, and Sarmatia, which developed independently prior merging around 1.8-1.7 Ga [1-8] (Fig.1a). Seismic and paleomagnetic data indicate that Baltica played a pivotal role in forming the Nuna supercontinent, characterised by substantial latitudinal movement and interactions among arc systems, microcontinents, and magmatism [3, 6-7]. In Fennoscandia (Fig.1b), this is represented by the Paleoproterozoic Svecofennian Orogen (SO), which shows large-scale accretionary growth through multiple subduction-collision events, resulting in NW-SE-trending tectonic domains [1-12], and later Mesoproterozoic reactivation during Gothian and AMCG (Anorthosite-Mangerite-Charnokite-Granite) magmatism (1.6-1.4 Ga), mainly marked by rapakivi granites dated between 1.64 and 1.44 Ga, indicating a major intracontinental magmatic phase overlying the Svecofennian crust [13-18].

The Estonian Precambrian crystalline basement is a hidden continuation of the Fennoscandian Shield (Fig.1b), mainly consisting of Paleo- to Mesoproterozoic metamorphic and igneous rocks (around 1.9-1.5 Ga) buried under 100-900 metres of Neoproterozoic to Devonian sediments that thicken southward (Figs.1c-e), which structure, lacking outcrops, is inferred from drill cores, gravity and magnetic data, seismic surveys, and geochronology [2,8-12,18-26]. Six structural-petrological zones: Tallinn, Alutaguse, Jõhvi, West-Estonian, Tapa, and South-Estonian, are identified based on lithology, metamorphic grade, and geophysical signatures, each with distinct features (Fig.1c) [8-12,18-22].

Data from the Trans-Baltic region show the Paleoproterozoic crust beneath Estonia originated in the southern SO, a zone with successive accretion of arc complexes, microcontinents, and sedimentary basins forming the current mosaic of tectonic belts and oroclinal structures [1-8,18-26]. The Southern Svecofennian (SS) area in Finland, encompassing the Häme and Uusimaa belts and the Saimaa and Ladoga zones, exhibits interactions between juvenile arc crust and Archean basement, with arcuate contours interpreted as the Bothnian and Saimaa oroclines. Data suggest the Uusimaa belt extends beneath the Gulf of Finland toward the Tallinn zone. In contrast, the Ladoga zone continues southeast along the Archean boundary, marking the southeastern edge of the Svecofennian margin. These domains link south to the Bergslagen microcontinent and its ore province in south-central Sweden, forming a continuous Paleoproterozoic structural and magmatic corridor [2,4,8-12,18-29].

This corridor, spanning 1.91-1.87 Ga (Fig.1f), encompasses the Tallinn, Alutaguse, and Jõhvi domains [2, 18-28]. It features Fe-S-Si-rich gneisses, magnetite units, and polymetallic systems, which are linked to Bergslagen-type successions in Finland and Sweden [2, 8, 19-25, 27, 28]. The units are associated with 1.91-1.89 Ga felsic volcanic and subvolcanic

rocks, carbonates, turbidites, skarns, and later 1.79 Ga granites and migmatites linked to the late Svecofennian belt [2,8,11,12,19-25]. Notably, the Alutaguse region, comprising aluminous and graphitic gneisses, quartzites, carbonates, and metavolcanics, is thought to be a rifted back-arc segment within the Uusimaa-Tallinn arc system, later compressed and sealed, creating conditions for VMS-type Cu-Pb-Zn and polymetallic deposits (i.e., Critical Raw Minerals; CRM) [2,8,19-25].

In the Baltic countries, rapakivi intrusions are largely buried beneath the Phanerozoic cover; their distribution and architecture are inferred from seismic and potential-field data [2, 3, 6, 8, 10, 14-17]. In Estonia, the rapakivi suite comprises the large Riga Batholith and smaller, older intrusions, such as Märjamaa, Kloostri, Naissaare, Taebla, Neeme, and Ereda (Figure 1c), along with A-type stocks (e.g., Abja, Muhu, Virstu) [8, 10, 30]. Mainland plutons are mainly pink, coarse monzogranite-syenogranite, locally cut by aplitic and microsyenitic dykes. However, the classic rapakivi texture is rare; geochemical and isotopic signatures match those of the broader Fennoscandian rapakivi suite, emplaced along the Åland-PPDZ structural corridor. Geophysical, petrological, and geochemical work has shown that Märjamaa-Kloostri is a composite, structurally controlled system with three magmatic phases that display marked contrasts in petrophysical properties, oxidation state, fluorine content, and rare-earth enrichment, particularly in the early Fe-rich, silica-poor facies [13-17, 23, 24, 30].

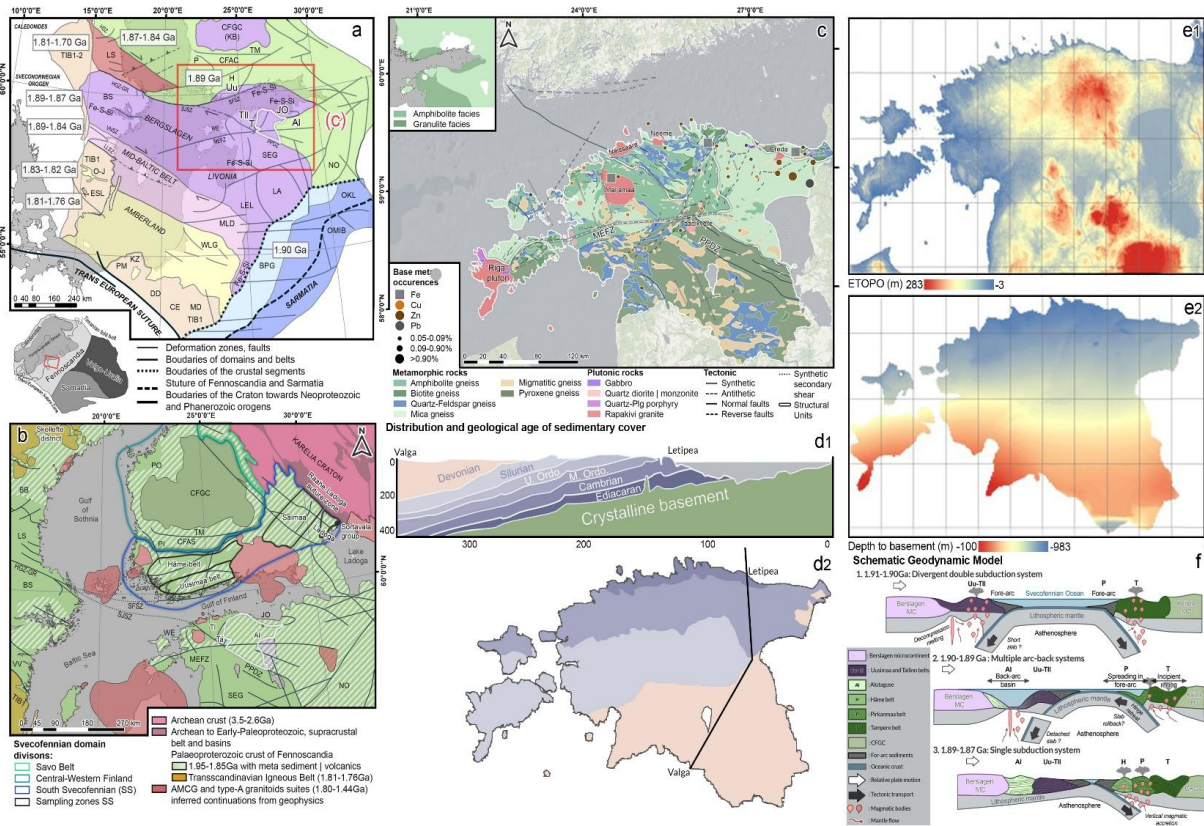


Figure 1: Geological setting of the study area. **(a)** Central and Southern Svecofennian crustal structure across the Baltic Sea, showing Bergslagen, Livonia, Amberland, Keitele, and Bothnia zones; the red box marks the study area. Ages indicate major accretionary events. Fe-S-Si marks iron-sulphide ore provinces. **(b)** Main Paleoproterozoic tectonic zones of the Fennoscandian region, including the South Svecofennian (SS) and Alutaguse domains, with Svecofennian sedimentary basins indicated by diagonal ruling. Key abbreviations: AL- Alutaguse, BB- Bothnian, BS- Bergslagen, CFAC- Central Finland Arc zone, CFGC- Central Finland Granitoid zone, JO- Jõhvi, KB- Keitele, LEL- Latvian-East Lithuanian, SEG- South Estonian granulite, TII- Tallinn, WE- West Estonian; deformation zones: MEFZ- Middle Estonian Fault Zone, PPDZ- Paldiski-Pskov Deformation Zone, SFSZ- South Finland Shear Zone. Figures (a) and (b) are adapted after Bogdanova et al., 2015 [2]; please refer there for further information. **(c)** Precambrian basement map of Estonia with geochemical anomalies after Soesoo et al., 2020 [8]. **(d)** Palaeozoic framework of Estonia: (d1) crustal cross-section; (d2) generalised map of sedimentary cover and Lower-Middle Palaeozoic outcrops. **(e)** Physical measurements: (e1) ETOPO1 topography-bathymetry model; (e2) crystalline basement depth from drill-core data. Figures (d) and (e) are modified from Solano-Acosta et al. (2023) [18]. **(f)** Schematic Svecofennian Orogeny geodynamic model for North Estonia-South Finland (1.92-1.90 Ga; 1.90-1.89 Ga; 1.89-1.87 Ga), after Solano-Acosta et al. (2025) [19].

During the past century, more than 500 holes have been drilled to explore the crystalline basement. A significant portion of these drill cores is preserved and available in the Geological Survey of Estonia (EGT) rock archive. However, between 1990 and 2018, few studies of the crystalline basement were conducted [8]. Recent investigations have significantly enhanced our understanding of metal genesis and the overall characterisation of Precambrian units. Based on several geological, geochemical, and geophysical studies, this research integrates their results into a unified framework for the concealed Paleoproterozoic to Mesoproterozoic North Estonian basement and its position within the eastern Fennoscandian crust. By combining interpretation of legacy drill-core datasets, petrological analysis, automated Geotek MSCL-XYZ scanning, and Bouguer and magnetic data, the study reevaluates the distribution of critical-element associations within the Tallinn-Alutaguse-Jõhvi corridor, synthesises the lithological and structural controls on these systems, and considers rapakivi intrusions (especially the Märjamaa-Kloostri) in a tectono-magmatic and metallogenic context. Within Baltica’s geodynamic evolution, this synthesis ought to refine the metallogenic model of Estonia and identify new exploration targets.

2. Geological Setting

2.1 Fennoscandian Context

Baltica The Paleoproterozoic crust of Fennoscandia consists of Archean blocks surrounded by Svecofennian supracrustal belts and granitoid complexes, segmented by major shear zones such as the Raahe-Ladoga system, the South Finland Shear Zone, and the Paldiski-Pskov Deformation Zone (PPDZ) [1-11,18-20] (Figs. 1a-c). Accretionary and collisional stages of the Svecofennian Orogen assembled arcs, microcontinents, and sedimentary basins into several provinces, including the SS zone (Häme, Uusimaa, Saimaa, Ladoga; Fig.1b), where juvenile arcs interacted with the Archean crust to develop oroclinal structures along the Bothnian and Saimaa trends [2,4,5]. Comparable processes shaped the Bergslagen province of Sweden, where 1.9-1.8 Ga volcanic-sedimentary-intrusive successions host extensive Fe-oxide and base-metal mineralisation [2, 18-24, 27, 28]. Seismic reflection-refraction experiments across the Baltic Sea, accompanied by gravity and magnetic modelling, reveal that the Svecofennian crust is organised into blocks bounded by steep, deeply rooted faults that remained mechanically weak into the Mesoproterozoic [3, 6, 18]. These inherited structures guided post-Svecofennian reactivation and, later, channelled AMCG magmatism along corridors such as the PPDZ [16,18], which continues to influence present-day intraplate seismicity [29].

2.2 Estonian Basement Domains

The Estonian crystalline basement represents the southeastern continuation of the Fennoscandian Shield. It is subdivided into structural-petrological domains bounded by long-lived shear systems such as the PPDZ and the Middle Estonian Fault Zone (MEFZ). These structures separate the amphibolite-facies crust in the north from granulite-facies blocks in the south and were repeatedly reactivated during the Mesoproterozoic [2,8-12,16]. In northern Estonia (Fig. 2), the Tallinn, Alutaguse, and Jõhvi domains host Paleoproterozoic volcanic-sedimentary successions, which were later intruded by granitoids dated at 1.92-1.88 Ga. These domains include Fe-Si-rich gneisses, magnetite-bearing units, and sulphide-graphite horizons forming a part of the Bergslagen-Southern Svecofennian metallogenic corridor, with metamorphism dominated by amphibolite facies, locally reaching granulite conditions at ~3-5 kbar [2,8].

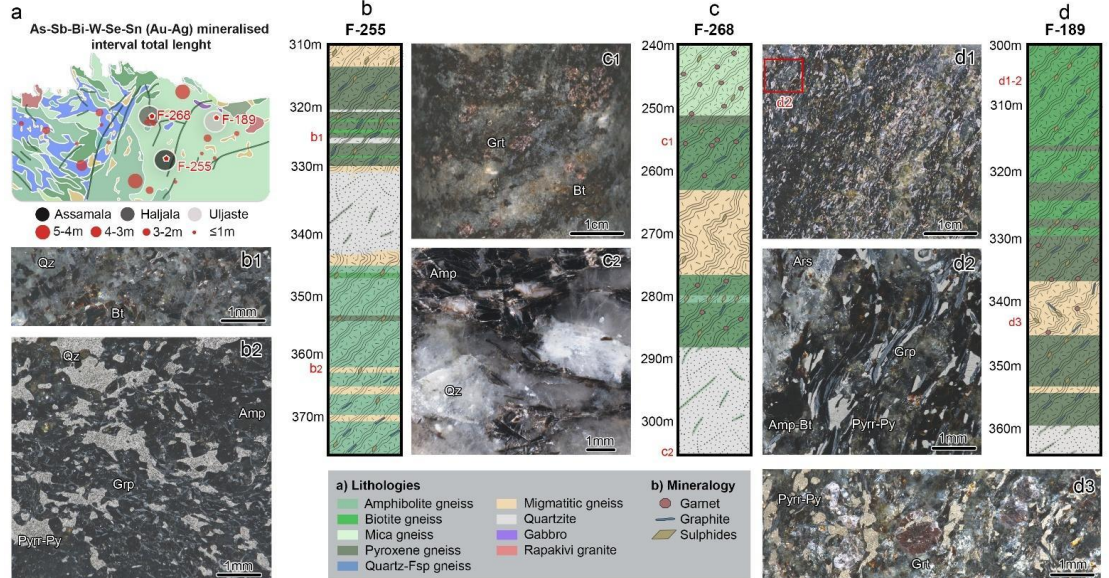


Figure 2: (a) Close-up of the Tallinn-Alutaguse domains. The red symbol size represents the length of As-Sb-Bi-W-Se-Sn mineralised intervals, identified through Geotek MSCL-XYZ scanning [25]. The map highlights prospective metalliferous

targets and representative drill cores at **(b)** Haljala, **(c)** Assamalla, and **(d)** Uljaste (see Fig. 2a). The ID number of each drill core is indicated on the map. Representative 10× photomicrographs of the analysed samples are provided for each locality, along with the sampling depth. Mineral abbreviations are as follows: Amp- amphibole, Ars- arsenopyrite, Bt- biotite, Qz- quartz, Pyr- pyrrhotite, Py- pyrite, Grp- graphite, Grt- garnet.

The Tallinn Domain features negative free-air gravity and magnetic anomalies, indicating a relatively felsic crust [8, 11, 19]. The PPDZ bounds it to the southeast and contains amphibolite-facies metavolcanic rocks, amphibole-biotite-plagioclase gneisses, quartz-feldspar and Al-mica gneisses (biotite-cordierite-garnet-sillimanite), along with local sulphide-graphite gneisses and magnetite quartzites of the Jägala Complex [8,11,12]. Migmatization is common, and structural-lithological correlations suggest that Tallinn is the southern extension of the Uusimaa Paleoproterozoic volcanic arc of Finland [2,8-12,16]. The Alutaguse Domain exhibits near-zero to slightly negative gravity and magnetic values, with local positive anomalies at Uljaste, Assamalla, and Haljala (Fig. 2). These anomalies are associated with sulphide-graphite gneisses, quartzites, skarnised carbonaceous rocks, and migmatized-pyroxene gneisses [8,11,12]. This domain is generally regarded as a rifted back-arc segment of the Uusimaa-Tallinn system, which was later compressed during post-Svecofennian deformation [8, 19-25]. The Jõhvi Domain is a narrower, distinct belt of alternating Fe- and S-rich garnet-pyroxene quartzites, high-Al garnet-cordierite-sillimanite gneisses, and Ca-rich to Ca-poor pyroxene, amphibole, and biotite gneisses. These rocks exhibit strong migmatization and granite intrusions dating back to around 1.8 Ga, which aligns with late- to post-orogenic phases in southern Finland [2,5,8,18-25,34-36]. The Jõhvi province also exhibits some of Estonia's strongest potential-field anomalies, mainly due to magnetite-rich units, highlighting it as a key part of the Bergslagen-type metallogenic corridor [24,25,31,32].

2.3 Mesoproterozoic AMCG and Rapakivi Magmatism

The Mesoproterozoic anorogenic magmatism of Fennoscandia consists predominantly of AMCG suites and rapakivi granites, emplaced between 1.64 and 1.44 Ga, within a broader 1.8-1.0 Ga intracontinental magmatic period [3, 7, 13-17]. These intrusions form batholiths, along with large plutons, that overprint the Svecofennian architecture. They are spatially linked to crustal-scale faults, horst-graben structures, and zones of mafic underplating [16,37-40]. Paleomagnetic reconstructions suggest that this magmatism evolved in a long-lived intracontinental environment associated with Nuna breakup, plume trapping, and superswell activity beneath a thick lithosphere, rather than through simple rift or hotspot processes [3, 7, 16, 37-40].

The Fennoscandian rapakivi granites are grouped into four AMCG age clusters: 1.67-1.62, 1.59-1.56, 1.55-1.53, and 1.53-1.44 Ga; represented by the Wiborg, Åland, Salmi, and Ragunda suites, which together extend from southern Finland beneath the Gulf of Finland into Estonia, Lithuania, and Poland [7,13-16,38-40]. Seismic reflection profiles and potential-field inversions across the Baltic Sea indicate that rapakivi magmas were channelled along listric faults and dilatational ramps within NW-SE Svecofennian shear corridors, especially the Åland-PPDZ trend, producing tabular, sheet-like plutons marked by strong internal density and magnetic contrasts [16, 41, 42]. Thermal modelling further indicates that high-heat-production granites and sustained lithospheric weakening governed the emplacement and longevity of AMCG magmatism [16, 35, 37].

A-type and rapakivi granites are increasingly recognised as important sources for economic mineral systems. The intracontinental rift-superswell environments that produced the AMCG suite globally favour SEDEX, stratabound Cu, and IOCG-style deposits, driven by oxidised, volatile-rich, mantle-derived or lithosphere-fertilising fluids [16,37,43]. Halogens, such as F, also enhance metal transport, promote silicate dissolution, increase permeability, and generate characteristic greisen and skarn assemblages [16, 37, 43]. In the Wiborg Batholith and related intrusions of southern Finland, mineralization includes In-bearing magnetite-sphalerite ores, Zn-Cu-Pb-Ag-In polymetallic veins, and Sn-W-Be greisen systems, controlled by late, evolved rapakivi phases and strike-slip/transensional structures [16, 37-39, 44].

In Estonia, the Wiborg Suite is represented by the concealed Riga Batholith (also in Saaremaa) as well as several smaller intrusions: Märjamaa, Kloostri, Naissaare, Taebala, Neeme, and Ereda, which are emplaced along the Åland-PPDZ corridor [13,16,39,42]. Geophysical and petrological studies indicate that the Märjamaa-Kloostri intrusions form a composite, structurally controlled plutonic system comprising three magmatic phases with pronounced contrasts in density, susceptibility, oxidation state, and trace-element enrichment [16, 17, 23]. Recent work documents anomalously elevated REE contents, up to several thousand parts per million (ppm), locally in early Fe-rich, silica-poor phases, together with elevated F and other volatiles. Moreover, a clear spatial relationship with inherited Svecofennian shear structures was highlighted [16, 17, 23].

3. Materials and datasets

3.1 Geochemical datasets

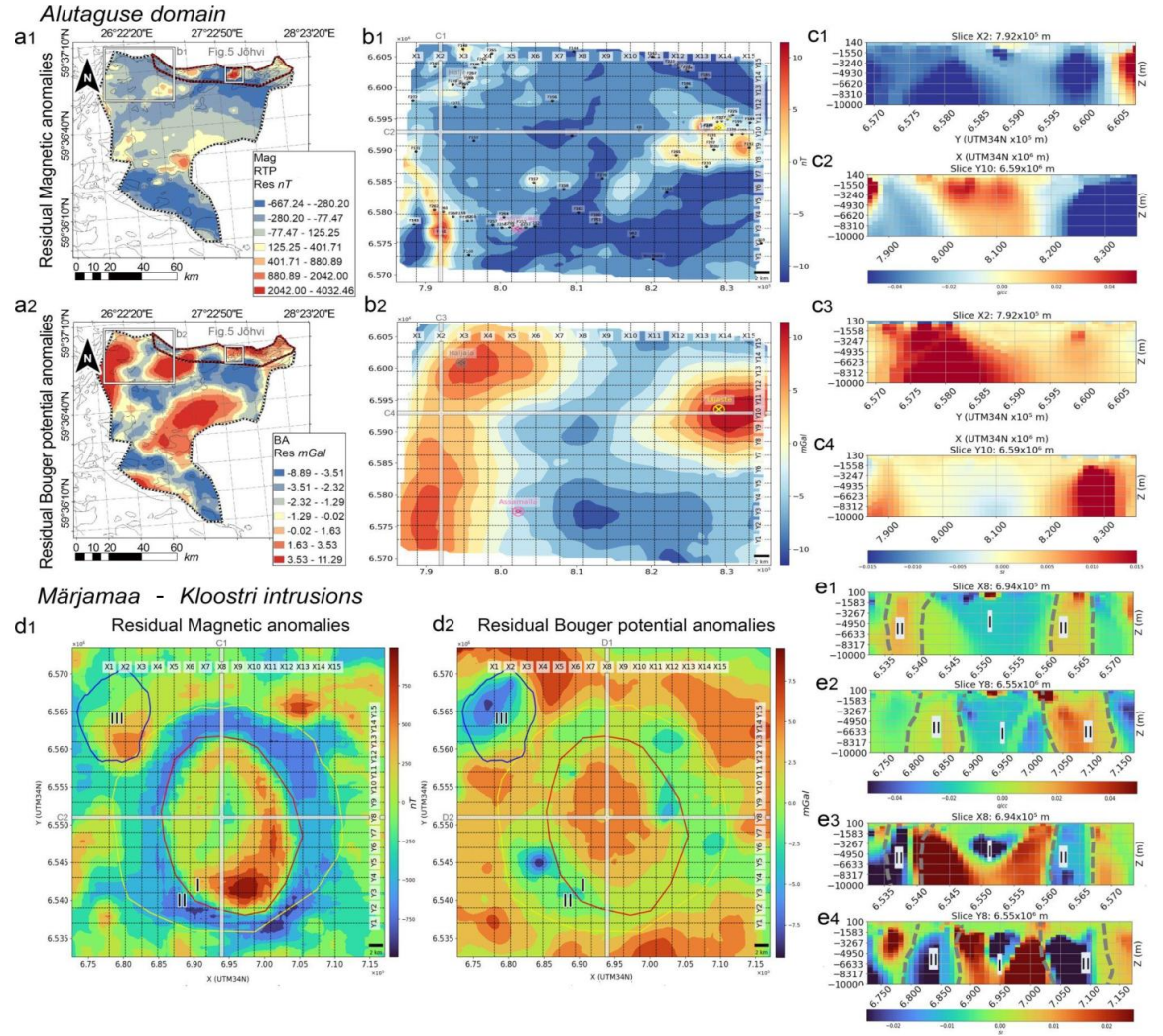
Alutaguse and SS metasedimentary and metavolcanic units' whole-rock compositions integrate three complementary datasets, analysed in previous articles [19,20,22-24]: (i) legacy major-element analyses (216 samples; Kivisilla et al. (1999) [33]); (ii) high-quality major-trace-REE data (16 samples; Solano-Acosta et al. [19]); (iii) new EGT trace-element data (149 samples) from Uljaste and adjacent cores [20,22-24]. Together, these define the major-, trace-, and REE variability of the Alutaguse units, supporting provenance and tectonic interpretations. Comparative Svecofennian data include 206 filtered major-trace samples from the GTK (Finnish Geological Survey) Rock Geochemical Database for Uusimaa, Häme, and Saimaa [44], supplemented by Ladoga siliciclastics (26) and metavolcanics (8), from Kotova et al. [26]. These datasets provide a robust regional framework for evaluating North Estonian arc/back-arc affinities with SS, as presented in Solano-

Acosta et al. (2025) [19, 20, 22]. CRM associations are derived from MSCL-XYZ multi-spectral scanning of 22 drill cores from NE Estonia (Tallinn-Alutaguse) conducted by the EGT as reported by Nirgi et al. [25] (Fig.2a). Inferred results reported certain elemental groups of polymetallic prospective systems, including Ni-Co-Cr, Cu-Ni, Ti-V-Fe, Mo-W-Bi, Sn-Zn-Cd, Nb-Y-P, K-Sn-Rb-Ga, and As-Sb-Bi-W-Se-Sn.

Rapakivi compositions are constrained using: (i) Legacy major-element data for Estonian rapakivi granitoids from Kivisilla et al. (1999) [33]. (ii) Märjamaa major-trace-REE dataset (62 samples) from Potagin (2024) [39], and (iii) Wiborg Suite data, extracted from the GTK Rock Geochemical Database [44].

3.2 Geophysical datasets

The geophysical component uses EGT Bouguer anomaly magnetic grids (<https://gis.egt.ee/>). Processed Residual Bouguer and RTP magnetic embbody studies for Alutaguse and Jõhvi (Figs . 3a-c) [23]. Additionally, the Jõhvi dataset incorporates a high-resolution ground magnetic survey acquired using a G-856AX proton-precession magnetometer [32]. For the Märjamaa-Kloostri intrusions, EGT gravity and magnetic grids were utilized [16, 23] (Figs. 3d-e).



Major-element trends (Fig. 4a) show that the Alutaguse metavolcanics define a tholeiitic array, with increasing MgO and P₂O₅ and decreasing Na₂O toward mafic compositions, consistent with juvenile magmas produced in an extensional arc-proximal to back-arc setting [19,20,22,26,27]. Relative to all SS domains, the Alutaguse suite is systematically enriched in TiO₂, Fe₂O₃, MgO, and CaO and depleted in Al₂O₃, Na₂O, K₂O, and P₂O₅, indicating derivation from a hotter, more depleted mantle source and subsequent hydrothermal modification. CaO and MnO are highest in the Alutaguse and Uusimaa metavolcanic systems. Low A/CNK values and Fe-Ti-rich compositions further support a back-arc tholeiitic affinity [19, 20, 22, 23, 26].

Trace-element signatures (Fig. 4b) reinforce these distinctions. Alutaguse metavolcanics exhibit low La/Yb and high Zr/Nb ratios, consistent with the partial melting of a depleted asthenospheric mantle. In contrast, Ladoga volcanics yield higher La/Yb and lower Zr/Nb, indicating a more slab-modified source [19,20,22,23,26]. Elevated Th/Nb and depleted Ba/Th and U/Th in Alutaguse indicate a minimal slab-fluid contribution, in contrast to Häme and Saimaa, which display stronger subduction signatures [19,20,22,23,26].

Base-metal concentrations in the Alutaguse suite: Ni 210 ppm, Cu 2230 ppm, Zn 2650 ppm, and Pb 4030 ppm, are among the highest in the entire corridor and exceed both Alutaguse metasedimentary maxima and all SS metavolcanic fields [19-23,25,27,31-36]. This combination of tholeiitic major-element chemistry, asthenospheric mantle signatures, and high Cu-Zn-Pb enrichment parallels those of Bergslagen-type back-arc volcanic centers [21,22,23,27,28].

REE patterns (Fig. 4c) indicate moderate LREE enrichment and subdued HREE in the Alutaguse metavolcanics, with average REE contents exceeding most SS suites except Saimaa, consistent with melts derived from amphibolitised sources and later modified by hydrothermal circulation [19,20,22]. A positive Eu anomaly in Alutaguse may reflect high-temperature hydrothermal processes under reducing conditions; however, further sampling is required to confirm this observation. In contrast, SS metavolcanics, particularly from Saimaa and Häme, are enriched in Zr, Hf, and Y (Fig. 4b), consistent with more evolved intra-arc or continental-margin magmatic sources.

4.2. MSCL-XYZ element associations and implications for targeting

The MSCL-XYZ scanning of selected drill cores from the Geological Survey of Estonia's archive refines the understanding of critical metal distribution in the Tallinn-Alutaguse domains (Fig. 2a) [22, 25]. Core selection relied on digitised historical data and prioritised CRM [21, 22, 25]. Because basement lithologies differ strongly in background chemistry, anomalous Ni-Co in felsic or metasedimentary rocks, or REE, are particularly diagnostic of hydrothermal or metasomatic processes [21, 22, 25].

Scanning at 5-15 cm spacing along selected intervals allowed identification of key element associations: Ni-Co-Cr, Ti-V-Fe, Mo-W-Bi, Sn-Zn-Cd (In), Cu-Ni (PGM), Nb-Y-P (REE), K-Sn-Rb-Ga (Li), and As-Sb-Bi-W-Se-Sn (Au-Ag) [25]. Ni-Co-Cr and Cu-Ni are mainly linked to mafic-ultramafic intrusions; Ti-V-Fe is concentrated in the Haljala area; Mo-W-Bi occurs in both Haljala and the Tallinn structural zone; and Sn-Zn-Cd anomalies cluster in Uljaste and Viru-Nigula [25]. Other associations (Nb-Y-P, K-Sn-Rb-Ga, As-Sb-Bi-W-Se-Sn) span multiple lithologies, reflecting varied magmatic and hydrothermal controls [21, 22, 25, 31].

Scanner data reveal stronger heterogeneity than suggested by legacy descriptions, exposing abrupt changes in rock character and metal content [25]. This directly improves exploration targeting by identifying previously unrecognised mineralised zones and linking them to specific geophysical anomalies and lithologies. To fully assess these signals, detailed laboratory geochemical and petrographic-mineralogical studies are required [19-25].

4.3 Exploration significance: Critical-metal systems in the Tallinn-Alutaguse-Jõhvi corridor

The Tallinn domain represents the felsic arc front of the 1.92-1.89 Ga Tallinn-Uusimaa-Bergslagen system. It also forms the metallogenic core of the corridor, hosting the highest Cu-Pb-Zn enrichments in Low-SiO₂ and metavolcanic units (Fig. 3b), consistent with sustained volcanogenic and intrusive inputs, as well as focused hydrothermal circulation, similar to fertile Bergslagen-type back-arc systems [19-22, 25, 27, 28]. At the eastern end of the corridor, the Jõhvi magnetite province provides the Fe-oxide endmember, with magnetite-rich gneisses (15-46 wt.% Fe; 1-6 wt.% Mn) and polymetallic sulphides representing metamorphosed volcanic-sedimentary protoliths rather than a typical skarn or BIF systems [8,21,31,32]. Magnetite compositions (Ti+V vs. Ca+Al+Mn) fall near the skarn-IOCG spectrum, and exceptionally strong remanent magnetisation reflects abundant sub-micrometre magnetite grains, a signature shared with Bergslagen and Orijärvi [2, 5, 21, 31, 32].

Geophysical enhancement of this scenario, with the residual Bouguer and RTP high-potential values, coincides with metal-rich lithologies across Alutaguse and Jõhvi (Fig. 2a-b; [8,18,21-23]), and cross-gradient SimPEG inversion (Fig. 3c) resolves ~10 km depth tabular zones of high susceptibility and low density corresponding to Fe-oxide-sulphide bodies at Haljala, Assamalla, and Uljaste; flanked by weakly magnetic graphite-rich horizons marking consolidated back-arc stratigraphy (e.g., density contrasts). The zones with lower magnetic signals are interpreted as graphite-rich gneisses, possibly formed during the consolidation of the back-arc basin. In contrast, the metalliferous Alutaguse zones are likely linked to intrusive bodies and hydrothermal-skarn enrichment, similar to those at Jõhvi [23, 31, 32].

Plado et al. (2020) [32] provide a geological insight into the Jõhvi magnetic anomaly in NE Estonia. The investigation revealed that the three magnetic peaks, with maximum amplitudes of 19,290 nT (western), 15,880 nT (eastern), and 8,080 nT (northern) (Fig. 5b), are primarily caused by strong remanent magnetisation that is directed roughly vertically and points downward. The direction aligns with the dip of the iron ore formation. Strong remanence indicates a significant presence of small magnetite grains (<1 µm) in the Jõhvi ore, which warrants future (magneto)mineralogical studies. Petrophysical

measurements of Mag-quartzites revealed that the residual magnetisation was about ten times greater than the induced magnetisation [32].

Together, Tallinn defines the felsic arc source, Alutaguse the metal-rich, tectonically focused back-arc basin, and Jõhvi the Fe-oxide-sulphide pole, establishing a continuous and concealed Cu-Pb-Zn-C-S-Fe-oxide metallogenic belt along the NW-SE Svecofennian structures. MSCL-XYZ scanning (5-15 cm resolution; Fig. 2a) further corroborates this framework by revealing heterogeneous Ni-Co-Cr, Ti-V-Fe, and Sn-Zn-Cd associations related to mafic influx and hydrothermal-metasomatic overprinting [21-25], thus refining exploration targets across the corridor.

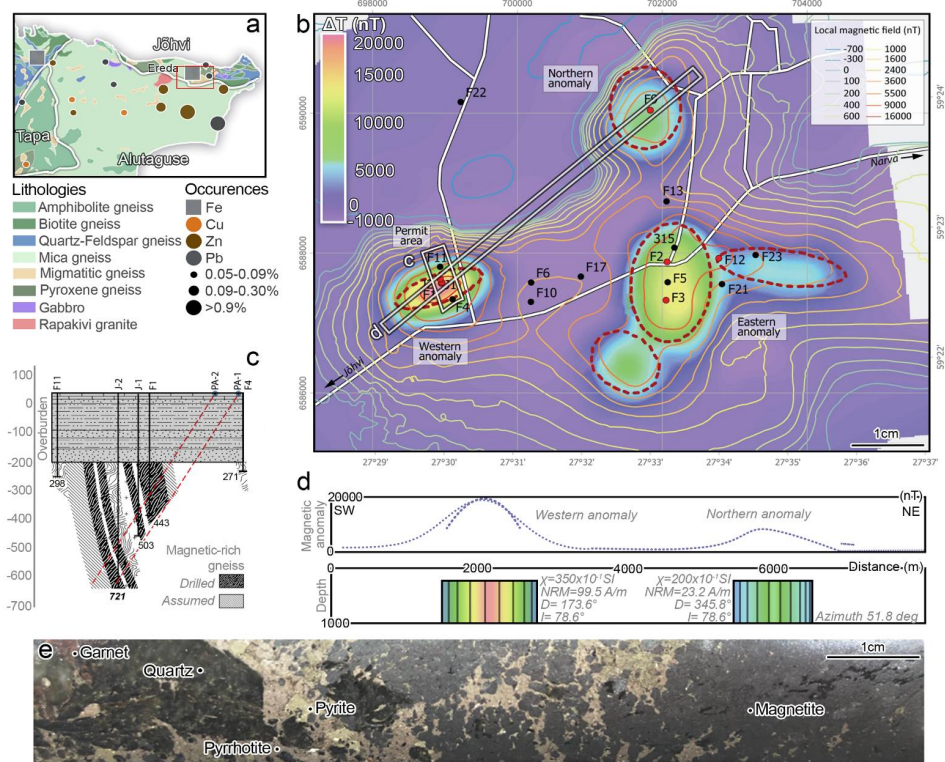


Figure 5. (a) Close-up of the Jõhvi zone; the red box indicates the analysed area. (b) Ground magnetic measurements in the Jõhvi area were conducted in 2019 using the G856AX proton precession magnetometer by Geometrics, Inc. A handheld GPS device (Garmin eTrex 20) was used to record measurement locations. An area of approximately 50 km² was surveyed, encompassing 2897 measurements. Results were processed by Plado et al. (2020) [32]. The Jõhvi Magnetic Anomaly is depicted with five elliptical cylinders (red dashed lines) representing the surface projection of modelled orebodies, as indicated by the partial anomalies. Borehole locations are marked as dots, with red dashed contours indicating areas of magnetite-rich gneiss. (c) Sketch map of historical (J-1, J-2, F1) and recent (PA-1, PA-2) boreholes drilled into the western part of the Jõhvi Magnetic Anomaly. (d) Cross sections of the elliptic cylinder anomalies. The response curve (solid line) and the original field measurements (over the elliptical modelled body) are shown. (e) An example of a drill core with a banded texture typical of magnetite quartzites in the Jõhvi iron formation.

4.3 Geophysical and Petrogenetic Insights from the Märjamaa-Kloostri Rapakivi Intrusions

4.3.1 Geophysical Architecture, Structural Controls, and Emplacement Framework

High-resolution residual Bouguer and RTP magnetic datasets (Fig. 3d) delineate a sharply zoned intrusive system characterised by strong density-susceptibility contrasts, which record the multiphasic Märjamaa-Kloostri assembly [13, 16, 17, 23]. Phase I forms a dense, magnetite-bearing core, expressed by coincident gravity and magnetic highs, surrounded by a lower-density, weakly magnetic Phase II shell, and an outer Phase III domain marked by a Bouguer low, along with a subtle NW RTP high. This is consistent with an evolved, oxide-poor, Na-rich composition [16,23]. SimPEG cross-gradient inversions to 10 km depth (Figs. 3d-e) resolve these three bodies as a steep, vertically elongated high-susceptibility Phase I column, a positive-density/low-susceptibility Phase II envelope, and a NW-shifted Phase III satellite with susceptibility increasing at depth. Linear susceptibility ridges at the Phase I-II boundary likely represent feeder zones or collapsed blocks, providing the first 3D evidence for piston-cauldron mechanisms beneath North Estonia [16,23].

The derivative grids define a composite structural pattern in which nested radial and arcuate gradients outline the internal roof collapse of the intrusive chamber [16], showing rims that mark contacts and ring-fault geometries. In contrast, lineament analysis reveals that the pluton is situated at the intersection of Riedel-type fabrics oriented NW-SE and NE-SW, which are linked to the PPDZ [2, 8, 16, 18]. Dense lineament clusters at the margins of Phases II and III correspond to reactivated strike-slip bands, whose organisation in R, R', and P geometries indicates dextral transtensional [16]. This shear regime enhances different deformation events, governs magma ascent pathways, facilitates roof fragmentation, and produces the NW-directed trapdoor behaviour observed in both derivative grids and inversion sections.

The combined geophysical and structural evidence supports a model in which Märljamaa-Kloostri was positioned as a steep, shear-guided magma chamber that developed ring-fault collapse and asymmetric subsidence along the PPDZ corridor. Its timing (1.67-1.53 Ga) places it within the AMCG superswell phase, characterised by mantle upwelling, slab stagnation, mafic underplating, and enhanced lithospheric heat flow along long-lived shear zones [3, 7, 16, 37-40]. These conditions favoured the generation of oxidised, volatile-rich A-type magmas. They facilitated the structural trapping of the pluton, with PPDZ reactivation providing both vertical permeability and lateral dilation required for cauldron-style subsidence. SimPEG sections showing a cylindrical high-susceptibility core at ~3 km (Fig. 3e) depth are consistent with a long-lived thermal-mechanical anchor and the progressive evolution of a resurgent cauldron [16,23].

4.3.2 Petrogenetic Evolution: Oxidised A-Type Signature and Trace-Element-REE Systematics

Major-element indices (Fig. 6) highlight the distribution of Estonian rapakivi intrusions relative to the Wiborg body, indicating a mostly oxidised character for the mainland suites and more reduced values in the Wiborg and Riga intrusions. Märljamaa shows the highest Ti-Fe-Mg-Ca-P contents among the bodies studied [30]. These signatures align with oxidised A-type magmatism (Fig. 6b) that crystallises near the NNO (nickel-nickel oxide) buffer [13, 16, 17, 23]. Märljamaa-Kloostri granitoids contain high levels of normative magnetite and titanite [45], along with biotite-hornblende assemblages that differentiate from the more reduced Finnish rapakivi Wiborg suite bodies (e.g., Wiborg, Bodom, Obbnäs, Ahvenisto), which are richer in fluorine yet less oxidised [7,13-17,39,44].

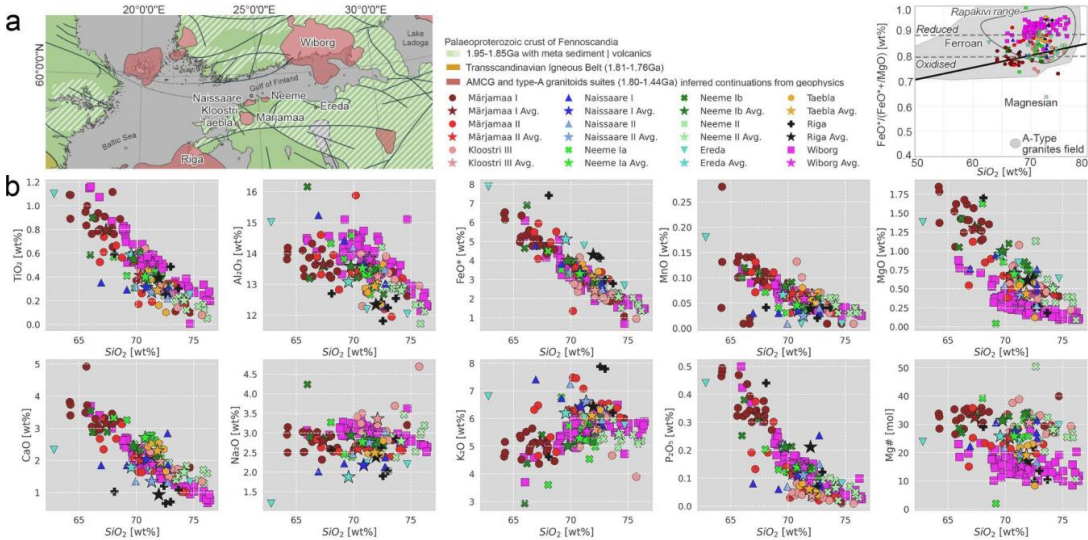


Figure 6: (a) Close-up of the Finnish-Estonian rapakivi bodies analysed, within Fennoscandia. **(b)** Hacker plots of the Wiborg suite rapakivi bodies and the Riga body. Estonian rapakivi samples were analysed by magmatic generation (see Klein et al. (1994) [45]). Estonian samples come from Kivisilla et al. (1999) [33] and Wiborg body data extracted from the GTK Rock Geochemical Database [44].

The trace-element and REE datasets (Fig. 7), reorganised by magmatic phase, reveal systematic differentiation throughout the intrusive sequence [16, 17, 23, 39]. Phase I is notably enriched in trace elements and LREE, with total REE content reaching approximately 3600 ppm in the most ferroan, silica-poor samples [16,17,23]. Phase II has intermediate compositions with peak K₂O, decreasing Fe-Ti-P, and declining oxide-phosphate modal fractions, while Phase III forms a Na-rich leucogranite with high SiO₂ and minimal Fe-Ti oxide content [16,17]. These trends suggest fractional crystallisation, varying degrees of crustal assimilation, and a change in oxygen fugacity during chamber development. All phases fall within the A₂-type granite field, characteristic of post-collisional to post-orogenic settings, where mantle-derived heating and crustal remelting interact within an intraplate, thick lithosphere environment [16, 17, 30, 45].

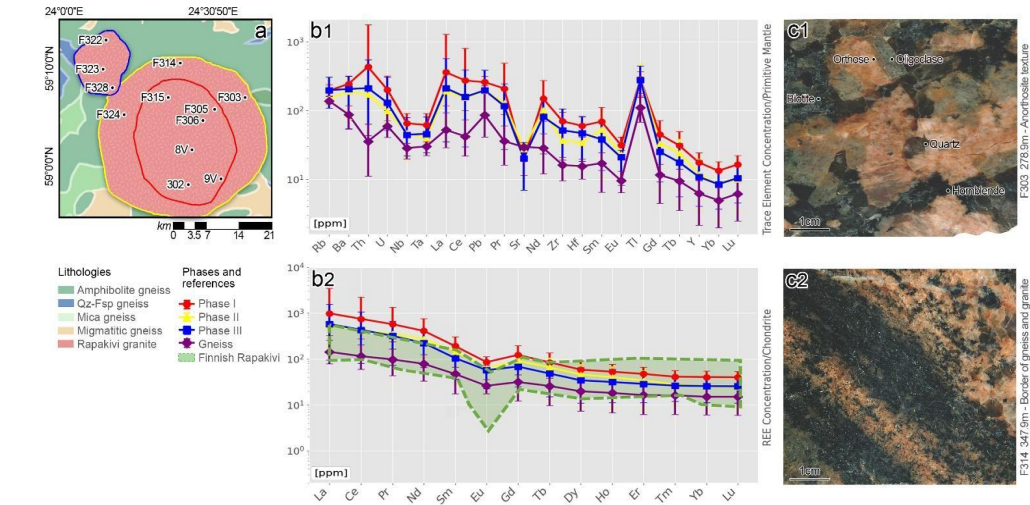


Figure 7: (a) Close-up of the Märdjamaa-Kloostri rapakivi bodies analysed, along with their respective phases. (b) Bivariate plots of the analysed Märdjamaa-Kloostri rapakivi bodies: (b1) trace and (b2) REE after Potagin (2024) [39] and Solano-Acosta et al. (2025) [30]. (c) Textural examples of (c1) rapakivi sample and (c2) interaction between rapakivi and gneiss host rock.

4.3.4 Metallogenic Zoning and Exploration

Similar rift and superswell environments to those that formed the Fennoscandian AMCG suite are common in SEDEX, stratabound Cu, and Fe Oxide-Cu-Au-U (IOCG) deposits, which involve lithospheric thinning, mantle upwelling, and magmatism with oxidized, volatile-rich fluids [9, 16, 37, 38, 45]. These systems originate from a fertilized lithospheric mantle and deep-sourced fluids, characterized by phases of F and S that enhance metal solubility, promoting wall-rock alteration and greisenization, which in turn form complex polymetallic assemblages [23, 27, 28, 32-34, 41, 42]. In southern Finland, rapakivi batholiths such as Wiborg, Åland, and Vehmaa, contain ores with magnetite-sphalerite, Zn-Cu-Pb-Ag-In veins, Sn-W-Be greisens, often in later phases (topaz granites, greisen caps, pegmatites) controlled by reactivated Svecofennian shear zones [12,17,19-21,29-31,36,41,42] (Fig. 3d3). In the Märdjamaa-Kloostri complex, SimPEG-derived density and susceptibility contrasts clearly distinguish phase-specific petrophysical signatures (Fig. 3e). In the X12-15 and Y15 zones of the northeastern sector (Fig. 3d), these anomalies suggest a structurally controlled subsurface feature that warrants targeted mineralogical and geochemical follow-up, potentially indicative of skarn-type alteration [16,17].

The Märdjamaa-Kloostri intrusion constitutes a high-potential metallogenic system within northern Estonia. Its oxidised A-type affinity, emplacement along a major shear zone, and pronounced REE-HFSE enrichment, particularly in Phase I (Fig. 7), are consistent with fertile AMCG magmatic provinces. Phase I ferroan granodiorites host the highest REE contents, abundant Fe-Ti oxides, phosphate-rich domains, and the strongest magnetic-gravity responses (Fig. 3d), defining the primary REE-HFSE investigation target. Phase II margins are associated with ring-faulting and shear reactivation, providing favourable pathways for hydrothermal fluid flow and potential greisenisation, REE enrichment, or Fe-oxide-apatite enrichment. Trapdoor structures in Phase III coincide with petrophysical anomalies indicative of late-stage fluid circulation. At a regional scale, superswell-driven extension and plume-related heating enhance the metallogenic fertility of the Wiborg Suite Estonian rapakivi intrusions, increasing the likelihood of concealed IOCG-like or HFSE-enriched A-type granite-related systems.

5. Conclusions

The Tallinn-Alutaguse-Jõhvi corridor defines a coherent arc-back-arc system, with Tallinn as the felsic arc front, Alutaguse as a base-metal-rich volcanic back-arc basin that represents the metallogenic core, and Jõhvi as a Fe-oxide-sulphide skarnised member. Elevated Cu-Pb-Zn enrichment in Alutaguse reflects sustained mafic intrusion and hydrothermal activity analogous to Bergslagen systems. Integrated geophysical signatures, including high susceptibility, low-density bodies, and strong remanent magnetisation, delineate this architecture and reveal the concealed geometry of mineralised zones. At the same time, MSCL-XYZ scanning and geochemical data identify multi-element associations. The High-Fe-Ti-P Märdjamaa-Kloostri granitoids exhibit oxidised A-type characteristics, HFSE-REE enrichment, and geophysical evidence for piston-cauldron emplacement along the PPDZ, thus consistent with Mesoproterozoic magmatism and localised host-rock alteration in the northeastern sector. Together, these results establish a robust crustal and metallogenic framework for northern Estonia, highlighting the concealed Precambrian basement as a significant target for CRM exploration and focused future studies.

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 Formal analysis: Juan David Solano Acosta and Sophie Graul.
 Investigation: Juan David Solano Acosta, Sophie Graul,
 Resources: Juan David Solano Acosta; Sophie Graul; Sirle Liivamägi; Johannes Vind
 Data curation: Juan David Solano Acosta and Sophie Graul.
 Writing-original draft preparation: Juan David Solano Acosta.
 Writing-review and editing: Juan David Solano Acosta and Sophie Graul.
 Visualisation: Juan David Solano Acosta and Sophie Graul.
 Supervision: Alvar Soesoo and Rutt Hints.
 Project administration: Sophie Graul, Johannes Vind.
 All authors have read and agreed to the published version of the manuscript.

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18. Solano-Acosta, J.D., Soesoo, A., & Hints, R. (2023). *New insights into the crustal structure across Estonia using satellite potential fields derived from WGM-2012 gravity data and EMAG2v3 magnetic data*. *Tectonophysics*, 846, 229656. <https://doi.org/10.1016/j.tecto.2022.229656>
19. Solano-Acosta, J.D., Soesoo, A., & Hints, R. (2025). *Geochemistry, provenance, and tectonic setting of Paleoproterozoic metasedimentary and metavolcanic units of the Estonian Alutaguse region, eastern Fennoscandia*. *Estonian Journal of Earth Sciences*, 74, 61–82. <https://doi.org/10.3176/earth.2025.05>
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21. Nirgi, S., & Soesoo, A. (2021). *Geology and geochemistry of a Paleoproterozoic iron mineralisation in North-Eastern Estonia*. *Proceedings of the Karelian Research Centre of the Russian Academy of Sciences*, (10), 25–43. <https://doi.org/10.17076/geo1492>
22. Solano Acosta, J.D., Soesoo, A., Hints, R., Graul, S., All, T., Vind, J., Maala, L., & Liivamägi, S. (2025). *Exploration of critical mineral resources in the North Estonian basement, Eastern Fennoscandia: Regional and local perspectives with analogies to the Southern Svecofennian Finnish terranes and the Swedish Bergslagen region*. 5th International Professional Geology Conference (IPGC 2025), Zaragoza, Spain. <https://doi.org/10.13140/RG.2.2.18847.01449/1>
23. Solano-Acosta, J.D. (2025). *Crustal structure, tectonics, and magmatic evolution of the Estonian Proterozoic basement: Geophysical and geochemical insights*. Doctoral thesis, Tallinn University of Technology.
24. Solano Acosta, J.D., Soesoo, A., Hints, R., & Graul, S. (2025). *Exploration of critical mineral resources in the North Estonian basement, Eastern Fennoscandia: A geochemical and geophysical investigation of a region akin to the Southern Svecofennian Finnish terranes and the Swedish Bergslagen Province*. *Goldschmidt 2025*, Prague. <https://doi.org/10.7185/gold2025.28444>
25. Nirgi, S., Maala, L., & Kont, R. (2024). *Kriitiliste toormete ja nendega asotsieeruvate elementide tuvastamine ning uuringupotentsiaali hindamine Eesti aluskorras*. *Geoloogiafond*. <https://doi.org/10.23679/1032>
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28. Beunk, F.F., & Kuipers, G. (2012). *The Bergslagen ore province, Sweden: Review and update of an accreted orocline, 1.9–1.8 Ga BP*. *Precambrian Research*, 216, 95–119. <https://doi.org/10.1016/j.precamres.2012.05.007>
29. Soosalu, H., Uski, M., Komminaho, K., & Veski, A. (2022). *Recent intraplate seismicity in Estonia, East European Platform*. *Seismological Research Letters*, 93, 1800–1811. <https://doi.org/10.1785/0220210277>
30. Solano Acosta, J.D., Soesoo, A., Hints, R., & Sula, C. (2025). *Oxidised rapakivi magmatism and crustal architecture in the earliest AMCG suite: Geochemical and geophysical insights from Estonian intrusions along the Paldiski–Pskov Shear Zone (PPDZ), Wiborg-Suite Complex, Fennoscandia*. 5th International Professional Geology Conference (IPGC 2025), Zaragoza. <https://doi.org/10.48726/1mv94-7qc33>
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32. Plado, J., Kiik, K., Jokinen, J., & Soesoo, A. (2020). *Magnetic anomaly of the Jõhvi iron ore, northeastern Estonia, controlled by subvertical remanent magnetisation*. *Estonian Journal of Earth Sciences*, 69(4), 189–199. <https://doi.org/10.3176/earth.2020.13>
33. Kivisilla, J., Niin, M., & Koppelman, H. (1999). *Catalogue of chemical analyses of major elements in the rocks of the crystalline basement of Estonia*. Eesti Geoloogiakeskus. <https://kirjandus.geoloogia.info/en/reference/21247>
34. Kähkönen, Y. (2005). *Svecofennian supracrustal rocks*. In *Developments in Precambrian Geology*, Vol. 14, 343–405. [https://doi.org/10.1016/S0166-2635\(05\)80009-X](https://doi.org/10.1016/S0166-2635(05)80009-X)
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36. Kara, J. (2021). *Evolution of the Svecofennian Bedrock in Southern Finland – Spatial and temporal changes in the mantle-derived magmatism and mantle-crust interaction*. Doctoral thesis, University of Turku. <http://urn.fi/URN:ISBN:978-951-29-8474-9>
37. Pirajno, F., & Santosh, M. (2015). *Mantle plumes, supercontinents, intracontinental rifting, and mineral systems*. *Precambrian Research*, 259, 114–142. <https://doi.org/10.1016/j.precamres.2014.12.016>
38. Rämö, O.T., & Haapala, I. (2005). *Rapakivi granites*. In *Developments in Precambrian Geology*, Vol. 14, 533–562. [https://doi.org/10.1016/S0166-2635\(05\)80013-1](https://doi.org/10.1016/S0166-2635(05)80013-1)
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40. Bonin, B. (2007). *A-type granites and related rocks: evolution of a concept, problems and prospects*. *Lithos*, 97, 1–29. <https://doi.org/10.1016/j.lithos.2006.12.007>
41. Karell, F., Ehlers, C., Airo, M.-L., & Selonen, O. (2009). *Intrusion mechanisms and magnetic fabrics of the Vehmaa rapakivi granite batholith in SW Finland*. *Geotectonic Research*, 96, 53–68. <https://doi.org/10.1127/1864-5658/09/96-0053>
42. Nordbäck, N., Skyttä, P., Engström, J., Ovaskainen, N., Mattila, J., & Aaltonen, I. (2024). *Mesoproterozoic strike-slip faulting within the Åland rapakivi batholith, southwestern Finland*. *Tektonika*, 2(1), 1–26. <https://doi.org/10.55575/tektonika2024.2.1.51>
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515 Estonian basement. *Proceedings of the Estonian Academy of Sciences: Geology*, 43(4), 165–176. Estonian Academy Publishers.
516 <https://doi.org/10.3176/geol.1994.4.01>

Curriculum Vitae

1. Personal data

Name	Juan David Solano Acosta
Date and place of birth	13 September 1995, Bogotá, Colombia
Nationality	Colombian

2. Contact information

Address	Tallinn University of Technology, School of Science, Department of Geology, Ehitajate tee 5, 19086 Tallinn, Estonia
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3. Education

- **09/2021 – 01/2026** — *PhD, Precambrian and Economic Geology*, Tallinn University of Technology, Estonia.
- **04/2023 – 02/2024** — *Specialization in Data Analysis*, Turing College (Remote, Vilnius, Lithuania).
Ranked top 5% (full scholarship). Focus on advanced data analytics, statistics, and interpretation of complex datasets. [Certificate link](#).
- **09/2021 – 01/2022** — *Specialization in GIS, Mapping, and Spatial Analysis*, University of Toronto (Remote).
GIS and remote sensing for interactive mapping. Certificate ID: [UXW4N664KCXL](#).
- **10/2020 – 03/2022** — *MSc in Geophysics and Meteorology*, University of Granada, Spain.
- **10/2019 – 07/2021** — *MSc in Geology Applied to Mineral and Energy Resources*, University of Granada, Spain.
- **01/2014 – 06/2019** — *BSc in Earth Sciences*, Los Andes University, Colombia.

4. Language competence

Spanish	native
English	fluent

5. Professional employment

- **2024–2027** — *Research Fellow*, Tallinn University of Technology, EU Horizon DEX-PLORE Project (Estonia).
Advancing the understanding of Estonia's potential for deep-seated critical raw materials (CRM) through integrated geophysical–geochemical modelling and mineral system analysis.
- **2021** — *Junior Geophysicist (Internship)*, Andalusian Institute of Geophysics and Prevention of Earthquake Disasters, Granada, Spain.
Analysed earthquake intensities and produced regional seismic intensity maps for southern Spain.

- **2020** — *Junior Geologist (Internship)*, Minera de Órgiva S.L., Órgiva, Spain. Conducted fieldwork and sampling at the Sierra de Lújar F–Pb–(Zn) deposit; prepared fluorite samples for XRF geochemical analysis.
- **2019** — *Research Assistant*, Los Andes University, Bogotá, Colombia. Participated in geological mapping, gravimetric (EGM-2008) data analysis, and field sampling for the Antioquian Batholith project under Dr. Ana Ibis Despaigne.
- **2018–2019** — *Junior Geologist (Exploration, Remote Sensing & GIS)*, Target Exploration S.A.S. / Pacífico Minerals (ASX:PMY), Colombia. Created geological maps, conducted spectral reconnaissance, and performed field sampling during Cu–Au–(Ag) porphyry and epithermal exploration projects in Urrao and Natagaima.

6. Computer and Technical Skills

- **General Skills:** Data Analysis, Report Preparation, Research Conducting, Exploration Completion, Diligence, Quick Learning.
- **Programming Languages:** Python, SQL, MATLAB, R.
- **Data and Analytical Tools:** Microsoft Office Suite, Power BI, Tableau, $\text{\LaTeX}$.
- **Geoscientific and GIS Software:** ArcGIS, Oasis Montaj, Leapfrog, QGIS, PCI Geomatics, ENVI, Surfer, Geoscience ANALYST, SimPEG, GemPy, SplitLab.

7. Honours and awards

- 2019, Finalist — *ESRI Young Scholars Award Colombia*.
- 2023–2024, Full scholarship — Specialization in Data Analysis, Turing College.
- 2025, Micro-degree — *Sustainable Exploration, Mining, and Circular Economy of Mineral Resources*, TalTech.

8. Defended theses

- 2021, MSc (Geology applied to Mineral and Energy Resources), University of Granada, Spain.
Thesis: Geochemical characterisation of fluorite from the F–Pb–(Zn) deposit of Sierra Lújar (Órgiva, Granada). <http://dx.doi.org/10.6084/m9.figshare.22047302.v1>.
- 2022, MSc (Geophysics and Meteorology), University of Granada, Spain.
Thesis: Seismic anisotropy beneath the Canary Islands from SKS/SKKS shear-wave splitting. <http://dx.doi.org/10.13140/RG.2.2.23154.79041>.
- 2019, BSc (Earth Sciences), Los Andes University, Colombia.
Thesis: Detection and analysis of morphotectonic features utilising SRTM 90m and GIS in Alta Guajira, Colombia. <http://hdl.handle.net/1992/45715>.

9. Field of research

- Precambrian geology, geophysics, and economic geology; critical raw materials (CRM).
- Integrated geophysical inversion and modelling; gravity and magnetic methods.
- Whole-rock geochemistry and mineral system analysis; sustainable exploration.

10. Scientific work

Publications

- **Solano-Acosta, J.D.**, Soesoo, A., Hints, R. (2025). Integrated geophysical and emplacement modelling of the Märjamaa and Kloostri rapakivi granitoids, Estonia: Insights into intrusion geometry and tectonic controls. *Precambrian Research* 430, 107938.
DOI:[doi:10.1016/j.precamres.2025.107938](https://doi.org/10.1016/j.precamres.2025.107938).
- **Solano-Acosta, J.D.**, Soesoo, A., Hints, R. (2025). Geochemistry, provenance, and tectonic setting of Paleoproterozoic metasedimentary and metavolcanic units of the Estonian Alutaguse region, eastern Fennoscandia. *Estonian Journal of Earth Sciences* 74, 61–82.
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- Senger, K., Ammerlaan, F., Betlem, P., Brönnner, M., Dumais, M.A., Gellein, J., Henningsen, T., Janocha, J., Johannessen, E.P., Liebsch, J., **Solano-Acosta, J.D.**, et al. (2025). Crustal heterogeneity onshore central Spitsbergen: Insights from new gravity and vintage geophysical data. *Geochemistry, Geophysics, Geosystems* 26.
DOI:[doi:10.1029/2024GC011563](https://doi.org/10.1029/2024GC011563).
- **Solano-Acosta, J.D.**, Soesoo, A., Hints, R. (2023). New insights of the crustal structure across Estonia using satellite potential fields derived from WGM-2012 gravity data and EMAG2v3 magnetic data. *Tectonophysics* 229656.
DOI: [doi:10.1016/j.tecto.2022.229656](https://doi.org/10.1016/j.tecto.2022.229656).
- Rodríguez-Corcho, A.F., Rojas-Agramonte, Y., **Solano-Acosta, J.D.**, Bayona, G., Cardona-Molina, A., Eglinton, B., Montes, C. (2021). The Colombian geochronological database (CGD). *International Geology Review*.
DOI:[doi:10.1080/00206814.2021.1954556](https://doi.org/10.1080/00206814.2021.1954556).
- **Solano-Acosta, J.D.**, Despaigne, A.I., Pearse, J. (2020). Morphotectonic analysis of Upper Guajira, Colombia: A GIS and Remote Sensing approach. *MDPI Preprints Research Letters*.
DOI:[doi:10.20944/preprints202010.0476.v1](https://doi.org/10.20944/preprints202010.0476.v1).

Conference Presentations

1. Juan David Solano Acosta, Alvar Soesoo, Rutt Hints, Sophie Graul, Tarmo All, Johannes Vind, Lennart Maala, Sirle Liivamägi. Exploration of Critical Mineral Resources in the North Estonian Basement, Eastern Fennoscandia: Regional and Local Perspectives with Analogies to the Southern Svecofennian Finnish Terranes and the Swedish Bergslagen Region. 5th International Professional Geology Conference (IPGC 2025), November 2025, Zaragoza, Spain.
DOI:[10.13140/RG.2.2.18847.01449/1](https://doi.org/10.13140/RG.2.2.18847.01449/1).
2. Juan David Solano Acosta, Alvar Soesoo, Rutt Hints, Carina Sula. Oxidised Rapakivi Magmatism and Crustal Architecture in the Earliest AMCG Suite: Geochemical and Geophysical Insights from Estonian Intrusions along the Paldiski–Pskov Shear Zone (PPDZ), Wiborg-Suite Complex, Fennoscandia. 5th International Professional Geology Conference (IPGC 2025), November 2025, Zaragoza, Spain.
DOI: [10.48726/1mv94-7qc33](https://doi.org/10.48726/1mv94-7qc33).

3. Juan David Solano Acosta, Alvar Soesoo, Rutt Hints, Sophie Graul. *Exploration of Critical Mineral Resources in the North Estonian Basement, Eastern Fennoscandia: A Geochemical and Geophysical Investigation of a Region Akin to the Southern Svecofennian Finnish Terranes and the Swedish Bergslagen Province*. Goldschmidt 2025, July 2025, Prague, Czech Republic.
DOI: [10.7185/gold2025.28444](https://doi.org/10.7185/gold2025.28444).
4. Juan David Solano Acosta, Alvar Soesoo, Rutt Hints, Sophie Graul. *Is the Estonian Alutaguse Section of Eastern Fennoscandia a Continuation of the Southern Svecofennian Finnish Terranes, or is it Akin to the Swedish Bergslagen Region?* EGU General Assembly 2025, April 2025, Vienna, Austria.
DOI: [10.5194/egusphere-egu25-7212](https://doi.org/10.5194/egusphere-egu25-7212).
5. Alvar Soesoo, Siim Nirgi, Juan David Solano Acosta. *Critical Raw Materials of the Estonian Precambrian Basement: Any New Targets?* Thirteenth Symposium on the Structure, Composition and Evolution of the Lithosphere, November 2024, Espoo, Finland.
6. Sophie Graul, Siim Nirgi, Alvar Soesoo, Juan David Solano Acosta, Rutt Hints, Triinu Hand, Camilo Hernandez-Moreno. *DEXPLORE Horizon EU Project: Investigation of CRM Potential of Estonian Basement*. Lithosphere 2024 Symposium, November 12–14, 2024, Espoo, Finland.
DOI: [10.13140/RG.2.2.18365.47840](https://doi.org/10.13140/RG.2.2.18365.47840).
7. Juan David Solano Acosta, Alvar Soesoo, Rutt Hints, Sophie Graul. *Geophysical and Geochemical Characterisation of the Märjamaa Rapakivi Granite, Estonia*. Lithosphere 2024, November 2024.
DOI: [10.13140/RG.2.2.27044.08326](https://doi.org/10.13140/RG.2.2.27044.08326).
8. Juan David Solano Acosta, Rutt Hints, Alvar Soesoo. *Insights on the Tectonic Styles Across Estonia Using Satellite Potential Fields Derived from WGM-2012 Gravity Data and EMAG2 Magnetic Data*. EGU General Assembly 2022, May 2022, Vienna, Austria.
DOI: [10.5194/egusphere-egu22-11398](https://doi.org/10.5194/egusphere-egu22-11398).

11. Courses and extracurricular activities

- Finalist in the contest '[Young Scholars Award Colombia 2019](#)' developed by the company ESRI, for the work 'Morphotectonic Analysis of Upper Guajira: A GIS and Remote Sensing Approach' - (Bogotá, Colombia) 2019.
- Apprenticeship in the [arQus Challenge-Based Program "Rethinking Climate Change"](#), endorsed by The Arqus European University Alliance - 2020-2021.
- Course '[Innovative Exploration Drilling and Data Acquisition Research School - Entrepreneurship for Scientists](#)', endorsed by the EIT RawMaterials Academy - (Online) 2021.
- Participation in the [Mining summer school 'TIMREX'](#), endorsed by EIT Raw Materials Academy and the University of Miskolc - (Telkibánya, Hungary), 2022.
- Participation in the [Digital manufacturing: Multidisciplinary, holistic product view - from molecules to products and back](#), known as the 'Digiman' summer school, endorsed by NTNU, TalTech, and ProfMOF AS. Project ID funded by EEA/Norway higher education cooperation program - (Kohtla-Järve, Estonia), 2022.
- Participation in the [Sustainability summer school 'SustainMV'](#), organized by all universities in Mecklenburg-Western Pomerania - (Germany), 2022.

- Participation in the course [Arctic Tectonics and Volcanism](#), organized by the University Centre in Svalbard (UNIS) - (Svalbard), 2022.
- Participation in the ADMA3 course [Stakeholders of sustainable development](#), organized by the University of Oulu and EIT RawMaterials, 2023.
- Participation in the [Open Innovation Challenge for Young Talents with Atlantic Copper for sustainable solutions](#), organized by Atlantic Copper, University of Huelva and EIT RawMaterials, in Huelva (Spain) 2024.
- Micro-Degree [Sustainable Exploration, Mining, and Circular Economy of Mineral Resources](#), endorsed by TalTech – Tallinn University of Technology, 2025.

Elulookirjeldus

1. Isikuandmed

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Sünniaeg ja -koht	13. september 1995, Bogotá, Colombia
Kodakondsus	Colombia

2. Kontaktandmed

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3. Haridus

- **09/2021 – 01/2026** — *PhD, proterosoikumi ja majandusgeoloogia*, Tallinna Tehnikaülikool, Eesti.
- **04/2023 – 02/2024** — *Andmeanalüüsi spetsialiseerumine*, Turing College (kaugõpe, Vilnius, Leedu).
Vastuvõetute parim 5% (täisstipendium). Keskenduti arenenud andmeanalüütikale, statistikule ja suurandmete tõlgendamisele. [Tõendi link](#).
- **09/2021 – 01/2022** — *GIS-i, kaardistamise ja ruumianalüüsi spetsialiseerumine*, Toronto Ülikool (kaugõpe).
GIS ja kaugseire interaktiivsete kaartide loomiseks. Tunnistuse ID: [UXW4N664KCXL](#).
- **10/2020 – 03/2022** — *MSc geofüüsikas ja meteoroloogias*, Granada Ülikool, Hispaania.
- **10/2019 – 07/2021** — *MSc geoloogias (maavarad ja energiaressursid)*, Granada Ülikool, Hispaania.
- **01/2014 – 06/2019** — *BSc maateadustes*, Los Andese Ülikool, Colombia.

4. Keelteoskus

hispaania keel	emakeel
inglise keel	kõrgtase

5. Teenistuskäik

- **2024–2027** — *Teadur (uurijastipendiaat)*, Tallinna Tehnikaülikool, EL Horisont DEXPLORE projekt (Eesti).
Eesti sügaval paiknevate kriitiliste toormete (CRM) potentsiaali uurimine integreeritud geofüüsika-geokeemia modelleerimise ja mineraalsüsteemide analüüsi abil.
- **2021** — *Noorem geofüüsik (praktika)*, Andalusia Geofüüsika Instituut ja Maavärinate Ennetuse Keskus, Granada, Hispaania.
Maavärinate intensiivsuste analüüs ja Lõuna-Hispaania regionaalsete intensiivsuskartide koostamine.
- **2020** — *Noorem geoloog (praktika)*, Minera de Órgiva S.L., Órgiva, Hispaania.
Välitööd ja proovivõtt Sierra de Lújari F–Pb–(Zn) leiukohal; proovide ettevalmistus XRF-analüüsiks.
- **2019** — *Teadusassistent*, Los Andes Ülikool, Bogotá, Colombia.
Geoloogiline kaardistamine, gravimeetria (EGM-2008) andmeanalüüs ning proovivõtt Antioquia batoliidi projektis (juhendaja dr Ana Ibis Despaigne).
- **2018–2019** — *Noorem geoloog (otsingud, kaugseire & GIS)*, Target Exploration S.A.S. / Pacifico Minerals (ASX:PMY), Colombia.
Geoloogilised kaardid, spektraalluure ja proovivõtt Cu–Au–(Ag) portföürsete ja epotermiliste protsektide (Urrao, Natagaima) uuringutes.

6. Arvuti- ja tehnilised oskused

- **Üldoskused:** andmeanalüüs, aruannete koostamine, uurimistöö läbiviimine, otsingutöö lõpetamine, põhjalikkus, kiire õppimisvõime.
- **Programmeerimiskeeled:** Python, SQL, MATLAB, R.
- **Andme- ja analüüsitööriistad:** Microsoft Office, Power BI, Tableau, ~~Excel~~.
- **Geoteadus- ja GIS-tarkvara:** ArcGIS, Oasis Montaj, Leapfrog, QGIS, PCI Geomatics, ENVI, Surfer, Geoscience ANALYST, SimPEG, GemPy, SplitLab.

7. Autasud ja tunnustused

- 2019 — finalisti staatus, *ESRI Young Scholars Award Colombia*.
- 2023–2024 — täisstipendium, andmeanalüüsi spetsialiseerumine (Turing College).
- 2025 — mikrokraad *Säästev maavarade otsing, kaevandamine ja ringmajandus* (TalTech).

8. Kaitstud lõputööd

- 2021 — MSc (geoloogia: maavarad ja energiaressursid), Granada Ülikool, Hispaania. Lõputöö: fluoriidi geokeemiline iseloomustus Sierra de Lújari F-Pb-(Zn) leiukohal (Órgiva, Granada). <http://dx.doi.org/10.6084/m9.figshare.22047302.v1>.
- 2022 — MSc (geofüüsika ja meteoroloogia), Granada Ülikool, Hispaania. Lõputöö: SKS/SKKS nihkelainete kaheksustumise analüüsist lähtuv Kanaari saarte aluspõhja seismitiline anisotroopia. <http://dx.doi.org/10.13140/RG.2.2.23154.79041>.
- 2019 — BSc (maateadused), Los Andese Ülikool, Colombia. Lõputöö: morfoloogiliste struktuuride tuvastamine ja analüüs (SRTM 90 m, GIS) — näide Ülem-Guajirast, Colombiast. <http://hdl.handle.net/1992/45715>.

9. Teadustöö põhisuunad

- Proterosoikumi geoloogia, geofüüsika ja majandusgeoloogia; kriitilised toormed (CRM).
- Integreeritud geofüüsikaline inversioon ja modelleerimine; gravitatsiooni- ja magnetmeetodid.
- Täiskoostise geokeemia ja mineraalsüsteemide analüüs; säästev otsing.

10. Teadustegevus

Artiklid

- **Solano-Acosta, J.D.**, Soesoo, A., Hints, R. (2025). Integrated geophysical and emplacement modelling of the Märjamaa and Kloostri rapakivi granitoids, Estonia: Insights into intrusion geometry and tectonic controls. *Precambrian Research* 430, 107938. DOI:[doi:10.1016/j.precamres.2025.107938](https://doi.org/10.1016/j.precamres.2025.107938).
- **Solano-Acosta, J.D.**, Soesoo, A., Hints, R. (2025). Geochemistry, provenance, and tectonic setting of Paleoproterozoic metasedimentary and metavolcanic units of the Estonian Alutaguse region, eastern Fennoscandia. *Estonian Journal of Earth Sciences* 74, 61–82. DOI:[doi:10.3176/earth.2025.05](https://doi.org/10.3176/earth.2025.05).
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- **Solano-Acosta, J.D.**, Despaigne, A.I., Pearce, J. (2020). Morphotectonic analysis of Upper Guajira, Colombia: A GIS and Remote Sensing approach. *MDPI Preprints Research Letters*. DOI:[doi:10.20944/preprints202010.0476.v1](https://doi.org/10.20944/preprints202010.0476.v1).

Konverentsiettekanded

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2. Juan David Solano Acosta, Alvar Soesoo, Rutt Hints, Carina Sula. Oksüdeerunud rapakiivimagmatism ja kooriku ehitus AMCG kompleksi varaseimas arengufaasis: geokeemilised ja geofüüsikalised tulemused Paldiski–Pihkva nihkevööndi (PPDZ) äärsetest Eesti intrusioonidest, Wiborgi koosseisus, Fennoskandia. 5. Rahvusvaheline Professionaalse Geoloogia Konverents (IPGC 2025), november 2025, Zaragoza, Hispaania. DOI: [10.48726/1mv94-7qc33](https://doi.org/10.48726/1mv94-7qc33).
3. Juan David Solano Acosta, Alvar Soesoo, Rutt Hints, Sophie Graul. *Exploration of Critical Mineral Resources in the North Estonian Basement, Eastern Fennoscandia: A Geochemical and Geophysical Investigation of a Region Akin to the Southern Svecofennian Finnish Terranes and the Swedish Bergslagen Province*. Goldschmidt 2025, juuli 2025, Praha, Tšehhi. DOI: [10.7185/gold2025.28444](https://doi.org/10.7185/gold2025.28444).
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5. Alvar Soesoo, Siim Nirgi, Juan David Solano Acosta. *Critical Raw Materials of the Estonian Precambrian Basement: Any New Targets?* 13. litosfääri struktuuri, koostise ja arengu sümposium, november 2024, Espoo, Soome.
6. Sophie Graul, Siim Nirgi, Alvar Soesoo, Juan David Solano Acosta, Rutt Hints, Triinu Hand, Camilo Hernández-Moreno. *DEXPLORE Horizon EU Project: Investigation of CRM Potential of Estonian Basement*. Lithosphere 2024 sümposium, 12.–14. november 2024, Espoo, Soome. DOI: [10.13140/RG.2.2.18365.47840](https://doi.org/10.13140/RG.2.2.18365.47840).
7. Juan David Solano Acosta, Alvar Soesoo, Rutt Hints, Sophie Graul. *Geophysical and Geochemical Characterisation of the Märjamaa Rapakivi Granite, Estonia*. Lithosphere 2024, november 2024, Espoo, Soome. DOI: [10.13140/RG.2.2.27044.08326](https://doi.org/10.13140/RG.2.2.27044.08326).
8. Juan David Solano Acosta, Rutt Hints, Alvar Soesoo. *Insights on the Tectonic Styles Across Estonia Using Satellite Potential Fields Derived from WGM-2012 Gravity Data and EMAG2 Magnetic Data*. EGU General Assembly 2022, mai 2022, Viin, Austria. DOI: <https://doi.org/10.5194/egusphere-egu22-11398>.

11. Kursused ja lisategevused

- **Young Scholars Award Colombia 2019** (ESRI) — finalist staatus tööga „Morphotectonic Analysis of Upper Guajira: A GIS and Remote Sensing Approach“; Bogotá, Kolumbia (2019).
- Öpipoisiõpe programmis **arQus Challenge-Based Program “Rethinking Climate Change”** (Arqus Euroopa Ülikoolide Liit), 2020–2021.

- Osalemine [suveülikoolis „TIMREX“](#) (EIT RawMaterials, Miskolci Ülikool) — Telkibánya, Ungari (2022).
- Osalemine [suvekoolis „SustainMV“](#) (Saksamaa Mecklenburg-Ees-Pommeri ülikoolid), 2022.
- Kursus [„Arctic Tectonics and Volcanism“](#) (UNIS, Svalbard), 2022.
- Kursus ADMA3 [„Stakeholders of Sustainable Development“](#) (Oulu Ülikool & EIT RawMaterials), 2023.
- Osalemine [Open Innovation Challenge for Young Talents with Atlantic Copper](#) — Huelva, Hispaania (2024).
- Mikrokraad [„Sustainable Exploration, Mining, and Circular Economy of Mineral Resources“](#) (TalTech), 2025.

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