

Deep structure and geodynamics of the Sarmatia-Fennoscandia transition zone based on geology and seismic tomography

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Received 24 November 2025

We examined the geological structure of the crust and mantle within the transition zone between the Sarmatia and Fennoscandia miniplates. The zone encompasses the central and southern parts of the Svecofennian orogen and the northwestern segment of the Ukrainian Shield. It is extended south-eastward to include several northeast-striking fault zones of the Ukrainian Shield (from the Horyn to the Teteriv fault zones, inclusive). The interpretation of the crustal structure is based on recent publications by Swedish, Polish, Estonian, Lithuanian, and Ukrainian scientists. The mantle to depths of 850—2500 km was investigated using a three-dimensional velocity model of Eurasia, developed by V.S. Geyko at the S. Subbotin Institute of Geophysics of the NAS of Ukraine based on seismic tomography using the Taylor approximation to the eikonal equation and the wave equation.

It is shown that the Sarmatia-Fennoscandia transition formed in the Paleoproterozoic (2.10—1.75 Ga) through complex geodynamic processes evidenced by multiple subducted mantle slabs. A key process was the subduction of Fennoscandia beneath Sarmatia, recorded by a south-dipping slab from the Keitele microcontinent beneath the Bergslagen microcontinent and the Mid-Baltic Belt, and by an east- to southeast-dipping slab beneath Sarmatia from the Belarus-Podlasie Granulite Belt beneath the Osnytsk-Mikashevychi Igneous Belt and the area of the present-day Korosten Pluton.

Subduction occurred with interruptions, one of which (ca. 1.89—1.84 Ga) coincided with formation of the Baltic part of the Svecofennian orogen via additional subduction of opposite polarity (north and northeast), as confirmed by slabs from the Amberland microcontinent beneath the Bergslagen microcontinent and from the Central Finland Arc Complex beneath the Karelian craton.

Thus, reconciling upper-mantle seismic tomography observations with the geological seismic tomography constraints on crustal evolution provides compelling evidence for a plate-tectonic origin of the processes that led to the formation of the East European Craton.

Key words: Sarmatia, Fennoscandia, Svecofennian orogen, Ukrainian Shield, transition zone, mantle, subduction.

Introduction. Modern geodynamics' core objective is a study of zones of convergence and divergence of lithospheric plates, continents, and terranes, as well as the collisional

structures and rock complexes formed in these settings.

Ukraine occupies a large area of the southwestern East European Craton (EEC). The

Citation: Gintov, O.B., Mychak, S.V., & Babynin, O.O. (2026). Deep structure and geodynamics of the Sarmatia-Fennoscandia transition zone based on geology and seismic tomography. *Geofizychnyi Zhurnal*, 48(1), 96—109. <https://doi.org/10.24028/gj.v48i1.344412>.

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area includes the Sarmatia miniplate and its convergence zones with several neighbouring tectonic entities: the West European Plate to the west and southwest, the Scythian Plate and the Black Sea-Azov region to the south, and the Fennoscandian miniplate to the northwest (see Fig. 1 in [Starostenko et al., 2024a]). While the transition zones between Sarmatia and the West European Plate, as well as the Scythian Plate and the Black Sea-Azov region, are comparatively well studied at both crustal and mantle levels [Starostenko, Gintov, 2018], the deep — particularly mantle — structure of the ancient convergence zone between the Sarmatia and Fennoscandia miniplates (hereafter SFCZ) remains insufficiently constrained. This issue is addressed only in part by [Starostenko et al., 2024b].

Previous 3D mantle velocity models for Sarmatia [Geyko et al., 2005; Tsvetkova et al., 2016, 2021] and for Fennoscandia [Tsvetkova et al., 2009, 2010, 2015; Bugaenko et al., 2015; Tsvetkova, Bugaenko, 2016] were derived from seismic tomography using a Taylor approximation to the eikonal and wave equations [Geyko, 2004; Tsvetkova, 2015]. They describe the mantle structure to 850—1700 km depth. However, the SFCZ itself lies outside those studies — as already noted by [Tsvetkova et al., 2010] — and has not subsequently been treated as an independent research target. This gap motivates the present paper.

Study area and geological setting according to recent literature data. In Fig. 1, the region is bounded by 63° N (north), 33° E (east), 50° N (south), the Trans-European Suture Zone (TESZ) to the southwest, and 18° E to the west. It includes the present-day northwestern part of Sarmatia (the northwestern edge of the Ukrainian Shield), the Belarus-Podlasie Granulite Belt (BPG), the Okolovo terrane (OKL), and a significant portion of the Svecofennian orogen, which was traced south of Finland and east of Sweden into the South Baltic region by [Gorbatshev, Bogdanova, 1993]. A detailed trans-Baltic correlation is provided by [Bogdanova et al., 2015]. Following [Baltybaev, 2013], the orogen can be subdivided into an Outer (northern) and

Inner zones separated approximately along 61° N. The Inner zone includes nearly all of the southern (Baltic) Svecofennian and lacks Archean rocks. To the Outer zone belong the Bothnian microcontinent, the Central Finland Arc Complex, and the Keitele microcontinent. The northwestern Ukrainian Shield within the SFCZ comprises Palaeoproterozoic deep fault zones of northeast strike azimuth — Horyn, Lutsk (Strelsk), Sushchany-Perha, and Teteriv — by which we extend the zone ~200 km to the southeast. The overall dimensions of the SFCZ are ~1400×900 km.

Outside the Ukrainian Shield and its flank, the crystalline basement of the SFCZ is covered by a Meso- to Neoproterozoic and Phanerozoic sedimentary cover exceeding 1—2 km. Key investigative tools are deep drilling, geochronology, and geophysical methods. The basement consists predominantly of Palaeoproterozoic (2.0—1.75 Ga) granulite (both magmatic and sedimentary) and amphibolite complexes: charnockitoids, tonalitic (north) and potassic (south) migmatites, granodiorites, calc-alkaline granite plutons, metagabbroids, metabasalts, metadiorites, meta-effusives, etc. Granulite complexes occur within the BPG (Sarmatia) and in Latgalia, West and East Estonia, and elsewhere in the Baltic Svecofennian. The remaining territory beneath the thick sedimentary cover is underlain by amphibolite-facies metabasites, gneisses, and granitoids. Within the Ukrainian Shield, there have developed gneisses of the Teteriv Series, Archean and Palaeoproterozoic granitoids (including rapakivi), and sedimentary rocks of the Ovruch and Prypiat basins. Numerous fault zones in the SFCZ host mylonites slightly younger than their country rocks. At the eastern margin of the study area, at the boundary between the OKL and the Osnytsk-Mikashevychi Igneous Belt (OMIB) near 56°N/32°E, lies the Slabodka tectono-geodynamic node [Garetsky, Karataev, 2011]; its nature remains controversial and is not considered here.

The Baltic Svecofennian within the study area comprises the Bothnia, Keitele, Bergslagen, Livonia, and Amberland Domains, which have dimensions comparable to the Ukrainian

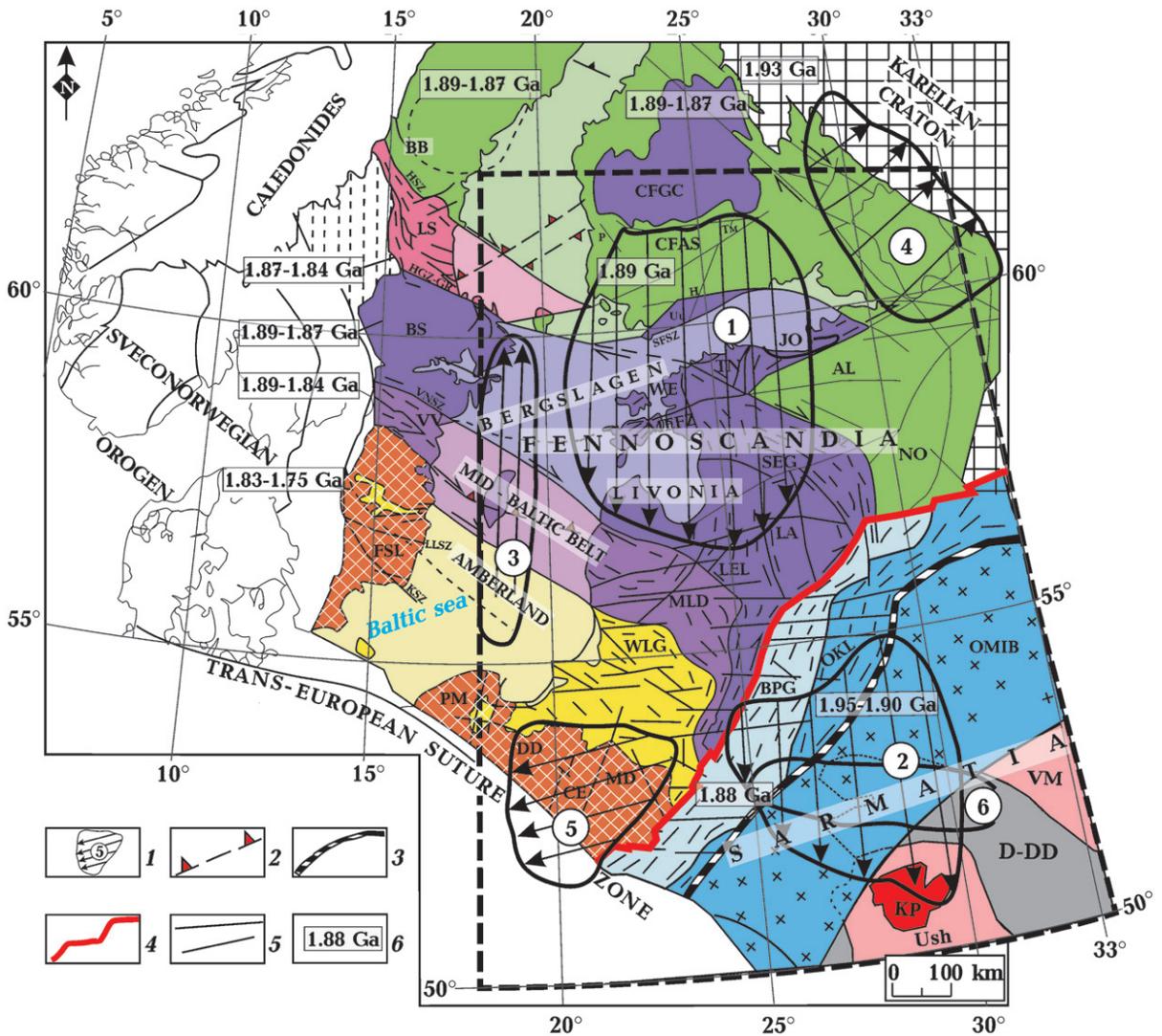


Fig. 1. Tectonic scheme of the central and southern Svecofennian orogen integrated across the Baltic Sea (after [Bogdanova et al., 2015], modified): 1 — subduction slabs (arrows indicate the direction of subduction); 2 — upper contours of dipping mantle reflectors, according to [Lahtinen et al., 2009]; 3 — the former boundary between Fennoscandia and Sarmatia, after [Bogdanova et al., 1996 a, b]; 4 — the current boundary between Fennoscandia and Sarmatia, after [Bogdanova et al., 2015]; 5 — deformation zones, faults; 6 — ages in boxes indicating the timing of major accretionary events in the regions.

Abbreviations for tectonic domains: AL — Alutaguse, AMBERLAND (Amberland microcontinent), BB — BOTHNIAN (Bothnia microcontinent), BS — BERGSLAGEN (Bergslagen microcontinent), BPG — Belarus-Podlasie Granulite Belt, CE — Ciechanow, CFAS — Central Finland Arc Complex, CFGC — Central Finland Granitoid Complex (KB — KEITELE microcontinent), DO — Dobrzyn, ESL — East Småland, JO — Jöhvi, LA — Latgalia, LEL — Latvian-East Lithuanian, LIVONIA microcontinent, LS — Ljusdal, MD — Mazowsze, MLD — Mid-Lithuanian domain, NO — Novgorod, OKL — Okolovo terrane, OMIB — Osnytsk-Mikashevychi Igneous Belt, PM — Pomorze, SEG — South Estonian granulite domain, TN — Tallinn, WE — West Estonian Domain, WLG — West Lithuanian granulite domain, USh — Ukrainian Shield, VM — Voronezh Massif. Abbreviations for deformation zones: HGZ—GR — Hagsta-Gävle-Rättvik Zone, HSZ — Hassela Shear Zone, KSZ — Karlskrona Shear Zone, LLSZ — Linköping-Loftahammar Shear Zone, MEFZ — Middle Estonian Fault Zone, PPDZ — Paldiski-Pskov Deformation Zone, SFSZ — South Finland Shear Zone, VNSZ — Vingåker-Nyköping Shear Zone. Abbreviations for volcanic belts and sedimentary basins: H — Häme, Mk — Monki, O-J — Oskarshamn-Jönköping, P — Pirkankaa, Pk — Poceai, Sc — Salcia, Tm — Tampere, D-DD — Dnipro-Donets Depression. AMCG and A-type granitoid intrusions: MZ — Mazury Pluton, RP — Riga Pluton, KP — Korosten Pluton.

Shield domains and are separated by dextral strike-slip fault zones of transpression (Hassela, Paldiski-Pskov, Vingåker-Nyköping, and others). To the south, the Baltic Svecofennian is bounded by the TESZ. The orogen formed north-to-south later than the Sarmatian segment of the SFCZ (1.89–1.75 Ga versus 1.95–1.90 Ga), but prior to the Fennoscandia-Sarmatia collision (1.82–1.80 Ga). Notably, recent isotopic data [Shumlyansky et al., 2018] place the formation of the Sarmatian SFCZ segment no later than 1.99–1.97 Ga (rather than 1.95–1.90 Ga), corresponding to the age of the Osnytsk granitoid complex that dominates in the OMIB.

According to [Bogdanova et al., 2015], the Bergslagen, Livonia, and Amberland microcontinents — like Keitele and Bothnia — bear magmatic belts of an active continental margin along their southwestern flanks. The widespread calc-alkaline magmatism dated to 1.86–1.84 Ga also suggests that the NW-trending Central Lithuanian block is part of an active continental margin. During this time (or slightly earlier), oceanic lithosphere subducted in a NE direction. In Fig. 1, this subduction is indicated by dipping mantle reflectors within the Mid-Baltic Belt. Northward subduction beneath Bergslagen is also supported by seismic tomography data (see below).

The Fennoscandia—Sarmatia collision (1.82–1.80 Ga) reorganized structures of the Baltic Svecofennian and induced dextral strike-slip motion along faults both in the eastern SFCZ and within the orogen itself [Bogdanova et al., 2015]. This is inferred from the southward vergence of its southeastern structures and evidence of N-S shortening. We note, however, some reservations regarding this interpretation.

The Boundary between Sarmatia and Fennoscandia. For a long time, the boundary between these miniplates was drawn along a narrow suture (the Central Belarus Suture Zone) with the Minsk fault in its axial part [Bogdanova et al., 1996a, 2006; Garetsky, Karataev, 2011] (see Fig. 1, black-white dashed line). To its southeast (all geographic data are in present coordinates) adjoins the 2.0–1.95 Ga

OMIB (Sarmatia), and to its northwest the 1.95–1.90 Ga BPG and the ~1.90 Ga OKL with numerous ~2.0 Ga terranes (Fennoscandia).

New data on the age, composition, and origin of the lithotectonic complexes in the Baltic Svecofennian [Bogdanova et al., 2015] prompted a reassessment of SFCZ geodynamics, including the Sarmatia-Fennoscandia boundary. Because the BPG and OKL are coeval and geochemically akin to complexes of the northwestern Ukrainian Shield, they were reassigned from Fennoscandia to Sarmatia. The decisive argument is their NE strike azimuth (parallel to the OMIB), whereas the Baltic Svecofennian structures trend predominantly NW. Consequently, the Sarmatia boundary in the study area shifts 100–250 km to the northwest. The new axial suture separating Sarmatia and Fennoscandia extends from the Grodno-Białystok fault zone in Poland across Belarus, Lithuania, and Latvia to the east.

Based on PolandSPAN seismic and magnetic surveys across the BPG, OKL, and OMIB, Polish geophysicists [Mężyk et al., 2021], did support the main arguments of [Bogdanova et al., 2015]. However, they concluded that the Sarmatia-Fennoscandia boundary is not a localized lithospheric rupture but a diffuse cryptic suture ~150 km wide in which complexes of both miniplates are intermingled, forming a single continental crust. They further argue that the Ivanovo-Borysivka segment of the OMIB is a thrust wedge complex rooted into the OMIB crust during subduction. Although the entire OMIB was considered the active continental margin of Sarmatia based on its geochemical features [Shumlyansky, 2014], it is still unclear where to draw the south-eastern boundary of the oceanic basin that subducted under Sarmatia.

Problems of oceanic basin reconstruction. Despite challenges posed by a thick sedimentary cover that preclude palaeomagnetic work over much of the SFCZ, there is no doubt that an ocean existed within the zone. This has long been indicated by palaeomagnetic data from the Baltic Shield and Ukrainian Shield [Elming, 1985; Elming et al., 1993; Bakhmutov, Iosifidi, 2010; Bakhmutov et al., 2023,

among others]. If the OMIB was an active continental margin, then the BPG and OKL — now attached to Sarmatia — were likely ocean-facing between 2.0 and 1.80 Ga.

In the Baltic Svecofennian, remnants of oceanic crust (ophiolites) are documented in Finland between the Karelian craton and the Bergslagen Domain. Geochemistry of metasedimentary rocks in the Central Lithuanian, Latvian-East Lithuanian, and West Lithuanian granulite blocks suggests the presence of oceanic lithosphere between the Livonia and Amberland Domains and its subduction beneath Livonia and Bergslagen [Bogdanova et al., 2015]. Seismic refraction (WARR) and seismic tomography provide further constraints on the mantle structure of the region. Additionally, magnetic models of the junction zone between Fennoscandia and Sarmatia offer further clarity. In particular, the magnetic model of the Earth's crust in the junction zone of the Fennoscandian and Sarmatian microplates along the «EUROBRIDGE» geotraverse corresponds to the theoretical and modern analogue models of an island arc. This generally supports the authors' conclusions regarding plate tectonic processes during the formation of the EEC [Bogdanova et al., 1996b; Orlyuk, 2000; Orlyuk, Pashkevich, 2012; Orlyuk et al., 2017].

Mantle Structure of the Sarmatia-Fennoscandia Convergence Zone. The SFCZ spans virtually all of Belarus and the Baltic states, as well as parts of Ukraine, Poland, Finland, Sweden, adjacent eastern areas, and a large part of the Baltic Sea (Fig. 2, white outlines). In this paper, the study area extends eastward beyond the Baltic shoreline (unlike [Bogdanova et al., 2015]) to encompass all available seismic tomography data. This broad extent reflects two primary factors:

- a revised crustal architecture for the SFCZ based on new geological and geophysical information [Bogdanova et al., 2015; Mężyk et al., 2021], with emphasis on the Palaeoproterozoic Baltic Svecofennian structures (1.89—1.84 Ga) later deformed during Fennoscandia-Sarmatia convergence;
- the recognition of northeast-striking deep fault zones in western Ukrainian Shield

[Gintov et al., 2017; Entin et al., 2020; Mychak, Farfuliak, 2021], which allow the SFCZ to be traced far to the southeast.

Seismic tomography data within the 3D P -wave velocity model of the Eurasian mantle constructed using the Taylor approximation method. The theoretical foundations of the method were established by V.S. Geyko [Geyko, 2004] and are also detailed in [Tsvetkova, 2015]. The practical application of the method has been described in numerous studies, the most recent of which include [Bugaenko et al., 2015; Tsvetkova et al., 2016; Gintov et al., 2022; Starostenko et al., 2024b]. Therefore, we will not provide a full description of the methodology here, focusing only on certain specifics.

The velocity structure of the mantle (see Fig. 2) and southward-dipping mantle subduction slabs, overlooked until this publication (Figs. 1—3), are recorded in the horizontal and vertical seismic tomography cross-sections [Geyko, 2004; Tsvetkova et al., 2021; Gintov et al., 2022].

At the 100 km depth, three V_P tiers are recognized. Northeast of the TESZ within the EEC, two tiers occur (8.20—8.30 and 8.10—8.20 $\text{km}\cdot\text{s}^{-1}$); southwest of the TESZ in the West European Platform, a 8.10—7.95 $\text{km}\cdot\text{s}^{-1}$ tier prevails. The highest velocities (and hence highest densities; $V_P \geq 8.30 \text{ km}\cdot\text{s}^{-1}$) occur beneath Estonia and Latvia, corresponding to slab № 1 (see Fig. 1). Westward beneath the Baltic Sea, this anomaly reaches slab № 3. A second-tier high- V_P anomaly (8.10—8.20 $\text{km}\cdot\text{s}^{-1}$), separated from the first by a belt of high- V_P gradients, is observed in Belarus and Poland. It crosses the TESZ and correlates with slabs № 2, № 5, and № 6 (see Fig. 1). All slabs in Fig. 1 exhibit elevated V_P , but their identification as slabs is confirmed only by analysis of vertical ΔV_P sections.

In the seismic tomography model, vertical sections are constructed every 1° of latitude and longitude (Fig. 3). To better visualize the mantle structure, velocities are displayed as isolines of $\Delta V_P = V_P - V_{P_aver}$, where V_{P_aver} is the mean P -wave velocity at each depth. At 100 km (see Fig. 2), $V_{P_aver} = 8.082 \text{ km}\cdot\text{s}^{-1}$. Method accuracy is $\delta = \pm 0.015 \text{ km}\cdot\text{s}^{-1}$ [Geyko,

2004]; principal ΔV_p contours are drawn every $0.05 \text{ km}\cdot\text{s}^{-1}$ with auxiliaries at $0.025 \text{ km}\cdot\text{s}^{-1}$.

Note that the subduction of Fennoscandia beneath Sarmatia is among the oldest within the EEC (initiating near 2.0 Ga). It was first recognized in the crust and immediately beneath the Moho (to $\sim 80 \text{ km}$) [Bogdanova et al., 2006; Gintov, Pashkevich, 2010]. Mantle slabs have previously been discussed mainly in the context of Palaeozoic subduction within the TESZ [Gintov et al., 2022], where they are more conspicuous. Slabs related to the Fennoscandia-Sarmatia convergence became evident only when examining the SFCZ and its deep structure.

Six subduction slabs are projected to the surface in Fig. 1 (visualized in Figs. 2, 3). These slabs are less distinctly expressed on vertical cross-sections than their Phanerozoic counterparts and are likely not entirely preserved. The clarity of the visualization and the

depth of immersion permit an approximate estimation of the relative age of the mantle slabs. Most of the mantle slabs (see Fig. 1) are considered within the structural context of the Baltic part of the Svecofennian Orogen, which formed subsequent to the initiation of subduction beneath Sarmatia. This particular juxtaposition was chosen for the convenience of further description and analysis.

Slab № 1 (sections SN $21^\circ\text{--}28^\circ$, segment $63^\circ\text{--}57^\circ \text{ N}$), $\sim 500 \text{ km}$ long, is the least distinct. It appears to begin beneath the southern margin of the Keitele microcontinent and extends southward to the southern boundary of the Livonia microcontinent. Its base lies shallower than $\sim 250 \text{ km}$.

Slab № 2 (sections SN $25^\circ\text{--}30^\circ$, segments $54^\circ\text{--}55^\circ\text{--}51^\circ \text{ N}$) lies entirely beneath Sarmatia (BPG, OKL, OMIB, Korosten Pluton) and dips to nearly 300 km . It may connect to slab № 1 (the boundary between them is

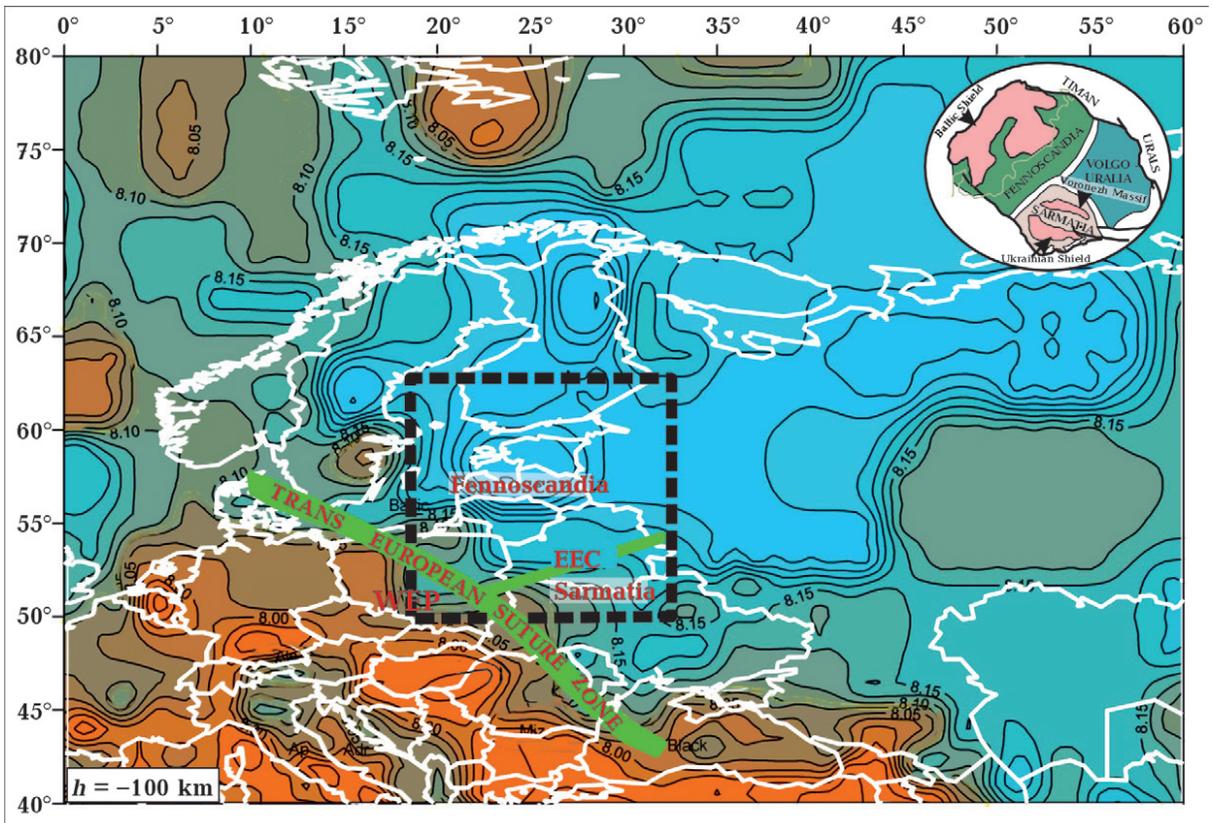
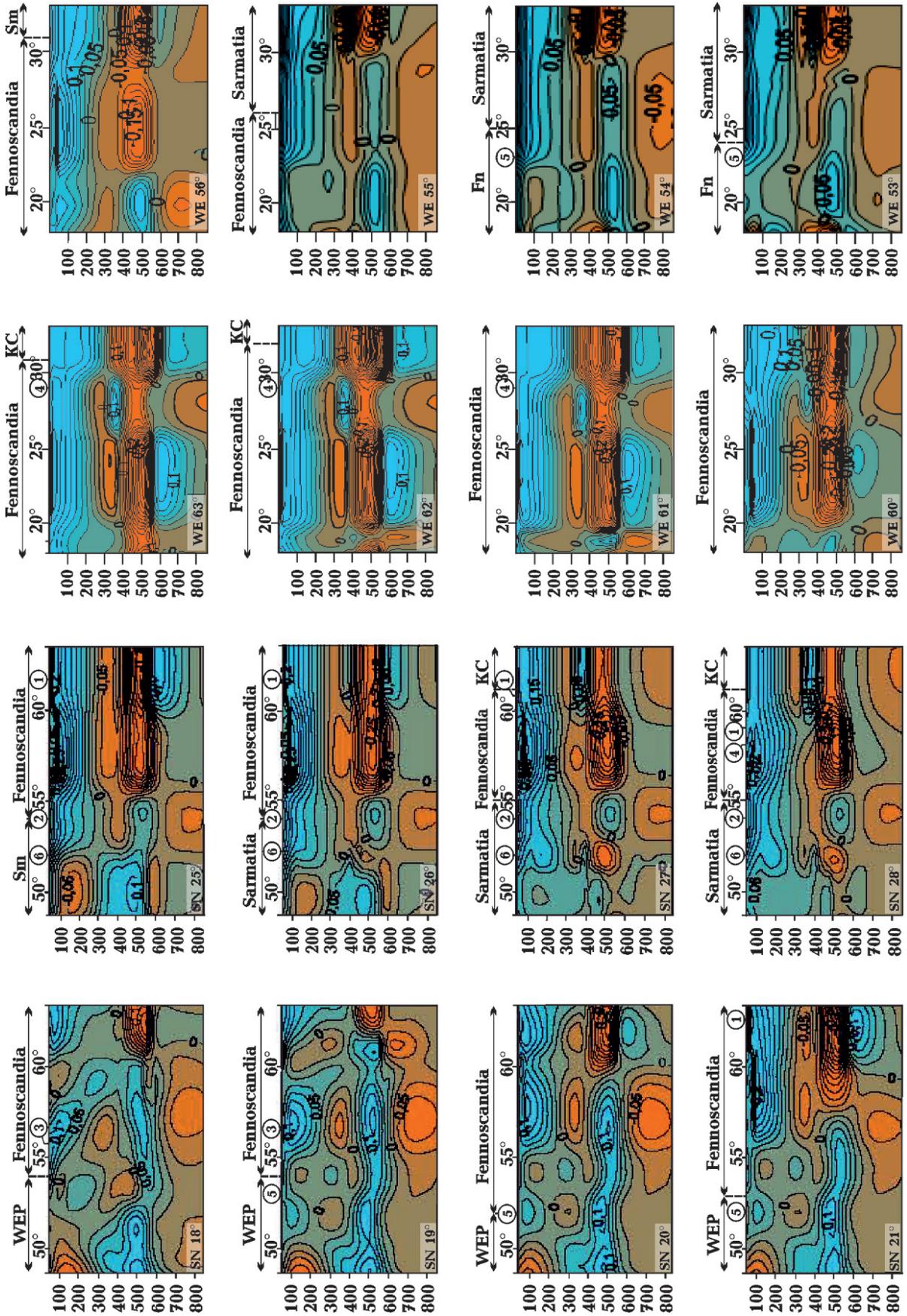


Fig. 2. Horizontal cross-section of the 3D P -wave velocity model of the East European Craton and West European Platform mantle at 100 km depth, after [Gintov et al., 2022]. The isoline $V_p=8.082 \text{ km}\cdot\text{s}^{-1}$ (reference velocity for this depth) marks the velocity boundary between elevated (blue tones) and reduced (red tones) velocities. Isolines are drawn at $0.025 \text{ km}\cdot\text{s}^{-1}$ intervals. Double green line — TESZ; white lines — state borders.



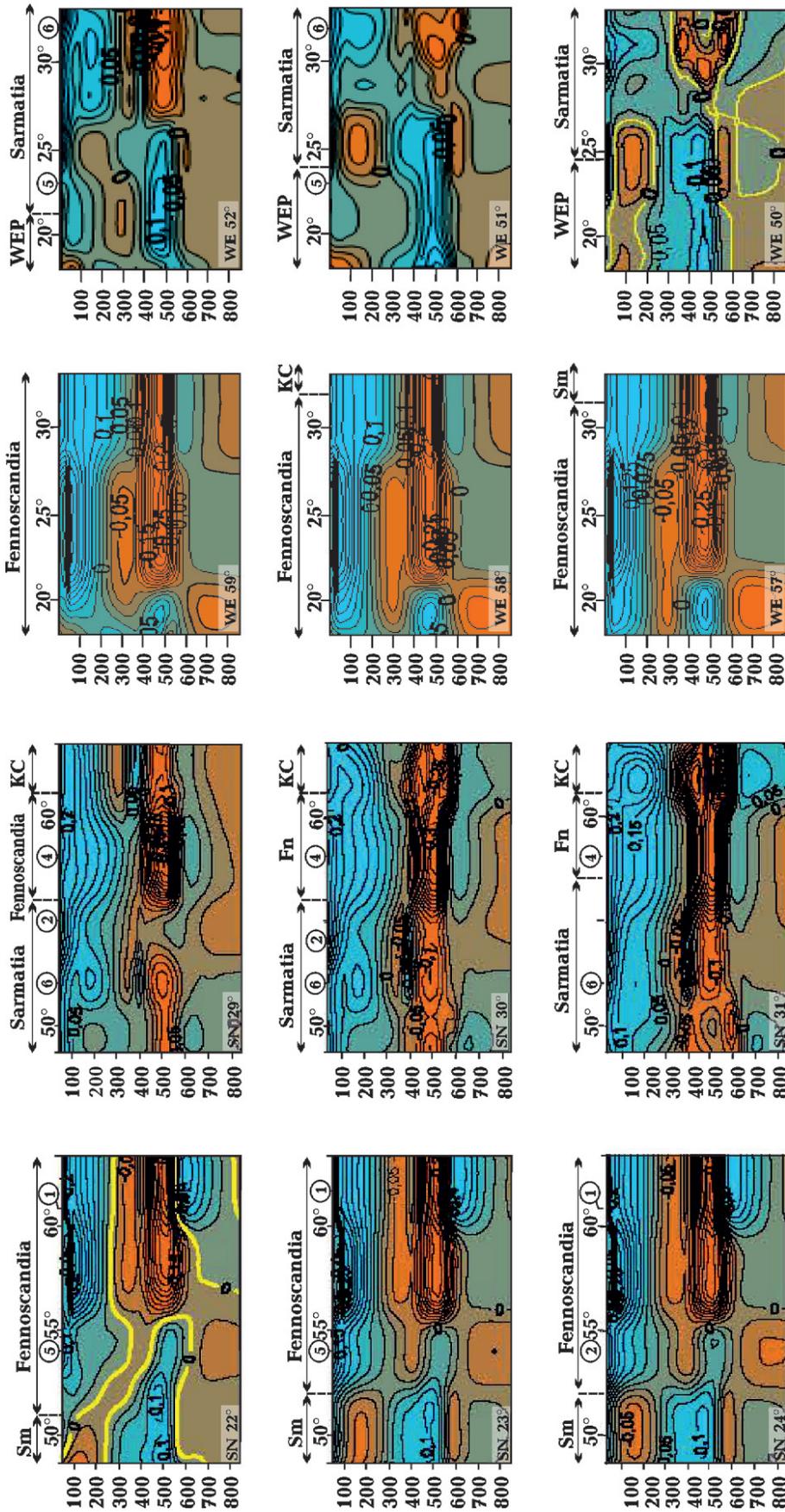


Fig. 3. Vertical longitudinal (meridional) (a) and vertical latitudinal (west-east) (b) sections through the study area. Circles with numbers indicate the position of the beginning of the near-surface part of the slab. Abbreviations: WEP — West European Plate, Sm — Sarmatia, KC — Karelian craton, Fn — Fennoscandia.

indistinct on SN 26°—28°), thereby delineating a complete mantle transition between the miniplates: a composite ~1000-km long with $\Delta V_p \geq 0.1$ km/s slab subducted, with interruptions, from ~2.0 to 1.80 Ga, initiating north of the present Gulf of Finland and dipping S-SE beneath the Ukrainian Shield to ≤ 300 km. Slab № 3 (sections SN 18°, segments 56°—60°N; SN 19°, segments 57°—60°N) dips northward within the Baltic Svecofennian, from the Amberland microcontinent beneath the northern margin of Bergslagen, to ≥ 400 km. It corresponds to the subduction zone of the Mid-Baltic Belt (see Fig. 1) identified by [Bogdanova et al., 2015] during formation of the Baltic Svecofennian (1.86—1.84 Ga).

Since slabs № 1 and № 2 are subducting (or dipping) to the south, and slab № 3 is subducting (or dipping) to the north, they are visible only on longitudinal cross-sections (or profiles), while on latitudinal cross-sections (or profiles) they correspond simply to areas of elevated ΔV_p .

Slab № 4 (sections SN 30°—31°, segments 60°—63°N; WE 61°—63°, segments 29°—31°E) lies partly beyond our study window but indicates subduction of the Finnish and Baltic Svecofennian beneath the Karelian craton along the Raahe-Ladoga suture zone at ~1.85—1.80 Ga [Baltybaev, 2013] (not labeled in Fig. 1), complementing knowledge of Baltic Svecofennian subduction.

Slab № 5 (sections SN 19°—23°—24°, segments 54°—52° N; WE 51°—53°, segments 18°—24° E) pertains to the TESZ [Tsvetkova et al., 2021; Gintov et al., 2022]. On longitudinal sections it nearly joins slab № 2, though it is distinct on latitudinal sections.

Slab № 6 (latitudinal sections WE 52°, segments 25°—32° E — clear; WE 51° — less clear), ~500 km long, follows the Ukraine-Belarus border, dipping westward to ~200 km beneath the OMIB and BPG. It corresponds to subduction imaged by EUROBRIDGE (1994—1997) in the middle and lower crust of Sarmatia beneath Fennoscandia [Bogdanova et al., 2006] and intersects slab № 2. The younger, east-born underthrusting beneath the OMIB and BPG implies continental subduction, possibly Devonian, during westward

development of the Prypiat-Dnipro-Donets aulacogen.

Discussion. The amalgamation of Fennoscandia, Volgo-Uralia, and Sarmatia into the EEC has been debated since the advent of plate tectonics, and inconsistencies remain. Pertinent to this study:

It has long been argued [Bogdanova et al., 1996a; Elming et al., 1997; Shumlyansky, 2014, and others] that the ~2.0 Ga OMIB represents the active continental margin of Sarmatia; the onset of the southward subduction of Fennoscandia beneath Sarmatia thus could not be younger, even after the reassignment of the BPG and OKL to Sarmatia.

In [Bogdanova et al., 2015], the collision occurred at 1.82—1.80 Ga. For a subduction system, a 0.18—0.20 Ga span is lengthy, implying one or more pauses.

The Baltic Svecofennian formed largely between 1.89 and 1.84 Ga under subduction of the opposite polarity (northward), i.e., opposite to the Fennoscandia-beneath-Sarmatia system. During a lull in the main subduction, a secondary, reverse-polarity system likely operated. Investigators of the Baltic Svecofennian [Rutland et al., 2004; Hermansson et al., 2008, among others] noted frequent alternations of compression and extension during its evolution.

A similar situation was considered in the mechanical model of early Precambrian plate tectonics of the Ukrainian Shield [Gintov, 2019], where a pulsing mantle plume superposed on plate motions was invoked to explain pauses and short-period reversals in lithospheric block displacements. Such a mechanism might apply to the Fennoscandia-Sarmatia convergence, although currently, the data are insufficient.

The south-southeastward dipping of slabs № 1 and № 2 implies that structures of the SFCZ developed in that direction. Consequently, the northeast-striking fault zones would initially have formed as sinistral strike-slip faults (left-lateral). Tectonophysical studies in the western Ukrainian Shield, a part of the SFCZ, confirm that the Horyn, Lutsk (Strelsk), Sushchany-Perha, and Teteriv fault zones — of the Nemyriv faulting stage

(≤ 2.0 Ga) — were initiated as sinistral strike-slip faults and reactivated later as dextral strike-slip faults [Gintov et al., 2017; Entin et al., 2020; Mychak, Farfuliak, 2021]. Regional compression during their inception was approximately meridional. This preceded the accretionary events in the Baltic Svecofennian (~ 1.89 Ga, Fig. 1) and thus did not cause the southward vergence of its structures.

Beginning at ~ 1.80 Ga, the Nemyriv stage was replaced by the Subbotsi-Moshoryne stage (1.80—1.77 Ga), characterized by a NE extension (strike azimuth $\sim 45^\circ$) and complementary compression at $\sim 315^\circ$. In this stress field, Sarmatia rotated 45° — 50° counterclockwise between 1.80 and 1.77 Ga, as indicated by palaeomagnetic [Kravchenko, 2005; Bogdanova et al., 2013; Bakhmutov et al., 2023] and tectonophysical data [Gintov, Mychak, 2014]. This rotation, followed by docking against the Baltic Svecofennian, likely induced the south-directed underthrusting (southward vergence) of its southeastern structures along the contact — a terminal stage of the Fennoscandia-Sarmatia collision. Immediately thereafter, the Korosten stage of an E-W extension (1.76—1.73 Ga) triggered the main magmatic pulse of the Korosten Pluton and reactivated dextral strike-slip motion along the NE-striking SFCZ faults [Gintov, Mychak, 2014].

Conclusions. 1. The Fennoscandia-Sarmatia convergence zone extends across the Baltic states, parts of Finland, Sweden, Poland, Ukraine, and adjacent northeastern territories, as well as much of the Baltic Sea. Tectonophysical and seismic tomography evidence shows that it continues into the northwestern Ukrainian Shield via a network of NE-striking fault zones (Horyn, Lutsk/Strel'sk, Sushchany-Perha, Teteriv).

2. Tectonophysical and seismic tomography results support the view of [Mężyk et al., 2021] that the Sarmatia-Fennoscandia boundary is not a localized lithospheric rupture but a diffuse zone ~ 150 km wide in which complexes of both miniplates are intermixed and form a continuous continental crust. The «old» boundary (Central Belarus Suture Zone with the axial Minsk fault) and

the «new» boundary continuing the Grodno-Białystok fault zone to the northeast [Bogdanova et al., 2015] are elements of this diffuse structure.

3. The southward vergence (underturning) of the southeastern structures of the Baltic Svecofennian at its junction with Sarmatia results from the counterclockwise rotation of Sarmatia during the final stage of Fennoscandia-Sarmatia collision (1.80—1.77 Ga).

4. Seismic tomography data reveal six subduction slabs within the SFCZ which, together with tectonophysical evidence, reflect the following geodynamic events:

- ≤ 2.0 Ga to 1.82—1.80 Ga — interrupted subduction of oceanic lithosphere separating Fennoscandia and Sarmatia beneath the latter to the south (slabs № 1 and № 2), formation of an active continental margin (OMIB) and NE-striking fault zones beneath the ocean and along the NW margin of the Ukrainian Shield; collision of the miniplates;

- ≤ 1.95 Ga — incorporation of the BPG and OKL into Sarmatia (slabs № 1 and № 2);

- 1.89—1.84 Ga — a pause in subduction beneath Sarmatia and formation of the Baltic Svecofennian (slabs № 3 and № 4) via subduction of local oceanic basins to the north;

- Ordovician-Silurian — onset of TESZ formation (slab № 5);

- Late Devonian? — west-directed subduction beneath the OMIB and BPG during the development of the Prypiat Trough (slab № 6).

5. Accumulating detailed geological and geophysical data paradoxically increases inconsistencies in interpretations of Fennoscandia-Sarmatia convergence. Earlier, when little was known about the crystalline basement south of the Gulf of Finland, it was widely accepted [Claesson et al., 2001; Shumlyansky, 2014, among others] that the OMIB formed as an active continental margin of Sarmatia due to a north-to-south subduction and collision at 2.0—1.98 Ga. Modern palaeomagnetic data [Bakhmutov et al., 2023], however, indicate that at 1.76 Ga the Ukrainian Shield (and Sarmatia) lay ~ 400 km farther from the Gulf of Finland than at present. Either Fennoscandia and Sarmatia separated for a time after

collision and later re-approximated, or the nature of the OMIB requires re-evaluation. Additional uncertainties persist in analyses of the Baltic Svecofennian structure [Bogdanova et al., 2015].

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Глибинна структура та геодинаміка перехідної зони Сарматія—Фенноскандія на основі геологічних і сейсмотомографічних даних

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Розглянуто геологічну будову земної кори та мантії в межах зони переходу між мініплитами Сарматія та Фенноскандія, що охоплює центральну та південну частини Свекофенського орогену і північно-західний сегмент Українського щита. Цю зону переходу розширено на південний схід за рахунок низки північно-східних розломів Українського щита (від Горинського до Тетерівського включно). Інтерпретація будови земної кори ґрунтується на сучасних публікаціях шведських, польських, естонських, литовських та українських вчених. Мантію до глибин 850—2500 км досліджено із застосуванням тривимірної швидкісної моделі Євразії, розробленої В.С. Гейком в Інституті геофізики ім. С.І. Субботіна НАН України на основі сейсмічної томографії з використанням наближення Тейлора до рівняння ейконала та хвильового рівняння.

Показано, що зона переходу між Сарматією та Фенноскандією сформувалася в палеопротерозої (2,10—1,75 млрд років тому) внаслідок складних геодинамічних процесів, про що свідчить наявність кількох субдукційних мантійних слебів. Ключовим процесом була субдукція Фенноскандії під Сарматію, зафіксована слебом південного занурення від мікроконтиненту Кейтеле під мікроконтинент Бергсларген і Середньо-Балтійський пояс, а також слебом східного—південно-східного занурення під Сарматію від Білорусько-Підляського гранулітового поясу під Осницько-Мікашевицький магматичний пояс і район сучасного Коростенського плутону.

Субдукція відбувалася з перервами. Під час однієї з них (близько 1,89—1,84 млрд років тому) сформувалася Балтійська частина Свекофенського орогену внаслідок додаткової субдукції протилежної полярності (північного та північно-східного напрямків занурення), що підтверджується слебами від мікроконтиненту Амберленд під мікроконтинент Бергсларген і від Центрального Фінляндського дугового комплексу під Карельський кратон.

Таким чином, узгодження сейсмотомографічних даних про структуру верхньої мантії з геолого-геофізичними обмеженнями щодо еволюції земної кори надає переконливі докази плитотектонічної природи процесів, що призвели до формування Східноєвропейського кратону.

Ключові слова: Сарматія, Фенноскандія, Свекофенський ороген, Український щит, перехідна зона, мантія, субдукція.