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Gneiss protolith ages and tectonic boundaries in the NE part of the Bohemian Massif (Fore-Sudetic Block, SW Poland)

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Published geochronological data, petrology, geochemistry and geological context of orthogneisses in the Strzelin and the Stachów complexes (NE-part of the Fore-Sudetic Block), together with structural observations help to locate the northern extension of the boundary between the East and West Sudetes within the poorly exposed NE margin of the Bohemian Massif. The Strzelin complex, in the east, comprises the Strzelin gneiss, with zircon ages of 600±7 and 568±7 Ma, and the Nowolesie gneiss with a mean zircon age of 1020 1 Ma. The Stachów complex to the west, which forms several tectonic klippen in the Strzelin Massif and in the Lipowe Hills Massif, contains the Gościęcice gneiss and pale Stachów gneiss, both yielding Late Cambrian zircon ages (~500±5 Ma). The orthogneisses in both complexes correspond to peraluminous S-type granites, but have different inherited zircon ages and display contrasting trace element characteristics, indicating different sources and petrogenetic histories. Based on the ages, petrology and overall geological context, the Strzelin orthogneiss is similar to the Keprník orthogneiss of the East Sudetes, whereas the orthogneisses of the Stachów complex correspond to rocks known from the West Sudetes (e.g. the Izera and Śnieżnik orthogneisses). The Stachów and the Strzelin complexes are separated by the Strzelin Thrust, which may be interpreted as the northern extension of the boundary between the East and West Sudetes, i.e. part of the boundary between the Brunovistulian and Moldanubian terranes of the NE part of the Bohemian Massif.

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Key words: Bohemian Massif, East-West Sudetes boundary, zircon ages.

INTRODUCTION

The Bohemian Massif, an important part of the Variscan Belt in Central Europe, is commonly considered to be a collage of tectono-stratigraphic units such as the Saxo-Thuringian and Moldanubian zones, defined by Kossmat (1927). These units have recently been interpreted as terranes (Matte *et al*., 1990; Oliver *et al*., 1993; Cymerman and Piasecki, 1994; Cymerman *et al*., 1997; Cymerman, 2000; Franke, 2000; Franke and Żelaźniewicz, 2000; Tait *et al*., 2000; Aleksandrowski and Mazur, 2002; Winchester *et al*., 2002) of Armorican or more generally — of Gondwana affinity. The terranes, better recognised in the western, central and southern parts of the Bohemian Massif, are believed to have their prolongations in the Sudetes, i.e. in its NE, marginal part. However, the terrane divisions, the terrane affinities of rock complexes and the location of the terrane boundaries in the NE part of the massif are still poorly constrained (Cymerman, 1998).

Along the eastern margin of the Bohemian Massif (Fig. 1), the Moldanubian Terrane adjoins the Brunovistulian Terrane (Bruno-Vistulicum, after Dudek, 1980), the latter being interpreted as a microcontinent (Matte *et al*., 1990) of Avalonian affinity (Moczydłowska, 1997; Friedl *et al*., 2000; Finger *et al*., 2000). According to Schulmann and Gayer (2000), the western part of the Brunovistulian Terrane is composed of Neoproterozoic, high-grade metamorphic rocks with their Devonian-Carboniferous envelope. It was thrusted and piled up NE-ward into a nappe sequence, during oblique Variscan collision between the Moldanubian and Brunovistulian Terranes, creating the Moravo-Silesian Zone.

The Moravo-Silesian Zone can be divided into three sections (Fig. 1). In the southern, Moravian section, high-grade gneisses, granulites and eclogites of the Moldanubian Terrane were thrusted on to the Moravo-Silesian Zone, which emerges from below these rocks in the Thaya and Svratka windows (Suess, 1912; Misař *et al*., 1983). The border between the Moldanubian and Moravo-Silesian zones was defined by Suess

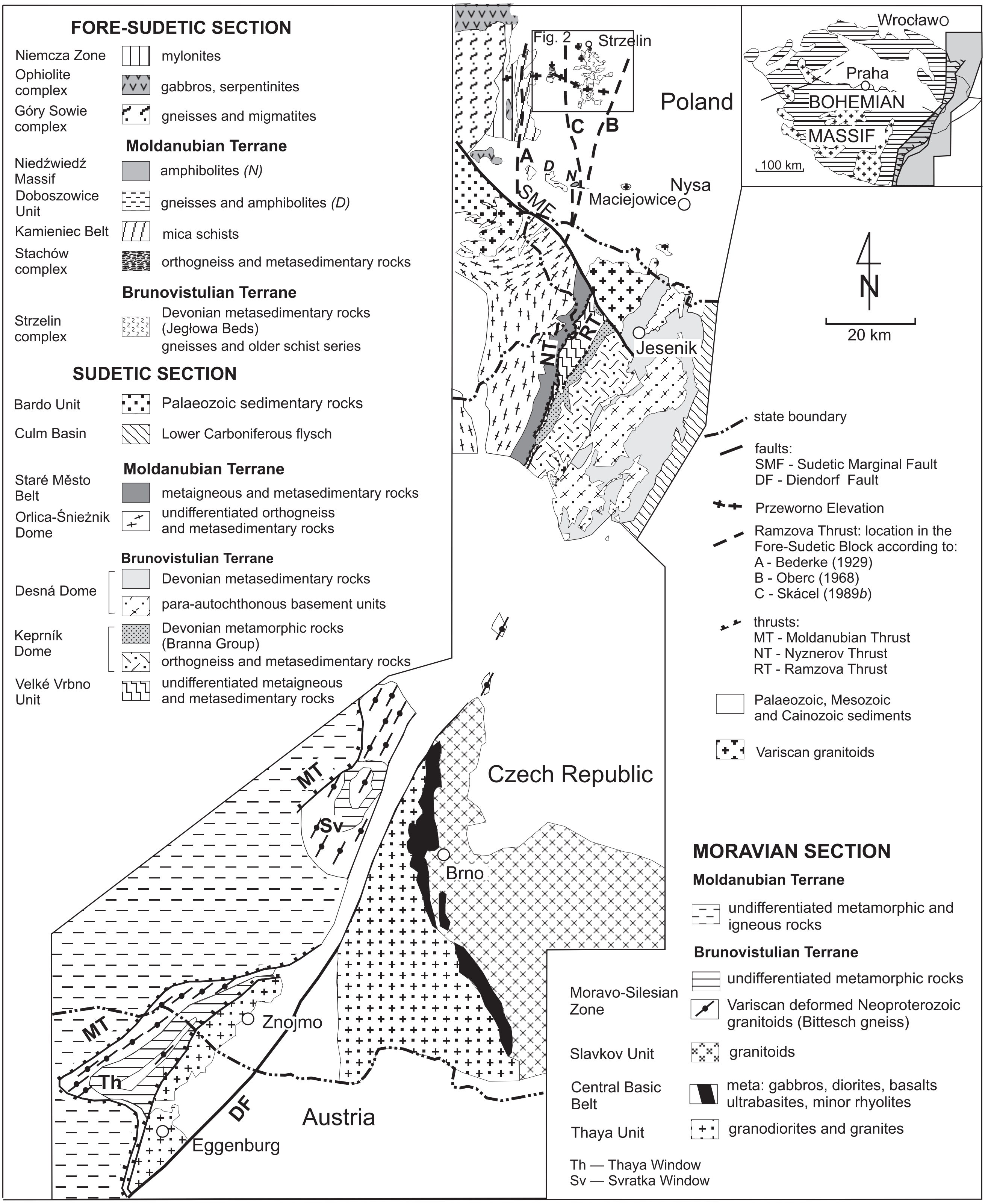

Fig. 1. Eastern margin of the Bohemian Massif, compiled by Oberc-Dziedzic and Madej (2002, modified) from Oberc *et al.* (1988); Puziewicz *et al.* (1999); Finger *et al.* (2000); Schulmann and Gayer (2000)

(1912) as the Moldanubian Overthrust. The granitoids cropping out in the Thaya and Svratka windows (Fig. 1) and further east in the Brunovistulian Terrane, were dated at around 580–590 Ma (Table 1; van Breemen *et al*., 1982; Finger *et al*., 2000 and references therein).

In the Sudetic section (Fig. 1), the Moravo-Silesian Zone (part of the East Sudetes) adjoins the Lugian (Suess, 1912) domain, which belongs to the West Sudetes. The western, structurally higher, part of the Lugian domain, the Orlica-Śnieżnik Dome, consists of medium-grade metasedimentary rocks, and gneisses derived from *ca*. 500 Ma Early Palaeozoic granites (Turniak *et al*., 2000), with small bodies of granulites and eclogites. The eastern, structurally lower part, the Staré Město Belt, is represented by an Early Palaeozoic sequence of metavolcanic rocks, sheared gabbros and metasediments, interpreted to have been formed during Cambro-Ordovician rifting (Schulmann and Gayer, 2000). The Moravo-Silesian Zone in the Sudetic section can be subdivided, from W to E, into the Velké Vrbno Unit (meta-igneous and metasedimentary rocks), the Keprník Nappe (orthogneiss with its inner high-grade metamorphic envelope and Devonian outer envelope; Cháb *et al*., 1994*a*), and the Desná Dome (gneiss with Devonian metasedimentary cover, bordered in the E by Culm foreland sediments; Schulmann and Gayer, 2000). The orthogneisses of the Sudetic section of the Moravo-Silesian Zone (except for *ca*. 500 Ma old small dykes) have Neoproterozoic ages (Kröner *et al*., 2000; Table 1).

The boundary between the West and East Sudetes was previously placed along the Ramzova Thrust and considered as the NE continuation of the Moldanubian Overthrust and the boundary between the Lugian domain and the Moravo-Silesian domain (Suess, 1912, 1926; Bederke, 1929, 1931; Skácel, 1956; Oberc, 1957; Misař, 1960). More recently it is interpreted to follow the Nyznerov Thrust along the eastern side of the Staré Město Belt (Skácel, 1989*a*; Schulmann and Gayer, 2000; Fig. 1). The extensional shearing which developed on both sides of the Ramzova-Nyznerov line (Cymerman, 1993*a*; Cháb *et al*., 1994*b*), sinistral strike-slip movement and brittle deformation along the Ramzova line (Aichler *et al*., 2002) were related to the later deformation history of the West–East Sudetes boundary (Dumicz, 1995; Mazur and Józefiak, 1999; Schulmann and Gayer, 2000; Szczepański and Mazur, 2004).

In the Fore-Sudetic section, the Kamieniec Metamorphic Belt in the western part of the area (Fig. 1) is composed mainly of metasedimentary rocks and was previously considered as a West Sudetic Unit. It has now been interpreted as part of the Saxo-Thuringian Terrane (Franke and Żelaźniewicz, 2000) or part of the Moldanubian Terrane (Cymerman and Piasecki, 1994; Aleksandrowski and Mazur, 2002). In contrast, the Strzelin Massif, composed of gneisses and metasedimentary rocks intruded by Variscan granitoids, was traditionally correlated with the East Sudetic units (Bederke, 1929, 1931; Oberc, 1966, 1972). Its Brunovistulian Terrane affinity has recently been further corroborated by SHRIMP zircon ages of the Strzelin orthogneiss (Oberc-Dziedzic *et al*., 2003*a*).

The prolongation of the boundary between the East and West Sudetes further north, in the Fore-Sudetic Block (Fig. 1), has been variously interpreted. Bederke (1929) placed it along the eastern border of the mylonitic Niemcza Zone, Oberc (1968) east of the Strzelin Massif, Skácel (1989*b*) west of this massif, Cwojdziński and Żelaźniewicz (1995) inside it, whereas Cymerman contested its presence altogether not only in the Fore-Sudetic Block (Cymerman, 2000) but also in the Sudetes (Cymerman, 1993*a*). The ambiguities exist mainly because of poor exposure and uncertain ages of rocks in this area.

This contribution refers to SHRIMP zircon ages for a gneiss from the Lipowe Hills (Oberc-Dziedzic *et al*., 2003*b*; Klimas, 2005, in press), and discusses these data within the context of previous geochronology. By combining the structural, petrographic and geochemical observations we aim at defining the location of the main tectonic boundaries in the NE part of the Variscan collage in the Sudetes.

GEOLOGY OF THE EASTERN PART OF THE FORE-SUDETIC BLOCK

The eastern part of the Fore-Sudetic Block (Fig. 1) consists of several, N–S elongated, tectono-metamorphic units, including, from west to east: the Niemcza Zone, the Kamieniec Metamorphic Belt, the Lipowe Hills Massif and the Strzelin Massif. The Doboszowice Metamorphic Unit and the Niedźwiedź Amphibolite Massif are situated south of the latter two units. Biotite gneisses, mica schists, phyllites and metagreywackes were also found under Cainozoic cover in boreholes east of the Strzelin Massif (Cymerman, 1991; Sawicki, 1995).

The Niemcza Zone, the Kamieniec Metamorphic Belt, the Doboszowice Metamorphic Unit and the Niedźwiedź Amphibolite Massif geologically belong to the West Sudetes. The Lipowe Hills Massif and the Strzelin Massif contain elements of both the East and West Sudetes affinities, so the tectonic boundary separating the East and West Sudetes may be located in these massifs (Oberc-Dziedzic and Madej, 2002; Oberc-Dziedzic *et al*., 2003*b*, and see below).

THE WESTERN AREA: THE WEST SUDETES UNITS

The Niemcza Shear Zone is interpreted as a sinistral strike-slip ductile shear belt (Mazur and Puziewicz, 1995), composed of high- and low-temperature mylonites derived from the Góry Sowie gneisses (Scheumann, 1937; Mazur and Puziewicz, 1995). Alternatively, these rocks are interpreted as metagreywackes containing clasts of mylonitised Góry Sowie gneisses, bedded cherts and small serpentinite bodies (Bederke, 1929; Franke and Żelaźniewicz, 2000). The Niemcza Shear Zone also contains small bodies of syenite/diorite and granodiorite dated (U-Pb, zircons) at about 340 Ma (Oliver *et al*., 1993). According to Aleksandrowski and Mazur (2002), sinistral, strike-slip motion along the Niemcza Shear Zone was superimposed on an earlier fabric, related to Early Carboniferous top-to NE thrusting and dextral shearing.

The Kamieniec Metamorphic Belt, east of the Niemcza Shear Zone, comprises medium-grade mica schists with minor quartzites, marbles, amphibolites and felsic metavolcanic rocks. The belt also contains small relics of eclogite bodies (Achramowicz *et al*., 1997).

Table 1

Selected geochronological data from the Moldanubian-Moravian and West-East Sudetes areas

Section	Terrane	Unit	Rock type and locality	Method	Age (Ma) and interpretation: (1) Crystallisation of magmatic protolith (2) Igneous emplacement (3) Source rock (4) Anatectic melting	Age (Ma) and interpretation: Inherited (1) cores (2) Xenocrysts (3) Detrital grains	Age (Ma) and interpretation: HT metamorphism, partial melting	Reference number
MORAVIAN	BRUNO-VISTULIAN	Thaya Window	Bittesch gneiss	SHRIMP (zircon)	578±7	2535±6, 2025±4, 1775±6, 1720–1695, 1651±9, 1520, 1250–1190 (1)		1
MORAVIAN	BRUNO-VISTULIAN	Thaya Unit	diorite, Dolni Kounice	U-Pb (zircon)	584±4			2
MORAVIAN	BRUNO-VISTULIAN	Thaya Unit	granite, Eggenburg	U-Pb (zircon)	583±11			3
SUDETIC	MOLDANUBIAN	Western part of the Orlica-Śnieżnik Dome	fine-grained granite-gneiss, G-type	Vapor Transfer Technique	509±11 (2)	555, 764 (2)		4
				single zircon evaporation	511±1 (2)			
			felsic, metavocanics and metatuffs	Vapor Transfer Technique		564±41–1939±5 (3)		4
				single zircon evaporation		842.1±1.7, 1497.2±1.6		
			granite-gneiss (G-type), Zakouti	single zircon evaporation	507.1±1.3 (2)	1066.2±3.2 (2)		4
			coarse-grained augengneiss, Zdobnice	single zircon evaporation	503.2±1.1 (2)	613.0±2.2, 778±2.7		4
				Vapor Transfer Technique	504±52 (2)	762±17 (2)		
			microgranite dyke, Zdobnice	single zircon evaporation	491.7±1.0			4
			augengneiss, Jablonnén.O.	Vapor Transfer Technique	508±40			4
				single zircon evaporation	504.6±1.0			
			augengneiss, Boková	Vapor Transfer Technique	503±15	564±15, 826±11 (2)		4
				single zircon evaporation	503.2±1.1 (2)	692±2.5, 867.2±1.1, 1184±1.5 (2)		
			augengneiss, Bartošovice	Vapor Transfer Technique	510±24 (2)	782±23, 1298±10 (3)		4
					509.9±1.3 (2)	1372±5 (3)		
			augengneiss, Mostowice-Spalona	Vapor Transfer Technique	510±37			4
				single zircon evaporation	510.5±1.1 (2)			
			Duszniki Zdrój gneiss	U-Pb (zircon)	506			5
		Eastern part of the Orlica-Śnieżnik Dome	Śnieżnik gneiss	U-Pb (zircon)	504±3, 488+4, 488–7			5
			Śnieżnik gneiss	SHRIMP (zircon)	495±7 (1)	530 –540, 565, 2600 (1)	340	6
			Gierałtów gneiss	SHRIMP (zircon)	495±14 (1)	530–540 (1)		6
			augengneiss, Vysoký Potok	Vapor Transfer Technique	512±11			4
				single zircon evaporation	510.8±1.3 (2)			
			migmatitic orthogneiss, Rači valley	single zircon evaporation	507.1±1.0 (1)	565.5±0.9 (3), 665.6±0.9 (3)		7
			migmatitic orthogneiss, Rači valley	single zircon evaporation	504.3±1.0	858.2±0.8 (3)		7

SUDETIC	MOLDAMUBIAN	Staré Město Belt	tonalitic gneiss (Hanušovice)	vapour digestion	504±4			8
				single zircon evaporation	504.1±1.0, 503.2±1.0, 502.5±1.0 (2)			
			leucocratic gneiss (Hanušovice)	vapour digestion	507±19			8
				single zircon evaporation	506.7±1.7	818±16 (2)		
			in situ melt patches (Hanušovice)	vapour digestion	503±7 (4)			8
				single zircon evaporation	502.1±1.0, 501.8±1.0 (4)	641±4, 684±6 (1)		
	BRUNOVISTULIAN	Desna Dome	light grey plagioclase-biotite schits	single zircon evaporation	598±0.9 (2 or 3)			8
			dark grey plagioclase-biotite schits	single zircon evaporation	664.8±0.8 (3) to 628±1.5			8
			leucocratic gneiss, Ludvikov	single zircon evaporation	684.5±0.9 (1)	864.0±1.6 (2)		8
			dyke of porphyritic granite	single zircon evaporation		594–911 (2)		8
				vapour digestion	517±12 (1)			
			migmatite (Kouty)	single zircon evaporation	506.7±1.7 (2)	604, 1106 (2)		8
			pegmatitic granite	single zircon evaporation	502.3±1.0 (2)			8
		ČSB*	mylonitic orthogneiss	single zircon evaporation	612.1±1.0 (2)			8
		Keprník Dome	Keprník gneiss	vapour digestion	584±8			8
				single zircon evaporation	583.8±0.9	802±7, 1073±9 661.3±0.9, 746.1±3.3 (2)		
			fine-grained orthogneiss	vapour digestion	553±7 (2)			8
				single zircon evaporation	554.6±1.2 (2)	781.5±1.0 (2)		
		Velké Vrbno Unit	tonalitic orthogneiss	vapour digestion	570±6			8
				single zircon evaporation	574.3±1.0 (2)			
FORE-SUDETIC	MOLDANUBIAN	Stachów complex	Gościęcice gneiss	U-Pb (zircon)	504±3 (1)			5
			Gościęcice gneiss	single zircon evaporation	513±1 (1)	1301, 1096 (2)		9
			pale Stachów gneiss	SHRIMP (zircon)	500±5 (3)	1916±25, 636.3±8.4, 630.9±6.4, 619.4±7.9 567.0±6.1, 560.3±6.9 (1)		10
			Maciejowice gneiss	single zircon evaporation	501±1 (1)	1694		9
	BRUNOVISTULIAN	Strzelin complex	Strzelin orthogneiss	SHRIMP (zircon)	600±7 (1)	1879±29 1785±21, 1775±38, 1532±17, 1496±18, 1353±20, 1248±23, 1230±8 (1)	568±7	11
			Nowolesie migmatitic gneiss	single zircon evaporation	1020±1 (1)	1767–1135 (2)		9

ČSB* — Červenohorské Sedlo Belt; reference numbers: 1 — Friedl *et al.*, 2000; 2 — van Breemen *et al.*, 1982; 3 — Finger *et al.*, 2000; 4 — Kröner *et al.*, 2001; 5 — Oliver *et al.*, 1993; 6 — Turniak *et al.*, 2000; 7 — Štípská *et al.*, 2004; 8 — Kröner *et al.*, 2000; 9 — Kröner, Mazur, 2003; 10 — Oberc-Dziedzic *et al.*, 2003*b*; 11 — Oberc-Dziedzic *et al.*, 2003*a*

The Doboszowice Metamorphic Unit is composed mainly of leucocratic two-mica orthogneiss (Puziewicz *et al*., 1999), considered to be a syntectonic intrusion (Mazur and Puziewicz, 1995) and dated at 379±1 Ma (U-Pb evaporation method on single zircons; Kröner and Mazur, 2003). Two-mica paragneisses with intercalations of mica schists, amphibolites and amphibole gneisses crop out in the eastern part of the Doboszowice Unit.

The Niedźwiedź Amphibolite Massif represents a ~3.8 km-thick sequence (Cymerman and Jerzmański, 1987) of predominantly MORB-type metagabbros and amphibolites (Awdankiewicz, 2001). These rocks experienced high-grade metamorphism and, locally, show evidence of partial melting (Puziewicz and Koepke, 2001).

According to Mazur and Józefiak (1999), the metamorphic complexes exposed between the Góry Sowie and the Niedźwiedź massifs are nappes which formed due to E-directed overthrusting (D_1 event) during the Variscan orogeny. A penetrative foliation S_1, dipping mostly to the NW at high to moderate angles, and a stretching lineation L_1 parallel to the thrust direction, were coeval with overthrusting. The subsequent D_2 deformation produced NE–SW-trending F_2 folds, stretching and intersection lineation L_2, as well as axial cleavage of F_2 folds, transposed, in places, into a new penetrative, subhorizontal foliation S_2 overprinting the older foliation S_1. The D_2 deformation was probably accompanied by a syntectonic granite intrusion, subsequently transformed into the Doboszowice gneiss (Mazur and Józefiak, 1999). The deformation D_3 was related to WSW-directed extensional collapse, recorded by the development of low-angle dip-slip zones. The sinistral strike-slip Niemcza Shear Zone formed along the eastern margin of the Góry Sowie Massif during the same D_3 extensional deformation (Mazur and Józefiak, 1999).

THE EASTERN AREA: THE STRZELIN AND LIPOWE HILLS MASSIFS

The Strzelin and Lipowe Hills massifs are separated by a several kilometres wide belt of Cainozoic sediments (Fig. 2). Two rock complexes are distinguished in this area: the structurally lower Strzelin complex, more widespread in the Strzelin Massif, and the upper Stachów complex, dominating in the Lipowe Hills Massif.

THE STRZELIN COMPLEX

The Strzelin complex is composed of a core unit, an inner envelope (older schist series), and an outer envelope (younger schist series = the Jegłowa Beds) (Figs. 2 and 3).

The core unit comprises several varieties of Neoproterozoic gneisses (see below for discussion on age):

1. Fine- to medium-grained, porphyritic biotite-muscovite Strzelin gneiss, with conformable, several centimetres to several metres thick bodies of amphibolite (Szczepański and Oberc-Dziedzic, 1998); this gneiss is typical of the northern part of the Strzelin Massif;

2. The Nowolesie migmatitic, sillimanite gneiss, with numerous pegmatite bodies; this gneiss occurs in the southern part of the Strzelin and Lipowe Hills massifs;

3. The Bożnowice and Gromnik gneisses, transitional between (1) and (2), cropping out in the middle and southern part of the Strzelin Massif (Fig. 2, Table 2).

The inner envelope of the gneisses consists of the older schist series of Neoproterozoic or Early Palaeozoic (?) age, composed of amphibolites, mica schists, calc-silicate rocks and marbles. All contacts with the gneisses are parallel to the lithological boundaries and to the main foliation.

The outer envelope, the younger schist series (the Jegłowa Beds; Oberc, 1966), comprises quartzites, quartz-sericite schists and metaconglomerates, the protoliths of which were deposited in a continental margin setting during Early- to Mid-Devonian (Patočka and Szczepański, 1997). The Jegłowa Beds were correlated by Bederke (1931) and Oberc (1966) with the quartzite formation in the Jeseniki Mts. of the East Sudetes, containing Early Devonian fossils (Chlupač, 1975). The Jegłowa Beds form thin slabs overlying the Strzelin and Nowolesie gneisses and the nearly horizontal contacts between them are tectonic, oblique to the S_1 foliation in both complexes (Oberc-Dziedzic, 1995).

The Strzelin complex was deformed and metamorphosed during the Variscan orogeny, prior to the end of the Viséan. The rocks underwent polyphase deformation (D_1–D_4) and metamorphism (M_1–M_4), but their main structural features developed during the D_1 and D_2 deformations, and the main phase of metamorphism occurred before the D_2 deformation (Wojnar, 1995).

The granitic protoliths of the Strzelin gneiss as well as rocks of the inner envelope and the Jegłowa Beds were subjected to top-to-NNE/NE non-coaxial shearing during deformation D_1, which resulted in the formation of a penetrative foliation S_1, dipping to NW/N in the northern part of the complex, and in a NNE-oriented stretching lineation L_1. The axes of rare F_1 folds generally plunge to the ESE. In the southern part of the complex, the S_1 foliation in the gneisses and in the Jegłowa Beds dips to SE/S (Oberc-Dziedzic and Madej, 2002).

During the D_2, the S_1 foliation was deformed, producing asymmetric, isoclinal or disharmonic F_2 folds of variable scales. The axes of these folds plunge to the N, NNE and NE in the northern part of the massif, and to the SE, S and SSW in its southern part (Oberc-Dziedzic and Madej, 2002 and references therein).

Deformation D_3 produced broad, open or kink-type F_3 folds with steep axial planes S_3. The F_3 fold axes plunge to N and NW, or W–E, WNW–ESE and NW–NE in the northern and southern part of the Strzelin complex, respectively. In the southern part of the complex, they are more or less perpendicular to the F_2 fold axes and parallel to the Przeworno Elevation (Fig. 2; Oberc, 1966). The Przeworno Elevation affected the entire megastructure of the eastern part of the Fore-Sudetic Block (Oberc, 1966, 1972) and caused the linear structures to plunge generally to the N in the northern part, and to the S in the southern part of the Strzelin complex (Oberc, 1966).

The D_4 event produced an S_4 foliation defined as local, thin mylonitic bands in the Strzelin gneiss, dipping at 10–25° to the N (Oberc-Dziedzic, 1999), and as narrow shear zones in the Jegłowa Beds dipping to the S in the southern part of the complex (Szczepański, 2001; Szczepański and Mazur, 2004). Kinematic indicators in the Jegłowa Beds document top-to-NE shearing in the northern part of the massif and top-to-SSW

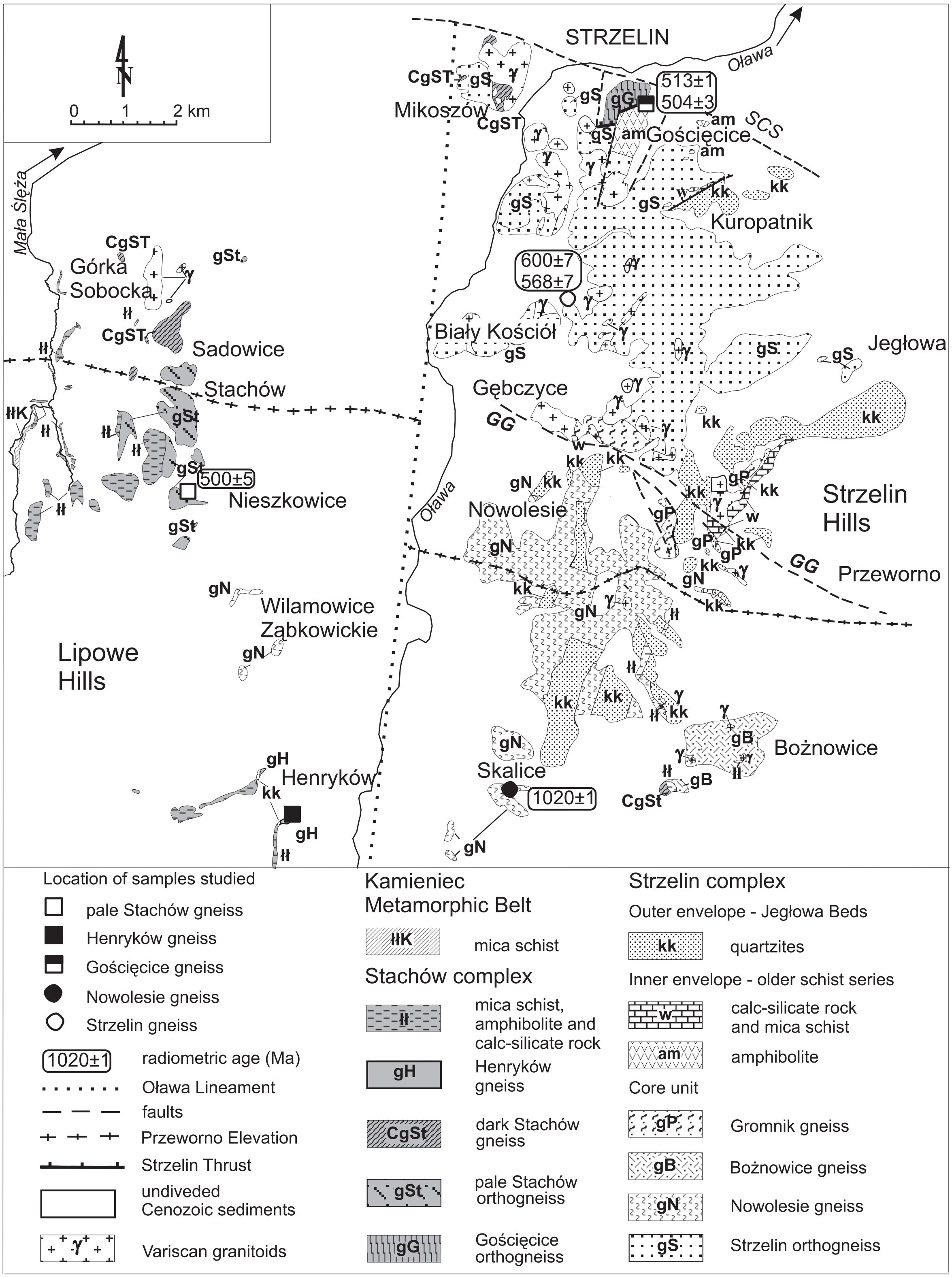

Fig. 2. Geological map of the Strzelin Massif; compiled by Oberc-Dziedzic and Madej (2002) from Oberc *et al*. (1988, simplified), Wójcik (1968), Wroński (1973), Badura (1979)

SCS — Sienice–Strzelin Fault, *GG* — Gębczyce–Gromnik Fault

shearing in its southern part (Szczepański, 2001). During the D_4 event, most of the geological boundaries were reactivated and modified; the Jegłowa Beds were detached from the gneisses and moved toward the NE from their original position.

The effects of M_1–M_3 Variscan metamorphic episodes differ between the thrust-bounded units, indicating different metamorphic paths. In all units, however, the M_1 metamorphic event was related to progressively increasing P-T conditions. The ef-

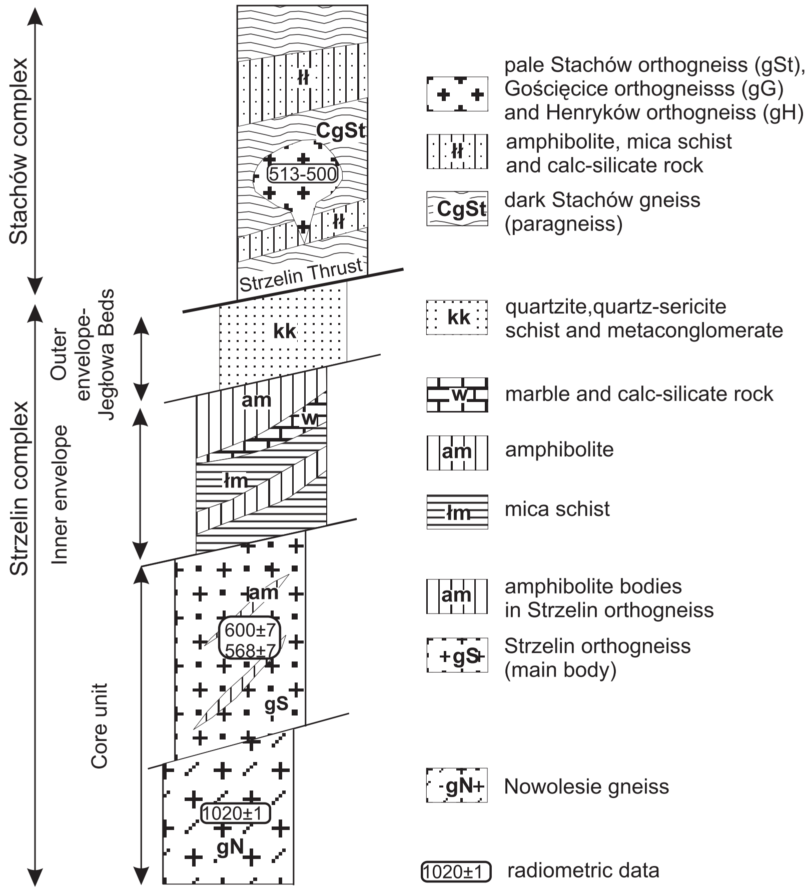

Fig. 3. Simplified tectonic subdivision and lithology of the Strzelin and Stachów complexes

fects of M_4 metamorphic event were similar in the whole Strzelin complex.

In the northern part of the Strzelin complex, the M_1 metamorphic conditions were typical of the greenschist facies in the Jegłowa Beds and of the amphibolite facies in the Strzelin gneiss. The rocks of the older envelope went through transitional, greenschist to amphibolite facies conditions (Oberc-Dziedzic and Madej, 2002). The M_1 metamorphic episode corresponds to nappe stacking syn- and post-dating the D_1 deformation but prior to the D_2 event. The P-T conditions of the M_2 metamorphism during the D_2 deformation were similar to that of M_1. The M_3 metamorphism, coeval with the D_3 and D_4 deformation events, took place under amphibolite- to greenschist facies conditions and caused localized retrogression.

In the southern part of the Strzelin complex, the Nowolesie gneiss attained anatectic conditions during the M_1 metamorphic event. The first stage of anatexis during M_1 was followed by an M_2 decompressive event, related to the formation of pegmatites and leucocratic granites.

The Jegłowa Beds in the southern part of the Strzelin complex were metamorphosed under greenschist facies conditions, with increasing temperature and pressure during the M_1 episode and decreasing T and P during the M_2 event (Szczepański and Józefiak, 1999).

The metamorphic conditions of the M_3 event were probably uniform in the entire southern part of the Strzelin complex and similar to those of the M_2 event in the Nowolesie gneisses. The latter did not experience any significant changes, whereas the Jegłowa Beds attained HT-LP amphibolite facies conditions (Szczepański and Józefiak, 1999).

The M_4 episode of regional metamorphism led to the crystallisation of post-kinematic cordierite and the formation of flecky gneisses in both the N and S parts of the Strzelin complex (Oberc-Dziedzic, 1999).

THE STACHÓW COMPLEX

The Stachów complex is defined here to comprise the Gościęcice augen gneiss (~500 Ma) and the Stachów gneisses. The Stachów gneisses have two varieties: fine-grained gneiss, referred to as the dark Stachów gneiss, and flaser gneiss, *ca*. 500 Ma in age, termed the pale Stachów gneiss (formerly referred to as the "light-coloured Stachów gneiss"). A strongly deformed variety of the latter is called the Henryków gneiss; (Madej, 1999). The dark Stachów gneiss alternates with mica schists and amphibolites. All these intercalations are interpreted as Neoproterozoic or Early Palaeozoic metasediments, representing the metamorphic envelope of the granitoid protolith of the pale Stachów gneiss and probably also, as suggested by xenoliths, of the Gościęcice gneiss (Oberc-Dziedzic and Madej, 2002). The presence of the dark gneiss and the nearly complete absence of calc-silicate rocks distinguishes this envelope from the inner envelope of the Strzelin gneiss.

The Stachów complex, similarly as the Strzelin complex, was deformed and metamorphosed during the Variscan orogeny. The tectonic position of the Stachów complex exposed in the northern and central part of the Lipowe Hills Massif is unknown. In the southern part of the Lipowe Hills, this complex forms a klippe (the Henryków klippe) composed of the Henryków gneiss and resting on the Strzelin complex (Oberc-Dziedzic and Madej, 2002). In the Strzelin Massif, a large klippe (the Gościęcice klippe), of the Gościęcice augen gneiss is found in the northern part of the area, and small klippen composed of the dark Stachów gneisses were identified in both northern and southern parts of the massif (Fig. 2).

In the pale and dark Stachów gneisses, exposed in northern and middle part of the Lipowe Hills, the poorly visible foliation S_1 formed during the earliest D_1 deformation was folded into F_2 folds during the D_2 event and transposed into the S_2 penetrative foliation dipping mainly to the SW–W–WNW. The axes of the F_2 folds plunge to the SW at 35–45°. The L_2 lineation (intersection of foliation S_1 and S_2) plunges to the W, SW and S. The S_2 foliation was reactivated during the extensional event D_3, forming locally the S_{2+3} complex foliation. The L_3 lineation, expressed by mica alignment on the S_{2+3} foliation, displays constant orientation throughout the entire area, plunging to the SW–W and being oblique or parallel to the L_2 lineation. Kinematic indicators suggest a SW sense of shear.

Table 2

Diagnostic features of gneisses

	Type of gneiss	Locality	Colour	Fabric	Foliation and mineral lineation	Other features
Strzelin complex	Strzelin gneiss	N and central part of the Strzelin Hills	light grey to grey	fine to medium-grained, porphyritic texture, streaky, layered or random structure	poorly elongated, several grain quartz aggregates; biotite and muscovite streaks	porphyroclasts of microcline and/or plagioclase, faint planar structure
	Nowolesie gneiss	S part of the Strzelin Hills and S part of the Lipowe Hills	grey to nearly white	fine-grained, stromatite and nebulite structures, often polygonal shape of grains; quartz inclusions in plagioclase	biotite streaks; several-grain domains of quartz	uniform distribution of biotite, sillimanite nodules
	Bożnowice gneiss	S part of the Strzelin Hills	light grey to grey	similar to the Strzelin gneiss but contains needles of sillimanite dispersed in the matrix		
	Gromnik gneiss	middle part of the Strzelin Hills	light grey	similar to the Nowolesie gneiss but without sillimanite nodules		
Stachów complex	Gościęcice gneiss	N part of the Strzelin Hills	grey to grey-blue	augen, augen-layered or random structure	foliation defined by mica concentrations, feldspar and quartz layers; lineation defined by elongated concentrations of biotite	characteristic nearly idiomorphic grey-blue microcline augen, rimmed by plagioclase
	dark Stachów gneiss	N part of the Lipowe Hills Massif but also found in the Strzelin Hills	dark grey	fine-grained texture, often polygonal shape of grains, streaky, stromatite structure	foliation defined by mica, feldspar and quartz layers; lineation by streaks of biotite	sillimanite nodules, association with mica schists and amphibolites
	pale Stachów gneiss	Lipowe Hills but also found in the Strzelin Hills	light grey, nearly white	coarse-grained structure, layered, pseudo-augen	foliation defined by mica concentrations, feldspar and quartz layers; lineation defined by elongated concentrations of biotite	thick concentrations of biotite
	Henryków gneiss	S part of the Lipowe Hills	greenish-grey	very fine-grained and fine layered	foliation defined by thin layers of feldspars, quartz , sericite and chlorite; lineation pronounced as streaks of sericite	strong mylonitization

The metamorphic history and P-T conditions in the Stachów complex in the northern and middle part of the Lipowe Hills were similar to those in the Nowolesie gneiss. However, the metamorphism differs between individual klippen of the Stachów Complex, but is generally compatible with that observed in the nearest rocks belonging to the Strzelin complex. This suggests that the peak metamorphic grade of the Stachów complex was established during the M_1 metamorphic event at the time of, and shortly after, thrust stacking.

THE TECTONIC CONTACT BETWEEN THE STRZELIN AND THE STACHÓW COMPLEXES

The contact between the Strzelin and the Stachów complexes is exposed in the Gościęcice klippe, where the Gościęcice gneiss was thrusted over amphibolites of the inner envelope, and in the Henryków klippe, where the Henryków gneiss was thrusted over the Jegłowa Beds and Nowolesie gneiss. Along the contacts, the rocks of the adjoining complexes were strongly mylonitised within zones, several tens of metres in thickness (Oberc-Dziedzic and Madej, 2002).

The Gościęcice gneiss in the Gościęcice klippe shows N-dipping S_1 foliation and NNW to N and NE plunging L_1 lineation, the effects of top to the N–NNE shearing in the course of thrusting during the D_1 deformation. A younger, striae-resembling lineation also plunging to the N, is connected with the reactivation of the S_1 foliation and the thrust zone during extensional deformation, corresponding to the D_4 event in the Strzelin complex (Oberc-Dziedzic and Madej, 2002). The thrust zone here comprises several bodies 0.1 to 1 m in size of strongly mylonitised, garnet bearing gneisses, and bodies of garnet-bearing, relatively HP amphibolite. These rocks differ from the Gościęcice gneiss and from the amphibolites of the inner envelope in their contrasting P-T path and thus are interpreted as tectonic lenses (Oberc-Dziedzic and Madej, 2002). Reactivation of the thrust zone is indicated by kinematic indicators showing a top-to-N sense of shear. The two deformation events D_1 and D_4 took place under amphibolite- and greenschist facies conditions, respectively.

In the Henryków klippe, effects of D_1–D_5 deformation events are recorded. The main feature of the Henryków gneiss is extremely strong mylonitisation, which took place during deformation D_1, under metamorphic conditions within the biotite stability field. The D_1 event gave rise to a mylonitic S_1 foliation. During the D_2 event, the S_1 foliation was deformed by isoclinal F_2 folds with axes plunging to the SW. The S_2 penetrative foliation and L_2 stretching lineation are parallel to the F_2 fold axial planes and axes, respectively. During the D_3 deformation event, the S_2 foliation was reactivated and complex S_{2+3} foliation was produced locally. It dips to the SW at a low angle and is parallel to the axial planes of F_3 folds with N–NW plunging axes. The L_3 mineral lineation, defined by muscovite and chlorite, plunges at 25–35° to the SW. Kinematic indicators show top-to-SW sense of shear during D_3 deformation under a greenschist facies conditions. The D_4 deformation produced non-penetrative S_4 shear zones, dipping at 30–35° to the SW and related F_4 folds. Their axes plunge to the SW. The last D_5 deformation produced kink folds with axes trending NW–SE and plunging at angles of 10–20°.

The contact zone of the Henryków klippe and the underlying Jegłowa Beds is not exposed, but the observed strong mylonitisation on both sides of the boundary (Oberc-Dziedzic and Madej, 2002) points to its tectonic character.

The above evidence, in particular the wide, mylonitic contact zone with tectonic inclusions of relatively high-grade rocks, different metamorphic histories of the adjoining rocks in the Gościęcice klippe, as well as strong mylonitisation of the Henryków gneiss and quartzites in the Henryków klippe, all strongly suggest that the boundary between the Strzelin complex and the Stachów complex is a tectonic contact. This structure was defined by Oberc-Dziedzic and Madej (2002) as the Strzelin Thrust. Its footwall and hanging wall comprise rocks of the Strzelin complex and the Stachów complex, respectively.

VARISCAN GRANITOIDS

The Strzelin and the Stachów complexes were intruded by four groups of Variscan granitoids:

— granodiorites,

— tonalites and quartz diorites,

— medium- and fine-grained biotite granites (347 12 Ma, Rb-Sr whole rock),

— two mica granites (330 6 Ma, whole rock) (Oberc-Dziedzic *et al.*, 1996; Oberc-Dziedzic and Pin, 2000). All these granitoids show a weak magmatic lineation, indicating that their emplacement occurred at the very end of the Variscan deformation. This also indicates that the Variscan deformation in this area, i.e. the NE part of the Fore-Sudetic Block, was not younger than Viséan.

THE GNEISSES OF THE STRZELIN AND THE STACHÓW COMPLEXES: PETROGRAPHY, GEOCHEMISTRY AND AGES

PETROGRAPHY

The gneisses of the Strzelin and Stachów complexes are composed of quartz, plagioclase, microcline, biotite and muscovite in various proportions. The mineralogy of these rocks is typical of peraluminous S-type granites or granodiorites; their diagnostic features are presented in Table 2. Petrogenetic studies of zircons also document that the gneisses are derived from protoliths of crustal affinity (Klimas *et al.*, 2001, 2002, 2003).

GEOCHEMISTRY

Five samples of the main types of orthogneiss were analysed for major, trace and rare earth elements in Actlabs, Canada, using combined *ICP-OES* and *ICP-MS* techniques (Actlabs code "4Lithores"). The analysed orthogneisses (Table 3) are high-potassic, calc-alkaline rocks, containing 70–76% of SiO_2 and variable amounts of TiO_2. Their peraluminous character is indicated by the molar proportion $Al_2O_3/(CaO+Na_2O+K_2O)$, ranging between 1.1 in the pale Stachów gneiss and 1.23 in the Strzelin gneiss, and by the CIPW normative corundum values of 1.54 and 2.65, respectively.

The gneisses from the Strzelin complex show lower absolute abundances of total REE (the Strzelin gneiss — 108 ppm and the Nowolesie gneiss — 73 ppm) and higher Eu/Eu* (0.8, 0.79) than gneisses from the Stachów complex (144–223 ppm; Eu/Eu* — 0.31–0.52) (Table 3).

The chondrite-normalised REE plots (Fig. 4A) reveal further considerable geochemical variation of the gneisses and enables the following subdivision:

— Type 1, including the Gościęcice, pale Stachów and Henryków gneisses from the Stachów complex and, possibly, the significantly distinct Nowolesie gneiss from the Strzelin complex;

— Type 2, represented only by the Strzelin gneiss.

Type 1 is characterised by two-sectional distribution pattern of REE, i.e. a flat, nearly horizontal HREE section, and a considerably enriched LREE section, indicating strong fractionation of the lightest REE (up to *ca*. 100 times chondrite values for La). The $(La/Yb)_N$ ratio ranges between 5.0 and 6.5. A distinct feature of these rocks is a strong negative Eu anomaly and a weak positive Tm-anomaly.

The Nowolesie gneiss shows a REE pattern nearly parallel to that of the Type 1 samples, but the concentration of REE is much lower at 5–6 times chondrites for HREE, and *ca*. 50 times chondrites for La. The negative Eu-anomaly is much weaker, but the positive Tm is also noticeable. The $(La/Yb)_N = 8.2$ is similar to the other Type 1 samples.

Type 2, the Strzelin gneiss, shows a very different REE pattern, with strong and uniform differentiation, from La through Lu. The $(La/Yb)_N = 38.3$ is much higher than for the Type 1 gneisses. The Eu anomaly is very weak, and no Tm anomaly can be seen.

The above differences in the REE patterns of the gneisses are also observed in their multi-element primitive-mantle-normalised plots (Fig. 4B), although the latter are not so distinct. Worth noticing is a very low concentration of Y in the Strzelin gneiss (6 ppm) and pronounced negative anomalies for Nb, Sr, P and Ti in most samples. The Nowolesie gneiss shows some similarities to the Strzelin gneiss on this diagram: nearly parallel distribution of elements on the left-hand side (for Rb, Th, U and K), strong negative U- and no distinct Sr anomalies.

Type 1 gneisses could have originated from a common crustal source and developed through roughly similar petrogenetic processes which produced strong LREE differentiation and practically no HREE fractionation. The flat HREE pattern may indicate the presence of pyroxene and/or hornblende in the source (Hanson, 1978). The combined negative Eu- and Sr anomalies in Type 1 gneisses suggest that plagioclase was either present in the source or was (partly?) removed during fractional crystallisation (Green, 1980). The high Th contents are typical of the upper crust materials. Also, the negative anomalies for U, Nb, P and Ti are characteristic of rocks derived from the continental upper crust (Wilson, 1991).

The Nowolesie gneiss may have developed from a similar crustal source, but experienced different processes which resulted in lower concentrations of incompatible trace-ele-

Table 3

Major (wt %) and trace-element (ppm) whole rock analyses of gneisses

Age	Neoproterozoic		Late Cambrian/Early Ordovician		
Sample	Strzelin gneiss	Nowolesie gneiss	Gościęcice gneiss	Stachów gneiss	Henryków gneiss
SiO_2	74.96	75.73	69.74	70.44	72.04
TiO_2	0.119	0.06	0.598	0.434	0.29
Al_2O_3	14.21	13.41	14.68	14.63	14.36
Fe_2O_3	0.87	1.87	3.92	3.84	3.04
MnO	0.018	0.02	0.05	0.035	0.03
MgO	0.52	0.5	1.28	0.77	0.59
CaO	0.37	0.63	1.79	1.46	0.89
Na_2O	3.23	3.38	3.03	3.36	3.74
K_2O	5.14	4.41	4.06	4.8	4.14
P_2O_5	0.05	0.04	0.19	0.16	0.15
LOI	0.66	0.22	0.79	0.18	0.96
Total	100.13	100.27	100.11	10013	100.23
A/CNK	1.23	1.17	1.16	1.09	1.17
A/NK	1.3	1.3	1.56	1.36	1.35
Ba	1260	916	902	785	803
Rb	112	110	148	147	116
Sr	242	272	131	99	115
Y	5.8	12	33.9	54.8	37
Zr	76	58	221	223	139
Nb	12.7	6.7	14.2	13.1	8.6
Th	6.06	5.88	12.9	20.5	13.4
Pb	14	16	33	27	20
Ga	16	16	19	21	18
Zn	–	0	54	44	35
Cu	–	20	15	16	15
Ni	–	0	–	0	0
V	–	0	61	37	27
Cr	–	0	42	28	23
Hf	2.2	2	6.3	6.9	5.4
Cs	2.2	0.9	5.4	4.6	1.3
Sc		2	–	8	6
Ta	6.73	0.6	3.35	0.89	0.9
Co	15	2	12	6	5
Li		1.91	–	1.54	2.35
Be	2	1	2	–	2
U	–	1.04	–	–	2.33
W	–	0	–	–	2.1
Sn	–	2	–	–	4
Mo	–	0	–	–	0
La	26.2	17.1	33.4	41.4	27.2
Ce	50.1	31	70.1	89.3	55
Pr	5.16	3.12	7.7	9.45	6.51
Nd	18.1	11.4	29.9	37.3	25.9
Sm	2.94	2.1	6.06	7.96	5.56
Eu	0.649	0.527	0.986	0.854	0.709
Gd	2.13	1.99	5.6	8.86	5.98
Tb	0.25	0.34	0.97	1.67	1.11
Dy	1.23	1.97	6	10.6	6.28
Ho	0.2	0.42	1.17	2.2	1.33
Er	0.55	1.31	3.47	6.39	4.25
Tm	0.073	0.22	0.554	0.956	0.683
Yb	0.49	1.4	3.43	5.42	3.65
Lu	0.073	0.228	0.5	0.741	0.55
REE	108.15	73.13	169.84	223.10	144.71
La_N/Yb_N	35.75	8.17	6.51	5.11	4.98
Eu/Eu*	0.8	0.79	0.52	0.31	0.38

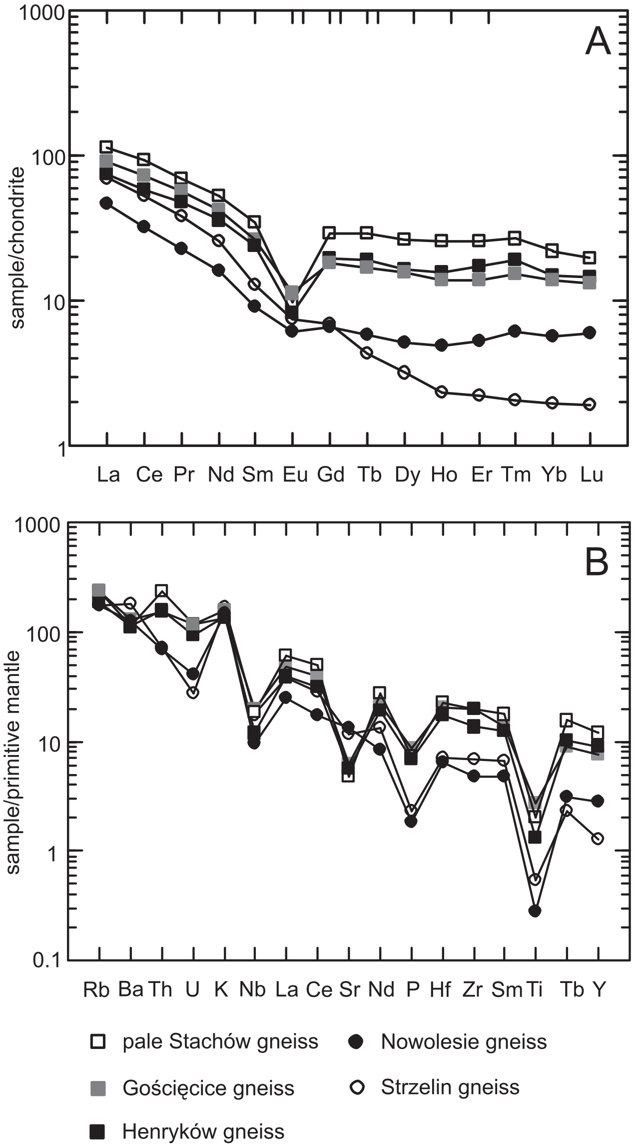

Fig. 4. A — chondrite normalized multi-element variation diagram; B — "primitive-mantle"— normalized REE plot for the main types of gneisses from the Strzelin complex and the Stachów complex

The normalizing values from Sun and McDonough (1989)

ments (e.g. more intensive partial melting or less evolved fractionation).

The Strzelin gneiss shows considerably different trace element patterns indicating a different source compared to that of Type 1 gneisses. The observed low contents of Y in the Strzelin gneiss suggests presence of garnet in the source (Drummond and Defant, 1990).

AGES

GNEISSES OF THE STRZELIN COMPLEX

SHRIMP zircon ages of 600 7 and 568 7 Ma were obtained for the Strzelin gneiss which belongs to the gneissic core (Oberc-Dziedzic *et al*., 2003*a*). The older age was interpreted as the time of magmatic crystallisation of the gneiss protolith, whereas the younger age may refelect a partial melting event (*op. cit.*). The SHRIMP study also indicated the presence of inherited zircon cores with Palaeo- to Mesoproterozoic $^{206}Pb/^{238}U$ ages between 1879 29 Ma and 1230 8 (Table 1). The Nowolesie gneiss, also from the gneissic core, yielded a mean $^{207}Pb/^{206}Pb$ age of 1020 1 Ma, using the single grain evaporation method (Kröner and Mazur, 2003, data for the Skalice migmatitic gneiss). Xenocrysts in that gneiss vary in age between 1135 and 1767 Ma, roughly corresponding to the age of xenocrysts from the Strzelin gneiss. This implies that the gneissic core, being part of the footwall of the Strzelin Thrust, is composed of at least two different intrusive phases, but derived from similar sources. However, more detailed geochemical and isotopic data are necessary to better constrain genetic links or differences between the gneisses.

SHRIMP zircon ages of 600 7–568 7 Ma for the Strzelin gneiss (Oberc-Dziedzic *et al*., 2003*a*) are similar to ages of gneisses from elsewhere in the Brunovistulian Terrane (Table 1). They confirm previous interpretations of the Strzelin gneiss to correlate with gneisses of the Moravo-Silesian Zone (Bederke, 1929, 1931; Oberc, 1966), pointing to the Keprník gneiss (584±8 Ma, Kröner *et al*., 2000) as an equivalent. Therefore, these data support an East Sudetes affinity for the Strzelin Massif. Apart from that, the Keprník gneiss of the Moravo-Silesian Zone resembles the high-K granitoids of the western part of the Brunovistulian Terrane further south (Finger *et al*., 2000). It thus appears that fragments of the Brunovistulian Terrane can be traced not only into the East Sudetes but also across the Sudetic Marginal Fault into the eastern part of the Fore-Sudetic Block (Oberc-Dziedzic *et al*., 2003*a*).

GNEISSES OF THE STACHÓW COMPLEX

The U-Pb and Pb-Pb zircon ages of 504±3 Ma (Oliver *et al*., 1993) and 513±1 (Kröner and Mazur, 2003), interpreted as the crystallization ages of the magmatic protolith, were obtained for the Gościęcice biotite gneiss (Table 1). The minimum ages for xenocrysts range between 1096 and 1301 Ma (Kröner and Mazur, 2003).

Our SHRIMP zircon dating of the pale Stachów gneiss (Oberc-Dziedzic *et al*., 2003*b*) indicates the presence of inherited zircon cores of Palaeo- to Neoproterozoic $^{206}Pb–^{238}U$ ages of 1916±25, 636.3±8.4 and 560.3±6.9 Ma, as well as mostly euhedral and zoned crystals of Late Cambrian age, with a mean of 500±5 Ma (Table 1), interpreted as the emplacement age of the magmatic precursor of the gneiss.

The ages of inherited zircons from the Stachów and Gościęcice gneisses suggest that the source material for these two gneisses and for the Strzelin gneiss (having inherited zircon cores of 1230–1870 Ma) were different and that the Strzelin gneiss was probably not the source material for the Stachów and Gościęcice gneisses.

The late Cambrian ages for the augen Gościęcice gneiss and the pale Stachów gneiss are similar to those of the West Sudetes orthogneisses, e.g. the gneisses in the Orlica-Śnieżnik Dome (Oliver *et al*., 1993; Turniak *et al*., 2000; Kröner *et al*., 2001; Table 1).

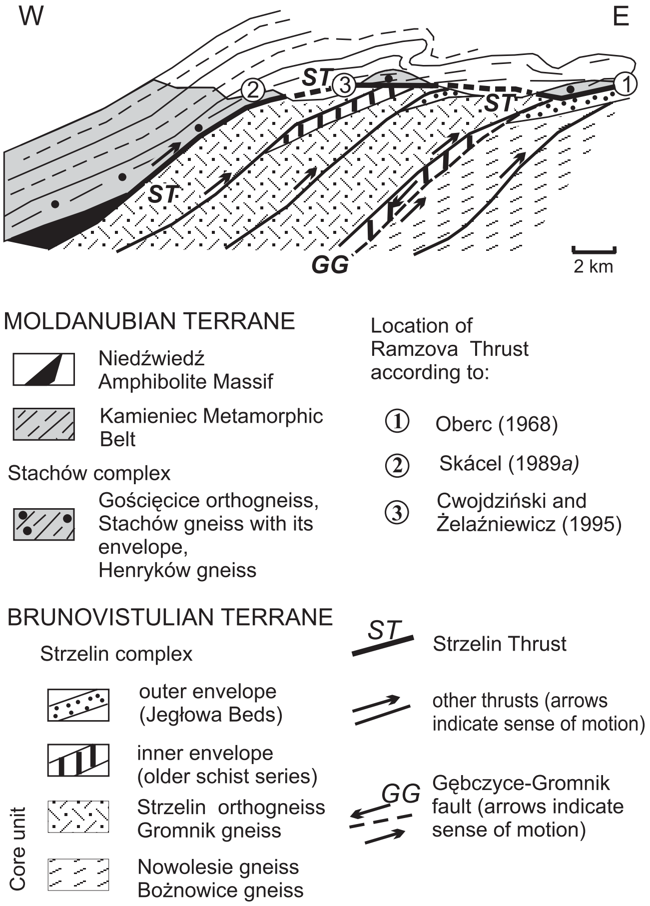

Fig. 5. Schematic section across eastern part of the Fore-Sudetic Block showing position of the Strzelin Thrust — the postulated extension of the boundary between the West and East Sudetes (after Oberc-Dziedzic and Madej, 2002, modified)

DISCUSSION

The boundary between the West and East Sudetes forms part of a major tectonic contact, possibly a fragment of the suture zone between two large tectono-stratigraphic units, recently referred to as the Moldanubian Terrane and the Brunovistulian Terrane. The Sudetic and the Fore-Sudetic parts of these terranes, apart from important similarities (see Aleksandrowski and Mazur, 2002 and data presented above), show significant differences.

The Moldanubian Terrane in its Fore-Sudetic part, in contrast to the Orlica-Śnieżnik Dome in the Sudetes, does not contain Early Palaeozoic gneisses (except for small bodies, such as the pale Stachów or Gościęcice gneiss) but, instead, Late Devonian syncollisional granites with an age of 380±1 Ma, transformed into the Doboszowice granite-gneiss (Kröner and Mazur, 2003). Apart from that, there is no clear equivalent of the Staré Město Belt (which extends along the eastern margin of the Orlica-Śnieżnik Dome) in the Fore-Sudetic Block.

However, the Brunovistulian Terrane in its Fore-Sudetic part (in the Strzelin Massif), again in unlike the East-Sudetic section, contains only minor Neoproterozoic metasedimentary rocks in the inner envelope, and Early- to Middle Devonian outer envelope is represented by monotonous, mostly quartzitic sedimentary rocks (the Jegłowa Beds). Furthermore, the rocks of the Strzelin Massif experienced regional metamorphism at relatively high T and low P, compared to the rocks from the East Sudetic part of the Brunovistulian Terrane.

The collisional zone between the two terranes in the Fore-Sudetic segment is intruded by relatively small bodies of late-orogenic Variscan granitoids, the distribution of which is controlled by shear zones and faults. In the western Moldanubian part, these are represented by the Niemcza granodiorites and diorites (338±3 Ma; Oliver *et al.*, 1993), and in the eastern Brunovistulian part (in the Strzelin Massif) by diorites, tonalites and biotite granites (347±12 Ma) and two-mica granites (330±6 Ma; Oberc-Dziedzic *et al.*, 1996; Oberc-Dziedzic and Pin, 2000).

The sequence and orientation of the Variscan structures, although generally similar in both terranes, also show some differences. In the Kamieniec Metamorphic Belt of the Moldanubian part of the Fore-Sudetic Block, the first tectonic event D_1 gave rise to nappe stacking due to W–E or WNW–ESE contraction (Mazur and Józefiak, 1999). In the Strzelin Massif (part of the Brunovistulian Terrane), the W–E contraction and thrusting can only be inferred from the presence of fragments of the Stachów complex inside the Strzelin Massif. The sense of tectonic transport cannot be established from kinematic indicators which in both the Strzelin complex and in the Stachów complex, persistently point to top-to-N–NNE shearing during the D_1 event (Cymerman, 1993*b*; Oberc-Dziedzic and Madej, 2002). The WNW–ESE or W–E contraction, the E–ESE thrusting and top-to-N–NNE shearing occurred at the same time and suggest a bulk triclinic transpressional deformation regime involving components of pure shear contraction and oblique simple shear (Holdsworth *et al.*, 2002; Oberc-Dziedzic and Madej, 2002) during collision of the Moldanubian and Brunovistulian terranes. The following D_2 deformation event produced F_2 folds trending N–S and NNE–SSW–SW in both terranes. The subsequent $D_{3–4}$ events involved extensional collapse directed to the NNE in the northern part of the Strzelin Massif, and to the SW in its southern part and in the Lipowe Hills Massif (Madej, 1999; Szczepański, 2001; Szczepański and Mazur, 2004), as well as in the Kamieniec Metamorphic Belt (Mazur and Józefiak, 1999).

In the Sudetic mountainous section, the suture zone between the Moldanubian and Brunovistulian terranes, i.e. the boundary between the East and West Sudetes (the Ramzova/Nyznerov Thrust), follows the eastern side of the Staré Město Belt, a composite unit with *ca*. 500 Ma ages of igneous protoliths (Kröner *et al*., 2000; Štípská *et al*., 2004), deformed and metamorphosed during Carboniferous collision (Skácel, 1989*a*; Schulmann and Gayer, 2000; Aleksandrowski and Mazur, 2002). No clear extension of this belt into the Fore-Sudetic Block has been found. This may be explained by deeper erosion of the Fore-Sudetic part compared with the recently mountainous section (Oberc, 1968; Guterch *et al*., 1975; Skácel, 1989*b*). Skácel (1989*a*) placed the boundary between the East and West Sudetes in the Fore-Sudetic Block, along the eastern margin of the Niedźwiedź Amphibolite Massif (Fig. 1) which may be equivalent to the Staré Město Belt. Further north of the Niedźwiedź Massif, the suture may follow the Strzelin

Thrust, separating the Strzelin complex and the Stachów complex (Fig. 5).

The gneisses of the Strzelin and Stachów complexes differ in their fabric, geochemistry, and ages. The tectonic juxtaposition of the Strzelin complex with the Neoproterozoic Strzelin and Nowolesie gneisses and the structurally overlying Stachów complex, containing Early Palaeozoic Gościęcice gneiss and the pale Stachów gneiss, resembles the situation along the Ramzova/Nyznerov Thrust further south. The latter is interpreted as the boundary separating two domains with contrasting gneiss protolith ages, namely the Moldanubian and Brunovistulian terranes. Consequently, the Strzelin Thrust can be considered as a continuation of the tectonic boundary between the East and West Sudetes within the Fore-Sudetic Block. At the present erosional level, the hanging wall of the Strzelin Thrust is represented by the Stachów complex, preserved in the form of klippen recognised in the Lipowe Hills and Strzelin massifs. However, other klippen may be present beyond these massifs, such as the Maciejowice gneiss of Early Palaeozoic age (501±1 Ma, Kröner and Mazur, 2003) further SSE of the Strzelin Massif. The longest distance between the klippen in the Lipowe Hills and Strzelin massifs is over 10 km, and this is probably the minimum transport distance along the Strzelin Thrust. Oberc (1968) previously estimated the minimum amplitude of the Ramzova Thrust at 17 km.

The surface of the Strzelin Thrust has a dome-like shape which was established during the origin of the Przeworno Elevation and extensional deformation, corresponding to the D_4 event in the Strzelin complex. At the base of the Gościęcice klippe, the thrust surface dips to NNE at *ca*. 25° (Oberc-Dziedzic and Madej, 2002) but generally, it has subhorizontal orientation, as inferred from similar hypsometric positions of the klippen. This is in contrast with the Ramzova/Nyznerov Thrust, which dips to the W at an angle of 40–50° (Misař *et al*., 1983). The Strzelin Thrust surface becomes steeper east of the Strzelin Massif, where it is hidden beneath the Cainozoic sediments, and west of the Lipowe Hills Massif, where it follows the eastern border of the Kamieniec Metamorphic Belt. The strongly mylonitised mica schists exposed along that border (the so-called Mała Ślęza river section, Fig. 2) may represent the root zone of the hanging wall. Changes in the inclination of the generally shallow dipping thrust surface may partly explain the previous difficulties in locating the extension of the boundary between the East and West Sudetes in the Fore-Sudetic Block.

CONCLUSIONS

Geochronological, petrological, geochemical and structural data suggest that the Strzelin Thrust is the northern extension of the boundary between the East and West Sudetes, i.e. it forms a part of the boundary between the Brunovistulian and Moldanubian terranes within the poorly exposed NE part of the Bohemian Massif.

The footwall and the hanging wall of the thrust are represented by the Strzelin complex and the Stachów complex, respectively. The Strzelin complex comprises Proterozoic gneisses: the Strzelin gneiss, with zircon ages of 600±7 and 568±7 Ma, and the Nowolesie gneiss with a mean zircon age of 1020 1 Ma. The Stachów complex contains the Gościęcice gneiss and the pale Stachów gneiss, both yielding Late Cambrian zircon ages (~500±5 Ma).

The orthogneisses in both complexes correspond to peraluminous S-type granites, but have different inherited zircon ages and display contrasting trace element characteristics, indicating different sources and petrogenetic histories.

Based on the ages, petrology and overall geological context, the Strzelin orthogneiss is similar to the Keprník orthogneiss of the East Sudetes, whereas the orthogneisses of the Stachów complex correspond to rocks known from the West Sudetes (e.g. the Izera and Śnieżnik orthogneisses).

The tectonic juxtaposition of the Strzelin complex, including the Neoproterozoic gneisses, and the structurally overlying Stachów complex, containing Early Palaeozoic gneisses, resembles contact between the Moldanubian and Brunovistulian terranes, i.e. the Ramzova/Nyznerov Thrust and the Moldanubian Overthrust to the south.

The hanging wall of the Strzelin Thrust, represented by the Stachów complex, is preserved in the form of a klippen. The minimum transport distance along the Strzelin Thrust is estimated at over 10 km.

The surface of the Strzelin Thrust has a dome-like shape. Along the contacts, the rocks of the adjoining complexes are strongly mylonitised within zones several tens of metres in thick.

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