The Variscan overthrust of the Lower Palaeozoic gneiss unit on the Cadomian basement in the Strzelin and Lipowe Hills massifs, Fore-Sudetic Block, SW Poland; is this part of the East-West Sudetes boundary?

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Abstract   The problem of the position of the boundary between the geological structures of the West and the East Sudetes has been a topic of discussion since 1912, when F.E. Suess developed the concept of the Moldanubian overthrust as a boundary between the Moldanubian and Moravo-Silesian zones. The West Sudetes comprise gneisses of Cambro-Ordovician protolith age with inclusions of high pressure metamorphic rocks. The Cadomian basement, referred to as the Brunovistulian and overlain by Devonian rocks, is characteristic of the East Sudetes. The location of the East-West Sudetes boundary is well-defined in the mountainous part of the Sudetes but still a matter of debate in the Fore-Sudetic Block. This paper puts forward a new approach to this problem.

The Proterozoic Strzelin gneiss with its Proterozoic (the older schist series) and Devonian envelope (the Jegłowa beds) are tectonically overlain by the Early Palaeozoic Gościęcice gneiss and the light Stachów gneiss with its envelope. The former occurs in the footwall and the latter in the hanging wall rocks of the Strzelin Thrust. This juxtaposition resembles the situation along the East-West Sudetes boundary separating two domains with contrasting protolith ages. Consequently, the Strzelin Thrust is considered part of the border zone between the East and West Sudetes, i.e. the northern continuation of the Ramzova/Nyznerov thrust to the Fore-Sudetic Block. At the present erosion level, the hanging wall rocks of the Strzelin Thrust are separated from their roots and form klippen. The minimum transport distance along the thrust is estimated at 10 km. The Strzelin Thrust forms a generally shallowly dipping domed surface. It becomes steeper east of the Strzelin massif, where it is hidden beneath Cenozoic sediments, and west of the Lipowe Hills, where it follows the eastern border of the Kamieniec Ząbkowicki Metamorphic Complex. The hanging wall is probably rooted in the strongly mylonitised mica schists exposed along the Mała Ślęza river. The thrust zone is a wide mylonitic belt carrying relatively HT and HP garnet-bearing amphibolites in the northern part of the Strzelin massif and the strongly mylonitised Henryków gneiss and quartzites in the southern part at the Lipowe Hills.

This paper discusses the problem of the East-West Sudetes boundary, mainly in the framework of the Strzelin massif. The attitude of this boundary in other parts of the Fore-Sudetic Block is still unclear because of poor exposure and numerous faults of E–W and NW–SE orientation that displace it from its original position.

INTRODUCTION

The eastern part of the Variscan belt comprises a collage of several tectonostratigraphic units like the Saxothuringian, Moldanubian or Moravo-Silesian Zones, mostly defined after Kossmat (1927). They were more recently interpreted as terranes (Matte *et al.*, 1990; Franke & Żelaźniewicz, 2000). Suess (1912, 1926) was the first to develop the concept of the Moldanubian overthrust at the boundary between the Moldanubian and Moravo-Silesian zones. According to the presently-held view (Schulmann &

Gayer, 2000), the Moravo-Silesian zone is a NE–SW-trending belt which resulted from the oblique Variscan collision between the Moldanubian-Lugian terrane (Moldanubicum and Lugicum *sensu* Suess, 1912) and the pan-African Brunovistulian domain (Bruno-Vistulicum, Dudek, 1980), interpreted as a microcontinent (Matte *et al.*, 1990; Finger *et al.*, 2000). During the collision, the western part of the Brunovistulian domain, composed of Neoproterozoic, high-grade metamorphic rocks and granitoids over-

FORE-SUDETIC SECTION

Strzelin massif
- Devonian metasedimentary rocks (the Jegłowa beds)
- gneisses
- Niedźwiedź massif amphibolites
- Doboszowice gneisses, amphibolites
- Niemcza-Kamieniec mica schists
- Niemcza Zone mylonites

Ophiolitic complexes
- gabbros, serpentinites
- Góry Sowie gneisses and migmatites

SUDETIC SECTION

Bardo unit
- Palaeozoic sedimentary rocks

Culm Basin
- Lower Carboniferous flysch

Staré Město belt
- metaigneous and metasedimentary rocks

Orlica-Śnieżnik complex
- undifferentiated orthogneiss and metasedimentary rocks

Desna dome
- Devonian metasedimentary cover rocks
- para-autochthonous basement units

Keprnik nappe
- Devonian low-grade rocks (Branna Group)
- metasedimentary rocks
- orthogneiss

Velké Vrbno unit
- undifferentiated metaigneous and metasedimentary rocks

faults:
SMF - Sudetic Marginal Fault,
DF - Diendorf Fault
Przeworno elevation
Ramzova Thrust: location in the Fore-Sudetic block according to:
A - Bederke (1929),
B - Oberc (1968),
C - Skacel (1989 b)
thrusts:
MT - Moldanubian Thrust,
NT - Nyznerov Thrust,
RT - Ramzova Thrust - location in the Sudetes
Mesozoic and Cenozoic sediments
Variscan granitoids

MORAVIAN SECTION

Moldanubian
- undifferentiated metamorphic and igneous rocks

Moravian
- undifferentiated metamorphic rocks
- Variscan deformed granitoids: Bi-Bittesch, Sv-Svratka, W-Weitersfeld gneiss

Brunovistulian Slavkov terrane
- granodiorites / leucogranites type Stupava
- granitoids type Slavkov
- granitoids type Zdanice

Brunovistu;ian Central Basic Belt
- metamorphosed: gabbros, diorites, basalts, ultrabasites, rhyolites

Brunovistulian Thaya terrane
- granodiorites and granite / type (Mikulov)
- tonalites

Wrocław
Praha
100 km

Strzelin
Poland
C B
A
SMF
Nysa
RT
NT
Jesenik
N
20 km

Bi
MT
Bi
Sv
Brno
Czech Republic
MT
Znojmo
Bi
W
Austria
Eggenburg
DF
Krems

Fig. 1. Eastern margin of the Bohemian Massif. Shadowed area near SMF refers to the part of the Ramzova thrust discussed by Oberc (1968).

Fig. 2. Regional setting of the Strzelin massif within the eastern part of the Fore-Sudetic Block. Compiled on the basis of Puziewicz *et al.* 1999 and Oberc *et al.* 1988.

lain by a Devonian–Carboniferous cover, was sheared, metamorphosed and piled up NE-ward into a nappe sequence giving rise to the Moravo-Silesian zone.

The Moravo-Silesian zone can be divided, from the south to the north, into the Moravian, Sudetic and Fore-Sudetic sections, which show some differences (Fig. 1).

In the Moravian section, the Moldanubian terrane, which is composed of high grade gneisses and minor bodies of granulites and eclogites, was thrust onto the Moravo-Silesian zone, which emerges from it in the Thaya window and the Svratka window. The border between the Moldanubian and Moravo-Silesian zones was defined by Suess (1912) as the Moldanubian overthrust. The Brunovistulian domain, situated east of the Thaya and Svratka windows, consists of the large granitoid complex of the Thaya Batholith and the Brno Batholith. The Brno Batholith is cut in a north-south direction by a narrow Central Basic Belt, which is interpreted as a boundary between two terranes showing different characteristics of the continental crust: the Thaya terrane comprising the Thaya Batholith and the western part of the Brno Batholith and

the Slavkov terrane containing the eastern part of the Brno Batholith (Finger & Pin, 1997; Finger *et al.*, 2000). The Brunovistulian granitoids have an age of around 580–590 Ma; the age of the basic rocks has been established as 725 ± 15 Ma (Finger *et al.*, 2000 and references therein).

In the Sudetic section (Fig. 1), the Moravo-Silesian zone adjoins the Lugian (*sensu* Suess, 1912) domain. The upper part of the Lugian domain is composed of medium-grade metasedimentary rocks and gneisses derived from c. 500 Ma Early Palaeozoic granites (Turniak *et al.*, 2000). The gneisses contain small bodies of granulites and eclogites. The lower part of the Lugian domain is represented by the Staré Město belt, which comprises an Early Palaeozoic sequence of a leptino-amphibolite complex associated with tonalitic gneisses and metasediments intruded by granodiorites. According to Schulmann & Gayer (2000), it documents the Cambro-Ordovician rifting. The Sudetic section of the Moravo-Silesian zone can be subdivided into a low-grade eastern Desna dome passing eastwards into the Culm foreland basin, medium-grade western Keprnik nappe and Velke Vrbno unit (Schulmann &
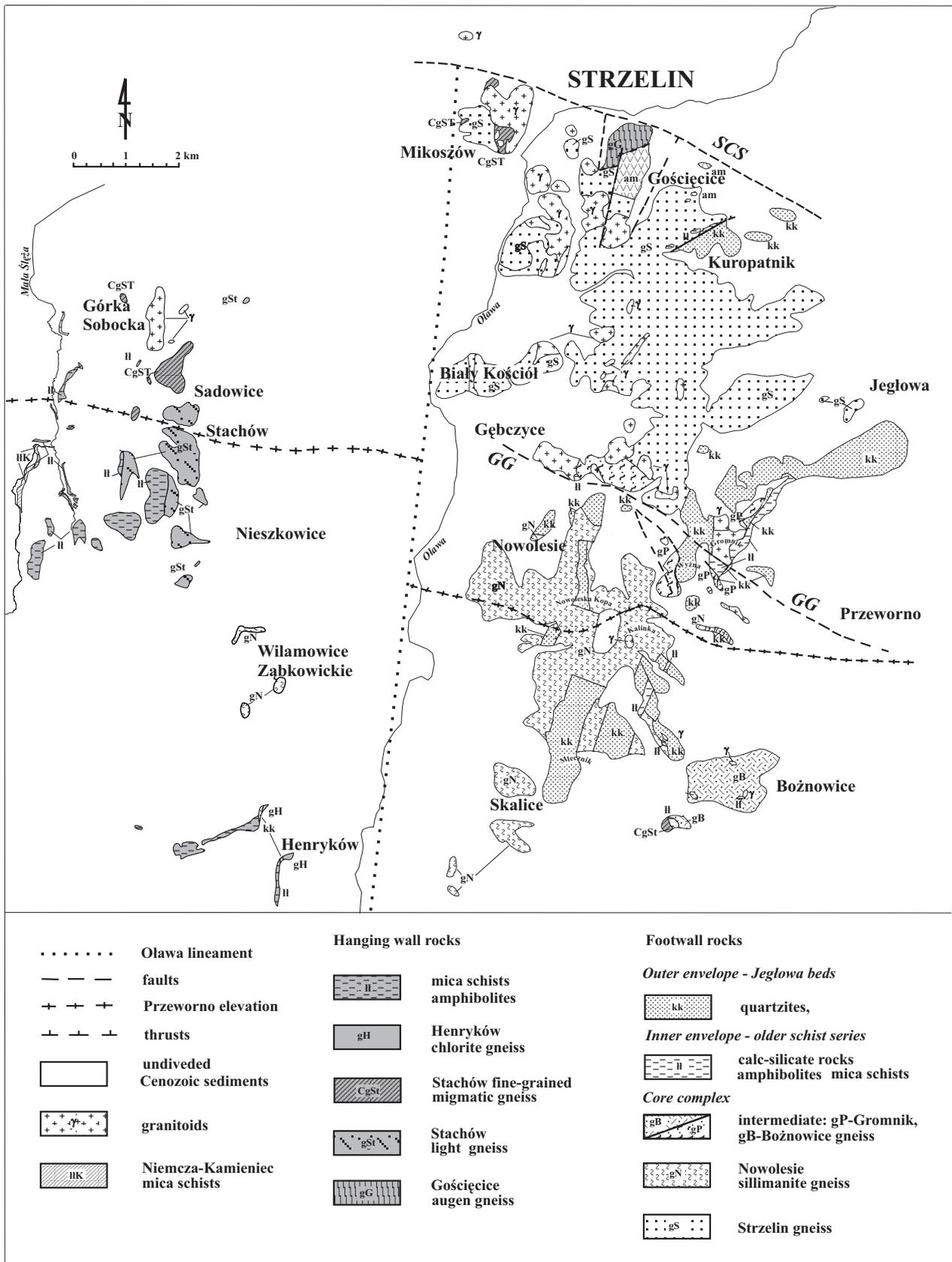
Fig. 3. Geological map of the Strzelin massif (according to Oberc *et al.* 1988, simplified) and the Lipowe Hills massif (Wójcik 1968, Wroński 1973, Badura 1979). The shaded areas show the position of the probable equivalents of the Strzelin Thrust. *SCS* – Sienice-Strzelin fault; *GG* – Gębczyce-Gromnik fault.
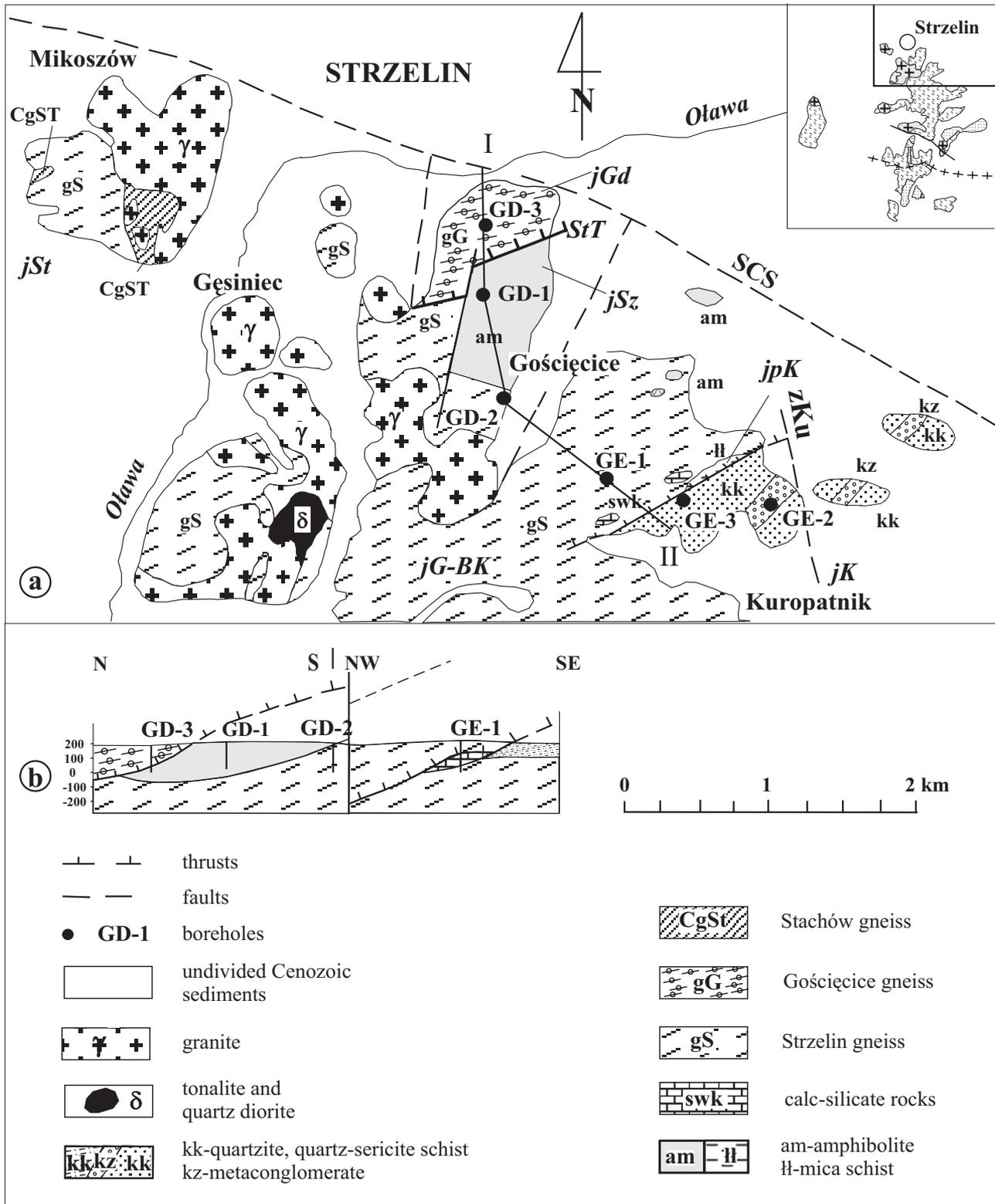
Fig. 4. Geological map of the northern part of the Strzelin massif. b) Section across the northern part of the Strzelin massif. *jSt* – Strzelin unit; *jSz* – Szańcowa Góra unit; *jG-BK* – Gościęcice-Biały Kościół unit; *jGd* – Gościęcice Dolne unit; *jpK* – northern Kuropatnik unit; *jK* – southern Kuropatnik unit; *StT* – Strzelin thrust; SCS – Sienice-Strzelin fault; zKu – Kuropatnik fault.

Gayer, 2000). All the orthogneiss of the Sudetic section of the Moravo-Silesian zone have Neoproterozoic ages (Kröner *et al.*, 2000). The Moravo-Silesian zone was recently interpreted as a continental accretionary wedge developed by an oblique collision (Schulmann & Gayer, 2000). The Ramzova overthrust was considered a NE continuation of the Moldanubian overthrust and the boundary between the Lugian domain (belonging to the West Sudetes) and

the Moravo-Silesian domain (part of the East Sudetes) (Suess 1912, 1926; Bederke,1929, 1931; Skácel, 1956; Oberc, 1957; Misař, 1960). However, recently, the boundary between the Lugian and the Moravo-Silesian domains, i.e. between the West and East Sudetes, is considered to lie along the Nyznerov thrust (Skácel, 1989 a; Schulmann & Gayer, 2000), i.e. along the eastern side of the Staré Město belt. The presence of the basic rocks of the Staré Město belt along the west boundary of the Moravo-Silesian zone differentiates the Sudetic section from the Moldanubian and the Fore-Sudetic sections.

In the Fore-Sudetic section, the location of the East-West Sudetes boundary – the Ramzova/Nyznerov overthrust – is still a matter of debate. It has been variously located: along the eastern border of the mylonitic Niemcza zone (Bederke, 1929), east of the Strzelin massif (Oberc, 1968), west of this massif (Skácel, 1989 b) (Fig. 2) or inside it (Cwojdziński & Żelaźniewicz, 1995). According to Cymerman (1993), the Ramzova overthrust does not exist. The ambiguity of the East-West Sudetes boundary position was because of the uncertain protolith age of the Strzelin gneiss and the lack of stratigraphic data for the metasedimentary rocks of the Jegłowa beds (Fig. 3), which were regarded as equivalents of the Moravo-Silesian rocks – the Keprnik gneiss (584 ± 8 Ma, Kröner et al., 2000) and its Devonian cover, respectively (Bederke, 1929, 1931; Oberc, 1966). The age of the gneisses, mica schists, phyllites and metagreywackes found in boreholes situated east of the Strzelin massif is also unknown. The gneisses and mica schist have been interpreted by Sawicki (1995) as equivalents of the Odra Fault Zone metamorphic rocks, whereas the phyllites and metagreywackes are seen as equivalents of the Andelská Hora beds (Culm foreland basin according to Schulmann & Gayer, 2000). The age of

the gneisses and metasedimentary rocks exposed west and south of the Strzelin massif are also unknown, except for the Doboszowice gneiss (379 ± 1 Ma, Hanžl et al., 1998), so it is not clear if they represent part of the Moravo-Silesian zone (East Sudetes) or a part of the Lugian structure (West Sudetes).

New SHRIMP zircon ages of 600–568 Ma obtained for the Strzelin gneiss (Oberc-Dziedzic et al., in prep.) confirmed earlier interpretations of this gneiss as an equivalent of the Keprnik gneiss from the Moravo-Silesian zone (Bederke, 1929, 1931; Oberc, 1966) and therefore, the affinity of the Strzelin massif with the East Sudetes. The Strzelin gneiss and the Keprnik gneiss not only show a similar age of 590–600 Ma, but they are also similar geochemically. In turn, the Keprnik gneiss resembles the high-K granitoids of the western part of the Brunovistulian domain further southwards (Finger et al., 2000). It may mean that the fragments of the Brunovistulian domain can be traced not only in the East Sudetes, but also across the Sudetic Marginal Fault in the eastern part of the Fore-Sudetic Block to the north (Oberc-Dziedzic et al., in prep.). However, another type of gneiss in the Strzelin massif, the augen Gościęcice gneiss, yields a late Cambrian age (504 ± 3 Ma, Oliver et al., 1993), similar to the Izera or Śnieżnik gneisses in the West Sudetes (Borkowska et al., 1980; Oliver et al., 1993; Turniak et al., 2000). The Gościęcice gneiss and its equivalents with their envelope were thrust onto the Strzelin gneiss and its metamorphic envelope along the thrust which we refer to as the Strzelin thrust (Fig. 3, 4). This would suggest that the contact between the two types of gneiss occurring within the Strzelin massif is a part of the boundary zone between the East and West Sudetes, i.e. between the Moldanubian/Lugian and Moravo-Silesian zones.

THE STRZELIN THRUST

The Strzelin Thrust produces a juxtaposition of the Proterozoic Strzelin gneiss with its Proterozoic (the older schist series) and Devonian (the Jegłowa beds) envelopes, which form the footwall rocks, and the Early Palaeozoic Gościęcice gneiss and the light Stachów gneiss with its envelope, which belong to the hanging wall rocks. The contact between the Strzelin gneiss and the Late Cambrian Gościęcice gneiss was initially described as the Gościęcice Dolne overthrust (Oberc-Dziedzic & Szczepański, 1995; Oberc-Dziedzic, 1999). According to the present view of the authors, the overthrust of the Early Palaeozoic rocks on the Cadomian basement is not confined to the vicinity of Gościęcice Dolne but has a greater, regional extent (Fig. 3). In this sense, the Gościęcice Dolne overthrust is part of the Strzelin Thrust.

The footwall of the Strzelin Thrust

The footwall of the Strzelin Thrust is composed of gneissic core complex, its inner envelope (older schist se-

ries), and outer envelope (= younger schist series = the Jegłowa beds) (Fig. 3).

Lithology

The core complex

The core complex comprises: (1) in the north, the fine- to medium-grained, porphyritic biotite-muscovite Strzelin gneiss (Fig. 5), with conformable, several centimeter to several meter thick, intercalations of amphibolites – probably former mafic dykes (Szczepański & Oberc-Dziedzic, 1998); (2) in the south, the Nowolesie migmatitic, sillimanite gneiss (Fig. 6), rich in pegmatites, but with no amphibolite intercalations; (3) transitional types of gneisses, the Bożnowice and Gromnik gneisses, related to (1) and (2). The granitic protolith of the Strzelin gneiss is dated at 600–568 Ma by U-Pb SHRIMP analyses (Oberc-Dziedzic et al., 2001). The age of the Nowolesie gneiss protolith is still unknown.

The 3 types of gneiss are composed of quartz, plagioclase, microcline, biotite and muscovite in various propor-

tions. The Nowolesie gneiss additionally contains sillimanite nodules and garnet. Chemically, all the gneisses are predominantly peraluminous and medium to highly potassic granites or granodiorites (Oberc-Dziedzic, 1999). Their compositions and proportions of significant rare and trace elements point to greywacke as a source for the granitic protolith of gneisses (Szczepański, 1999). The peraluminousity of the gneisses, their monotonous, fine- to medium grained, porphyritic fabric, and the lack of mafic enclaves also suggest that an S-type granite was a precursor of the gneisses.

Various types of gneiss were transformed into the flecky gneisses (Oberc-Dziedzic, 1988). The flecks are composed of idiomorphic cordierite crystals, 0.5–1 cm in size, or pseudographic cordierite-quartz intergrowths, very often rimmed by medium-grained quartz-plagioclase-muscovite aggregates, up to several cm thick.

The inner envelope – the older schist series

The inner envelope – the older schist series of Proterozoic or Early Palaeozoic (?) age – is composed of amphibolites, mica schists, calc-silicate rocks and marbles. In the field, these rocks are closely connected with the Strzelin gneisses or their equivalents. Contacts between the older schist series and the Strzelin gneisses are parallel to the lithological boundaries and foliation plane. In the northern part of the massif, gneisses and amphibolites of the older schist series are mylonitised along their contacts, whereas in the middle and southern part of the massif, the contacts of gneisses and calc-silicate rocks of the older schist series are affected by alkalic metasomatism. The rocks of the older schist series almost nowhere coexist with the Nowolesie gneiss.

The outer envelope – the younger schist series

The outer envelope – the younger schist series, the Jegłowa beds (Oberc, 1966), consist of quartzites, quartz-sericite schists and metaconglomerates with granitic pebbles (Scheumann, 1937). The Jegłowa beds were interpreted by Patočka & Szczepański (1997) as sediments deposited along a continental margin during Early and Mid-Devonian times. According to Bederke (1931) and Oberc (1966), the Jegłowa beds correspond to the quartzite formation in the Jeseniki of the East Sudetes, containing Early Devonian fossils (Chlupač, 1975). The Jegłowa beds appear as thin slabs overlying the Strzelin and Nowolesie gneiss. The contact surfaces between the Jegłowa beds and the gneisses are nearly horizontal and disconformably oblique to the foliation both in the gneisses and in the Jegłowa beds. Such contacts were tectonically modified in all known cases.

The footwall rocks were deformed and metamorphosed during the Variscan orogeny, before the end of the Viséan. The orientation of the tectonic structures and the metamorphic grade are different in the northern and southern part. The Przeworno elevation (Fig. 3) is the border between the generally north- and south-oriented lineations and fold axes, whereas the Gębczyce-Gromnik fault (Fig. 3) divides two differently metamorphosed domains.
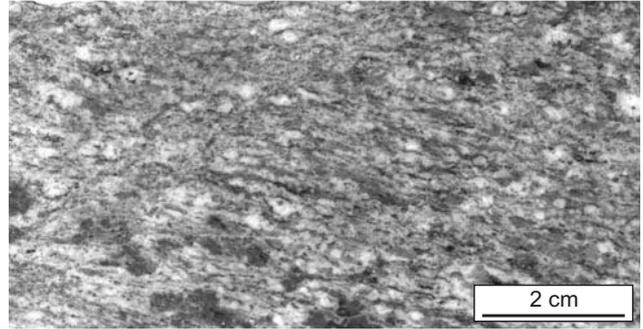
Fig. 5. The Strzelin gneiss. Dark spots – pinitised cordierite. Outcrop east of Biały Kościół.
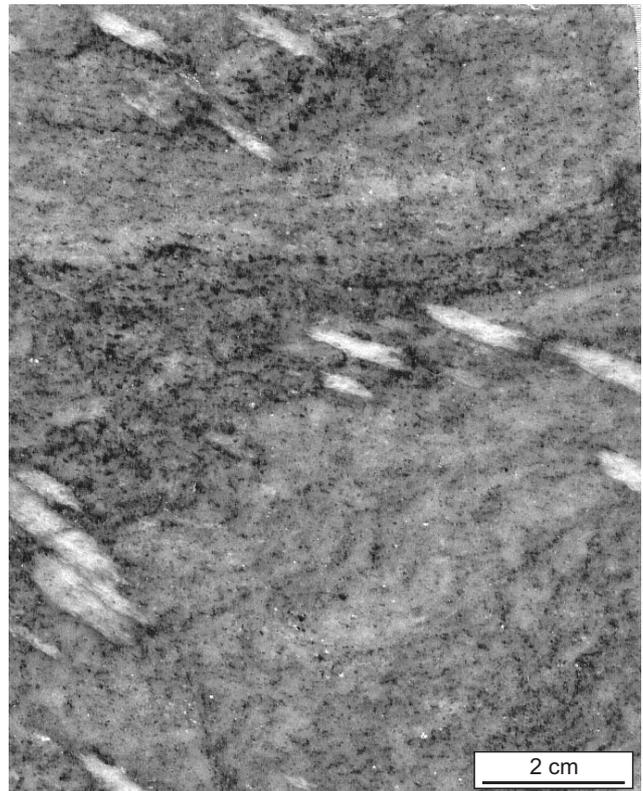
Fig. 6. The Nowolesie gneiss. Nodules of sillimanite are parallel to the axial plane of the F_2 folds. Borehole N-5, 106,5, NE of Nowolesie.

Deformation of the footwall rocks

The Pre-Variscan structures of the gneisses and the older schist series are mostly unknown. Very poor exposure, especially of the older schist series, hampers their identification, particularly they were largely obliterated by Variscan deformations. However, it seems probable that the first foliation and mutual relationships between the Strzelin gneiss and the older schist series were mostly established during pre-Variscan deformations and later they were only tectonically modified.

Four Variscan deformation events D_1-D_4 produced mesoscopic F_1-F_3 folds, L_1-L_3 lineations and S_1-S_4 planar structures showing similar orientation in the gneisses and in the older and younger envelopes (Oberc, 1966; Wojnar, 1995).

The primary S_0 sedimentary bedding is never visible in the older envelope but its relics have been described from the Jegłowa beds (Szczepański, 2001). During the D_1 deformation event, metasedimentary rocks, i.e. the older schist series and the Jegłowa beds, were deformed into very tight, isoclinal and intrafolial, and often rootless, F_1 folds (Wojnar, 1995; Szczepański, 2001). The F_1 folds are poorly preserved. Their axes generally plunge towards the ENE or E (Wojnar, 1995; Szczepański, 2001). The S_1 foliation is parallel to the axial planes of these folds. The lithological boundaries within the older schist series and the Jegłowa beds are usually parallel or nearly parallel to the S_1 foliation which is the only penetrative foliation in nearly all the rocks.

The deformation of the Jegłowa beds was characterised by Szczepański (2001). Contrary to other authors' views (Oberc-Dziedzic & Szczepański, 1995; Cymerman, 1993; Wojnar, 1995) he regards the S_2 planes as a dominant foliation in the Jegłowa beds. This paper presents Szczepański's data on the deformation of the Jegłowa beds but their interpretation is that of the present authors. According to Szczepański (2001), kinematic indicators such as S-C structures, asymetric pressure shadows around quartz clasts, mica fishes and extensional crenulation cleavage which point to the top-to-the-NE shearing, as well as quartz $<c>$ axis preferred orientation were all related to the D_3 deformation event. We reinterpret these structures as connected with the D_1 deformation (cf. Wojnar, 1995). Consequently, the quartz c-axis patterns of small girdle type around the poles to the foliation and the (I) type of crossed girdles documenting a coaxial component (Szczepański, 2001) are taken as the effects of the D_1 deformation. However, this deformation also included a component of top-to-the-NE simple shear, recorded by the above-mentioned kinematic indicators and quartz $<c>$ axis pattern type of a single girdle inclined to the foliation.

The granitic or gneissic protoliths of the Strzelin gneiss also were subjected to the top-to-the-NNE/NE non-coaxial shearing during the D_1 deformation. It resulted in the formation of the penetrative foliation S_1 and stretching lineation L_1. The top-to-the-NNE shearing is documented by σ type porphyroclasts and S-C structures, accompanied by F_1 folds, rarely preserved in the quartz layers. Their axes generally plunge to the ESE.

In the Nowolesie gneiss, there is no evidence of shearing connected with D_1 deformation. Its S_1 foliation is parallel to the axial planes of very rare isoclinal F_1 folds.

In the northern part of the Strzelin massif, the S_1 foliation moderately or shallowly dips to the NW and N in the gneisses and to the N and NE in the Jegłowa beds. In the southern part of the massif, the foliation in gneisses and the Jegłowa beds dips to the SE and S at varying angles of 10–45°. The orientation of the foliation in the calc-silicate rocks in the northern part of the massif displays a great-circle distribution around the axis plunging gently to the NNE, i.e. parallel to the F_2 fold axes (Oberc, 1966).

The S_1 planar structures were deformed during the D_2 event, resulting in asymmetric, isoclinal or disharmonic F_2 folds of variable scale. The axes of the F_2 folds plunge to the N, NNE and NE in the northern part of the massif and to the SE, S and SSW in the southern part. In the Strzelin gneiss and the Jegłowa beds in the northern part of the massif, the F_2 fold axes are parallel to the L_1 lineation. This suggests that during the D_2 deformation, itself a continuation of the D_1 event, the top-to-the-NNE shearing was replaced by a coaxial WNW–ESE shortening. The L_2 lineation of the same orientation as the fold axes was developed as a result of the intersection between the S_1 foliation and the S_2 axial cleavage. In very well foliated varieties of the gneisses and calc-silicate rocks and in the Jegłowa beds the S_2 planes may be represented by crenulation cleavage. The S_2 planes are generally non-penetrative structures. In the Strzelin gneiss, the crenulation cleavage dips to the NW at an angle of 20°. In the Nowolesie gneiss and calc-silicate rocks, the S_2 cleavage is also non-penetrative and its presence can be proved only in the hinges of the folds where it is marked by the alignment of new minerals: sillimanite nodules and biotite, respectively (Oberc-Dziedzic, 1999; Wojnar, 1995). However, in the mica-rich quartzites of the Jegłowa beds, the S_2 foliation is penetrative and dips to the E, ESE (Wojnar, 1995) and NW (Oberc-Dziedzic & Szczepański, 1995) in the northern part of the massif and to the E in the southern part of the massif (Wojnar, 1995; Szczepański, 2001).

The D_3 deformation event produced an F_3 kink type or broad, open folds, several centimeters in amplitude, with steep axial planes S_3. The F_3 fold axes plunge to the N and NW in the rocks of the northern part of the massif, but in the southern part of the massif, they trend W–E, ESE–WNW or NW–NE, more or less perpendicularly to the F_2 folds and parallelly to the Przeworno elevation (Oberc, 1966).

The D_4 event produced a zonally localised S_4 foliation defined as thin mylonitic bands inclined at an angle of 10–15° to the S_1 foliation and dipping to the N at an angle of 10–25° in the Strzelin gneiss (Oberc-Dziedzic, 1999) and as narrow shear zones dipping to the S in the Jegłowa quartzites in the southern part of the massif (Szczepański, 2001). In the Jegłowa beds, kinematic indicators such as extensional crenulation cleavage and asymmetric pressure shadows around quartz segregations document top-to-the-NE shearing in the northern part of the massif and top-to-the-SSW in the southern part (Szczepański, 2001). During the D_4 event, most of the geological boundaries were tectonically modified; the Jegłowa beds were detached from the gneisses and moved NE-ward from their original position.

The four deformation events led to the formation of small thrust units (Oberc-Dziedzic & Szczepański, 1995; Oberc-Dziedzic, 1999) bounded by mylonitic zones and differing in terms of theirmetamorphic grade. Some of them will be characterised in the next chapter.

Metamorphism of the footwall rocks

Up till now, no indisputable evidence has been found of pre-Devonian metamorphism of the gneisses and the older envelope. Nevertheless, the younger group of zircon ages (568 Ma) in the Strzelin gneiss is taken as a record of Late Proterozoic crystallisation during partial melting

associated with metamorphism which affected the c. 600 Ma protolith of the Strzelin gneiss (Oberc-Dziedzic et al., in prep.). It is possible that migmatitic structures preserved in the Strzelin gneiss document this event. It is also probable that inclusions of biotite and garnet in plagioclase porphyroblasts in the amphibolite of the older envelope represent the pre-Variscan assemblage. They indicate LT-HP conditions with $T = 500–530°$ (Grt + Bt thermometer of Ferry & Spear, 1978) and $P \sim 14$ kb estimated on the basis of the plagioclase-garnet-Al_2SiO_5-quartz barometer (Ghent, 1976; Ghent et al., 1979) assuming that Al_2SiO_5 is kyanite.

The effects of the M_1-M_3 Variscan metamorphic episodes differ between the thrust-bounded units, which implies they were derived from various metamorphic zones. In all these units however, the M_1 metamorphic event was related to progressing T-P. All the rocks achieved their peak of metamorphism before the second deformation phase.

Metamorphism of the northern domain

In the northern domain, metamorphic conditions M_1 were typical of the greenschist facies in the case of the Jegłowa beds and of the amphibolite facies in the case of the Strzelin gneiss. The rocks of the older envelope bear the record of continuous transition from greenschist facies conditions: $T \sim 500°C$ to amphibolite facies $T \sim 680°C$ under constant $P = 8 \pm 1$ kb during the M_1 metamorphic episode (Oberc-Dziedzic, 1999). The M_1 metamorphic episode corresponds to the nappe stacking during the D_1 deformation event and the period after it, but before the D_2 event. The temprature during the M_2 metamorphic event was probably similar to that of M_1. However, the presence of cummingtonite younger than the assemblage defining the M_1 conditions can indicate that the pressure decreased during the M_2 metamorphic event (Evans & Ghiorso, 1995). The M_3 metamorphic episode, which was coeval with the D_3 and D_4 deformation events, took place under lower amphibolite-greenschist facies conditions and caused the retrogressive changes detectable mainly in the contact zones.

Metamorphism of the southern domain

In the southern domain, the Nowolesie gneiss achieved anatectic conditions during the M_1 metamorphic event. The P-T conditions estimated on the basis of the Grt + Bt + Sil + Mc assemblage in the Nowolesie gneiss indicate $T = 720°C$ (Grt + Bt thermometer, Ferry & Spear, 1978) and $P = 6.5 \pm 1$ kb (geobarometer of Ghent, 1976; Ghent et al., 1979) during progressive metamorphism and $T = 730°C$ and $P = 5 \pm 1$ kb as the metamorphic peak (the same methods). The first stage of anatexis (M_1 after the D_1 and during the D_2) was followed by the M_2 decompressive event related to the beginning of tectonic denudation. It gave rise to the formation of pegmatites and leucocratic granites. The P-T conditions in the pegmatites were estimated on the basis of the Grt + Bt + Sil + Mc assemblage to be $T = 600°C$ (Grt + Bt thermometer, Ferry & Spear, 1978) and $P = 3$ kb (Pl + Bt + Grt + Ms geobarometer, Ghent & Stout, 1981).

According to Szczepański & Józefiak (1999), the M_1 event in the Jegłowa beds of the southern domain took place under progressive greenschist facies conditions reaching pressures up to 6–7 kb and temperatures up to 500°C. Here, the effects of the M_2 episode in the Jegłowa beds were obliterated by recrystallisation during the M_3 episode.

The P-T conditions of the M_3 metamorphic event were probably similar to those of M_2. They did not induce any significant changes in the Nowolesie gneisses, but the Jegłowa beds recorded a significant increase in temperature, up to 630°C, and a pressure decrease to about 3.8 kb at that time (Szczepański & Józefiak, 1999).

In both domains, the M_4 episode of a regional metamorphism led to the crystallisation of post-kinematic cordierite in the Strzelin and Nowolesie gneisses and the formation of the flecky gneisses (Oberc-Dziedzic, 1988 a, 1995). The cordierite flecks were formed just before the emplacement of the Variscan granitoids at 347–330 Ma (Oberc-Dziedzic et al. 1996). The last metamorphic event, M_5, was connected with the thermal influence of the granitoid bodies and caused crystallisation of the prismatic sillimanite and andalusite in their proximity.

Tectonometamorphic evolution of the footwall rocks

The footwall rocks underwent polyphase deformation and metamorphism. However, the main features of the tectonic structure of the footwall were established over the course of the D_1 and D_2 deformations and the main stage of metamorphic recrystalisation of the rocks was over before the second deformation phase (Wojnar, 1995).

The lithological boundaries in the northern part of the massif trend E–W or ENE–WSW; in the middle and the southern part, their trend is NNE–SSW. It seems that tectonic units in the northern part were formed due to the NNE- and NE-vergent shearing and thrusting during the D_1 deformation event. It also cannot be excluded that during this deformation the protolith of the Nowolesie gneiss was overthrust by the Strzelin gneiss (Fig. 7). The so-called Bożnowice gneiss is probably a more highly metamorphosed equivalent of the Strzelin gneiss situated south of the Przeworno elevation (Oberc, 1966). The tectonic units of the northern part of the footwall were only imperceptibly modified during weak D_2 folding. However, in the middle and southern part, F_2 folds are the dominant type of megastructure. The M_1 metamorphic event followed the D_1 deformation and lasted until after its end. During the M_1 metamorphic episode, the Nowolesie gneiss of the southern part achieved anatectic conditions. The calc-silicate rocks were metamorphosed under similar high temperature and pressure conditions to the gneisses (Wojnar, 1995). In the Strzelin gneiss however, there are no signs of the Variscan migmatitisation. Some premises suggest that this rock was metamorphosed under higher pressure and lower temperature than the Nowolesie gneiss. The M_2 metamorphic episode, which was coeval with the D_2 deformation and outlasted it, was characterised by decreasing temperature and pressure in the southern domain.
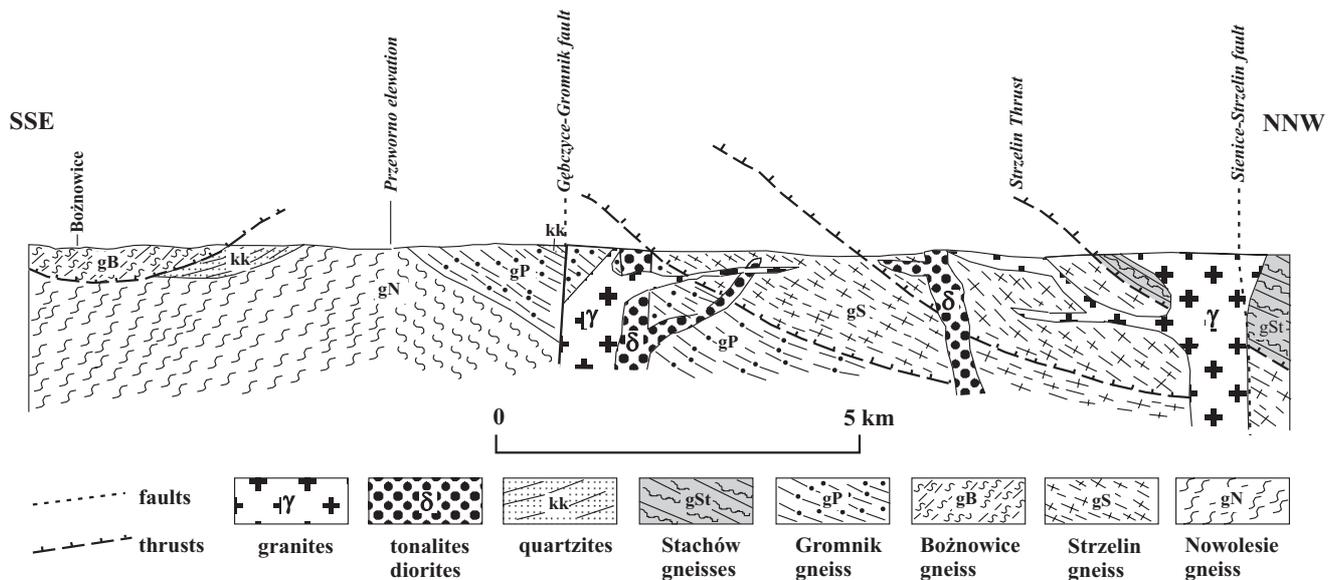
Fig. 7. Section across the Strzelin massif.

During the D$_3$ deformation event, the Przeworno elevation, trending WNW–ESE to NW–SE came into existence (Fig. 7). It affected the whole megastructure of the eastern part of the Fore-Sudetic Block (Oberc, 1966, 1972) and caused the linear structures in the Strzelin massif to dip generally to the N in the northern part of the massif and to the S in the southern part. In the Strzelin massif, the origin of the Przeworno elevation was preceded by an uplift of the migmatised masses due to decompression and the beginning of tectonic denudation. The origin of the Przeworno elevation caused the Jegłowa beds to become detached from their original position and moved to their present positions on the either side of the elevation. The transport distance was significant in the case of the Jegłowa beds in the north, which show a much lower metamorphic grade than the Strzelin gneiss and its older envelope. In the case of the southern part of the massif, the effects of the M$_2$ metamorphic event in the Jegłowa beds have been obliterated by the M$_3$ episode, and they show similar metamorphic degree as the Nowolesie gneiss. This may mean that the Jegłowa beds and the Nowolesie gneiss were metamorphosed together after the M$_2$ episode or that they did not come into contact until the M$_3$ metamorphic event. In the first case, the normal fault transport of the Jegłowa beds was small; in the second case it could have been significant. The movements subsequent to the origin of the Przeworno elevation were documented by the S$_4$ planes and shear indicators.

The northern and southern domains of the Strzelin massif, with their different structure and metamorphic history, are separated by the Gębczyce-Gromnik fault (Fig. 7). This is a very important NW–SE trending oblique-normal fault with a sinistral component, northern side down-thrown. It came into existence during the extensional stage of deformation in the Strzelin massif, i.e. probably during the D$_3$ and D$_4$ deformations. This deep-rooted dislocation controlled the emplacement of the

Variscan two-mica granitoids of the Gębczyce and Gromnik intrusions (Oberc-Dziedzic, 1991) and probably also the Górka Sobocka intrusion in the Lipowe Hills. The two domains on either side of the Gębczyce-Gromnik fault also differ in their distribution of granite and tonalite-diorite bodies. The granites are more frequent in the northern part of the massif, whereas tonalites and diorites dominate in the southern part (Oberc-Dziedzic, 1991). Such an uneven distribution of magmatic rocks matches that of the anatectic rocks and of the flecky gneisses characterised by broad leucocratic rims around cordierite cores (Oberc-Dziedzic, 1988).

The hanging wall of the Strzelin Thrust

The hanging wall of the Strzelin Thrust is preserved as several klippen. The big klippe is situated in the northern part of the Strzelin massif (Fig. 3, 4). It is composed of the Gościęcice biotite gneiss (504±3 Ma, Oliver *et al.*, 1993) containing characteristic nearly idiomorphic grey-blue microcline augen, rimmed by plagioclase (Fig. 8). The Gościęcice gneiss must be also hidden beneath the Variscan granite body exposed west of Strzelin, as it appears as enclaves in the latter (Fig. 9).

Other klippen are situated near the Strzelin quarry in the N part of the Strzelin massif, south of Bożnowice in the S part of the Strzelin massif, and also near Nieszkowice, Stachów and Henryków in the Lipowe Hills, west of the Strzelin massif (Fig. 3). All these klippen are composed of two types of gneiss typical of the Lipowe Hills (Oberc-Dziedzic, 1988, 1995): flaser gneiss, referred to as the light Stachów gneiss (Fig. 10), and fine-grained gneiss, referred to as the dark Stachów gneiss. The dark Stachów gneiss, together with mica schist, amphibole schist and amphibolite which alternate with it, are all considered Proterozoic or Early Palaeozoic metasediments,
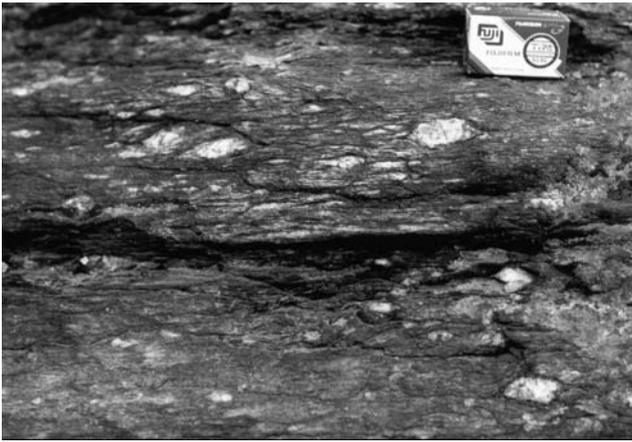
Fig. 8. The Gościęcice gneiss. Asymmetric pressure shadows around microcline porphyroclasts indicate top-to-the-NNE sense of shear. Box size is 55 x 38 mm. Outcrop north of Gościęcice.
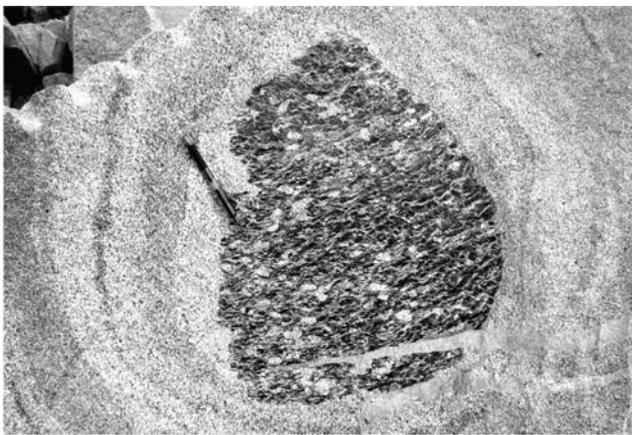
Fig. 9. Enclave of the augen Gościęcice gneiss in the Strzelin granite. Length of pen is 14.5 cm. The Strzelin quarry.

Fig. 10. The light, flaser Stachów gneiss. Borehole ST-2, 80,9, south of Stachów.

representing the metamorphic envelope of the granitoid protolith of the light Stachów gneiss. This envelope differs from the inner envelope of the Strzelin gneiss by the presence of the dark gneiss and by the nearly complete absence of calc-silicate rocks.

The light Stachów gneiss and its more deformed variety, the so-called Henryków gneiss (Madej, 1999) are probably the equivalent of the Gościęcice gneiss, as suggested by the geochemical similarity of the two gneisses and by the presence of the dark Stachów gneiss xenoliths in the Gościęcice gneiss. Supposedly, the dark Stachów gneiss formed the envelope not only of the granitic protolith of the light Stachów gneiss but also of the Gościęcice gneiss.

Structural and metamorphic evolution of the hanging wall rocks

The effects of the Pre-Variscan deformation and metamorphism are much highly obliterated. The Early Palaeozoic precursor of the Gościęcice gneiss probably intruded into an already deformed and metamorphosed envelope,

since it contains gneissic enclaves with two foliations. The older foliation in the enclaves is oblique to the Variscan foliation of the gneiss, whereas the younger one is parallel to the foliation in the host gneiss.

The sequence of the Variscan structural and metamorphic events in the hanging wall rocks are similar to those of the footwall rocks.

The granitic protoliths of the Gościęcice gneiss, similarly to the protoliths of the Strzelin gneiss, were subjected to non-coaxial shearing during the Variscan D_1 deformation. This resulted in the formation of the penetrative foliation S_1 and stretching lineation L_1. The generally top-to-the-N-NNE shearing is documented by σ porphyroclasts and S-C structures. The foliation in the Gościęcice gneiss dips to the N, the L_1 stretching lineation plunges to the NNW-N and NE. A younger lineation resembling tectonic striae dipping to the N probably came into existence during the D_4 event in the core complex, when the Strzelin Thrust was reactivated.

The tectonic structures of the hanging wall rocks (the dark and light Stachów gneisses) exposed in the Mikoszów

quarry (Fig. 4) were reoriented during the intrusion of the Variscan granite. The foliation dips to the NNW at an angle of 60–70°, and the lineation plunges to the NE at 45°.

In the light and dark Stachów gneiss of the Lipowe Hills, the main foliation S_1 dips to the NE, W and SW. During the D_2 event, it was folded into the F_2 folds. The F_2 axes plunge to the SW with their axial planes dipping gently to the NW and SE. The axial-plane cleavage S_2 is represented by crenulation planes. In the dark Stachów gneiss, it is marked by the alignment of sillimanite nodules. The L_2 lineation formed by the intersection of the foliation S_1 and cleavage S_2 plunges to the NNE, S, SW and W. The S_1 foliation was reactivated during the D_3 event. The stretching lineation L_3, defined by micas on the S_{1+3} foliation, remains constantly oriented throughout the whole Lipowe Hills. It plunges to the SW/W and is oblique or parallel to the lineation L_2.

The effects of the Variscan metamorphism differ between klippen but they are compatible with the metamorphic grade of the nearest footwall rocks. This may suggest that the metamorphic grade of the hanging wall rocks was established during the M_1 metamorphic event, which took place during and shortly after the thrust stacking. The dark gneisses exposed in the klippen near Mikoszów and Bożnowice do not show symptoms of the migmatitisation typical of the dark gneiss from the Lipowe Hills.

The footwall-hanging wall contact zone

Northern part of the Strzelin massif

The northern domain (Fig. 4a) of the Strzelin massif (Fig. 3) is composed of several thrust sheets. The biggest, the Gościęcice-Biały Kościół unit (jG-BK) comprises the Strzelin gneiss, with intercalations of amphibolites, and it probably represents the autochthonous, core part of the massif. The older schist series enveloping the Strzelin gneiss, was tectonically detached from the gneiss, or tectonically wedged into it. It gave rise to the formation of two independent tectonic units: the Szańcowa Góra unit (jSz), composed of amphibolites and mica schist, and the northern Kuropatnik unit (jpK), which consists of calc-silicate rocks, mica schists and amphibolites. The Gościęcice-Biały Kościół unit and the Szańcowa unit belong to the footwall of the Strzelin Thrust. They were overthrust by the Gościęcice Dolne unit (jGd), composed of the Gościęcice gneiss which belongs to the hanging wall of the Strzelin Thrust (ST, Fig. 4b).

The tectonic contact between the gneiss of the Gościęcice unit of the hanging wall and amphibolites of the Szańcowa Góra unit of the footwall was studied in borehole GD-3 (Fig. 4a, 10) and in outcrops in its vicinity. The thrust zone parallels the S_1 gneiss foliation and dips to the NNE at c. 25°.

In the gneisses (Fig. 8), muscovite contents increase toward the contact zone. In the same direction, elongation of microcline porphyroclasts increases while their amount decreases (Fig. 11, 12). Near the contact zone, porphyroclasts disappear and the rock changes into a finely banded mylonite. The Gościęcice gneiss recorded
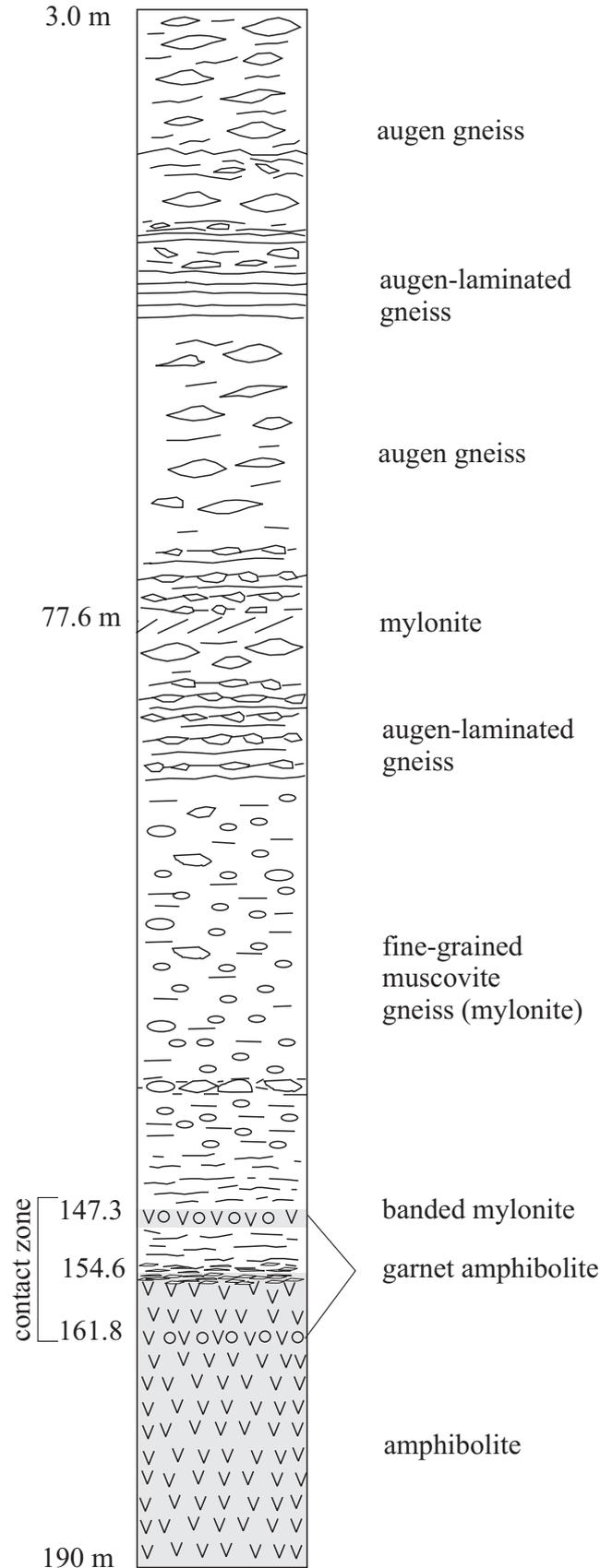
Fig. 11. The GD-3 borehole column.
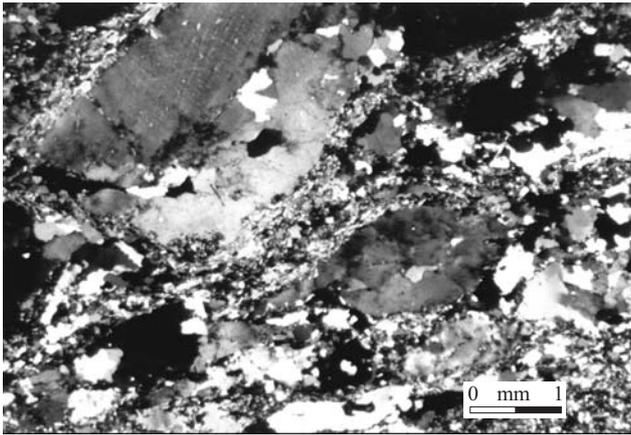
Fig. 12. Mylonitised augen gneiss. Crossed polars. GD-3 bore-hole. Depth 77. 6 m.
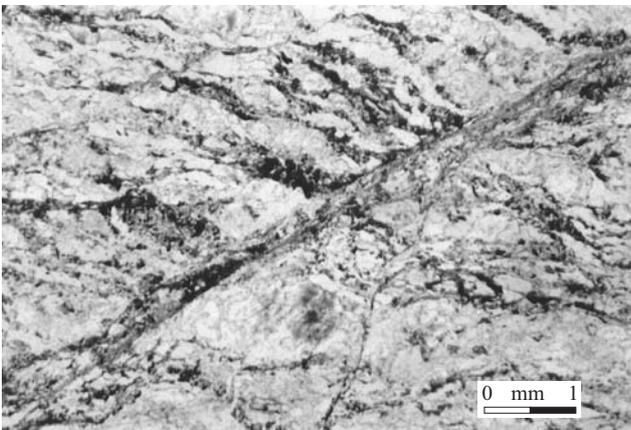
Fig. 13. Extensional crenulation cleavage in the mylonitised augen gneiss. Without analyser. GD-3 borehole. Depth 77.6 m.

Fig. 14. Quartz c-axis patterns in the Gościęcice gneiss. Equal-area net, lower hemisphere. Density contours at 1% intervals. Projection plane is the XZ plane of the strain ellipsoid. Foliation plane is perpendicular and lineation trend is parallel to the figure.

evidence of two deformations which led to the formation of the Strzelin Thrust. During the D_1 event, when a thrust stacking took place, the granitic protolith of the Gościęcice gneiss was deformed as way described above. The direction of the thrusting documented by the penetrative L_1 stretching lineation of the gneiss, σ porphyroclasts and older S-C structures was generally top-to-the-N-NNE. During the D_4 event, the Strzelin Thrust contact zone and the S_1 foliation of the Gościęcice gneiss were reactivated. The effects of this deformation are visible over a distance of several tens of meters as cross-cutting mylonitic streaks in gneisses, new S-C structures, extensional shear bands (Fig. 13) and younger lineation resembling tectonic striae plunging to the N. The quartz c-axis patterns in the Gościęcice gneiss (Fig. 14), scattered along two small circles, document a coaxial component of the D_4 deformation in the field of general flattening. The concentrations of the c-axis maximum in the periphery of the diagrams may suggest the activity of basal plane slip <0001> under greenschist facies conditions (Bouchez & Pecher, 1981; Schmid & Casey, 1986).

In the amphibolites of the Szańcowa Góra unit, mylonitisation caused a grain size reduction of plagioclase and hornblende. The strongly mylonitised amphibolites are

Fig. 15. Mylonitised amphibolite. Crossed polars. GD-3 borehole. Depth 160.0 m.

very fine-grained rocks, composed of new pale-green hornblende aggregates as well as relics of brown hornblende, plagioclase and opaque grains (Fig. 15).
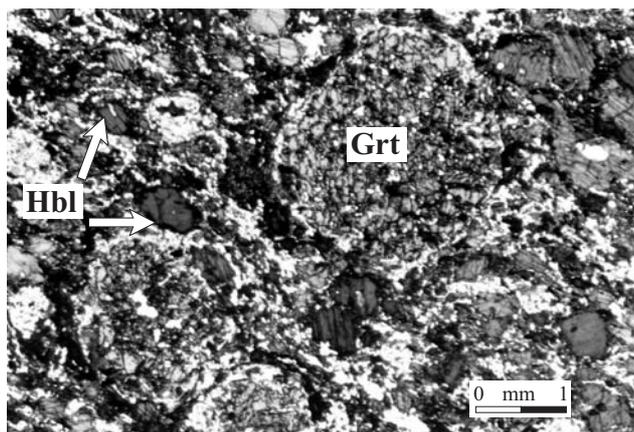
Fig. 16. Garnet-bearing amphibolite. Pressure shadows around garnet grains (Grt) are filled with plagioclase and green hornblende. Brown hornblende grains (Hbl) are rimmed by green hornblende. Crossed polars. GD-3 borehole. Depth 147.3 m.

Fig. 17. The Henryków gneiss. Asymetric σ-type feldspar porphyroclasts indicate a top-to-the-SW sense of shear during the D$_3$ deformation. Scale bar is 3 cm long. Outcrop 1.2 km west of Henryków.

The c. 10 m thick contact zone comprises several dm- to m-sized bodies of strongly mylonitised laminated gneisses, amphibolites and bodies of garnet-bearing amphibolite (Fig. 16) different from the amphibolite of the Szańcowa Góra unit, thus taken as tectonic xenoliths.

The metamorphic history of the tectonic units in the northern footwall-hanging wall contact zone

The Gościęcice gneiss is composed of the mineral assemblage Qtz+Kfs+Pl+Bt. representing the M$_1$ metamorphic episode. It was probably formed during the D$_1$ deformation and did not change during the D$_2$ deformation. The D$_4$ deformation was accompanied by a sericitisation of the feldspars and a crystallisation of the Pl+Chl+Ms assemblage, corresponding to the M$_4$ metamorphic episode. It defines the greenschist facies conditions of the D$_4$ deformation event.

The garnet bearing amphibolite (Fig. 16), incorporated into the thrust zone, contains relics of pyroxene accompanied by two other mineral assemblages corresponding to two distinct metamorphic events. The first assemblage: brown Hb+Grt+Pl$_{An27}$ grew at T=670°C (a very good match between the Hb-Grt thermometer of Graham & Powell 1984 and the Hb-Pl thermometer of Holland & Blundy 1994) and P=11±0.5 kb (Kohn & Spear 1990, geobarometer). The pressure shadows around the deformed grains of this assemblage are filled with the second, green Hb+Pl$_{An43}$ assemblage which corresponds to T=680°C (Hb-Pl thermometer of Holland & Blundy 1994) and P=8–9 kb (the best match between temperature and assumed pressure, Oberc-Dziedzic, 1999).

The rocks of the Szańcowa Góra unit bear the record of continuous transition from greenschist facies conditions: T=500–520°C and P= 8±1 kb to amphibolite facies involving a temperature increase of up to T=680°C under constant P= 8±1 kb during the M$_1$ metamorphic episode (Oberc-Dziedzic, 1999). The M$_1$ metamorphic episode corresponds to the nappe stacking during the D$_1$ deformation event, when the Gościęcice Dolne unit was thrust onto the Szańcowa Góra unit. The thermo- and geobarometry results indicate that the metamorphic conditions which affected the rocks of the Szańcowa Góra unit during the M$_1$ metamorphic episode correspond to the conditions which produced the second assemblage in the garnet-bearing amphibolite (Fig. 16). The whole P-T path of the Szańcowa Góra unit has not yet been established. However, some intercalations of amphibolite contain cummingtonite younger than the Hb+Grt+Pl assemblage defining the S$_1$ foliation. The presence of cummingtonite can indicate a pressure decrease during the M$_2$ metamorphic event (Evans & Ghiorso, 1995).
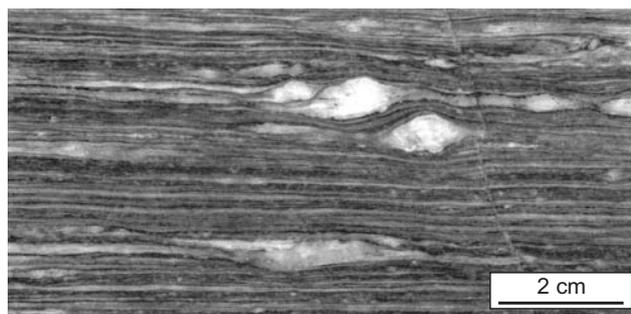
Southern part of the Lipowe Hills

The relationships between the footwall and hanging wall rocks of the Strzelin Thrust can be studied west of Henryków. The footwall rocks are represented by the sillimanite Nowolesie gneiss and quartzites. The mutual contacts between these rocks are not exposed anywhere, but their relationships must be similar to those in the Strzelin massif, where the quartzites overlie the gneisses.

The hanging wall rocks are represented by the very strongly mylonitised Henryków gneiss composed of quartz, K-feldspar, plagioclase, muscovite and chlorite. The effects of four deformation events have been documented within the Henryków gneiss. The first D$_1$ event gave rise to the mylonitic S$_1$ foliation defined by the alternation of quartz, feldspar and chlorite-muscovite layers up to 1.5 mm thick. It dips to the W at angles of 30–40°. The S$_1$ foliation was deformed by F$_2$ folds several millimeters to several centimeters in size. Their axes are oriented 170–210°/ ~20°. During the subsequent D$_3$ event, the S$_1$ foliation was reactivated and the S$_3$ foliation came into existence locally. It dips to the W at a high angle. The L$_3$ mineral stretching lineation defined by chlorite and muscovite plunges to the SW at angles of 25–35°. Kinematic indicators: S-C structures, mica fish, extensional crenulation cleavage, asymmetric σ-type feldspar porphyroclasts indicate a top-to-the-SW sense of shear during the D$_3$ deformation (Fig. 17). The D$_4$ deformation produced kink folds with their axes trending NW–SE and dipping at an angle of 10–20°.

The main feature of the Henryków gneiss is an extremely strong mylonitisation of the granitic protolith. It
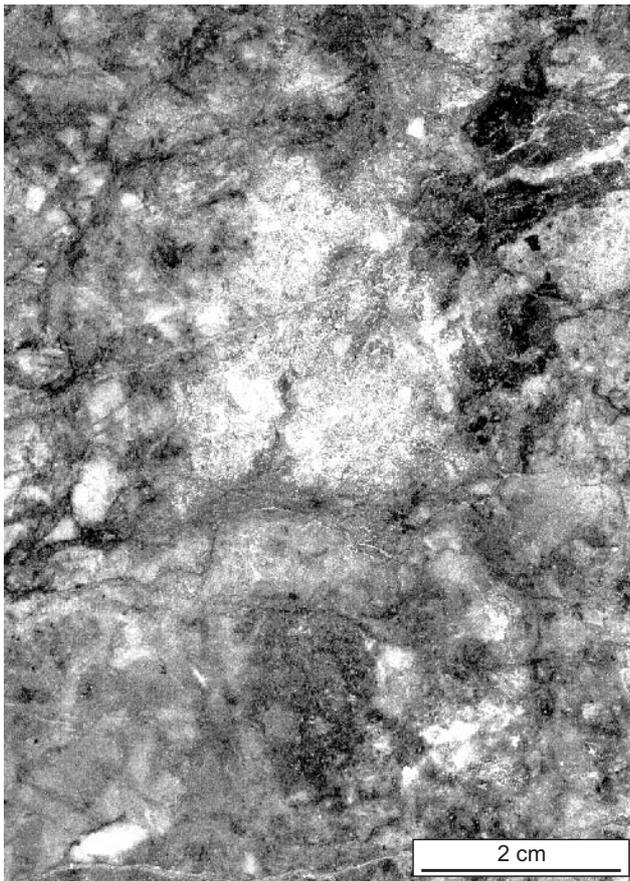
Fig. 18. Deformed quartzite from the footwall of the Strzelin Thrust. Outcrop 1.5 km west of Henryków.

Fig. 19. S-C structures (in the middle of the photo) and asymmetrical extensional shear bands (lower left corner) showing top-to-the-SW sense of shear within mylonitised gneiss from the footwall of the Strzelin Thrust. Crossed polars. Outcrop 1.5 km west of Henryków.

the high contents of Si^{4+} p.f.u. in the white micas. The effects of the earlier metamorphic event were completely obliterated by the metamorphism accompanying the D_3 deformation, which took place under greenschist facies conditiona but at lower pressures.

The contact between the footwall rocks and the Henryków gneiss is not exposed. The proximity of the contact zone is signalised by the strong mylonitisation of the Henryków gneiss and quartzites (Fig. 18) outcropping west of the gneiss together with the rock described by Badura (1981) as a mica schist. The true nature of the latter rock is not known yet, but it may be a gneissic ultramylonite (Fig. 19).

took place during the D_1 deformation event, under metamorphic conditions characterised by the stability of biotite and relatively high pressure estimated on the basis of

DISCUSSION AND CONCLUSION

The occurrence of a broad, mylonitic contact zone with tectonically incorporated relatively HT and HP garnet-bearing amphibolites, the contrasting P-T paths of the adjoining rocks (Fig. 20) in the northern part of the Strzelin massif, as well as the strong mylonitisation of the Henryków gneiss and quartzites in the southern part of the Lipowe Hills, indicate that the Strzelin Thrust is an important tectonic structure. It was formed during four tectonic events. The first tectonic event, D_1, gave rise to nappe stacking due to W–E or WNW–ESE contraction, as in the Kamieniec Ząbkowicki Metamorphic Complex farther to the west (Mazur & Józefiak, 1999). This contraction and thrusting can only be inferred from the presence of the Stachów gneisses, typical of the Lipowe Hills, inside the Strzelin massif, but the sense of transport cannot be proved by kinematic indicators. On the contrary, the penetrative L_1 stretching lineation, σ porphyroclasts and S-C structures in the footwall and hanging wall rocks all persistently point to a generally top-to-the-N-NNE direc-

tion of shearing during the D_1 deformation event. The W–E or WNW–ESE contraction, top-to-the-N-NNE shearing and the E-ESE thrusting occurring at the same time suggest a bulk triclinic transpressional deformation regime involving components of pure shear contraction and oblique simple shear (Holdsworth et al., 2002) during the collision of the West and East Sudetes.

During the D_2 event, which was a continuation of the D_1 event, the subsequent coaxial shortening produced F_2 folds trending N and NNE. The D_{3-4} tectonic event involved extensional collapse directed to the NNE in the northern part of the Strzelin massif and to the SW in the southern part of the Strzelin massif and Lipowe Hills (Szczepański, 2001; Madej, 1999), recorded by the development of the S_4 foliation and shear indicators. During the D_4 event, the contact zone between the footwall rocks and hanging wall rocks was reactivated and strongly mylonitised under greenschist facies conditions.

The tectonic juxtaposition of the Proterozoic Strzelin
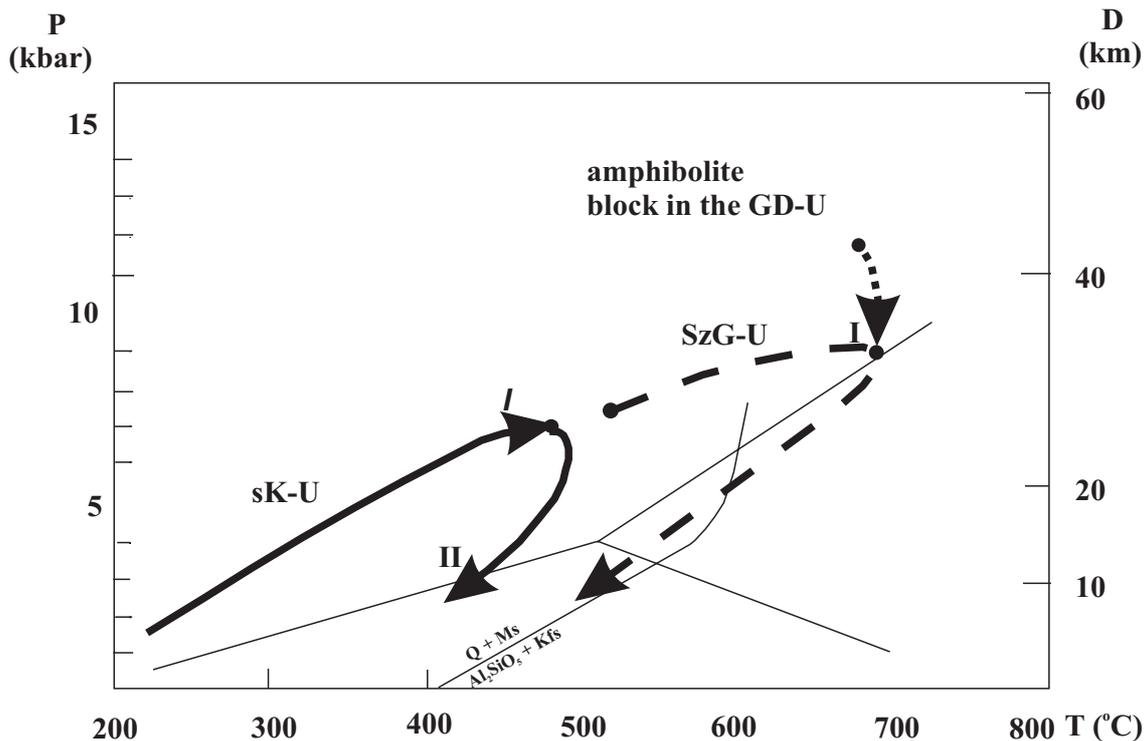
Fig. 20. Pressure-temperature diagram for the Szańcowa Góra unit (Sz G U – broken line), the amphibolite block in the Gościęcice Dolne unit (GD U – dotted line), and the northern Kuropatnik unit (nK U – solid line).

gneiss and the overlying Early Palaeozoic Gościęcice gneiss and its equivalent – the light Stachów gneiss with its envelope – resembles the situation along the East-West Sudetes boundary separating two domains with contrasting protolith ages. Consequently, the Strzelin Thrust can be considered a part of the tectonic boundary between the East and West Sudetes, i.e. the continuation of the Ramzova/Nyznerov thrust to the Fore-Sudetic Block. At the present erosion level, the hanging wall of the Strzelin Thrust is represented by the Gościęcice gneiss and its equivalents which are preserved in the form of klippen. It is also very probable that there are other klippen beyond the Strzelin massif and the Lipowe Hills. The Maciejowice gneiss may represent one of them. The longest distance between klippen in the Lipowe Hills and in the Strzelin massif is over 10 km. This value can be accepted as the minimum transport distance along the Strzelin Thrust. However, it should be mentioned that Oberc (1968) estimated the minimum amplitude of the Ramzova thrust at 17 km. The Strzelin Thrust surface has a dome-like shape and is generally shallowly dipping, as inferred from the similar hypsometric position of the klippen. This weakly inclined surface differs the Strzelin Thrust from the Ramzova/Nyznerov thrust, which dips to the W at an angle of 40–50° (Misař *et al.*, 1983). The change of the Ramzova thrust dip was predicted by Oberc(1968) for the area along the Sudetic Marginal Fault (shadowed figure (area) on Fig. 1). The Strzelin Thrust in the area of the Strzelin massif and the Lipowe Hills is another part of the East-West Sudetes boundary showing a weak inclination but exposed on the surface. The hanging wall of the thrust is preserved as klippen, so the East-West Sudetes bound-

ary should be located at the base of these klippen. The Strzelin Thrust surface becomes steeper east of the Strzelin massif, where it is hidden beneath Cenozoic sediments, and. west of the Lipowe Hills, where it follows the eastern border of the Kamieniec Ząbkowicki Metamorphic Complex. The strongly mylonitised mica schists exposed along the Mała Ślęza river may represent a root zone of the hanging wall.

The changes in the inclination of the thrust surface may explain the earlier mentioned difficulties in properly locating the East-West Sudetes boundary inside the Fore-Sudetic Block. In the light of these changes, the proposals of the location of the East-West Sudetes boundary inside and near the Strzelin massif given by Oberc (1968), Skácel (1989 b) and Cwojdziński & Żelaźniewicz (1995) can be accepted with the modification referring to the Skácel concept (Fig. 21). According to Skácel (1989 a), the boundary between the East and West Sudetes in the Sudetes follows the eastern side of the Stare Město belt. However, the Fore-Sudetic Block is 2.5 km (Oberc, 1968) to 5 km (Guterch *et al.*, 1975) or even 10 km (Skácel, 1989 b) more deeply eroded than the Sudetes and for this reason the Stare Město belt, similarly to many other Sudetic tectonic units, does not have its prolongation inside it. Skácel suggested that the Niedźwiedź massif situated in the southern part of the Fore Sudetic Block could be considered an equivalent of the Stare Město belt (1989 a). He placed the East-West Sudetes boundary along the eastern margin of the Niedźwiedź massif. Farther to the north, where the basic and ultrabasic rocks of the Niedźwiedź massif do not have their prolongation, the boundary follows the eastern margin of the Kamieniec Ząbkowicki Metamorphic Com-
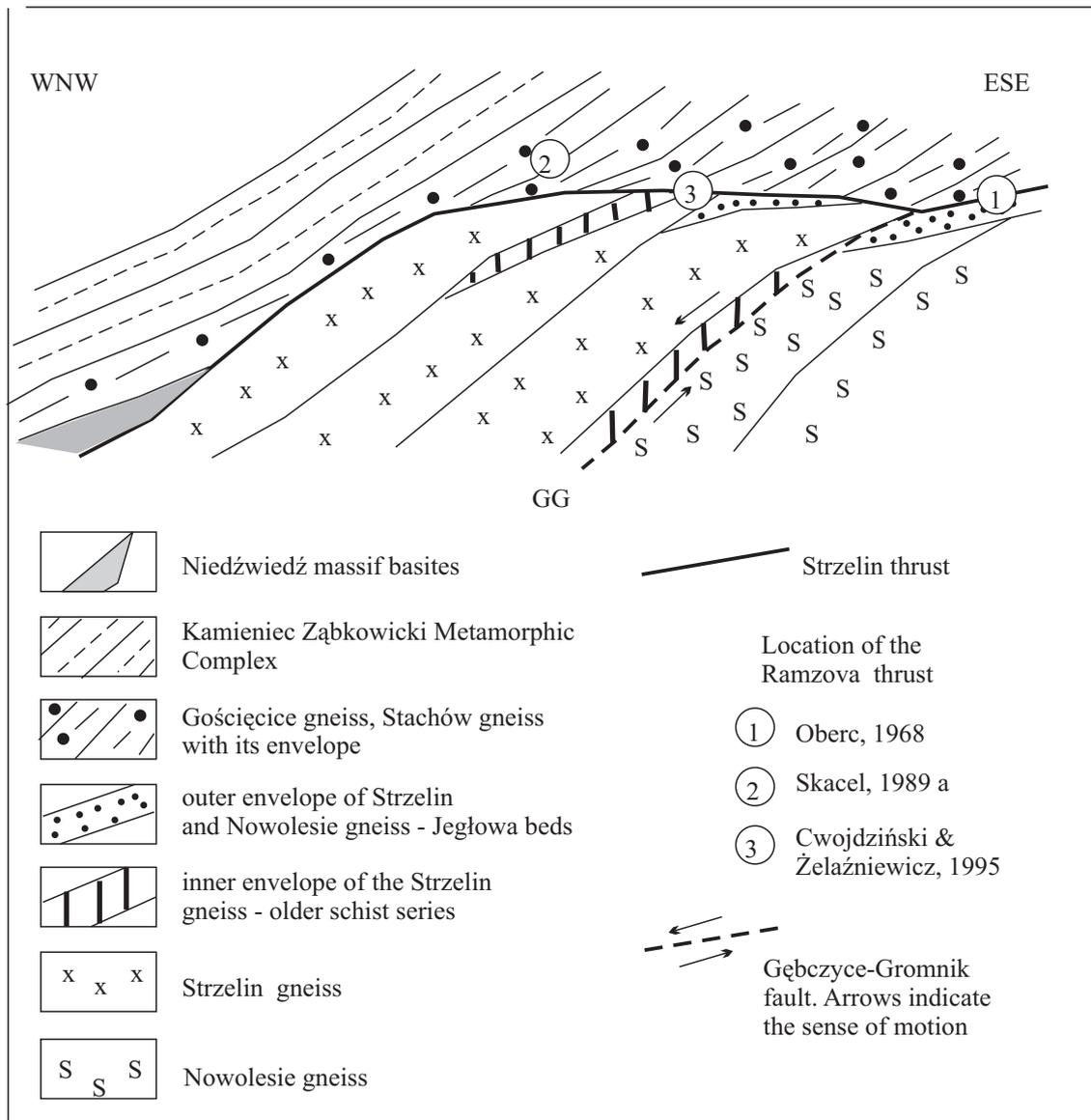
Fig. 21. Schematic section across the Lipowe Hills and the Strzelin massif showing position of the Strzelin Thrust.

plex along the Oława river (Skácel, 1989 a), i.e. east of the Lipowe Hills. The earlier-mentioned modification does not refer to this general concept but to the fact that from the geological point of view, the Lipowe Hills are a part of the Strzelin massif (Oberc-Dziedzic, 1995), so the East-West Sudetes boundary should be placed west of them, i.e. along the Mała Ślęza river.

In this paper we discussed the problem of the East-West Sudetes boundary mainly in the context of the Strzelin massif. The position of this boundary in other parts of the Fore-Sudetic Block is still unclear because of the very poor exposure of the area and numerous faults of E–W and NW–SE orientation, which probably change its position. New petrological, structural and age data are necessary to solve this problem.

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