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MULTISTAGE EVOLUTION OF THE SHEAR ZONE AT THE BASE OF THE GIEWONT UNIT, TATRA MOUNTAINS (POLAND)

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Abstract: The paper presents a mesostructural and petrotectonic analysis of rocks from the contact zone between the Giewont and Czerwone Wierchy Units (High-Tatric nappes). Rocks occurring in the vicinity of Siadła Turnia and Turnia Olejarnia earlier referred to the "brecciated Campilian" (Late Scythian), are in reality mylonites (mainly dolomitic mylonites), and their unique preservation was possible due to dilatant sites linked with faults developed in the basement of the thrusting Giewont Unit. The mylonitization process as well as the thrusting of the nappe was not a one-stage, but a multi-stage re-activated process. Its cyclicity was determined by the build up and drop of pore fluid pressure, leading to changes of rheological behaviour of the deformation process. Fluids released to the shear zone together with the brecciated rock formed a suspension with low friction values, which acted as a "water pillow" facilitating the movement of the nappe mass. Deformation and mylonitization processes, the temperature of which reached 300 °C in some cases, accompanied further stages of tectonic transport.

Key words: Tatra Mts, shear zone, dolomitic mylonites, fluid pressure, hydraulic fracturing, pressure solution.

Introduction

The Tatra Mountains are composed of a crystalline core, overlain by a Mesozoic sedimentary cover and two nappes: High-Tatric (Tatric Superunit) and Krížna (Fig. 1A). They represent the eastern prolongation of the Austroalpine tectonic system into the Central Western Carpathians. In the Tatra Mts, Alpine thrusting and folding took place after the Turonian and are linked with the Mediterranean phase (Andrusov 1965). Analysis of the geometry of the Alpine thrust folding in the Tatra Mts is difficult to conduct due to the high activity of pressure solution processes (Bac-Moszaszwili et al. 1981; Jaroszewski 1982), and thus with the lack of sufficient data of slip structures for statistic analyses. Therefore, earlier reconstructions of the thrust processes were based on slickensides from crystalline rocks of the so-called "Goryczkowa island" (Burchart 1963) and from the granitoid core of the High Tatra Mts (Jurewicz 2000a). Earlier papers were also devoted to temperatures and pressures during the Alpine thrust folding. According to Lefeld (1997), the pT condition of thrusting did not exceed 200 °C and 1 kbar at the contact of granitoid core and sedimentary cover boundary. At this time, the maximal temperatures for the crystalline rocks could reach 300–350 °C according to Janák (1994), whereas Putiš (1992) indicated that they did not exceed 300 °C at depths of 6–8 km. Investigations based on liquid + gas fluid inclusions on slickenside surfaces in the upper parts of the granitoid core of the High Tatra Mts indicate pressures of 1.4–1.7 kbar and temperatures not exceeding ca. 250 °C (Kozłowski & Jurewicz 2001). For the uppermost Krížna Nappe, Grabowski et al. (1999) suggest low temperature values (50–80 °C), which indicate a cold regime of folding.

This paper presents the conditions, under which the thrust of Giewont Unit on the Czerwone Wierchy Unit developed within the High-Tatric Nappe, on the basis of structures occurring within dolomitic mylonites of the so-called "Myophoria beds" of Late Scythian age in the vicinity of Siadła Turnia and Turnia Olejarnia (Figs. 1B, 2A,B).

Geological setting of the so-called "brecciated Campilian"

The investigation area is located on the western slopes of the Giewont Mt, dropping towards the Mała Łąka Valley (Fig. 2A). The thrust contact of dolomitic-marly breccia, considered by Kotański (1956, 1959a) to be of Late Scythian age, referred to the "Myophoria beds" (*Myophoria costata* Zenk) and included in the Giewont Unit, with the Urgonian limestones of the Czerwone Wierchy Unit (Organy subunit — *op. cit.*) can be observed in several couloirs in the vicinity of Turnia Olejarnia and Siadła Turnia. In the upper part of the section the breccias are overlain by black shales interbedded with black bituminous limestones and yellow-weathering dolomites. Platy dolomites considered by Kotański (1959a) as the so-called "supra-Myophoria beds" occur above. Older members of the Giewont Unit can be observed under the Upper Scythian breccias to the southeast of Siadła Turnia (vicinity of Kondratowa Pass). These include Upper Scythian cellular dolomites, Lower Scythian sandstones and crystalline rocks of the "Goryczkowa island". Deposits overlying these dolomites represent the Anisian and begin with a regional-scale basal conglomerate (cliff breccia), comprising poorly rounded pebbles of yellow-weathering dolomites and grey-green shales.

Interbedded complexes of blue-black limestones and yellow dolomites, with total thicknesses of 70 to 160 m represent younger Anisian members (Kotański 1956, 1959a). The uppermost, up to 200-m thick, part of the sequence comprises carbonates of Doggerian, Malmian and Neocomian age and Albian marls. The total thickness of beds in the Giewont Unit reaches ca. 500 m and is smaller than in other structural elements of the High-Tatric Unit.

The Giewont Unit probably represents the remains of an initially larger structure (Bac-Moszaszwili et al. 1984), with beds displaying steep northern dips (Fig. 1C). After reversal to positions prior to the young Tertiary uplift of the Tatra Mts (Jurewicz 2000b), they most commonly attain 150/45N (Fig. 1D), thus pointing to a more north-western direction of tectonic transport in relation to the northern directions in other units of the High-Tatric Nappe (Bac-Moszaszwili et al. 1984; Jurewicz 2000b).

This paper will focus on the so-called "brecciated Campilian", representing the Late Scythian, in the vicinity of Siadła Turnia (Fig. 2B,C) and Turnia Olejarnia. When describing similar breccias from other sections in the Giewont Unit, Kotański (1956, 1959a) noted that they comprise poorly rounded, dark grey, yellow-weathering dolomite fragments. The binding material is composed of numerous very fine dolomite fragments, in some cases with visible lamination. The breccias are of marine origin and were developed as intraformational breccias during intense storms when the wave base reached the seabed. They are cut by epigenetic calcite veins and contain ferruginous mineralization (pyrite, limonite). It is worth noting that besides sedimentary breccias, tectonic breccias also occur: the rock is tectonically fractured and the fractures are filled with calcite. There is a lack of transport between the fragments. In relation to their structure, these kinds of breccias represent "crackle breccia" (Kotański 1954).

According to Jaroszewski (1982), the contact of the Krížna Nappe with the High-Tatric Unit, which is generally the contact of Middle Triassic dolomites thrust on the Malmian-Neocomian or Urgonian limestones, is a macrostylolite developed in the course of pressure solution. In other places, the dolomites of the Krížna Nappe penetrating by means of pressure solution mechanism into the High-Tatric limestones bear traces of deformation in the course of cataclastic flow. The described below contact of the Upper Scythian dolomites of the Giewont Unit with the Urgonian limestones of the the Czerwone Wierchy Unit is an example (Jaroszewski 1982).

Paulo (1997) suggested that the pyritic facies of the ferruginous Upper Scythian deposits in the High-Tatric and Krížna Nappes represent pre-salinary sediments. Brownish and black limonites with clusterous or clotty-colomorphic texture were described in them. X-ray diffraction analyses (Paulo 1997) indicate the presence of goethite, hydrogoethite and lepidocrosite. Zawidzka (1967), characterizing the Late Scythian of the Krížna series from a nearby site located in the western part of the Mała Łąka Valley (Sywarowa Pass), noted the presence of crystalline sulphur, forming 1-cm in diameter concentrations, developed in the course of sulphate reduction (anhydrite or gypsum) by organic matter.

Plašienka & Soták (1996) described carbonate tectonic breccias from the Central Western Carpathians formed through fracturing and crushing with a dominant role of percolating solutions under intricate fluid regimes by multiple pressure solution, chemical alternations, leaching, weathering with concentration of Fe-hydroxides and karstification; these rocks in German terminology are referred to as Rauhwacke or Zellendolomite. According to Plašienka & Soták (1996), Rauhwacke typically comprise carbonates (dolomites, rarer limestones) and accompanying sulphate (gypsum and anhydrite), in some cases also salt. The radically different mechanical behaviour of these rocks causes that dolomites undergo brittle disintegration, as well as sulphate ductile flow, starting already at temperatures of 100 °C (Schmid 1982). Transformation of gypsum into anhydrite and the reverse process induce stress responsible for brecciation of the carbonate rocks. The developed intraformational breccias are referred to as dilation breccias. According to Plašienka & Soták (1996), such breccias occur in many units of the Western Carpathians in red shales of Permian-Scythian and Norian (Keuper) age. From the area of the Northern Calcareous Alps, Spötl & Hasenhüttl (1998) describe evaporate rocks from a tectonic mélange (Haselgebirge), the distribution of which is restricted largely to the topmost thrust unit (Juvavicum). According to Warren (1999), "... in older studies, the Rauhwacke itself was considered to have facilitated décollement during Alpine tectonics, but now it is known that the evaporite-lubricated protoliths of the Rauhwacke acted as detachment horizons during thrusting and folding".

The rocks described below, the unique character of which is a result of tectonic processes, are a source of information on processes linked with the Alpine nappe thrusts. The described sites from the vicinity of Siadła Turnia and Turnia Olejarnia are not exceptional in the Tatra region, where the "Myophoria beds" act as a "lubricant" for the thrust plane; on the contrary, according to Kotański (1959b) this is a regional phenomenon and the "Myophoria beds" commonly occur in the lowermost part of the thrust of the upper limb of the Czerwone Wierchy fold. When describing the contact of the High-Tatric flake with the Krížna Nappe, Zawidzka (1967) also characterized the Upper Scythian breccias of the Krížna Nappe as strongly tectonically deformed with a texture resembling that of augengneisses.

Meso- and microstructural characteristics of the shear zone

The tectonic contact between the Giewont and Czerwone Wierchy Units in the vicinity of Siadła Turnia and Turnia Olejarnia is an uneven, several tens of centimetres to 2–3 m thick zone with textures parallel to its local orientation, numerous slip, mineralization and recrystallization planes. The contact of both units is not a plane and is not clearly delimited from the surrounding rocks; its morphology is rather complex, and it is further deformed by small faults. The zone lacks slickensides that would enable geometric analysis of the thrust and tectonic transport directions.

Rocks occurring at the base of the Giewont Unit are variously tectonically deformed. Typically, they are strongly folded, and the folds have features of disharmonic folds developed in ductile deformation conditions (Figs. 2C, 3A,B). In most cases the rocks are mylonites — where the mylonites are de-

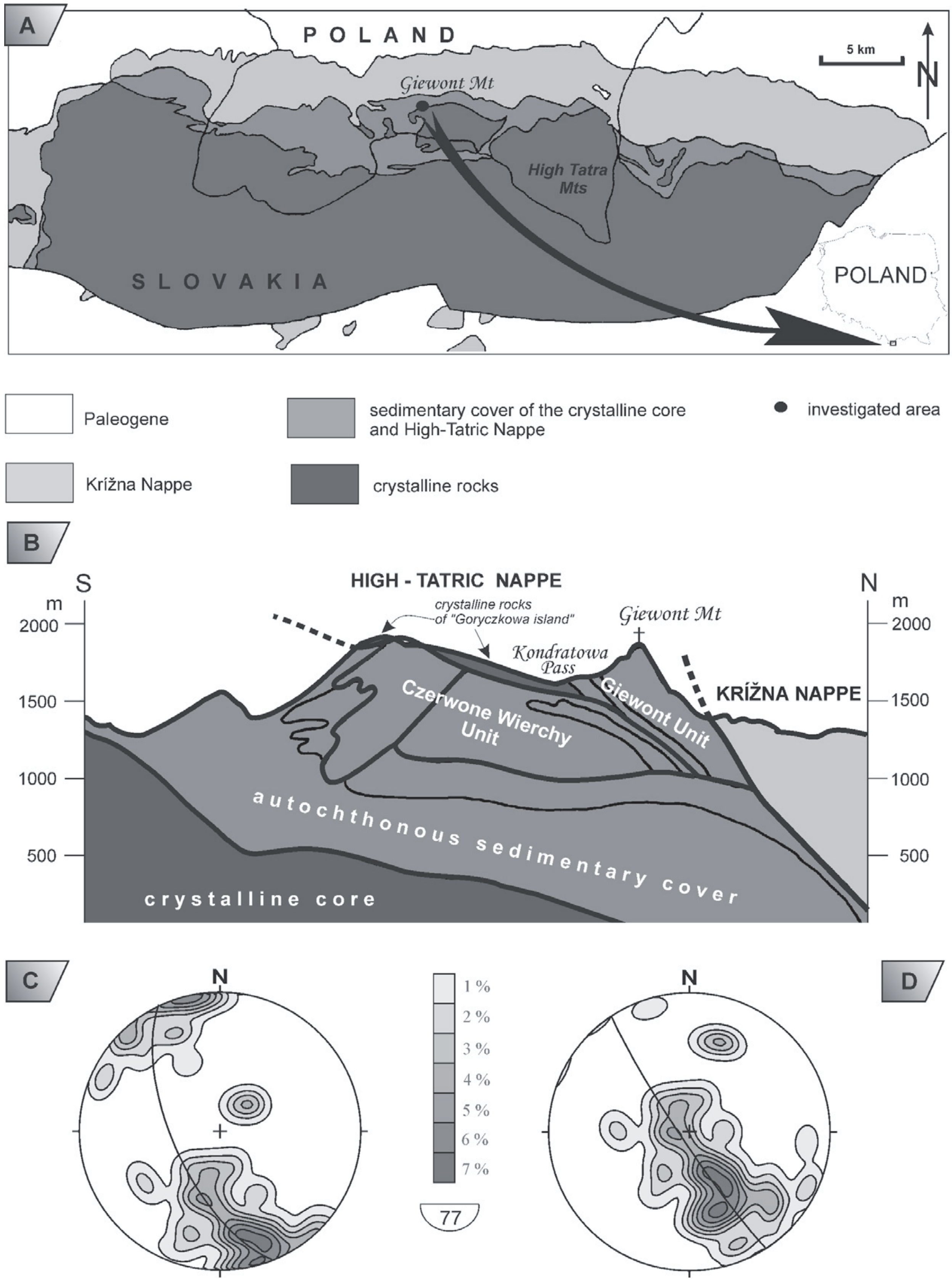


Fig. 1. A — Study area in relation to the main geological tectonic structures of the Tatra Mts (after Bac-Moszaszwili et al. 1979). **B** — Schematic geological cross-section through the Giewont Unit (after Bac-Moszaszwili et al. 1979). **C**, **D** — attitude of bedding in the Giewont Unit (after SteroNet software; pole to planes): C — present day position, D — after vertical rotation (40° southwards around the 90/0 axis) to the pre-Late Tertiary position.

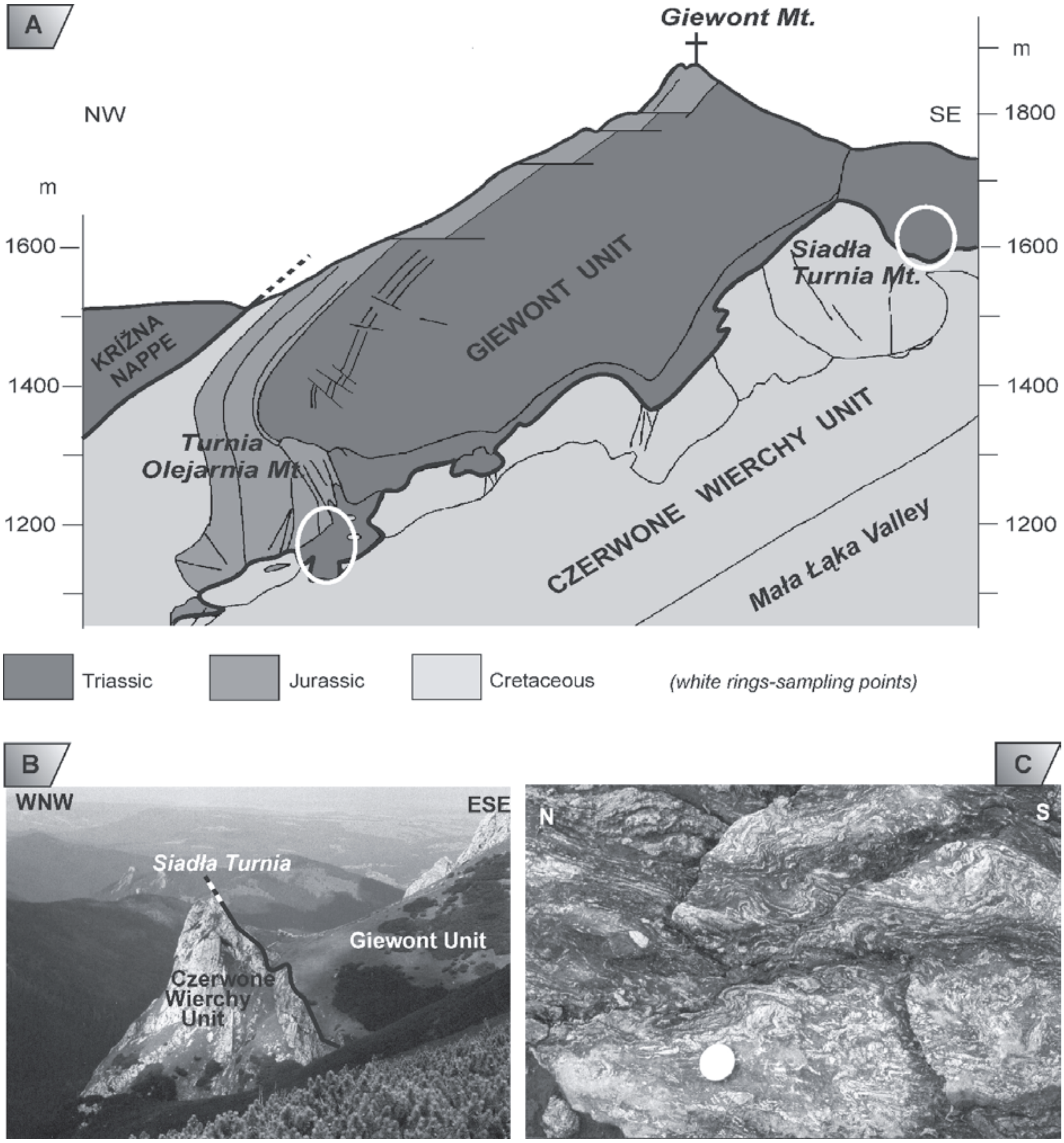


Fig. 2. A — View of the SW slope of Giewont Mt. Photograph and its geological interpretation after Gąsienica-Szostak (1973). **B** — tectonic contact of the Giewont and Czerwone Wierchy Units in the vicinity of Siadła Turnia. **C** — fold deformation at the base of Giewont Unit in the vicinity of Siadła Turnia (scale — 10 groszy coin).

fined as foliated and lineated rocks showing evidence for strong and ductile deformation and are understood as a strictly structural term referring only to the fabric of the rock (White et al. 1980; Passchier & Trouw 1998). Locally, that is in the vicinity of Turnia Olejarnia (Figs. 3E, 5D), although macroscopically the rock resembles strongly folded schists, the degree of mylonitization is high enough to observe the prevalence of the matrix in relation to the porphyroclasts; when the matrix exceeds 90 % of the rock (Sibson 1977) and the grain size of the recrystallized matrix is typically smaller than 10–20 μm (Hippert & Hong 1998), they can be referred to as ultramylonites.

Despite the different degree of tectonic deformation (Figs. 4, 4D–H, 5), the structural features of the described rocks allow to distinguish them as mylonites; in most cases they are represented by dolomitic mylonites. They comprise clasts of dolomite (black, red and grey-yellow) of various sizes, shapes and degree of rounding (Figs. 3A–D, 4). The breccias bear traces of various stages of textural transformation, linked with tectonic processes in the thrust zone. Most clasts

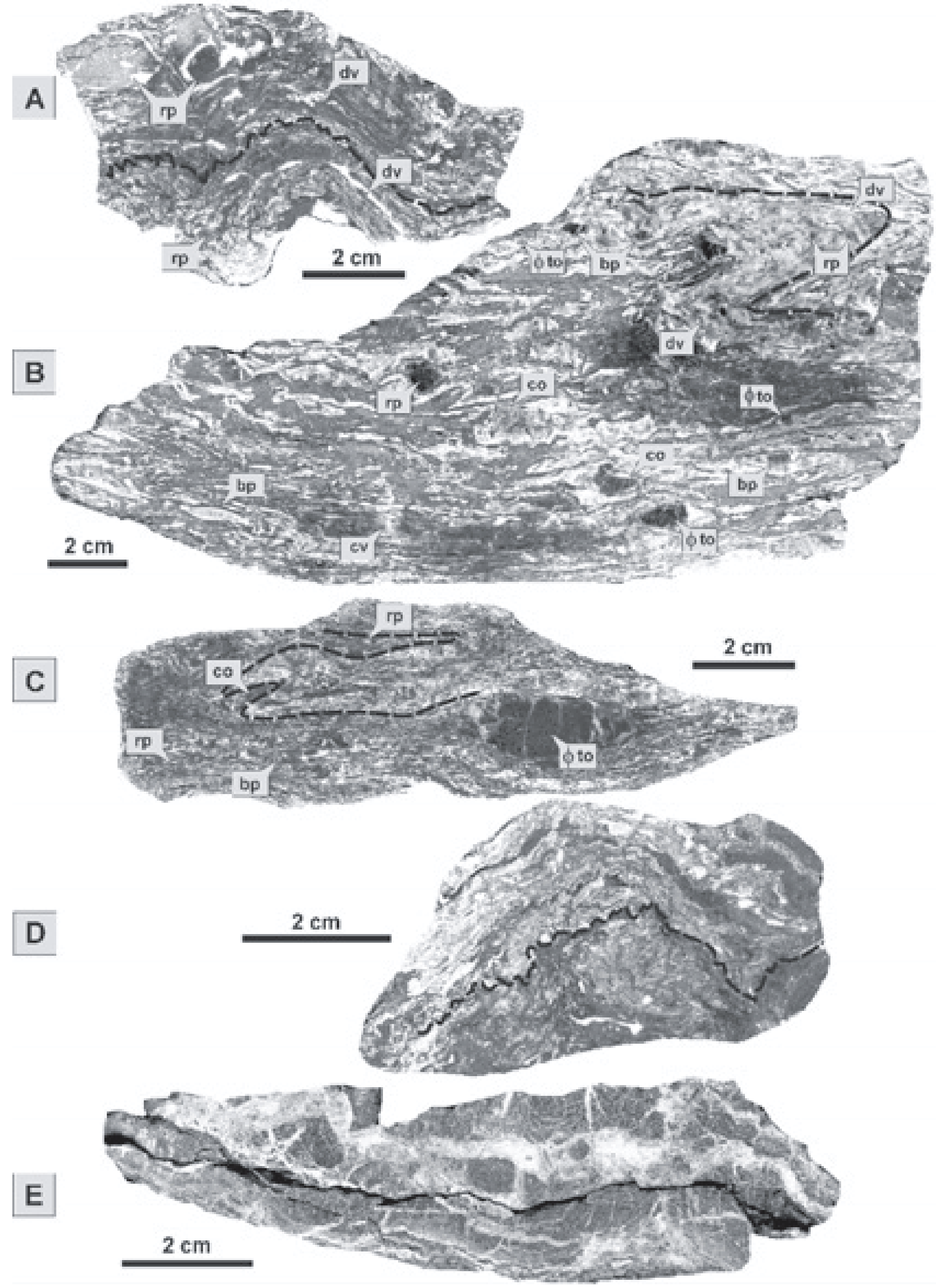


Fig. 3. Different degrees of deformation in dolomitic rocks from the shear zone near Siadła Turnia (A–C) and Turnia Olejarnia (D, E); polished surfaces. **A, B** — microfolds in foliated mylonite and dolomite veins parallel to foliation; **C** — strongly folded mylonite with well-developed foliation and single rounded porphyroclasts; **D** — microfolds within ultramylonite; **E** — ultramylonite with numerous extensional fracturing (dolomite veins; axial part of wider veins is filled with calcite). This is evidence that the dolomite mineralization is earlier than the calcite one. dv — dolomite vein, cv — calcite vein, bp — boudinaged porphyroclast, rp — rotated porphyroclast; mantled porphyroclast: θ–type object, Φ–type object, co — complex object; dashed line — contour of microfolds.

do not exceed 2–3 cm in diameter; they are typically elongated and stretched, with the longer axis parallel to the local elongation of structures. Lamination is in some cases quite distinct in the clasts, which to a certain degree is an inheritance of sedimentary structures, but also of dynamic recrystallization (Figs. 4E, 5G). The elongation degree, resulting from simple shear extension taking place within the thrust zone, is sometimes so large that the structure resembles that described by Davis & Reynolds (1996) as stretched-pebble conglomerate. The elongated clasts are frequently boudinaged or form augen-structures in a finer-grained, foliated matrix (Fig. 3B–D). Dolomite porphyroclasts are commonly flanked by fine-grained dolomite aggregates forming mantled porphyroclasts (Φ-type objects, Figs. 3B,D, 4D–F, 5D, terminology after Passchier & Trouw 1998). In the Turnia Olejarnia region, flanking aggregates around dolomite porphyroclasts consist of SiO_2 minerals, deforming into wings and forming δ-type objects (Fig. 5D). In many cases round-shaped porphyroclasts bearing features pointing to rotation, and forming complex objects (Fig. 3B,C), can be observed macroscopically.

The mylonite matrix is composed of grain fragments with a wide range of grain size, in which the smallest ones have diameters of several to several tens of microns. The flat-parallel textural arrangement can be observed in the matrix (Fig. 4E,F). Spaces between the laminae are typically filled by several millimetre thick veins of dolomite (Figs. 3A–D, 5F). The mylonitized rocks also bear signs of later strong tectonic deformation. Folds resembling disharmonic folds developed in flow conditions (Figs. 2C, 3A–D) can be observed. The folds are typically of small sizes, and their amplitudes generally do not exceed 10 cm. The flow shows a local arrangement resulting from the geometry of the contact with the underlying Czerwone Wierchy Unit. In direct vicinity of the thrust surface the small folds are isoclinal. In some cases folds with amplitudes up to several tens of centimetres can be observed. The folding is not only distinguishable in the field but also in microscopic scale, and includes both elongated clasts and foliation in mylonites, as well as veins of dolomite within the laminae (Fig. 4G).

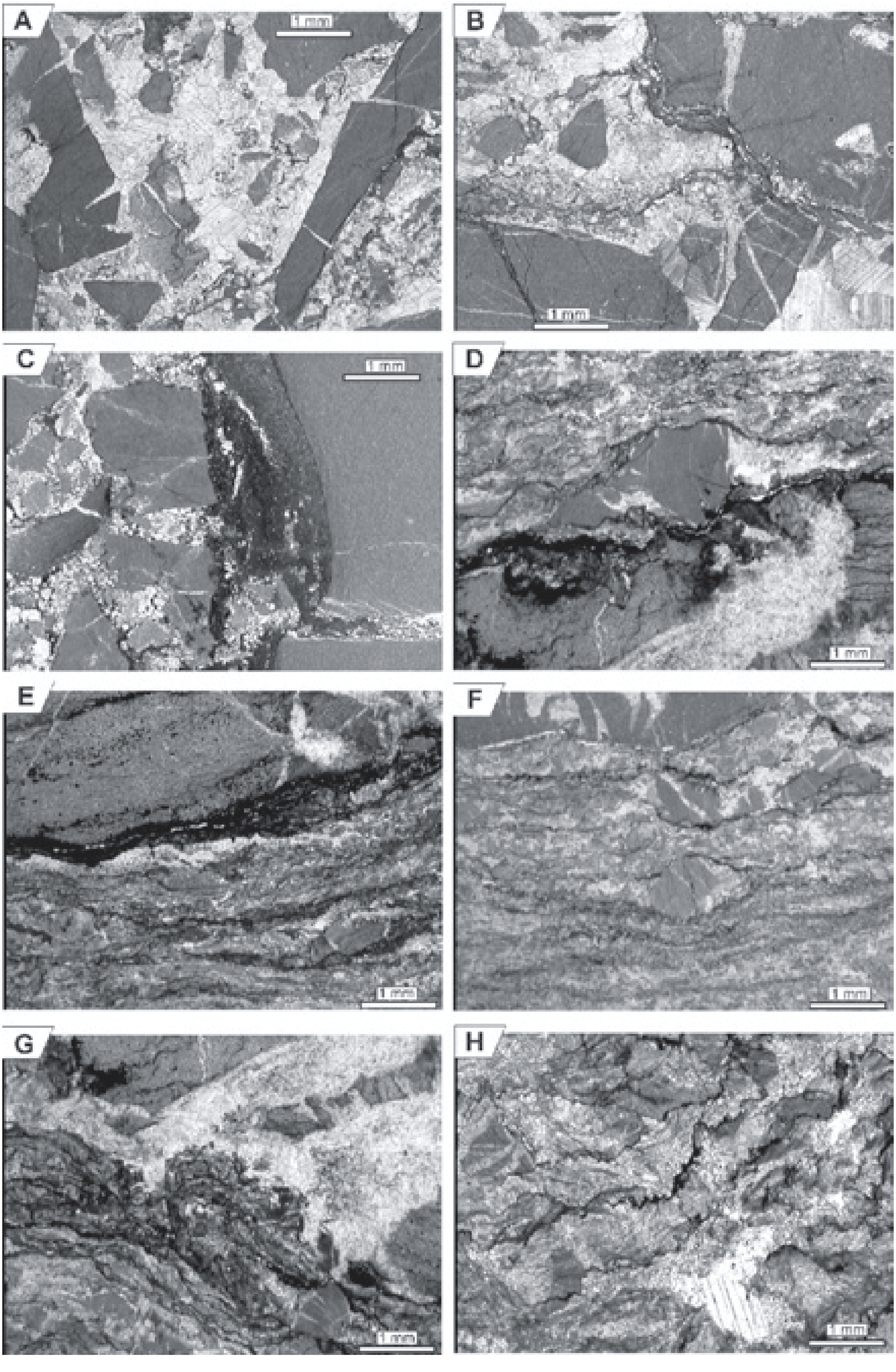
Later deformation of the mylonitized rocks includes not only folding but also extensional fractures filled with coarse-crystalline dolomite, rarer with calcite. Such mineralization in some cases uses the S-C fabric (Fig. 5E) and in others — fractures developed in course of hydraulic fracturing (Fig. 3E).

Although the described rocks commonly contain structures, which can be used to determine the sense of shear (e.g. Figs. 3B, 4D), these are insufficient for interpretations. Their occurrence only proves the multistage character of the movement and the ductile type of deformation responsible for their creation. Similarly, due to the non-planar character of the thrust surface linked with the presence of faults in its lowermost part, common participation of pressure solution processes and most probably its secondary folding deforming the geometry of the thrust zone, the orientation of the directional textures and structures is local in character, has a large variability and thus cannot be used in reconstructions of the tectonic transport direction. Analyses of stress fields responsible for the nappe thrusts and of directions of tectonic transport were carried out on the basis of tectonic striae orientations on the surfaces of Alpine-age slickensides in the granitoid core of the Tatra Mts (Jurewicz 2000a).

At this stage of investigations it is difficult to give a definite statement whether the rocks occurring at the tectonic contact — mylonites or ultramylonites — initially represented sedimentary breccia (such as those commonly occurring in the Upper Scythian) or their present character is a result of tectonic processes. Although the degree of structural transformation is very high, it can be presumed that it was the sedimentary breccias that were the weakest lithological member, thus being most the susceptible to deformation. The latter feature was favoured by the presence of fractures and pores, which could develop both during the dolomitization process, as well as during the formation of sedimentary breccia. If the original chemical composition and rheological properties of the rock, which acted as a lubricant during the thrusting of the particular tectonic units, were known, the physical-chemical properties during deformation could be recognized more precisely. Unfortunately, the large participation of pressure solution processes caused the deterioration of such unstable minerals as halite or gypsum, the mechanical conditions of which could significantly influence the thrusting process. Water from fractures and pores, as well as that released, for example, during dehydration of gypsum, could influence the stress values. Knowledge of the initial lithology would also help in determining the chemistry of the fluids and evaluation of their influence on the course and magnitude of solution processes. For instance, in the vicinity of the Glarus thrust in the Alps of eastern Switzerland, a 36% volume decrease of flysch sandstones was noted in the process of Helvetic nappes formation (Ring et al. 2001), where the open-system behaviour was probably driven by dissolution and bulk removal of the more soluble components of the rock due to flow of a solvent fluid phase on a regional scale.

The character of the clast deformation (elongated and folded — Fig. 3B,C), lamination of fine material and its folding (Fig. 3A–D), and the means of recrystallization (dynamically recrystallized grains — Fig. 5G) point to ductile rheological behaviour. Lamination in the fragmented material may be linked with the fact that the freely soluble components (calcium carbonate, gypsum and halite) were removed, leaving

Fig. 4. Different degree of brecciation and mylonitization of rocks from the shear zone in thin sections; non-polarized light. **A**, **B** — brittle fracturing in brecciated dolomites, sharp-edged grains displaced by microfaults; spaces between clasts filled with crystals of dolomite; **C** — "vein" of mylonite within tectonic breccia, in which matrix is composed of dynamically deformed quartz; **D** — crystal-plastic deformation with stylolites parallel to foliation in mylonitized dolomites (upper part) and bedding parallel stylolites in the dolomite clast (on the right); mantled porphyroclast with wings of crystalline dolomite (in centre); **E, F** — well developed stretching lineation, with different degree of grain size reduction, stylolites parallel to foliation (both) and within clasts (centre of F); upper part of F — elongated and boudinaged clast; centre of F — mantled porphyroclast with wings of crystalline dolomite; **G** — microfolds of mylonitic textures; **H** — well developed stylolites occurring subparallel to foliation.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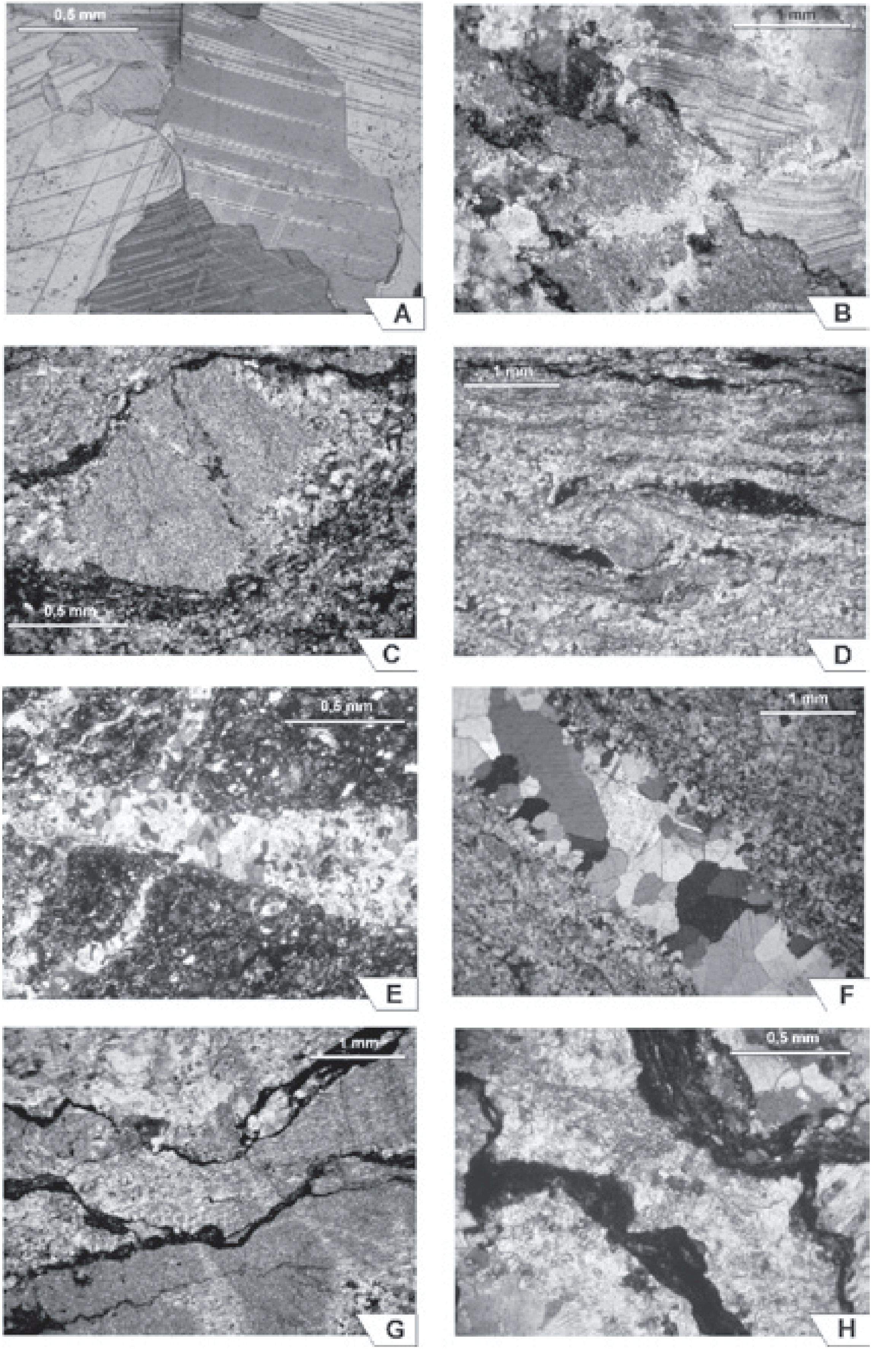
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fragments of dolomites as the residuum. Rounded porphyroclasts bear traces of selective pressure solution (Figs. 4D,F, 5C) in domains where the stress was relatively high (Knipe 1989). Overgrowth of silicate, dolomite and calcite grains in strain shadows can be observed in the case of dolomite porphyroclasts (Figs. 3A–C, 4D–E, 5C–D).

The next evidence of pressure solution is the presence of stylolites on the dolomite clast margins (Fig. 4D,F), between laminae in the mylonites (Figs. 4E,F, 5G), as well as cutting carbonate veins and crystals (Figs. 4H, 5B). Furthermore, processes of intracrystalline deformation (Fig. 5A,B), grain size reduction and preferred grain shape orientation (Fig. 5E,F) take place (cf. De Roo & Weber 1992).

Petrology and mineralogy

Analysis of thin sections in polarized light and microprobe determined the mineral content of the investigated rocks (both porphyroclasts and matrix), mineral veins, relations between dolomite and calcite, as well as the means of recrystallization of material during deformation. The petrological characteristics were based on the analysis of 24 thin sections from samples collected from beneath Siadła Turnia and Turnia Olejarnia. Cathodoluminescence analyses were also conducted. The investigated rocks, however, although containing numerous carbonate veins with large euhedral crystals, are hardly luminescent at all. This indicates the lack of distinct zones of crystal growth, which may be linked with the very fast growth of crystals at a relatively stable solution composition, or with their later homogenization. The latter possibility — due to the multistage development of mylonites — seems to be highly probable.

Microscopic investigations (Fig. 6) were carried out on three thin sections from Siadła Turnia from samples lying within 1 m from the contact with the Urgonian limestones (Kotański 1956) of the Czerwone Wierchy Unit, and on a sample taken from the direct vicinity of the thrust surface in region of Turnia Olejarnia. The earlier optical identifications of the minerals were confirmed by investigations in SEM connected to EDS detector. Microprobe images allowed recognition of the relations between the matrix components and distinguishing calcite and dolomite. Several chemical analyses of the composition of feldspars, of which most are potassium feldspars, were also carried out.

Fig. 5. Thin sections in crossed polarizers. **A**, **B** — dolomite crystals with well developed twins and their deformation (irregular shape of twin boundaries); **C** — mantled porphyroclast of dolomite (Φ-type object with tails of recrystallized dolomite); **D** — rotated porphyroclast of dolomite with tails of silica (δ-type object) in ultramylonites from Turnia Olejarnia; **E** — fragment of S-C structures in ultramylonites from Turnia Olejarnia; quartz in matrix is dynamically deformed; extensional fractures filled with dolomite crystals; **F** — thin dolomite vein parallel to foliation within fine-grained dolomite mylonites — fragment of microfold; **G** — in central part — dolomite porphyroclast fragment with traces of mimetic recrystallization, parallel to primary lamination, in lower part porphyroclast with fractures, along which recrystallization took place; bedding-parallel stylolite also visible; black veins-mylonitized matrix with Fe-hydroxides; **H** — black veins-mylonitized matrix with Fe-hydroxides (as in G).

Because the presence of sulphur was observed in the investigated rocks (Zawidzka 1967), which, according to Paulo (1997) can be considered pre-salinary, investigations in microimages were focused on the presence of minerals typical for evaporatic rocks (gypsum, halite). The negative result is not, however, an evidence of their initial absence in the sediment, as pressure solution processes could lead to their complete removal.

The bulk of the investigated rocks are porphyroclasts of dolomites (up to several centimetres in diameter) and matrix — also mainly dolomitic — filling the space between the fragments. The dolomite porphyroclasts are typically strongly fractured, and recrystallization can be observed along these fractures, which were not filled by later mineralization. The dolomite crystals are elongated and distributed semi-perpendicular to the fractures, which were migration paths for fluids favouring recrystallization in the direct vicinity of fractures. Similar recrystallization took place along sedimentary laminae, which could also become paths for fluid migration and cause the mimetic growth of dolomite crystals in the direction of foliation (Fig. 5G). The matrix filling spaces between the porphyroclasts, besides dolomites, the finest fragments of which are below 10 μm, includes:

a) aggregates of euhedral dolomite grains with sizes of several tens of μm (Figs. 6B, 7A) and easily visible twinnings (Fig. 5A); in some cases rombohedrons are rimmed with calcite, which may be an artefact of pressure solution;

b) subhedral crystals of calcite with sizes typically exceeding 100 μm, in some cases overgrowing euhedral dolomite (Fig. 6B);

c) crystals of K-feldspars, up to several tens of μm in diameter, in some cases with inclusions of romboedric dolomite (Fig. 6A,E) or anhedral quartz crystals (Fig. 6F);

d) sporadic euhedral crystals of plagioclase not exceeding 2–3 μm;

e) anhedral quartz crystals (occurring in aggregates), with sizes from several to several tens of μm, and hypereuhedral quartz crystals within a feldspar-silicate layers mass (Fig. 6D–F); some crystals reveal undulose extinction indicating crystalplastic deformation;

f) microcrystalline concentrations of SiO_2 filling spaces between the quartz grains (Fig. 5E) and forming tails in δ-type objects (Fig. 5D);

g) aggregates of feldspar-silicate layers comprising grains up to several μm in diameter (Fig. 6);

h) euhedral pyrite and iron oxides crystals up to several tens of μm (Fig. 6F);

i) single sheets of chlorites not exceeding 10 μm (Fig. 6C);

j) titanium oxides (<10 μm);

k) single euhedral apatite crystals (<10 μm).

The matrix is characterized by flat-parallel textures, developed in the course of cataclastic deformation, and resulting from:

a) distribution of flattened mineral grains (mechanical rotation), as well as their later flattening and elongation;

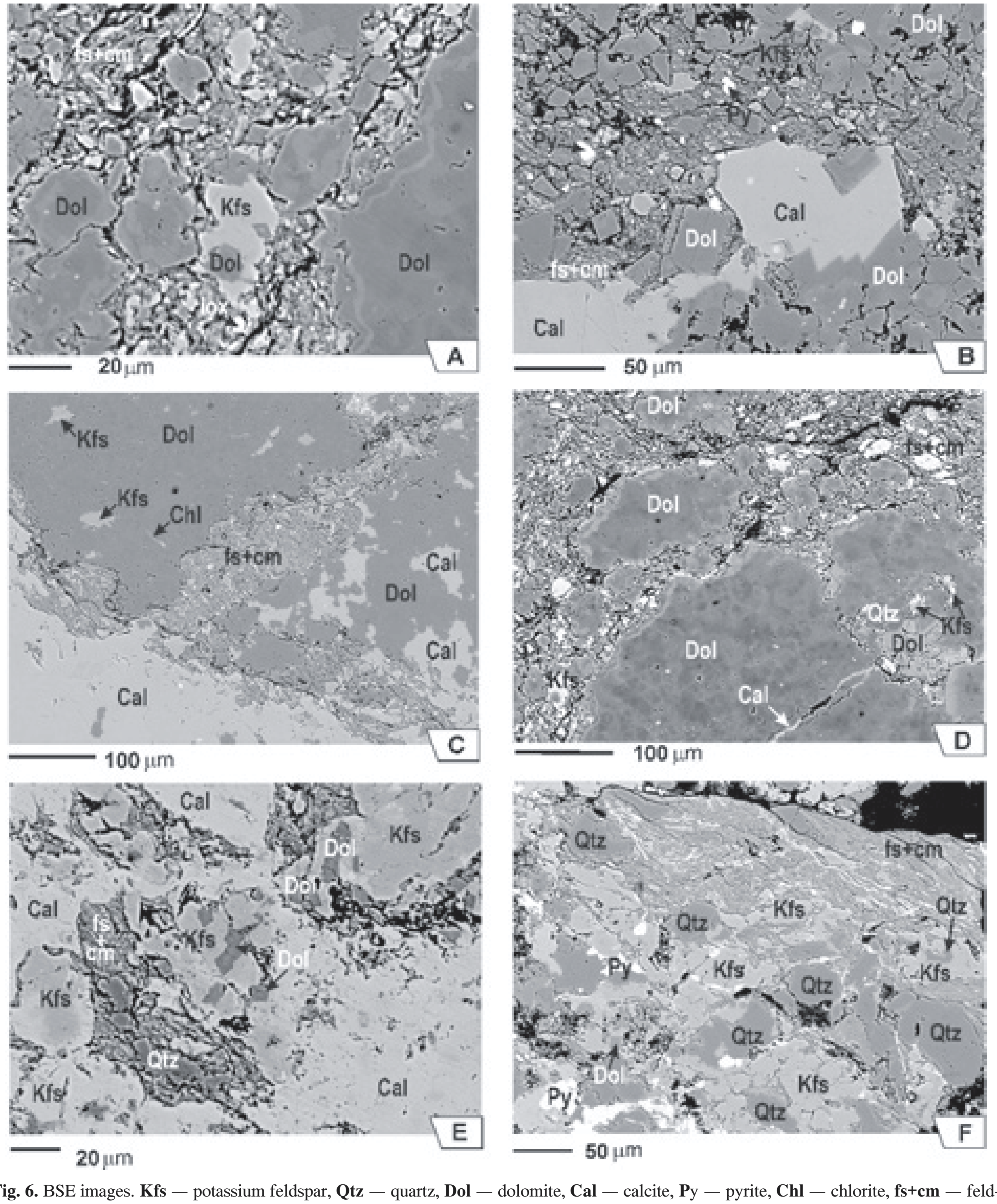

Fig. 6. BSE images. **Kfs** — potassium feldspar, **Qtz** — quartz, **Dol** — dolomite, **Cal** — calcite, **Py** — pyrite, **Chl** — chlorite, **fs+cm** — feldspar and clay minerals.

b) preferential grain growth of some crystals, for example, along foliation surfaces;

c) dynamic recrystallization linked with simple shear during thrust napping;

d) stress-induced solution transfer — preferential dissolution of poorly rounded clasts. In effect, stylolites are formed on the margins of clasts, along relict bedding within the clasts and parallel to foliation in mylonites (Fig. 5D,E,F,H).

Additionally, the rock also contains numerous veins of dolomites parallel to foliation, along small folds. The means of filling the space between laminae by large hypautomorhic crystals is evidence of growth in extensional conditions and a relatively small degree of later deformation (Fig. 5F), which only locally lead to grain boundary migration and twinning deformation (Fig. 5A,B). Besides veins of crystalline dolomite, veins of calcite with large interlobate crystals, unconformably cutting the fold structures, are also present. The calcite crystals are commonly twinned, often with traces of deformation. In some cases the twins attain serrated boundaries due to grain boundary migration.

The described rocks, although spatially not forming regular zones with a composition typical for mylonites, can be generally classified as such, both according to microtectonic definitions (e.g. Passchier & Trouw 1998), the main criterion of which is the degree and character of deformation, and according to petrographic definitions requiring not only the deformation, but also the dynamic recrystallization of grains, or the presence of neomorphic crystals (Yardley 1991; Lin 2001), to which in this case feldspars, recrystallized dolomite, layer silicates, quartz and other minerals of the silicate group can be included.

Stages of thrust-napping and microstructure evolution within the shear zone at the base of the Giewont Unit

Stage A

The stress values increase in conditions of horizontal compression preceding nappe development of the Tatra Mts (e.g. Andrusov 1965; Kotański 1961; Plašienka 1991), and thus the thrust of the Giewont Unit on the Czerwone Wierchy Unit. In the first stage (Fig. 7A) horizontal compression causes the formation of a symmetrical fold. Ductile deformation relaxes the stress.

Stage B

Further compression induces the development of simple shear. The fold becomes more asymmetrical, for example, due to summing up of the intrastratal slip. When beds become too steep in relation to the directions of tectonic transport to allow stress release through intrastratal slip, the fault plane can be formed by breaking the cohesion of the rocks (Fig. 7B). According to the model of Mitra (2002), forelimb shear thrusts form, due to rotation and layer-parallel extension of the steep forelimbs of folds, in the late stages of folding.

At the same time, with the increase of stress values, the pore fluid pressure builds up as well. In a fluid-saturated rock mass the build up of fluid pressure (P_f) causes the reduction of all normal stresses (σ_n) to give effective stresses (Fig. 8B), where $\sigma_{ef} = \sigma_n - P_f$ (Hubber & Rubey 1959; Sibson 1996).

In a deformed rock mass the presence of pores may be linked with the leaching of freely soluble components, such as gypsum (thus the term "cellular dolomites" — e.g. Kasiński 1981; Passendorfer 1983) and with the breccia character of the deposits (Kotański 1954). Fluids infilling the pores are of a meteoric type or originate from processes such as the dehydratation of gypsum. Assuming that differential stress is small, the Mohr circle passes to the left, to the tensile failure envelope (Hancock 1985). Thus a process leading to the development of a fault, which takes place due to hydraulic fracturing is responsible for the formation of tectonic breccia (De Roo & Weber 1992). It is accompanied by the release of fluids from the pores, thus the rock mass filling the space between the fault walls is a multiphase mixture (Treagus & Treagus 2002). It can be assumed that hydraulic fracturing spreads out gradually into the fault walls leading to their brecciation.

Stage C

The displacement accompanying brecciation causes stress release and the pore system within the shear zone becomes open. In this stage the thrust zone is a flow path for fluids. The direction of fluid flow through a brecciated rock mass is governed by the maximum hydraulic gradient (Sibson 1996). The shear zone was a transport passageway for the rock fragments and matrix as well as for fluids (Branquet et al. 1999). In the thrust zone mechanical crushing and gradual mylonitization take place during tectonic transport. The relatively larger content of dolomite in relation to calcite within the matrix and the porphyroclasts may be a result of the fact that dolomite is less susceptible to dissolution (Kennedy & Logan 1997). Strain within the shear zone involves stretching lineations, lattice preferred orientation of grains and foliation of the matrix (Fig. 7C; cf. Figs. 4E,F,G, 5D). According to Mandal et al. (2001) the flattening of structures may also be a result of the increasing viscosity contrast between the shear zone and the wall rocks.

These processes lead to pressure solution, dynamic recrystallization and grain-size reduction. According to Etheridge & Wilkie (1979) grain-size reduction is expected to progress until recrystallized grains become stable at a size that is in equilibrium with the flow stress.

Further processes of tectonic displacement along the shear zone are easier because of the presence in the nappe foot fluidized rocks of a tectonic lubricant with low viscosity, which is a suspension comprising salt solutions, dissolved gases, rock fragments and matrix. Wohletz & Sheridan (1979) defined fluidization as a process in which "the frictional force between the fluid and the particles counterbalances the weight of the particles and the whole mass behaves as a fluid". Gases can be released from the solution as a result of a temporary drop of pressure or build up of temperature. According to McCallum (1985), gas streaming is considered to be an advanced stage of fluidization. Movements of fluid and gases displaced on the tectonic surface along a hydraulic gradient (Sibson 1996) favour the tectonic transport of clasts. During tectonic transport friction drops (the angle of internal friction decreases to values close to zero). The composite failure envelope on the Coulomb-Mohr diagram is flat and the Mohr circle attains contact with the composite failure envelope even at low absolute values of σ_1 (Fig. 8C). Tectonic transport could take place at even inconsiderable differential stress values (σ_1–σ_3), irrespective of the high or low value of σ_1. This phenomenon is similar to the one taking place at the foot of an advancing glacier — a water film in the foot of the glacier allows horizontal transport for long distances (Piotrowski &

Kraus 1997). In the case of nappes, although the factor responsible for transport is different than in the case of glaciers, the mechanism favouring tectonic transport is similar; instead of a water pillow (cf. Plašienka & Soták 1996) there is a suspension composed of fluids, rock fragments and matrix. Small values of differential stress necessary for displacement may be responsible for the non straight-line character of the nappe tectonic movement, which is reflected in the oblique directions of tectonic transport of the Giewont Unit (from the SSE — Fig. 1D) in relation to the Czerwone Wierchy Unit (from the SSW — see Bac-Moszaszwili et al. 1984).

At the end of this stage pressure solution processes induce the formation of stylolites, which develop parallel to bedding within the clasts (Fig. 5G), at clast boundaries (Fig. 5D,F) and along textural surfaces (Fig. 5E,F). Stylolites in clasts may partially be a result of diagenesis (Smith 2000); they, however, generally originate during the tectonic stage (Newman & Mitra 1994).

Stage D

A re-increase of σ_1 takes place at a simultaneous drop of σ_3 (Fig. 7D), caused among others by the migration and disappearance of fluids, and the anastomosing character of the shear zone that consisted of compressional and dilational domains, for the presence of which were probably responsible faults developed in the foot of the thrust nappe (see: De Roo & Weber 1992). In effect, extension leading to the opening of space between laminae and formation of mineral veins is developed in the shear stress field (Figs. 3, 5F). The veins are filled with freely overgrowing dolomite crystals (calcite is usually subsequent and oblique in relation to earlier structures), which do not show traces of growth simultaneous with folding, as is the case, with for example turbiditic sandstone shale sequences from Australia, from where Jessell et al. (1993) described bedding-parallel laminated veins. A similar case is described by Kennedy & Logan (1997), who noted bedding-parallel calcite veins from mylonites of the McConnell thrust (Alberta), which are relatively undeformed. The process takes place during pure tectonic activity, in which the shear zone is gradually cemented and immobilized, thus causing the termination of the nappe movement. This leads to the closure of pores (although the rock reveals lower porosity than during the phase preceding deformation), and with the build up of stress — to the re-increase of fluid pressure.

Stage E

Temporary release caused by the formation of dislocations in the basement, removal of fluids and mineralization leads to the lithification of the deformation zone and its immobilization. At the beginning, deformation appearing during further stress build up is ductile in character. Meso- and micro folds develop (Figs. 3B–D, 4G). In consequence, pore fluid pressure builds up thus the deformation process becomes brittle. In the following stage the drop of stresses takes place, the Mohr circle is closer to the composite failure envelope and the rock cohesion is ruptured by brittle failure (Fig. 7E). Destruction due to hydraulic fracturing occurs both in the more porous rocks surrounding the mylonitic zone as well as in the mylonites themselves (Figs. 3E, 5E). Thus new parts of the rock adjacent to the fault walls undergo brecciation, and in a further stage — mylonitization. The process, along with pressure solution, which in the rock mass at the grain scale in the gauge is much faster than the pressure solution, for example, along stylolites and associated precipitation in veins (Renard et al. 2000), is responsible for considerable mass loss from the direct vicinity of the thrust. The large role of mass loss was already noticed by Vernon (1998), Ring et al. (2001) and other authors. In the case of the Tatra nappes the selective mass loss may be responsible for their geometric divergence from classic duplexes (Boyer & Elliot 1982).

Stage F

The newly developed shear zone becomes the migration path for fluids moving along the hydraulic gradient. In general, this stage is characterized by similar conditions to stage C (see Figs. 7C and 7F), that is fluids causing friction drop are released into the thrust zone, and nappe transport takes place even at low stress values. Extensional stresses, which are responsible for the formation of elongation structures, develop in a plane parallel to the thrust plane. Pressure solution processes, dynamic recrystallization and grain-size reduction take place. Mylonitization becomes more advanced and locally leads to the formation of ultramylonites. Temporal stress decrease may cause immobilization of the nappe movement and return to the previous stage (Fig. 7E), thus another cycle will begin.

Field and microscopic observations indicate that the co-occurrence of brittle and ductile deformation is a consequence of repeatable brittle and ductile conditions as a result of build up and drop of pore fluid pressure. Teixell et al. (2000) describe the role of the change of fluid pressure in the Larra Thrust, taking place within competent limestones in the Pyrenees. The large role of high fluid pressure and resulting hydraulic fracturing and large-scale pressure solution at the base of the Krížna Nappe in the Tatra Mts (Świerkule Range and Stoły Hill) was already stressed by Jaroszewski (1982). He proposed to determine all such phenomena as “hydrotectonic” (cf. Kopf 1982). Branquet et al. (1994), based on investigations of fluidized hydrothermal breccia in the Colombian Eastern Cordillera, state that brecciation during thrusting can be regarded as a multistage process. The combination of fluidization and hydraulic fracturing suggests that the pulses may be related to successive build up and drop of the fluid pressure. Kennedy & Logan (1997) observe a similar case of brittle failure cyclicity and ductile deformation determined by the change of pore fluid pressure. Gratier et al. (1999) indicated also that brittle and ductile deformation could interact in the upper crust.

Temperature during thrusting

Temperatures stated in earlier papers for Alpine thrust processes in the Tatra Mts vary from 300–350 °C for crystalline rocks (Janák 1994) to 50–80 °C for the upper Krížna Nappe (Grabowski et al. 1999).

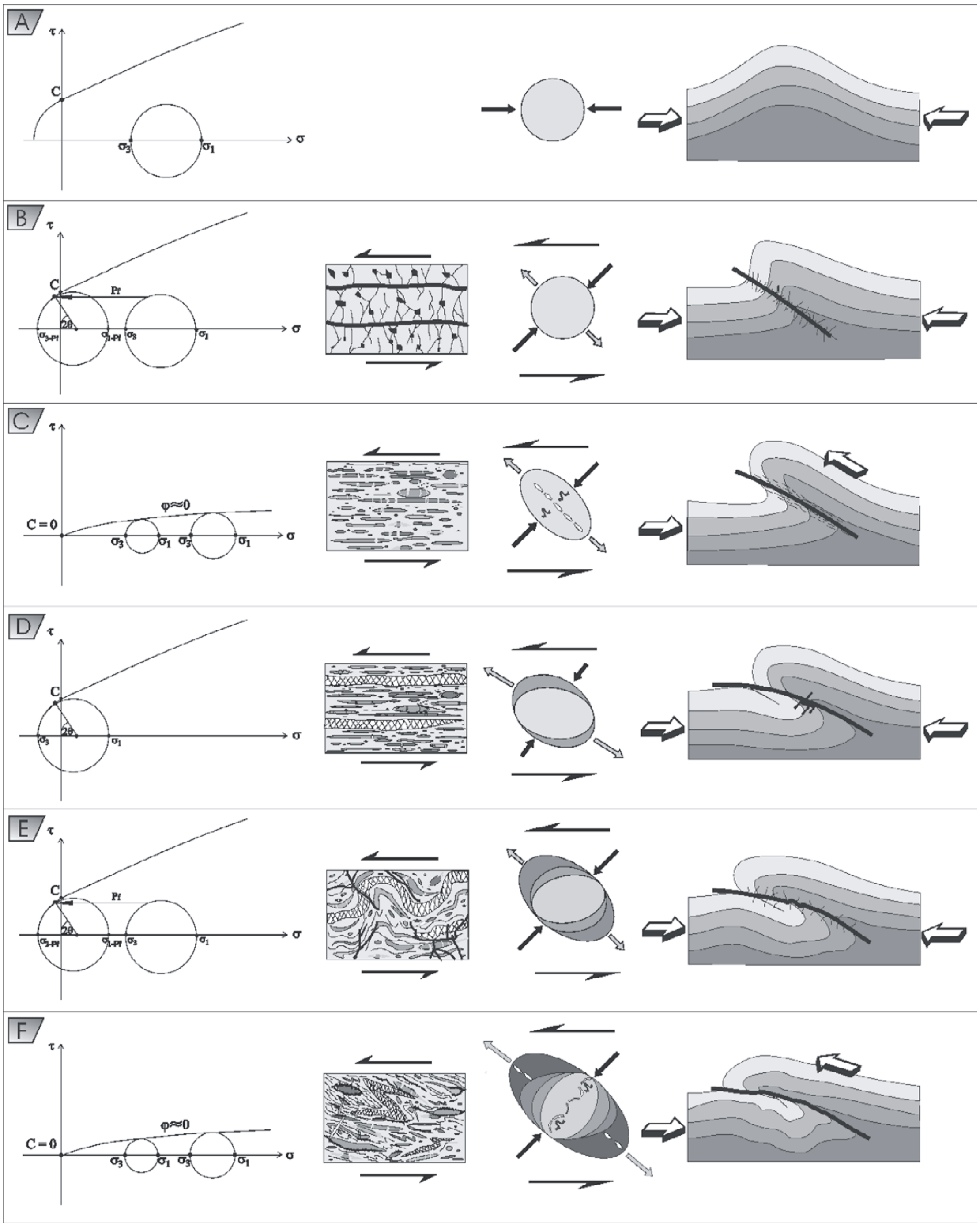

Fig. 7. Stages of tectonic evolution of the thrust between the Giewont and Czerwone Wierchy Units, illustrated (from the left): Coulomb-Mohr diagram (σ — normal stress, τ — shear stress, φ — angle of internal friction, θ — angle of shear, C — cohesive strength, P_f — fluid pressure), scheme of microstructure development within the thrust zone, deformation ellipsoid (note how the structures from the domain of extensional deformation rotated to the position of shortening and produced folding of boudinages and veins) and scheme of thrust development. See text for more explanations.

The problem of temperature may to some degree be solved by deformation observed in carbonate rocks. Experimental data indicate (Barber et al. 1981) that dolomite deforms by slip at temperatures <300 °C and that twinning prevails only at temperatures of 300–600 °C. Recrystallization of dolomite cannot occur at temperatures under 300 °C (Newman & Mitra 1994). According to Burkhard (1993), deformation of twinning in calcite in the form of bending takes place in temperatures over 200 °C. The irregular shape of the twin boundaries in calcite indicates (Vernon 1981) that these boundaries migrated after formation, which occurred at a temperature of about 300 °C.

The values of temperature presented in literature for the processes of cataclasis and mylonitization in carbonate rocks occurring in thrust zones have levels such as 300 °C for fine-grained mylonites of the McConnell thrust in Alberta based on the deformation mechanism map (Kennedy & Logan 1997), 300–350 °C for the Sesia Zone in Western Alps (Küster & Stöckhert 1999), and <300 °C for the Pioneer Landing fault zone in Tennessee based on geothermal gradients (Newman & Mitra 1994).

It can thus be assumed that the temperature within the thrust zone in the multistage process leading to the formation of the Giewont Unit was variable and could periodically attain 200–300 °C.

Conclusions

The so-called "brecciated Campilian" (Late Scythian) from sites located beneath Siadła Turnia and Turnia Olejarnia represents dolomitic mylonites developed at the base of the thrust of the Giewont Unit on the Czerwone Wierchy Unit (High-Tatric units). Its unique preservation in the form of dilatant sites is linked with faults, which originated in the basement of the overthrusting unit. The mylonitization process, which could take place in temperatures reaching 200–300 °C, as well as thrusting of the nappe, was not a one-stage step-like process, but a multistage re-activated process, in which brittle behaviour of deformation was frequently alternated by a ductile rheological condition. Its cyclicity depended on the build up and drop of pore fluid pressure, leading to the drop of stress to an effective value and to rupturing by faulting in the process of hydraulic fracturing (hydrotectonic phenomena). In consequence, the formation of an open pore-system took place, in which the contrast of viscosity between walls of the thrust-fault and the suspension filling the space between them, comprising fluids, rock fragments and matrix, was noted. The presence of an almost friction-less mass in the shear zone that acted as a "water pillow" moving along the pressure gradient, induced easier nappe transport. Stresses released by displacement reappeared when the "fusion" of the thrust limbs and pore closure took place. This caused the build up of fluid pressure, drop of stresses and hydraulic fracturing — mainly in the surrounding rock, more porous than the tectonically changed mylonitic zone, but also in the mylonites themselves, what made the mylonitization process more advanced. In effect, a shear zone revived, displacement and stress release took place, and the whole cycle began once again. This process caused the destruction of still larger and larger parts of the rock and was responsible along with pressure solution for considerable mass loss along the thrust zone. As a result, the Tatra nappes do not bear the characters of typical duplexes, but are their remainders difficult for geometric analysis.

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References

Andrusov D. 1965: Aperçu générale sur la géologie des Carpathes occidentales. *Bull. Soc. Géol. France* 1029–1062.

Bac-Moszaszwili M., Burchart J., Głazek J., Iwanow A., Jaroszewski W., Kotański Z., Lefeld J., Mastella L., Ozimkowski W., Roniewicz P., Skupiński A. & Westfalewicz-Mogilska E. 1979: Geological map of the Polish Tatra Mts, 1:30,000 scale. *Instytut Geologiczny,* Warszawa.

Bac-Moszaszwili M., Gamkerlidze I.P., Jaroszewski W., Schroeder E., Stojanov S. & Tzankov T.V. 1981: Thrust zone of the Križna Nappe at Stoły in Tatra Mts (Poland). *Stud. Geol. Pol.* 68, 61–73.

Bac-Moszaszwili M., Jaroszewski W. & Passendorfer E. 1984: On the tectonics of Czerwone Wierchy and Giewont area in the Tatra Mts., Poland. *Ann. Soc. Geol. Pol.* 52, 67–88 (in Polish, English summary).

Barber D.J., Heard H.C. & Wenk H.R. 1981: Deformation of dolomite single crystals from 20–800 °C. *Physics and Chemistry of Minerals* 7, 271–286.

Boyer S.E. & Elliot D. 1982: Thrust systems. *Amer. Assoc. Petrol. Geol. Bull.* 66, 1196–1230.

Branquet Y., Cheilletz A., Gilliani G., Laumonier B. & Blanco O. 1999: Fluidized hydrothermal breccia in dilatant faults during thrusting: the Colombian emerald deposits. In: McCaffrey K.J.W., Lonergan L. & Wilkinson J.J. (Eds.): Fractures, fluid flow and mineralization. *Geol. Soc. Spec. Publ.* 1–155, 1–321.

Burkhard M. 1993: Calcite-twins, their geometry, appearance and significance as stress-strain markers and indicators of tectonic regime: a review. *J. Struct. Geol.* 15, 351–368.

Burchart J. 1963: Remarks on the directions of the slickesides and fault striae in the crystalline rocks of the Goryczkowa "crystalline island" in the Tatra Mts. *Acta Geol. Pol.* 13, 27–40.

Davis G.H. & Reynolds S.J. 1996: Structural Geology. *John Wiley & Sons, Inc.,* 1–755.

De Roo J.A. & Weber K. 1992: Laminated veins and hydrothermal breccia as markers of low-angle faulting, Rhenish Massif, Germany. *Tectonophysics* 208, 413–430.

Etheridge M.A. & Wilkie J.C. 1979: Grain size reduction, grain boundary sliding and the flow strength of mylonites. *Tectonophysics* 58, 159–178.

Gąsienica-Szostak M. 1973: Geological setting of the northern slope of the Mała Łąka Valley. *M.Sc. thesis.. Arch. Wydz. Geol. Uniw. Warszaw* (in Polish, unpublished).

Grabowski J., Narkiewicz K. & Poprawa P. 1999: First results of paleomagnetic and paleothermal (CAI) investigations of the highest Sub-Tatric units in the Polish Tatra Mts. *Przegl. Geol.* 47, 153–158 (in Polish, English summary).

Gratier J.P., Renard F. & Labaume P. 1999: How pressure solution creep and fracturing process interact in the upper crust to make it behave in both a viscous and brittle manner. *J. Struct. Geol.* 21, 1189–1197.

Hancock P.L. 1985: Brittle microtectonics: principles and practice. *J. Struct. Geol.* 7, 437–457.

Hippert J.F. & Hongn F.D. 1998: Deformation mechanisms in the mylonite/ultramylonite transition. *J. Struct. Geol.* 20, 1435–1448.

Hubert M.K. & Rubey W.W. 1959: Role of fluid pressure in mechanics of overthrust faulting. Part 1. *Geol. Soc. Amer. Bull.* 70, 115–166.

Jessel M.W., Willman C.E. & Gray D.R. 1994: Bedding parallel veins and their relationship to folding. *J. Struct. Geol.* 16, 753–767.

Janák M. 1994: Variscan uplift of the crystalline basement, Tatra Mts, Central West Carpathians: evidence from $^{40}Ar/^{39}Ar$ laser probe dating of biotite and P-T-t paths. *Geol. Carpathica* 45, 239–300.

Jaroszewski W. 1982: Hydrotectonic phenomena at the base of the Križna nappe, Tatra Mts. In: M. Mahel' (Ed.): Alpine structural elements: Carpathian-Balkan-Caucasus-Pamir orogene zone. *Veda*, Bratislava, 137–148.

Jurewicz E. 2000a: Tentative reconstructions of the stress axes from the thrust-folding stage in the Tatra Mts. on the basis of slickensides in the granitoid core, southern Poland. *Przegl. Geol.* 48, 239–246 (in Polish, English summary).

Jurewicz E. 2000b: Tentative correlation of the results of structural analysis in the granitoid core and nappe units of the Tatra Mts., southern Poland. *Przegl. Geol.* 48, 1014–1018 (in Polish, English summary).

Kasiński J. 1981: Cellular dolomites in the High-Tatric Triassic, Polish Tatra Mts. *Przegl. Geol.* 10, 524–529 (in Polish, English summary).

Kennedy L.A. & Logan J.M. 1997: The role of veining and dissolution in the evolution of fine-grained mylonites: the McConnell thrust, Alberta. *J. Struct. Geol.* 19, 785–797.

Knipe R.J. 1989: Deformation mechanisms — recognition from natural tectonites. *J. Struct. Geol.* 11, 127–146.

Kotański Z. 1954: Tentative genetic classification of breccias on the basis of studies concerning the High-Tatric Triassic in the Tatra Mountains. *Rocz. Pol. Tow. Geol.* 24, 63–95 (in Polish, English summary).

Kotański Z. 1956: High-Tatric Campillian in the Tatra Mts. *Acta Geol. Pol.* 6, 65–73 (in Polish, English summary).

Kotański Z. 1959a: Stratigraphical sections of the High-Tatric Series in the Tatra Mts. *Biul. Inst. Geol.* 139, 1–139 (in Polish, English summary).

Kotański Z. 1959b: Contributions to the tectonics of the High-Tatric series. *Biul. Inst. Geol.* 149, 159–174 (in Polish, English summary).

Kotański Z. 1961: Tectogénèse et reconstitution de la paléogéographie de la zone haut-tatrique dans les Tatras. *Acta Geol. Pol.* 11, 187–467 (in Polish, French summary).

Kozłowski A. & Jurewicz E. 2001: Fluid inclusions in slickenside fault mineralisation and quartz veins from the Tatra Mts, Poland. *Mineral. Soc. Pol., Spec. Pap.* 19, 91–93.

Küster M. & Stöckhert B. 1999: High differential stress and sublithostatic pore fluid pressure in the ductile regime — microstructural evidence for short-term post-seismic creep in the Sesia Zone, Western Alps. *Tectonophysics* 303, 263–277.

Lefeld J. 1997: Tectogenesis of the Tatra Mts. The Alpine cycle. In: Lefeld J. & A. Gaździcki (Eds.): 68 meetig of PTGeol., Zakopane 2–4 XI 1997. *Guide to Excurssions,* 16–21 (in Polish).

Lin A. 2001: S–C fabrics developed in cataclastic rocks from Nojima fault zone, Japan, and their implication for tectonic history. *J. Struct. Geol.* 23, 1167–1178.

McCallum M.E. 1985: Experimental evidence for fluidization processes in breccia pipe formation. *Econ. Geol.* 80, 1523–1543.

Mandal N., Chakraborty C. & Samanta S.K. 2001: Flattening in shear zones under constant volume: a theoretical evaluation. *J. Struct. Geol.* 23, 1771–1780.

Michibayashi K. 1993: Syntectonic development of a strain-independent steady-state grain size during mylonitization. *Tectonophysics* 222, 151–164.

Mitra S. 2002: Fold-accommodation faults. *Amer. Assoc. Petrol. Geol. Bull.* 86, 671–693.

Newman J. & Mitra G. 1994: Fluid influenced deformation and recrystallization of dolomite at low temperatures along a natural fault zone, Mountain City window, Tennesee. *Geol. Soc. Amer. Bull.* 106, 1267–1279.

Oelkers E.H., Bjørkum P.A. & Murphy W.M.1996: A petrographic and computational investigation of quartz cementation porosity reduction in North Sea sandstone. *Amer. J. Sci.* 296, 420–452.

Passendorfer E. 1983: How did the Tatra Mts. develop? *Wydawnictwa Geologiczne,* Warszawa, 1–286 (in Polish).

Paulo A. 1997: Remarks on the mineralisation studies in the Tatra Mts. *Przegl. Geol.* 45, 908–909 (in Polish).

Passchier C.W. & Trouw R.A.J. 1998: Microtectonics. *Springer-Verlag*, 1–253.

Piotrowski J.A. & Kraus A.M. 1997: Response of sediment to ice sheet loading in northwestern Germany: effective stress and glacier-bed stability. *J. Glaciol.* 43, 495–502.

Plašienka D. 1991: Mesozoic tectonic evolution of the epi-Variscan continental crust of the Central Western Carpathians — a tentative model. *Miner. Slovaca* 23, 447–457.

Plašienka D. & Soták J. 1996: Rauwackized carbonate tectonic breccias in the West Carpathian nappe edifice: introductory remarks and preliminary results. *Slovak Geol. Magazine* 3–4, 287–291.

Putiš M. 1992: Variscan and Alpidic nappe structures of the Western Carpathian crystalline basement. *Geol. Carpathica* 43, 6, 369–379.

Renard F., Gratier J.P. & Jamtveit B. 2000: Kinetics of crack-sealing, intergranular pressure solution, and compaction around active faults. *J. Struct. Geol.* 22, 1395–1407.

Ring U., Brandon M.T. & Ramthun A. 2001: Solution-mass-transfer deformation adjacent to the Glarus Thrust, with implications for the tectonic evolution of the Alpine wedge in eastern Switzerland. *J. Struct. Geol.* 23, 1491–1505.

Schmid S. 1982: Microfabric studies as indicators of deformation mechanisms and flow laws operative in mountain building. In: Hsü K.J. (Ed.): Mountain building processes. *Academic Press*, London, 95–110.

Sibson R.H. 1977: Fault rocks and fault mechanisms. *J. Geol. Soc. London* 133, 191–213.

Sibson R.H. 1996: Structural permeability of fluid-driven fault-fracture meshes. *J. Struct. Geol.* 18, 1031–1042.

Smith J.V. 2000: Tree-dimensional morphology connectivity of stylolites hyperactived during veining. *J. Struct. Geol.* 22, 59–64.

Spötl C. & Hasenhüttl C. 1998: Thermal history of the evaporatic Haselgebirge melange in the Northern Calcareous Alps (Austria). *Geol. Rdsch.* 87, 449–460.

Treagus S.H. & Treagus J.E. 2002: Studies of strain and rheology of conglomerates. *J. Struct. Geol.* 24, 1541–1567.

Vernon R.H. 1998: Chemical and volume changes during deformation and prograde metamorphism of sediments. In: Treloar P.J. & O'Brien P.J. (Eds.): What drives metamorphism and metamorphic reactions? *Geol. Soc. London, Spec. Publ.* 138, 215–246.

Warren J. 1999: Evaporites. Their evolution and economics. *Blackwell Science,* Tokio, 1–422.

White S.H., Burrows S.E., Carreras J., Shaw N.D. & Humphreys F.J. 1980: On mylonites in ductile shear zones. *J. Struct. Geol.* 2, 175–187.

Wohletz K.H. & Sheridan M. 1979: A model of pyroclastic surge. *Geol. Soc. Amer., Spec. Publ.* 180, 177–194.

Yardley B.W.D. 1991: An introduction to metamorphic petrology. *Longman Singapore Publishers,* 1–248.

Zawidzka K. 1967: On the geology of the region of the Przełęcz Sywarowa Pass in the western Tatras. *Acta Geol. Pol.* 17, 623–645 (in Polish, English summary).