

Multi-stage development of the joint network in the flysch rocks of western Podhale (Inner Western Carpathians, Poland)

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ABSTRACT:

Ludwiniak, M. 2010. Multi-stage development of the joint network in the flysch rocks of western Podhale (Inner Western Carpathians, Poland). *Acta Geologica Polonica*, **60** (2), 283–316. Warszawa.

The geometry and morphology of joints have been examined in the flysch rocks in the western part of the Podhale synclinorium. They form a regular network, which has been developed in several stages connected with the structural evolution of the synclinorium.

The initiation of the oldest diagonal system (D_R , D_L sets), in the form of strength anisotropy in horizontal beds, took place during the successive increase in NNE–SSW compression. The formation of the majority of the L' set and of a small portion of the L set took place during gentle open folding connected with the setting of the general structural framework of the synclinorium. The beginning of the formation of the T set was related to WNW–ESE extension connected with the uplift of the synclinorium. The youngest joints – the majority of the L and a small portion of the L' set, were formed as the result of stress relaxation in the rock masses during progressive uplift lasting up to recent times. The formation of the joints proceeded in two stages: (I) their initiation in the form of joint-anisotropy and (II) opening of joints. These stages have often been significantly separated in time. Sometimes the process of joint opening continues up to recent times.

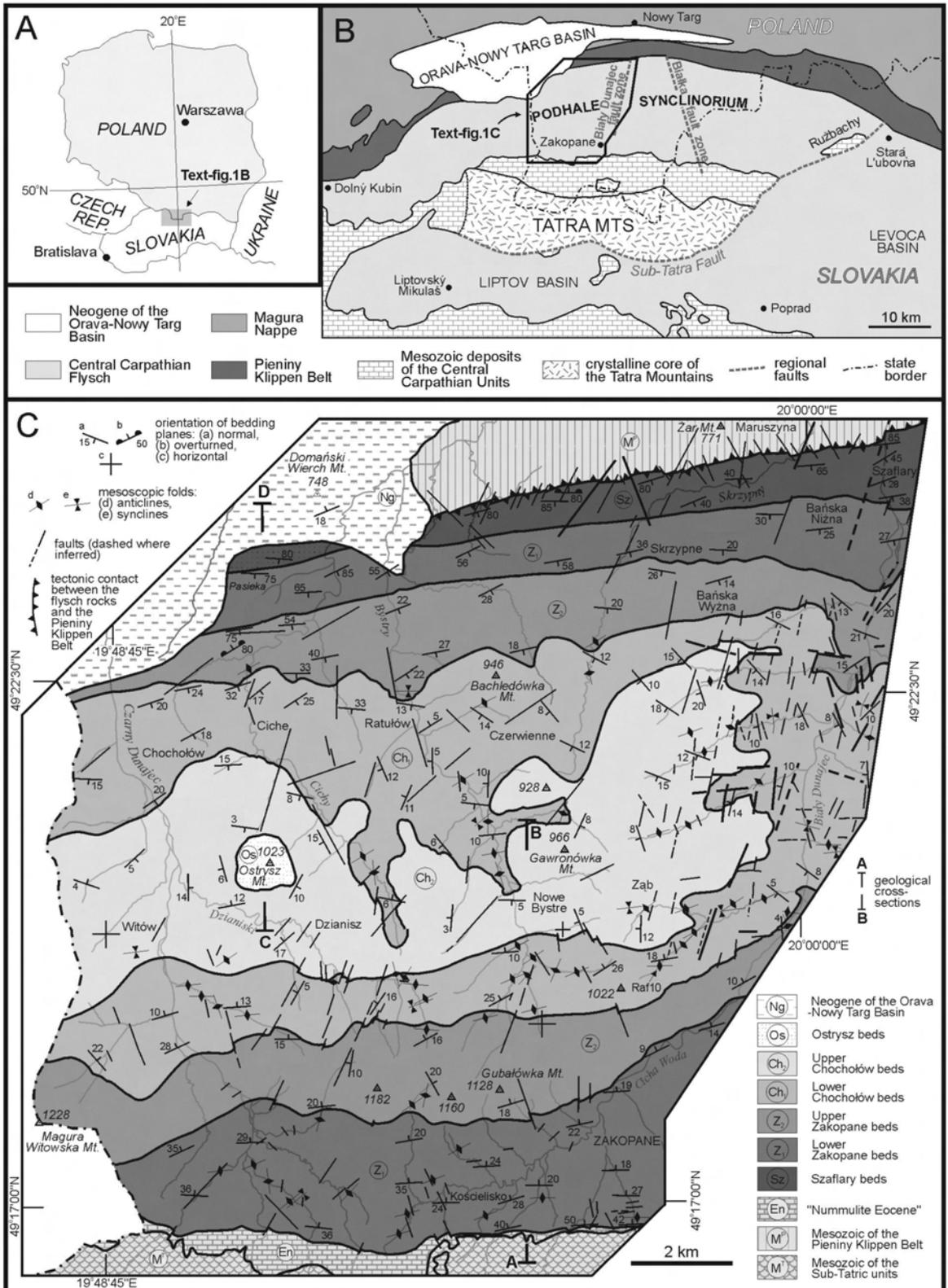
There is a regional tendency that the double shear angle (2Θ) values increase from the axial zone towards the marginal parts of the synclinorium, as well as downward in the flysch lithostratigraphic section. This is probably caused by the increase in confining pressure and rock ductility attributed to the increase in overburden load. The 2Θ values could also have been controlled by tectonic factors.

Key words: Joints; Podhale flysch rocks; Inner Carpathians; Palaeostress reconstruction.

INTRODUCTION

The aim of this work is to determine the origin of the joint network in the Western Podhale region and to reconstruct the palaeostresses responsible for its development. It continues the previous research on jointing in the Polish and Slovak Carpathians initiated by Plička

(1962), Boretti-Onyszkiewicz (1966, 1968a and b), Książkiewicz (1968), and then continued by many authors (e.g., Mastella 1972a and b; Tokarski 1975, 1977; Henkiel and Zuchiewicz 1988; Mastella 1988; Aleksandrowski 1989; Mastella *et al.* 1997; Zuchiewicz 1997a and b, 1998; Rubinkiewicz 1998; Mastella and Zuchiewicz 2000; Mastella and Konon 2002).



Text-fig. 1. A, B – Simplified geological map of the northern part of the Inner Western Carpathians showing location of the study area (after Biely *et al.* 1996; Żytko *et al.* 1989; modified). C – Geological map of the study area (compiled after Watycha 1974, 1976b; Mastella *et al.* 1988, 2000; Mastella and Klimkiewicz 2005; Bac-Moszaszwili *et al.* 1979; and the author's own investigations)

The term “joints” used here refers to systematic, roughly perpendicular to bedding (with the permissible deviation up to $\pm 15^\circ$) and penetrative bed-confined (Mastella 1972a; Ladeira and Price 1981; Gross and Eyal 2007; single-layer – Bahat 1999) fractures cutting beds without macroscopically measurable mode II offset or with only a marked tendency to offset and at spacing approximately proportional to the thickness of the host bed (e.g. Mastella 1972a). This definition of joints corresponds generally to the one used e.g., by Jaroszewski (1972), Hancock (1985), Dadlez and Jaroszewski (1994), Dunne and Hancock (1994) and Mastella and Konon (2002). Fissures and veins developed as a result of transformation of single joints within well-defined joint sets, as well as joint-related arrays of *en echelon* fractures and feather fractures have also been studied.

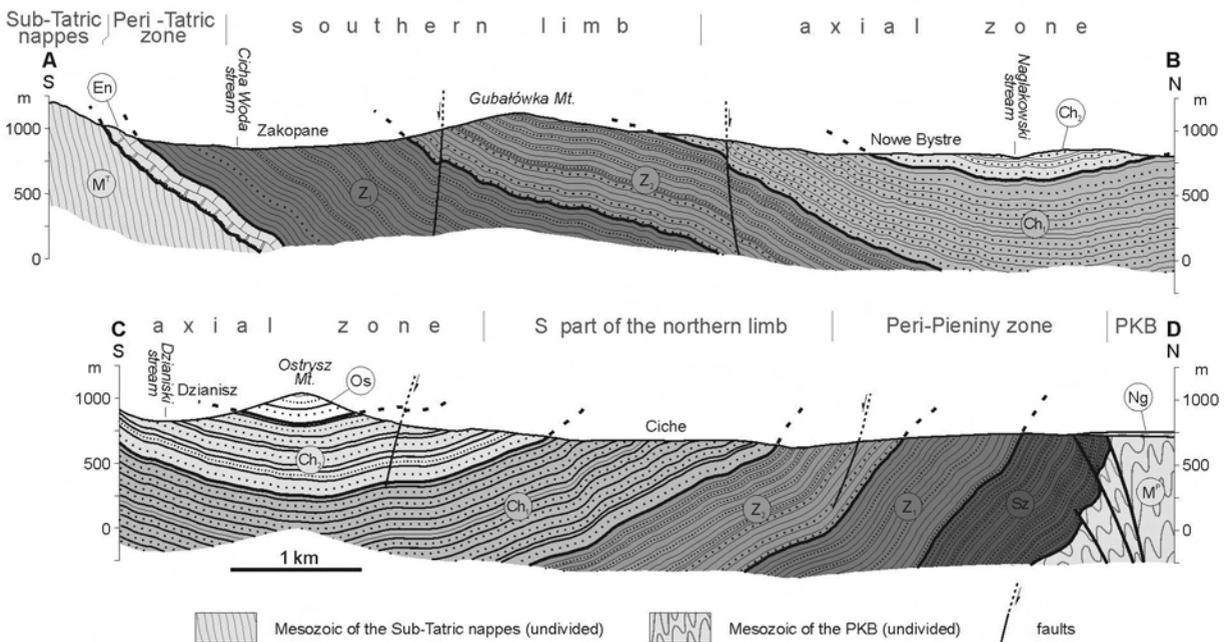
GEOLOGICAL SETTING

The Podhale flysch rocks belong to the Palaeogene cover of the Western Inner Carpathians (Text-fig. 1B; Fusán *et al.* 1967; Marschalko 1968). The age of the flysch deposits has been determined as earliest Oligocene–Late Oligocene (Gedl 2000a and b); however, the age of the highest part has been determined by some authors as Early Miocene (Olszewska and Wiczorek 1998; Garecka 2005). A narrow, discontinuous zone of the so-called “Nummulite Eocene” (Borové Formation *sensu* Gross *et al.* 1984) contacts with the

northern margin of the Sub-Tatric nappes (Text-fig. 1C; Sokołowski 1959). It consists of conglomerates, nummulitic limestones and siltstones of the Middle–Upper Eocene (Bieda 1959; Bartholdy *et al.* 1996), 200 m thick on average (Roniewicz 1969), covering the Tatra tectonic units transgressively (Limanowski 1910) and underlying the flysch series. The thickness of the flysch series is estimated to range from 2.5 km in the eastern to 4.0–4.5 km in the western part of Podhale synclinorium (Watycha 1959, 1976b, 1977; Gołab 1959; Ludwiniak 2006). Estimation based on the geological cross-sections shows that the flysch series is thicker in the northern part of the study area than in the southern part (Text-figs 2 and 3; Ozimkowski 1991).

LITHOSTRATIGRAPHY

The Podhale flysch rocks have been subdivided into several informal lithostratigraphic members (Gołab 1959; Watycha 1959, 1968; Text-fig. 1C and 3). The oldest, sandstone-conglomerate Szaflary beds (Sz; Šambron beds *sensu* Chmelik 1957) occur only in the northern part of the study area and are probably equivalents of the “Nummulite Eocene” from its southern part. The lower member (Z_1) of the overlying Zakopane beds (Huty Formation *sensu* Gross *et al.* 1984) is represented predominantly by thin-bedded claystones and mudstones with minor intercalations of thin-bedded sandstones. The upper member (Z_2) is composed of claystones and mud-



Text-fig. 2. Geological cross-sections through the study area (for other explanations see Text-figs 1C, 3; for location see Text-fig. 1C)

stones with more numerous intercalations of thin- to medium-bedded sandstones. Local intercalations of ferruginous dolomites occur within both members (Kosiorek-Jaczynowska 1959). The lower member (Ch_1) of the succeeding Chochołów beds (Zuberec Formation *sensu* Gross *et al.* 1984) consists of complexes of medium- and thin-bedded sandstones, mudstones and claystones intercalated with numerous thick-bedded sandstones. The upper member (Ch_2) is composed of complexes of sandstones and claystones intercalated by a few thick sandstone beds. Thin tuffite intercalations also occur in the Chochołów beds (Michalik and Wieser 1959; Roniewicz and Westwalewicz 1974). Thin- to thick-bedded sandstones, mudstones and claystones of the Ostrysz beds (Os; Biely Potok Formation *sensu* Gross *et al.* 1984) lie above the Chochołów beds in the western part of the study area, at the top part of the Ostrysz Mt. (Text-figs 1C and 2). A detailed description of the lithology of the particular subdivisions was presented by Watycha (1974, 1976a and b, 1977). The Neogene deposits in the north-western part of the area overlie the Podhale Flysch discordantly (Text-figs 1C, 2).

TECTONICS

The Palaeogene rocks form an asymmetrical synclorium (Halicki 1963; Watycha 1968; Text-fig. 2), within which several parallel tectonic zones are present (Mastella 1975). In the study area, five parallel tectonic zones have been distinguished (Text-fig. 4).

The tectonic contact between the flysch rocks and the Pieniny Klippen Belt (PKB) is a deep-seated, steep fault zone (Uhlig 1897, 1903; Birkenmajer 1958, 1960, 1964, 1965, 1968) with a variable dip (Mastella 1975; Mastella *et al.* 1988). It is composed of steeply dipping faults that are approximately parallel to the strike of the beds. Their southern blocks are usually downfaulted. These faults occur in the most northern part of the contact zone, which is only a few dozen metres wide. Aligned with it is a ca. 1.5–2 km wide zone of tectonic disturbances in which the beds dip relatively steeply. In the northern part of this zone, oblique faults occur in the vicinity of the contact. These are steeply-dipping dextral strike-slip faults (Mastella *et al.* 1988).

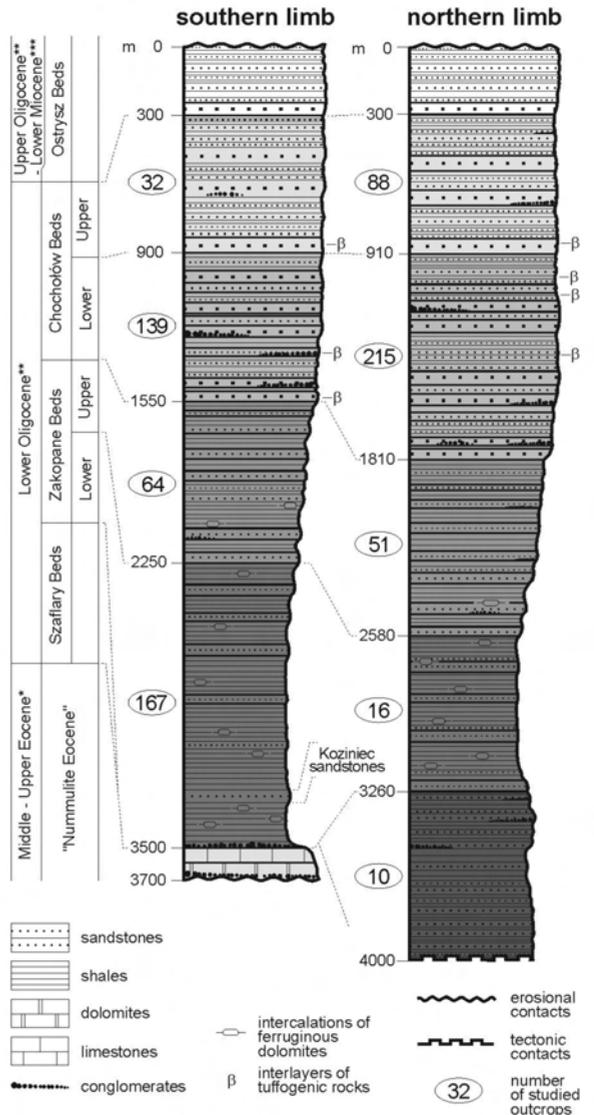
In the eastern Podhale, about 1.5 km south of the contact line, the so-called “Peri-Pieniny flexure” zone is situated (Pokropek 1960; Mastella 1975). It passes westward into a zone of steeply dipping beds. The Pasięka fold (Gołab 1952), situated in the western part of the study area, would be an equivalent of that zone.

Within **the northern limb zone**, the beds dip generally gently southward at ca. 10 to 25°. The ca. 150°

and 30° bedding strike orientations are connected with oblique fault zones situated along the Bystry and Skrzypny streams.

The axial zone is characterised by beds that lie horizontally or dip slightly in different directions, as well as by the common presence of mesofolds with approximately parallel-oriented axes (Halicki 1963; Mastella 1975). The axis of the synclorium plunges westward at several degrees into ca. 258° in the western and ca. 280° in the eastern part of the study area.

The southern limb zone has a monoclinical structure (Mastella and Mizerski 1977; Mastella and Ozim-



Text-fig. 3. Lithostratigraphic logs of the western Podhale flysch (compiled after Watycha 1974, 1976b, Mastella *et al.* 2000, Bac-Moszaszwili *et al.* 1979 and the author's own investigations; age of particular members after Bieda (1959)*, Gedl (2000a)** and Garecka (2005)***)

kowski 1979). The most common bedding strikes are 70–90° and dips change gradually southward from ca. 5°N to ca. 30°N. The flysch rocks within this zone are cut by numerous, mainly NNW–SSE and NE–SW oriented dip-slip faults (Mastella and Mizerski 1977).

A narrow zone of south-verging mesoscopic folds cut by reversed, north-dipping faults is placed in the northern part of the peri-Tatric zone, about 250–1000 m north of the boundary between the flysch rocks and the "Nummulite Eocene". In the southern part of this zone, beds striking approximately W–E lie monoclinally. Their dips reach 40–50°N in the vicinity of the flysch contact with the "Nummulite Eocene" rocks. NNW–SSE mesoscopic strike-slip faults have been also recognized (Mastella and Ozimkowski 1986).

The synclinorium is cut by two large oblique fault zones: the NNW–SSE-trending Białka zone and the NNE–SSW-trending Biały Dunajec zone (Text-fig. 1B; Mastella 1975; Mastella *et al.* 1988, 1996; Klimkiewicz 2008). A fragment of the Biały Dunajec fault zone (BDFZ) is situated in the eastern part of the study area (Text-fig. 4). The BDFZ is composed of numerous, mainly steeply-dipping normal faults, which are approximately parallel to the longer axis of the zone. Their western blocks are downfaulted. However, some of these faults show evidence of an additional small sinistral strike-slip component. NNW–SSE-trending faults also occur within that zone. They feather the main fault zone and make an *en echelon* arrangement. The dominant bedding strikes change gradually from ca. 90° in the northern to ca. 15–20° in the southern part of the BDFZ. The whole BDFZ is rotational in character (Mastella 1975): in the northern part, its eastern block is downfaulted and in the southern part the western block is downfaulted.

MATERIAL AND METHODS

The research reported in this paper was carried out at 782 natural outcrops, differing in dimensions and quality, located in stream and river beds. The orientation of

joints and bedding was measured at each outcrop. 74 outcrops were chosen for detailed analyses, which were made mainly within lithologically similar fine- and medium-grained sandstone beds of varying thickness. In these outcrops the number of joint orientation measurements range from 50 to 300. Following the earlier papers (e.g., Spencer 1959; Jerzykiewicz *et al.* 1974; Zuchiewicz 1998; Mastella and Konon 2002) this data set constitutes a representative sample for joint analysis. The resolution of these measurements was $\pm 2^\circ$. Bed thicknesses were measured at all outcrops wherever possible. Following the suggestions of Jaroszewski (1972), Hancock (1985), Price and Cosgrove (1990), Dunne and Hancock (1994) and Twiss and Moores (2001), cross-cutting relationships and the morphology of joints were observed, as well as structures occurring on their surfaces and joint traces on bedding surfaces. Joint apertures and filling of fissures were also studied. In the 149 selected outcrops joint density was analysed, but these data are not presented in this paper. Among the other tectonic structures, mesofaults and mesoscopic folds were also observed.

In each outcrop, the joints were divided into particular sets during fieldwork. A post-fieldwork procedure of dividing joint orientation data into sets on a statistical basis has not been used. This is justifiable, because, for example, for a small acute angle between two sets the strike dispersion in each set may exceed the half-value of this angle, leading to interference of these sets (Hancock 1994).

Following Price's model* (1959, 1966; Książkiewicz 1968; Jaroszewski 1972; Aleksandrowski 1989), it was assumed that the majority of joint sets have a pre-folding origin at an initial stage of their development. Thus, in order to restore the original orientations of the joints, the strata containing them were back-tilted to the horizontal, following e.g. Murray (1967), Al Kadhi and Hancock (1980) and Kibitlewski (1987). In the case of folds for which the plunge of the axes exceeds 10°, the fold axis was first rotated to the horizontal (Rubinkiewicz 1998). Then, the beds in the limbs were rotated to the hor-

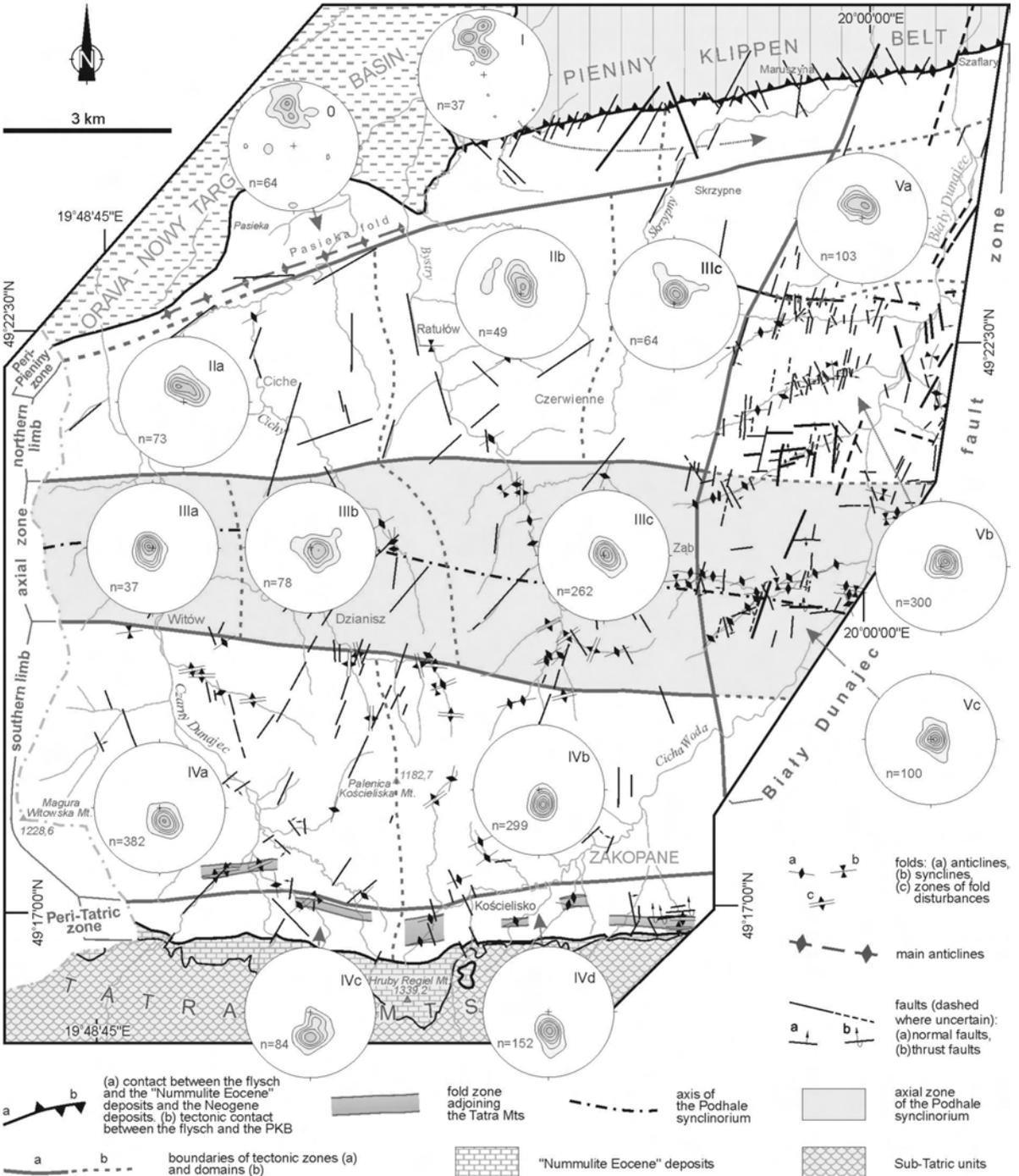
* There is no universal model of joint development in sedimentary rocks (see e.g. Pollard and Aydin 1988). This is because joints have been studied in areas that have different tectonic histories. Thus, rocks occurring in those areas have been deformed in different ways and to different degrees.

Joint formation has been often described as a process taking place during uplift (e.g. Price 1959, 1966; Engelder 1985). On the other hand, it is also interpreted as an effect of increase in pore fluid pressure (Secor 1965). Some authors considered the development of joints in poorly consolidated sediments (e.g. Hodgson 1961; Dehandschutter *et al.* 2005). Moreover, some joints have been described as genetically related to folding (e.g. Stearns and Friedman 1972; Reches 1976; Gutiérrez-Alonso and Gross 1999; Fischer and Wilkerson 2000; Bergbauer and Pollard 2004; Hanks *et al.* 2006; Stephenson *et al.* 2007).

Engelder (1985) distinguished four genetic classes of joints: tectonic, hydraulic, unloading and release. Bahat (1991) proposed the subdivision of joints into: burial, syntectonic, uplifted and post-uplifted.

horizontal about the bedding strike. As a result, orientations of individual joint sets were appreciably unified, which was proven by a fold test conducted on 15 mesofolds (Text-fig. 5). The dominant azimuths of the particular sets were determined on the basis of

contour and rose diagrams. This enabled determination of the directions of the stress axes. The rotations were made with the use of the spreadsheet by Śmigielski (2003), and the data were plotted using the Stereonet™ program.



Text-fig. 4. Tectonic sketch-map of the study area (based on the author's own investigations and after Watycha 1974, 1976b; Mastella *et al.* 1988, 2000; Mastella and Klimkiewicz (2005), and Bac-Moszaszwili *et al.* 1979). Contour diagrams of bedding orientation – lower hemisphere Schmidt net, isolines values are: 3, 7, 11, 15, 29, 31, 37% of all measurements (n)

In order to conduct a regional analysis of the joint network, the study area was divided into 14 smaller sub-areas, according to Jaroszewski's (1972) suggestions (Text-fig. 6). These sub-areas were made by dividing the W–E trending tectonic zones into smaller subzones with a relatively homogeneous the tectonic style.

For recognition of the manner of opening of the joint-related veins, microstructural observations of thin sections were undertaken. The thin sections were made from 13 selected oriented rock samples cut perpendicular to the surfaces of the veins and parallel to the bedding. When describing the mineral fillings of veins, the classifications used in papers by Durney and Ramsay (1973), Passchier and Trouw (2005) and Hilgers and Urai (2005) were applied.

DESCRIPTION OF JOINT GEOMETRY

The results of research conducted in the study area prove that the joint network is reasonably regular throughout Podhale (Halicki 1963; Boretti-Onyszkiwicz 1966, 1968b; Mastella 1972b; Morawski 1972; Mastella and Mizerski 1977; Mastella and Ozimkowski 1979). This regularity is visible both on a map scale (Text-fig. 6) and in individual outcrops (Text-fig. 7A, D). The regional joint network is composed of five sets. However, usually only two to four of these sets occur in a single outcrop. The sets were determined on the basis of their present orientation with respect to the regional extent of the Podhale synclinorium, i.e. in the similar way that was used for the Outer Carpathians by, e.g. Książkiewicz (1968), Aleksandrowski (1989), Mastella *et al.* (1997) and Rubinkiewicz (1998). Following this procedure, two oblique (D_R and D_L), transverse (T), longitudinal (L) and sub-longitudinal (L') sets were distinguished.

D_R and D_L sets

The orientation of the D_R and D_L sets is relatively consistent within particular outcrops; however, it displays a regional variability. The D_R joints strike at 143–173° (dominant class 156–160°) and the D_L set at 13–58° and 36–40° respectively, after bedding correction. The joint planes of both sets are generally perpendicular to the bedding, sometimes with a negligible deviation. Since dips and the sense of dipping of these planes depend on bedding position they change throughout different outcrops. In those outcrops in which both sets co-exist, joints cross at acute angles of 33–77°.

T set

This set is composed of joints which strike at 355–25°, clustering between 16° and 20° (Text-fig. 8B). Their surfaces tend to be mainly vertical or sub-vertical irrespective of bedding orientation. The joints of the T set occur rarely in comparison with those of other sets (they account for less than 5% of all joints) and they do not stand out significantly in the regional picture of the joint network. The sparse occurrence of T joints in the Central Carpathian Flysch has been recorded in the Levočské vrchy Mts, too (Imrich *et al.* 2007).

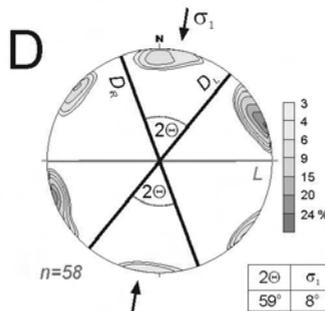
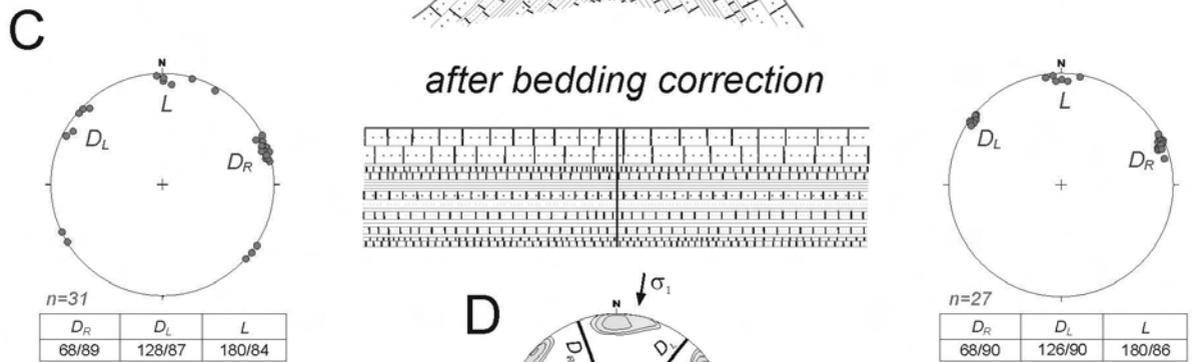
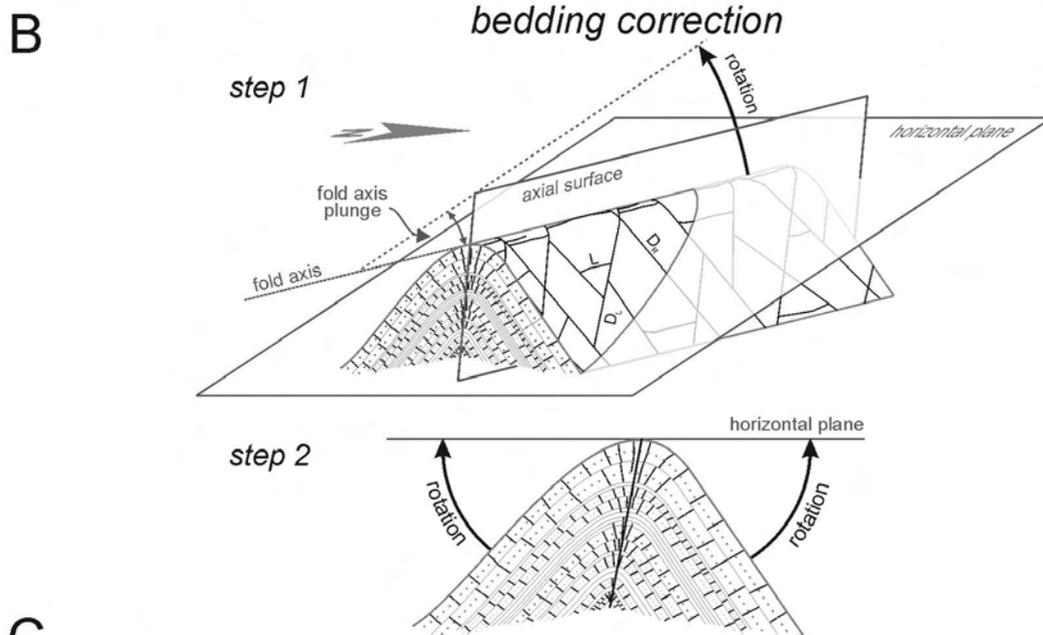
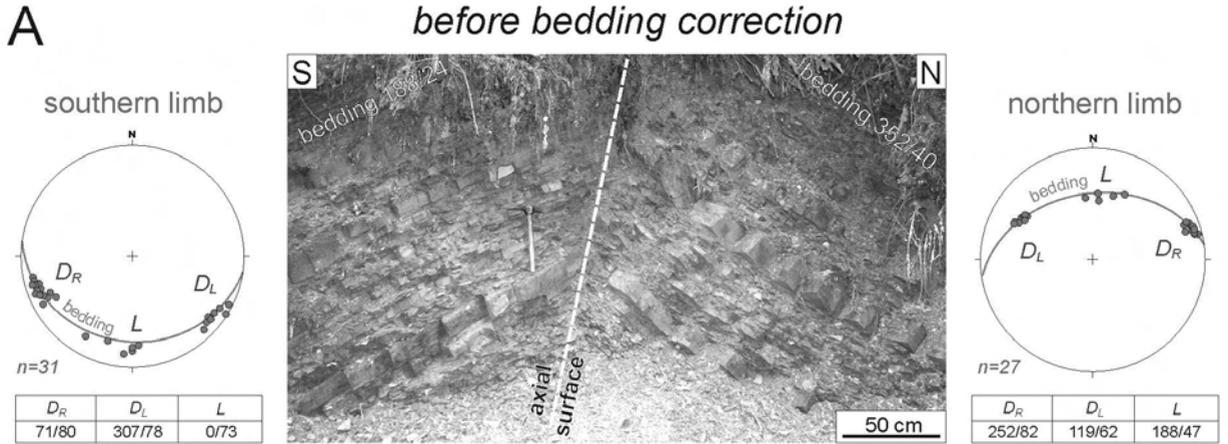
L and L' sets

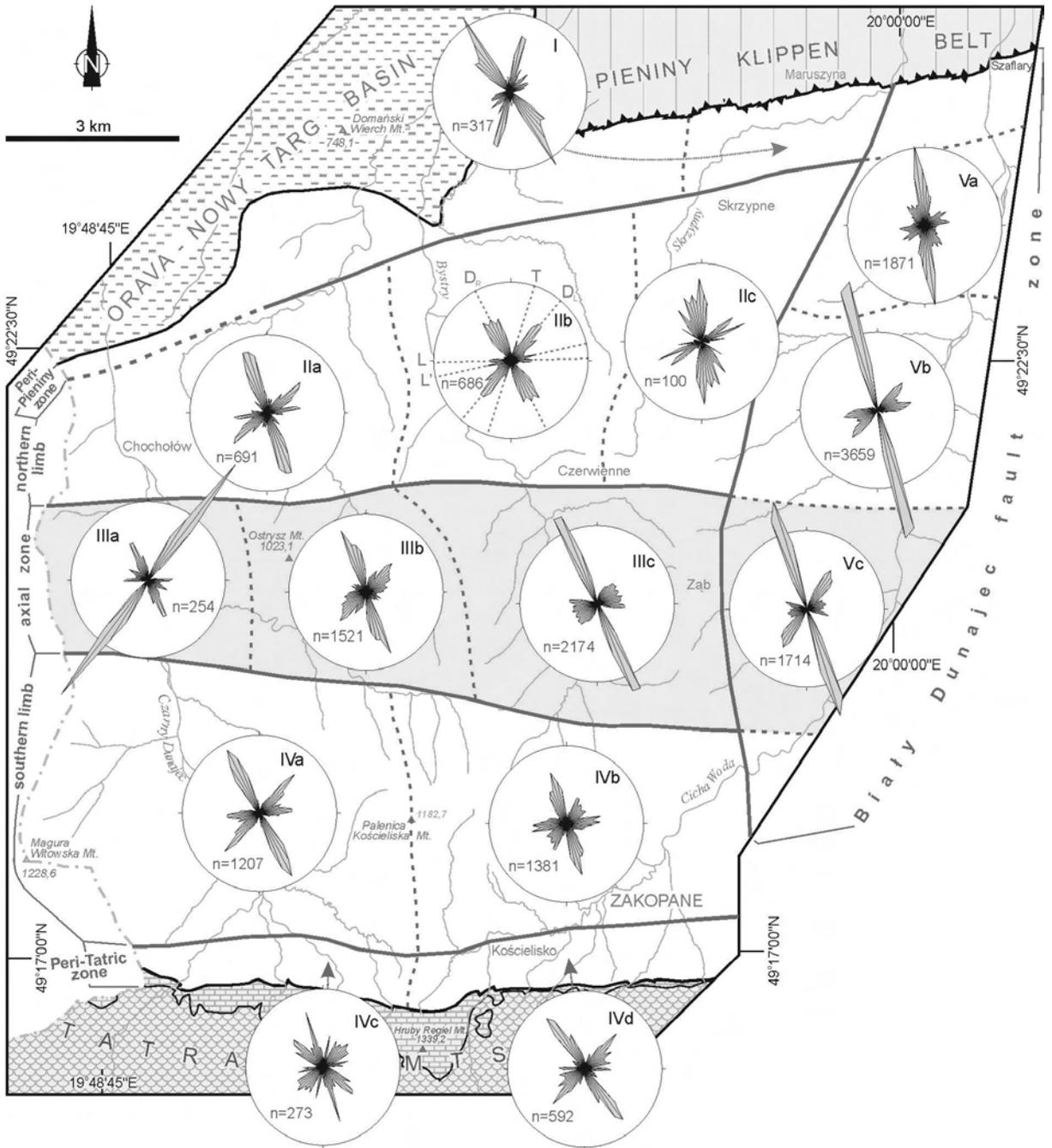
Joints of the L set strike at 65–103° (dominant class 86–90°), and the L' set at 48–78° and 66–70° respectively. In those outcrops in which both sets co-exist, the strikes of the L' set joints deviate on average by 20–25° from those of the L set. The joints of both sets are roughly vertical within horizontal beds. A spectrum of joint dips, from the perpendicular to bedding to the vertical position can be observed in tilted beds. However, their deviation from perpendicularity usually does not exceed 15°.

DESCRIPTION OF JOINT MORPHOLOGY AND CROSS-CUTTING RELATIONSHIPS

The morphologies of the D_R and D_L joints are very similar. Their surfaces are usually flat and smooth without any slickensides. Some joint surfaces are accompanied by plumose structures. In most cases they are S-type ones (Text-fig. 9A). However, occasionally rhythmic C-type plumose structures (Bahat and Engelder 1984) with subtle rib marks can be observed. The axes of the S-type structures are rectilinear and parallel or sub-parallel to the joint-bedding intersection line. Patterns of barbs forming these structures are often oriented in opposite directions even on joint surfaces belonging to the same set within a single outcrop.

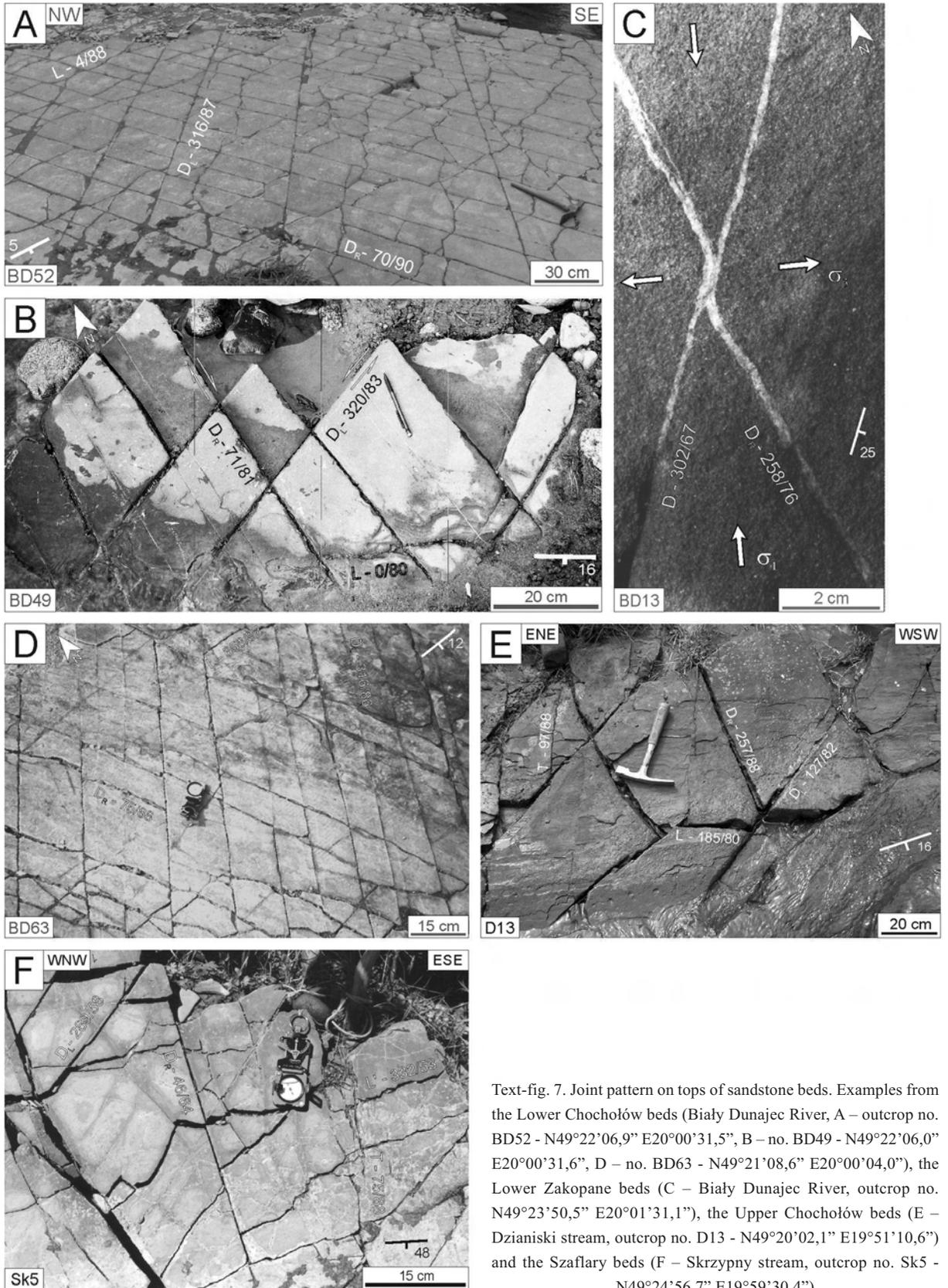
Surfaces of joints belonging to the T, L and L' sets are usually uneven and rougher than those of the D_R and D_L joints within a particular bed (Text-fig. 7A, F). This is consistent with the results of Domonik's (2005) experimental research, which shows a similarity of morphological features between artificially produced fractures and analogous natural joint sets. Both surfaces of a single, non-weathered and non-mineralized joint belonging to the T, L or L' sets are similar to a cast and mould respectively. They are accompanied by plumose structures, but more rarely than in the case of the D_R and D_L joints (Text-fig. 9C).





Text-fig. 6. Map of the spatial distribution of joint orientations. The diagrams present orientations of joints restored to their original position. Radius of the great circle of each rose diagram represents 10% of the total number of measurements (n). Five-degree intervals of azimuths were applied for each diagram. On rose diagram IIb symbols of particular joint sets are given, for example. For other explanations see Text-fig. 4 and text

Text-fig. 5. Method of restoration of original orientation of joints – the example based on measurements in a mesoscopic anticline from the outcrop of the Lower Chochółów beds located on the Rafaczański stream (Raf10 - N49°19'50,3" E19°58'13,7"), A – View of the fold and stereograms of the recent position of joints. B – Schemes showing the technique of bedding correction. C – Scheme showing traces of joints on the vertical wall of the outcrop and stereograms of the joint orientations after back tilting of beds to horizontal. The dominant directions of particular sets are given in the tables. D – Scheme of determining the dominant joint directions and selected parameters of the joint network. n – number of measurements; 2θ – double value of shear angle; σ_1 – maximum normal stress axis. For other explanations see text

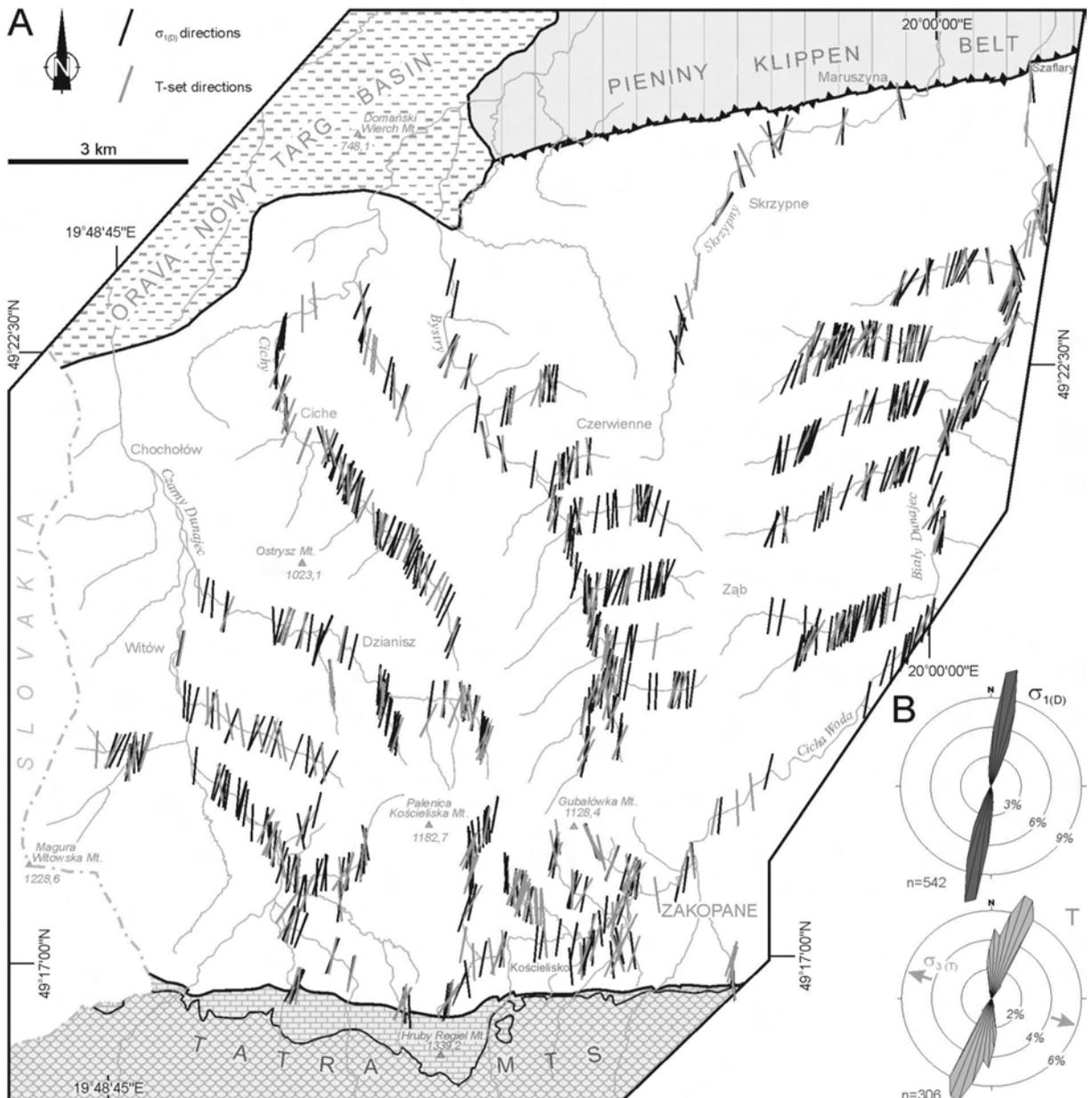


Text-fig. 7. Joint pattern on tops of sandstone beds. Examples from the Lower Chochołów beds (Biały Dunajec River, A – outcrop no. BD52 - N49°22'06,9" E20°00'31,5", B – no. BD49 - N49°22'06,0" E20°00'31,6", D – no. BD63 - N49°21'08,6" E20°00'04,0"), the Lower Zakopane beds (C – Biały Dunajec River, outcrop no. N49°23'50,5" E20°01'31,1"), the Upper Chochołów beds (E – Dziański stream, outcrop no. D13 - N49°20'02,1" E19°51'10,6") and the Szaflary beds (F – Skrzypny stream, outcrop no. Sk5 - N49°24'56,7" E19°59'30,4")

The morphology of joint traces on bedding surfaces varies significantly. Independently of the outcrop size, the D_R and D_L joints often cut the whole exposed fragment of a bed. The apertures of the D_R and D_L joints reach up to ca. 15 mm. Veins filled with calcite related to both sets also occur. Sparse occurrences of joint-related fissures filled with host-rock material have been recognized within the Szaflary beds.

If D_R and D_L joints coexist in one outcrop they either intersect (Text-fig. 7A, B, D) or alternately terminate on (Text-fig. 7D) each other. Sometimes an arc-like transi-

tion between calcite veins related to these sets can be observed (Text-fig. 7C). Traces of the D_R and D_L joints are usually continuous and rectilinear, in contrast to those of the other joint sets (Text-fig. 7A, D). Sometimes they are accompanied by minor feather fractures, and the transition between continuous traces and *en echelon* fracture arrays can also be observed (Text-fig. 10A, D, E, H). The angles between particular feather fractures and a main fracture plane, as well as between particular *en echelon* fractures and the general direction of the array (δ) are 2–27°, with a dominant value of 9°. Rela-



Text-fig. 8. A – Map of spatial distribution of the σ_1 axis direction for the diagonal joints system and T set joint orientations. B – Diagrams of $\sigma_{1(D)}$ and T set orientations. For other explanations see text

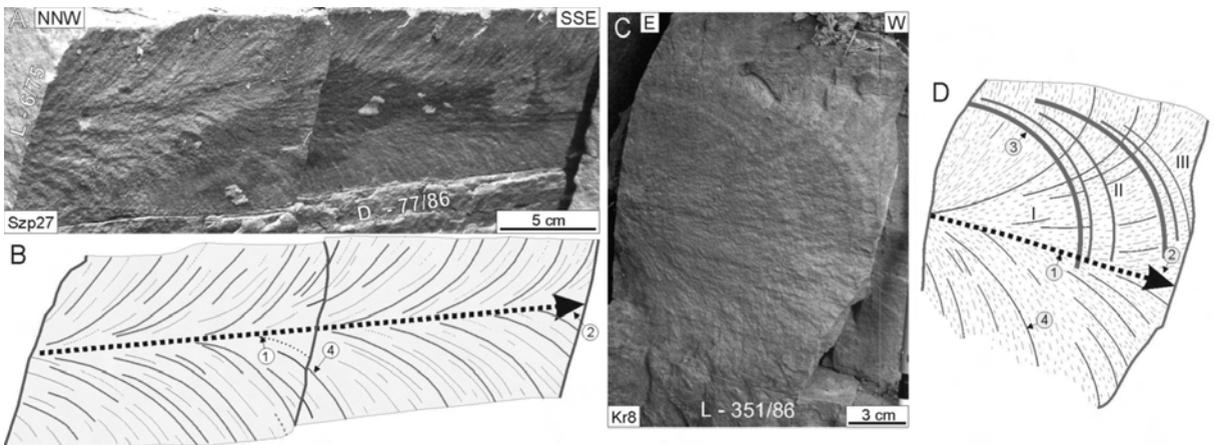
tive overlaps (z_T – Text-fig. 10A) between adjacent *en echelon* fractures within a particular array vary from 0.06 to 0.63, with a dominant value of 0.23. The features of the structures described above indicate that some of them correspond to low-angle Riedel shears (R – terminology after Riedel 1929; Twiss and Moores 2001), while the others form transtensional–extensional *en echelon* arrays (Rothery 1988).

Unlike those of the D_R and D_L sets, the traces of the T, L and L' joints are often curvilinear and sometimes discontinuous or fading. They terminate or change their direction sharply on the D_R and D_L joint traces (Text-fig. 7A, D). However, the veins related to the D_R and D_L sets are often cut by non-mineralized T, L and L' joints (Text-fig. 11C; see also Dunne and Hancock 1994). The traces of the T joints are shorter than those of the D_R and D_L joints. Their apertures reach up to ca. 20 mm. They are predominantly barren joints; veins related to the T set occur less abundantly than in the case of the D_R and D_L sets. The traces of the L and L' joints are considerably shorter than those of the D_R and D_L joints. They are barren joints and their apertures reach up to ca. 35 mm. The traces of the L and L' joints frequently terminate on the

T set joint traces, and in some cases the T set joint traces terminate on the L and L' set joint traces. Sometimes, in the hinges of mesoscopic folds, fold axis-parallel L' joints and, less frequently, L joints occur. The spacing between them is generally smaller in fold hinges than within unfolded beds (Text-fig. 12; see also Ghosh and Mitra 2009). They have typical features of radial fractures (Kettner 1952; Jaroszewski 1980).

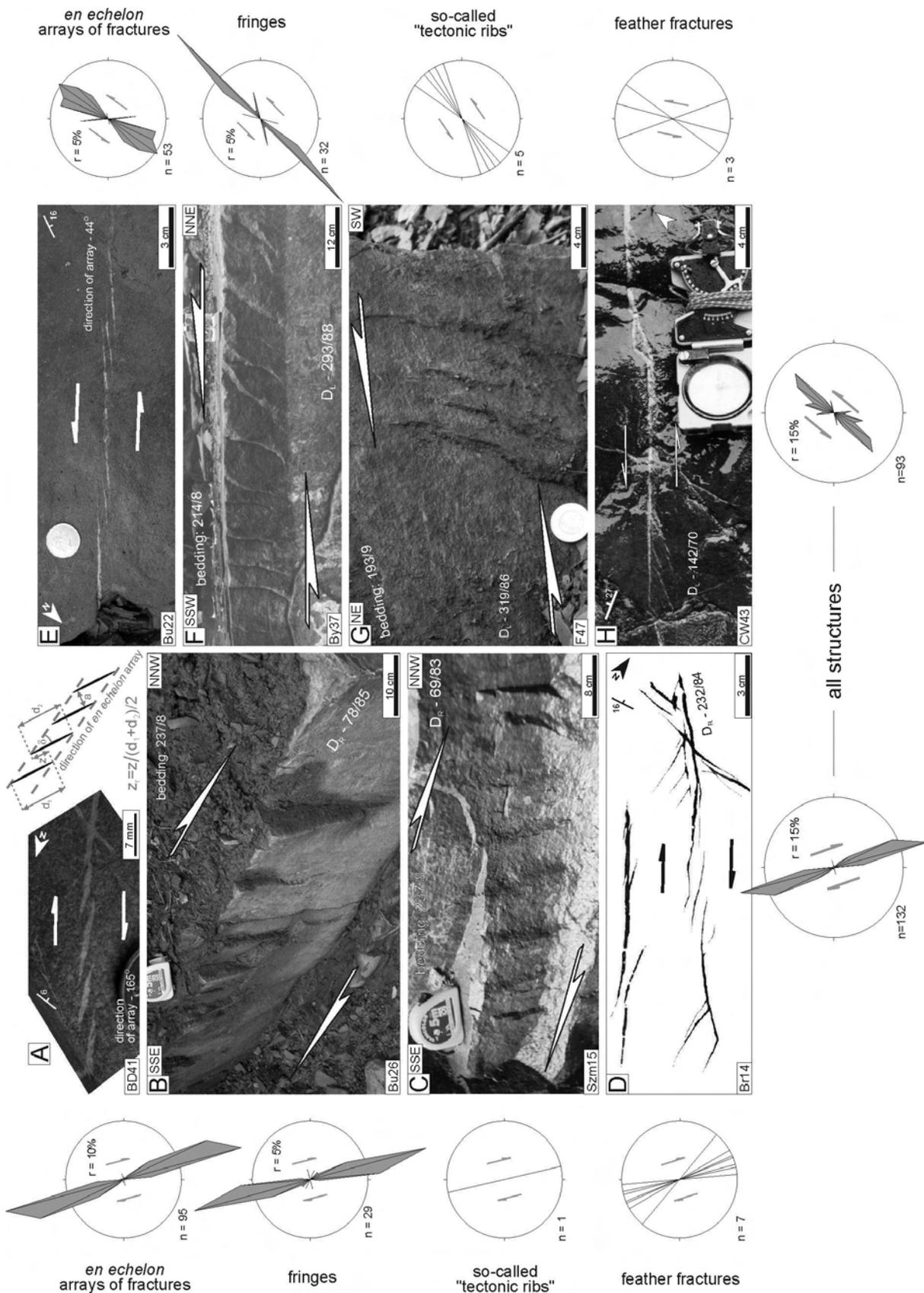
MINERAL VEINS HOSTED IN JOINTS

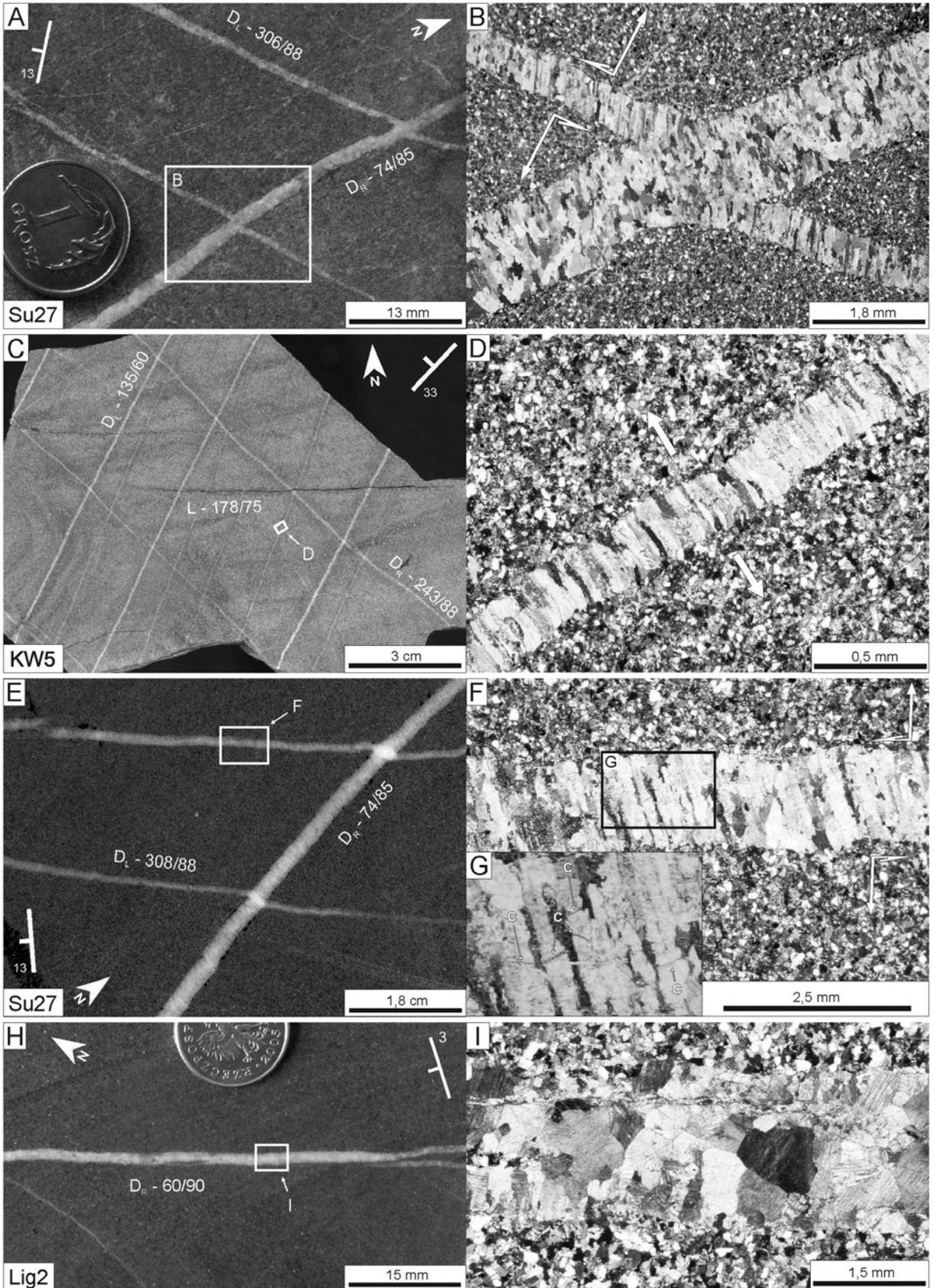
Joint-related calcite veins (JRCV; including veins and mineralized joints *sensu* Rawnsley *et al.* 1998) have been distinguished in the study area. They are hosted in the D_R , D_L and, more rarely, the T joint sets. The fieldwork-based estimation reveals that ca. 35% of the D_R , 20% of the D_L and 4% of the T set joints is exploited by the JRCV. Only a few cases of calcite veins related to L' set joints have been found and they have not been analyzed. The calcite fillings of the JRCV are less resistant to weathering than the host-rocks and consequently they are more easily destroyed in the near-surface part of the rock masses. Thus,

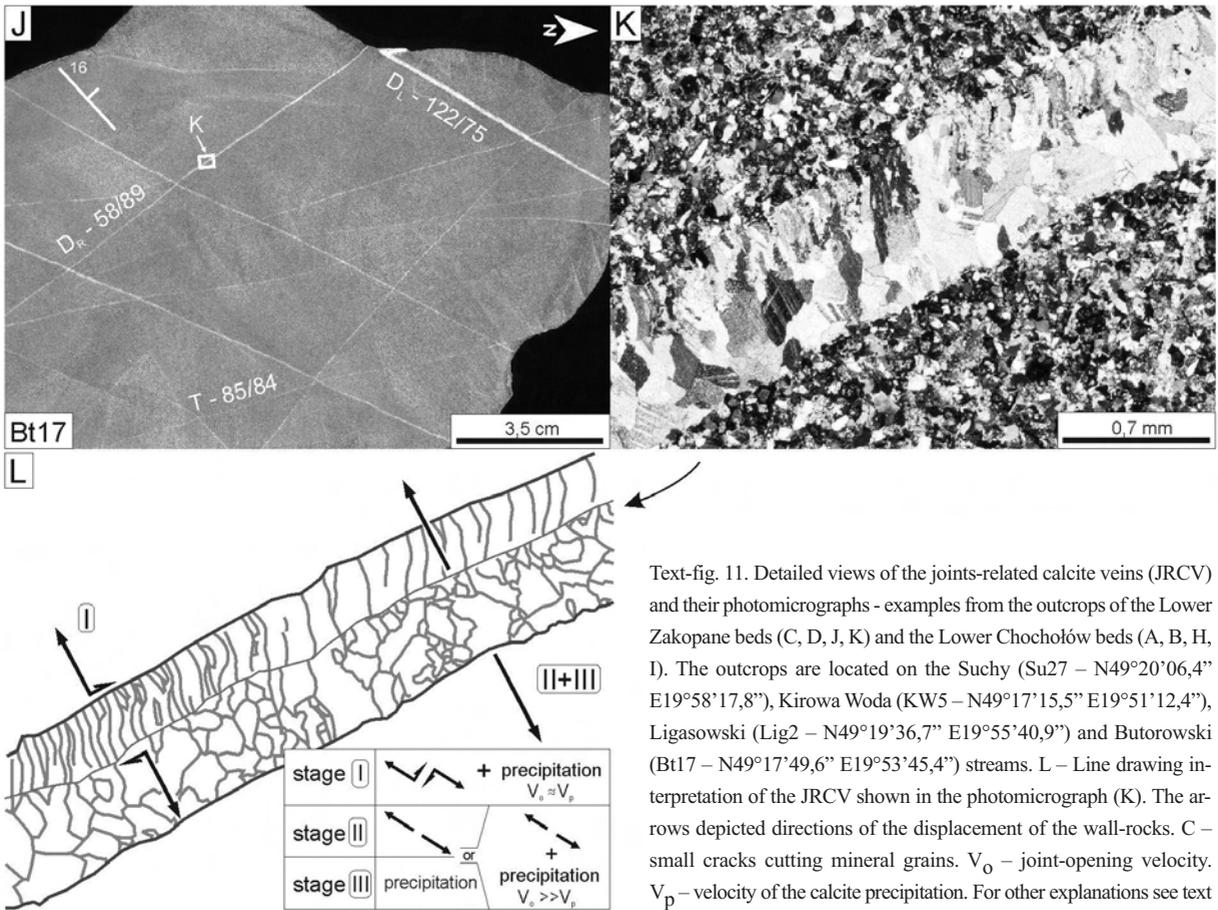


Text-fig. 9. A – S-type plumose structure – example from the Lower Chochółów beds (Szyposzów stream, outcrop no. Szp27 – N49°21'53,8" E19°58'29,9"). B – Line drawing interpretation of the plumose structure shown in photograph (A). C – Rhythmic C-type plumose structure – example from the Upper Zakopane beds (Krajowy stream, outcrop no. Kr8 – N49°23'24,7" E20°00'19,7"). D – Line drawing interpretation of the plumose structure shown in photograph (C) – based on the papers by Bahat (1979) and Bahat and Engelder (1984). 1–median line (axis of plumose structure), 2–direction of fracture propagation, 3–rib marks, 4–hackle-marks, I–III – steps of fracture propagation. For other explanations see text

Text-fig. 10. Detailed view of joint-bearing small-scale brittle structures with diagrams of their orientation throughout the study area, examples from the outcrops of the Lower Zakopane beds (C, D, H) and the Lower (A, B, E) and Upper (F, G) Chochółów beds. The outcrops are located on the Biały Dunajec River (no. BD41 – N49°22'20,6" E20°00'33,2") and on the Za Bramką (no. Br14 – N49°16'58,6" E19°55'42,9"), Cicha Woda (no. CW43 – N49°17'27,7" E19°55'50,1"), Szymoszków (no. Szm15 – N49°17'43,2" E19°56'04,1"), Bustryk (no. Bu22 – N49°21'19,2" E19°58'50,2" and Bu26 – N49°21'14,1" E19°58'35,2"), Bystry (no. By37 – N49°19'43,9" E19°55'35,4") and Florynów (no. F47 – N49°22'39,8" E19°58'47,4") streams. Structures depicted on photographs and line drawings point to the tendency to the dextral (A–D) or sinistral (E–H) shear movement (depicted by the arrows). The diagrams show azimuths of *en echelon* arrays or azimuths of joints surfaces on which the fringes and the tectonic ribs occur. In the case of feather fractures, the diagrams present azimuths of main joint traces which are linked to them. r – radius of the great circle of the rose diagrams which represents 5, 10 or 15% of the total number of measurements (n) for particular structures. For other explanations see text







Text-fig. 11. Detailed views of the joints-related calcite veins (JRCV) and their photomicrographs - examples from the outcrops of the Lower Zakopane beds (C, D, J, K) and the Lower Chochołów beds (A, B, H, I). The outcrops are located on the Suchy (Su27 – N49°20'06,4" E19°58'17,8"), Kirowa Woda (KW5 – N49°17'15,5" E19°51'12,4"), Ligasowski (Lig2 – N49°19'36,7" E19°55'40,9") and Butorowski (Bt17 – N49°17'49,6" E19°53'45,4") streams. L – Line drawing interpretation of the JRCV shown in the photomicrograph (K). The arrows depicted directions of the displacement of the wall-rocks. C – small cracks cutting mineral grains. V_o – joint-opening velocity. V_p – velocity of the calcite precipitation. For other explanations see text

these percentage values are most probably underestimated (see e.g. Mollema and Antonellini 1999).

The JRCV widths range from ca. 0.5 mm up to 5 mm; in sporadic cases they reach 15 mm. The dimensions of the calcite crystals, measured along the longest visible axis, are from several micrometres to 5 mm as seen macroscopically and in thin section samples. In some cases euhedral or subeuhedral crystals can be observed on the exposed JRCV surfaces. Microscopic observations show that the calcite grains have a blocky or columnar shape (Hilgers and Urai 2005; elongated blocky *sensu* Bons 2001; fibrous, with a relatively small grain aspect ratio - Passchier and Trouw 2005; Text-fig. 10B, D, F). The columnar crystals are usually straight and oriented perpendicular to or slightly deviated (up to a maximum 18°) from perpendicularity to the wall-rock. Occasionally, they are slightly curved sigmoidally into perpendicularity with the wall-rock. Wall-rock-parallel medial sutures are sometimes visible at both macro- and microscale (Text-fig. 11D). Blocky grains are approximately isometric, whereas the aspect ratio of the columnar ones ranges up to over a dozen. As a rule, the dimensions of the blocky grains increase with

the JRCV widths. Neither the blocky grains nor the columnar ones are noticeably internally deformed. In some cases small cracks cutting single or several columnar grains have been found (Text-fig. 11G). The majority of the JRCV described correspond to "simple veins" *sensu* Dadlez and Jaroszewski (1994), whereas a few of them correspond to asymmetrical "composite veins" (Durney and Ramsay 1973; Passchier and Trouw 2005; Text-fig. 11K).

Microstructural analysis of the fillings of mineral veins, particularly those composed of elongated grains (fibrous and columnar ones), can be helpful in the reconstruction of the kinematics of deformation in rocks (e.g. Ramsay 1980; Spencer 1991; Urai *et al.* 1991; Köhn 2000; Means and Li 2000; Hilgers and Urai 2005). The shapes of the columnar calcite grains and their attitude with respect to the veins' wall-rocks show that the JRCV were generally opened in a direction normal to their walls (as *tension gashes* – Passchier and Trouw 2005). These joint-related veins, therefore, correspond to mode I fractures. Sometimes, their opening proceeded with a very small component of mode II displacement. However, it is also possible

that they would have been opened askew to the wall-rocks.

The relatively high aspect ratio of the columnar grains indicates that the opening rate was close to or even slightly slower than the growth rate of crystals (Köhn 2000). The blocky microstructures, in turn, are highly ambiguous kinematic indicators. Their occurrence can point to calcite precipitation after the opening of a fracture. However, it is also possible that the fracture was being filled during opening. In this case the opening rate is faster than the maximum growth rate of calcite crystals (Köhn 2000).

INTERPRETATION OF THE JOINTS

Diagonal joint system (D_R and D_L sets)

The morphological and geometric features described above show that D_R and D_L sets are coeval and form a diagonal conjugate system. This is indicated by the fact that traces of these joint intersect one another and terminate on one another alternately (Stubbs and Wheeler 1975; Angelier 1984; Price and Cosgrove 1990). The arc-like transitions between the veins related to the D_R and D_L sets point to their synchronous opening, earlier than in the case of the other sets (compare also Jaroszewski's 1972 and Aleksandrowski's 1989 opinions).

It follows from theoretical considerations (Mandl 1988) that the D_R and D_L joints were formed during the initial stage of the diagonal system development as "potential shear surfaces" in a triaxial, compressive stress field ($\sigma_1 > \sigma_2 > \sigma_3$) (in the form of initial microcracks - Brace and Bombolakis 1963; Lajtai 1971, occurring in the zones extending along a potential shear surface - Scholz 1968). The occurrence of microcracks in rocks acts as a "release mechanism" under loading conditions



Text-fig. 12. Radial fractures in the anticline hinge - example from the Upper Zakopane beds (Biały Dunajec River, outcrop no. BD28 - N49°23'04,4" E20°01'05,1"). For other explanations see text

(Pinińska 1995) - their growth and coalescence leads to the opening of joints (Segall and Pollard 1983). The "shear" nature of the initiation of diagonal joints is indicated by smooth joint surfaces (see e.g. Mastella and Zuchiewicz 2000; see also the results of experimental studies by Ramsey and Chester 2004). Moreover, the geometry of joint-related small brittle structures points to a tendency to strike-slip movement along these joints. The arrangement of *en echelon* and feather fractures indicates that these sets would have corresponded to dextral (D_R) and sinistral (D_L) shear fractures (Hancock 1985). It should be mentioned, however, that some of the *en echelon* fractures are probably connected with the extensional opening of the joints (Pollard *et al.* 1982; Olson and Pollard 1991). The small values of the δ -angle in some *en echelon* arrays indicate a rather small contribution of the shear component in the development of the joints. Moreover, macroscopically visible mode II offsets along joint surfaces have not been distinguished. Thus, the D_R and D_L joints are not shear fractures. Hybrid joints (Hancock and Al Kadhi 1982) are only a small fraction of joints belonging to both sets.

The necessary condition for the formation of joints in sedimentary rocks is the lithification of deposits to a degree that makes the accumulation of stresses possible. However, the results of previous fieldwork studies (e.g. Beach and Jack 1982; Mastella 1988; Świerczewska and Tokarski 1998; Dehandschutter *et al.* 2005) and laboratory research (Maltman 1988) show that the formation of some joints is possible in poorly lithified sediments. The lack of wall-rock grains within the JRCV in the study area allows the inference that they opened while the sandstone beds were well lithified. Only a few cases of joints filled with host-rock material indicate their formation prior to lithification. As verified by the fold-test (Text-fig. 5), the diagonal system developed in horizontal beds. This is also proved by the more unified orientation of D_R and D_L joints at the map-scale after back-tilting of the bedding to a horizontal position. The high degree of perpendicularity to bedding (with deviation up to + 5°) within the tilted beds suggests a prefolding origin of the D_R and D_L sets (Jaroszewski 1972; Mastella *et al.* 1997; Mastella and Konon 2002; Bergbauer and Pollard 2004; Rubinkiewicz and Ludwiniak 2005; Bellahsen *et al.* 2006; Whitaker and Engelder 2006).

In the subsequent stage of its development, at the residual stress, the diagonal joint system was opened as mode I fractures (Atkinson 1987) (Price 1959, 1966; Billings 1972). The plumose structures occurring on the joints indicate that they opened perpendicular to their surfaces (Bankwitz 1965, 1966) and the directions of joint propagation were approximately parallel to the

bedding. The arrangement of barbs shows that the sense of propagation was random (DeGraff and Aydin 1987). Analysis of the JRCV filling connected with the D_R and D_L joints suggests that they opened in the same way (Text-fig. 11D), sometimes with only a negligible strike-slip movement component (0.1–0.4mm; Text-fig. 11B,F,K) or askew to their wall-rocks.

There are different populations of D_R and D_L joints and their ratio changes throughout the different outcrops analyzed. In some of them, both sets exist in relative quantitative equilibrium, but often one set prevails over the other (see e.g. Aleksandrowski 1989). However, in the whole area studied, there are more D_R joints than D_L joints (Text-fig. 6, 13D). Similar quantitative differences between two conjugate joint sets have also been found in the Outer Carpathians (Książkiewicz 1968) and in other regions by e.g. Parker (1942), Wilson (1952), Hancock (1964) and Hanks *et al.* (1997). A possible explanation involves a slightly faster formation of one of the conjugate fractures, so that in some cases the other one does not form at all. A similar phenomenon has been found in laboratory geomechanical studies (e.g. Hobbs 1960; Jaegger 1960; Donath 1961; Scholz 1968; Paterson 1958, 1978).

The cross-cutting relationships observed on the bedding surfaces show that the opening of a considerable portion of the D_R and D_L joints occurred earlier than in the other joint sets. Generally, younger T, L and L' joints do not cross older D_R and D_L joints (except for veins related to the D_R and D_L sets). The older, preexisting diagonal sets influenced the propagation of younger sets, which is shown by their curved traces close to the D_R and D_L joints. This is because the local stress field in the immediate vicinity of an older joint influences the local stress field of the propagating joint (Dyer 1988; Rogers *et al.* 2004). Moreover, the occurrence of the JRCV connected mainly with the D_R and D_L sets (and to a smaller degree with the T set) indicates a general tendency to open earlier than the other sets. The synkinematic character of part of their fillings suggests that they were opened as tension gashes under fluid pressure while the fracturing beds could have been covered by overburden up to several kilometres thick (Secor 1965).

T set

The surface morphology of the T joints and the lack of features indicating a tendency to shear show that they were developed as extensional fractures. T joints were formed in a triaxial stress field ($\sigma_1 > \sigma_2 > \sigma_3$), where the σ_3 axis was horizontal and perpendicular to the joint surface (the σ_1 and σ_2 axes were perpendicular to the σ_3 axis and parallel to the joint surface, but their unambiguous ori-

entation cannot be deduced – see Mastella *et al.* 1997). Taking into account their orientation approximately perpendicular to the regional extent of the Podhale synclinorium and Price's theoretical model (1959, 1966), the T set can be interpreted, in analogy with the Outer Carpathians (e.g. Aleksandrowski 1989; Zuchiewicz 1998; Mastella and Konon 2002), as being formed under progressively diminishing, horizontal NNE–SSW compression during the uplift of the Podhale flysch rocks, resulting in their extension parallel to the axis of the synclinorium. Within the Skrzypne stream fault zone (Mastella *et al.* 1988; sub-area IIc – Text-fig. 6), a more numerous population of the T set has been observed. This is probably due to fault-related stress concentration within this zone. The cross-cutting relationships and the fact that they are sometimes more deviated from perpendicularity to tilted bedding planes, suggest that the T set joints are younger than the D_R and D_L joints.

L and L' sets

The morphology of the L and L' joints, together with the absence of visible signs of shear movement, show that both sets were developed in an extensional mode. They were formed in a triaxial stress field ($\sigma_1 > \sigma_2 > \sigma_3$), where the σ_3 axis was horizontal and perpendicular to the joint surfaces. The σ_1 and σ_2 axes were perpendicular to the σ_3 axis and parallel to the joint surfaces. Considering the cross-cutting relationships, the absence of the JRCV and, in some cases, deviation of L and L' joints from perpendicularity to tilted beds it has been found that the L and L' sets are younger than the joints of the diagonal system. The small number of T joints in particular outcrops makes determination of the age relationships between the L, L' and T sets difficult. In most cases where T and L sets co-exist, the T joints appear to be older than the L joints, and also older than a few joints of the L' set. Only locally do T and L joints form an orthogonal pattern (referred to as *fracture grid-lock* – Hancock *et al.* 1987). In these cases they appear to be conjugate joints. As follows from observations of cross-cutting relationships, the majority of the L' and a small portion of the L-joints are older than the T-set (similarly as found by Aleksandrowski 1989). These joints are genetically connected with the development of folds. The radial fractures described earlier fall into this group. The latter ones can be classified as tensional fractures.

A similar succession of joint development to the one described above was found in some areas of the Polish Outer Carpathians (e.g. Książkiewicz 1968; Rubinkiewicz 1998; Mastella and Zuchiewicz 2000).

outcrop A										outcrop B									
outcrop	bedding orientation (dip dir'n/dip)	lithology	coordinates	bed thickness - M_A (cm)	$2\Theta_A$ - angle ($^\circ$)	outcrop	bedding orientation (dip dir'n/dip)	lithology	coordinates	bed thickness - M_B (cm)	$2\Theta_B$ - angle ($^\circ$)	distance between outcrops (m)	bed thickness proportion M_A/M_B (-)	2Θ -angle difference $\Delta 2\Theta = 2\Theta_A - 2\Theta_B$ ($^\circ$)	$\Delta 2\Theta/M_A/M_B$ ($^\circ/1$)	increase (\uparrow) / decrease (\downarrow) of 2Θ -angle with the increase of bed thickness			
BD7	157/20	fine-grained sandstone	N49°24'12.9" E20°01'37.5"	80	66	BD6	158/25	fine-grained sandstone	N49°24'13.7" E20°01'37.6"	10	60	26	8.00	6	0.75	\uparrow			
BD22	135/16	medium-grained sandstone	N49°23'24.0" E20°01'11.2"	55	48	BD21	150/10	fine-grained sandstone	N49°23'26.3" E20°01'09.4"	14	46	79	3.93	2	0.51	\uparrow			
BD26	172/16	fine-grained sandstone	N49°23'06.4" E20°01'01.9"	47	61	BD27	155/8	fine-grained sandstone	N49°23'05.5" E20°01'04.5"	7	65	59	6.71	-4	-0.60	\uparrow			
BD34	179/11	medium-grained sandstone	N49°22'32.7" E20°00'49.2"	65	67	BD33	188/12	fine-grained sandstone	N49°22'35.2" E20°00'49.3"	20	56	79	3.25	11	3.38	\uparrow			
BD38	211/11	fine-grained sandstone (medium-grained at the bottom)	N49°22'21.6" E20°00'35.4"	26	52	BD37	198/10	fine-grained sandstone	N49°22'22.8" E20°00'35.7"	10	57	35	2.60	-5	-1.92	\downarrow			
BD41	198/6	fine-grained sandstone	N49°22'20.6" E20°00'33.2"	56	52	BD40	197/9	fine-grained sandstone	N49°22'20.6" E20°00'34.7"	17	51	84	3.29	1	0.30	\uparrow			
BD50	205/12	fine-grained sandstone	N49°22'04.8" E20°00'28.2"	100	49	BD49	220/10	fine-grained sandstone	N49°22'06.0" E20°00'31.6"	20	72	76	5.00	-23	-4.60	\downarrow			
BD55	220/10	medium-grained sandstone	N49°21'53.9" E20°00'14.8"	70	57	BD56	225/8	fine-grained sandstone	N49°21'52.3" E20°00'14.9"	20	71	51	3.50	-14	-4.00	\downarrow			
BD58	215/8	fine-grained sandstone	N49°21'48.6" E20°00'15.7"	30	49	BD57	200/9	fine-grained sandstone	N49°21'50.5" E20°00'15.4"	5	60	58	6.00	-11	-1.83	\downarrow			
BD77	180/2	fine-grained sandstone	N49°19'51.5" E19°59'42.1"	30	65	BD76	230/2	fine-grained sandstone	N49°19'53.7" E19°59'45.4"	5	65	96	6.00	0	0.00	0			
BD78	190/2	medium-grained sandstone (coarse-grained at the bottom)	N49°19'50.2" E19°59'39.8"	100	61	BD77	180/2	fine-grained sandstone	N49°19'51.5" E19°59'42.1"	30	65	60	3.33	-4	-1.20	\downarrow			
F9	250/12	fine-grained sandstone (medium-grained at the bottom)	N49°22'42.0" E19°59'43.7"	72	71	F8	257/12	fine-grained sandstone	N49°22'41.9" E19°59'46.5"	27	62	56	2.67	9	3.38	\uparrow			
F21	240/20	medium-grained sandstone	N49°22'45.1" E19°59'18.1"	55	65	F20	235/16	medium-grained sandstone	N49°22'43.6" E19°59'11.1"	30	73	21	1.83	-8	-4.36	\downarrow			
Sap35	213/14	fine-grained sandstone (medium-grained at the bottom)	N49°21'49.3" E19°58'21.2"	63	65	Sap36	220/9	fine-grained sandstone	N49°21'48.0" E19°58'18.6"	25	57	65	2.52	8	3.17	\uparrow			
Bu14	300/10	medium-grained sandstone	N49°21'27.0" E19°59'10.3"	150	64	Bu15	288/11	medium-grained sandstone	N49°21'26.4" E19°59'09.1"	56	73	30	2.68	-9	-3.36	\downarrow			
Su8	256/6	medium-grained sandstone (coarse-grained at the bottom)	N49°20'12.3" E19°59'12.6"	80	53	Su7	243/10	fine-grained sandstone	N49°20'12.7" E19°59'14.8"	20	54	46	4.00	-1	-0.25	\downarrow			
B8	335/20	fine-grained sandstone	N49°18'05.8" E19°53'48.2"	22	66	B7	323/20	fine-grained sandstone	N49°18'08.4" E19°53'52.3"	9	59	114	2.44	7	2.86	\uparrow			
B19	335/18	fine-grained sandstone	N49°17'47.7" E19°53'45.1"	25	65	B17	337/16	fine-grained sandstone	N49°17'49.6" E19°53'45.4"	10	64	62	2.50	1	0.40	\uparrow			
B13	331/14	fine-grained sandstone (medium-grained at the bottom)	N49°17'47.3" E19°54'12.1"	40	60	B12	0/17	fine-grained sandstone	N49°17'48.0" E19°54'11.8"	11	60	22	3.64	0	0.00	0			
Dz11	343/18	medium-grained sandstone	N49°17'41.3" E19°55'03.7"	62	45	Dz10	345/12	medium-grained sandstone	N49°17'42.8" E19°55'02.0"	31	62	57	2.00	-17	-8.50	\downarrow			
Szm13	335/27	fine-grained sandstone (medium-grained at the bottom)	N49°17'42.4" E19°56'03.9"	50	72	Szm14	330/22	fine-grained sandstone	N49°17'42.2" E19°56'04.9"	22	66	21	2.27	6	2.64	\uparrow			
Kj17	280/20	fine-grained sandstone	N49°19'02.2" E19°52'40.7"	10	53	Kj18	270/15	fine-grained sandstone	N49°19'01.2" E19°52'41.4"	5	72	33	2.00	-19	-9.50	\downarrow			
C10	154/35	fine-grained sandstone	N49°22'23.0" E19°50'55.8"	50	69	C12	160/32	fine-grained sandstone	N49°22'22.6" E19°50'57.2"	13	67	81	3.85	2	0.52	\uparrow			
C69	270/8	fine-grained sandstone (medium-grained at the bottom)	N49°20'43.6" E19°52'51.8"	50	55	C71	280/10	fine-grained sandstone	N49°20'41.5" E19°52'52.0"	18	61	63	2.78	-6	-2.16	\downarrow			
By32	205/5	medium-grained sandstone	N49°20'27.3" E19°55'13.3"	100	50	By33	212/5	fine-grained sandstone (medium-grained at the bottom)	N49°20'24.1" E19°55'15.0"	25	48	107	4.00	2	0.50	\uparrow			

Table 1. 2Θ angle vs. bed thickness

Joints in different stages of their development are present in the Podhale flysch rocks. The “joint anisotropy” proven by geomechanical studies is the least advanced form of jointing (Boretti-Onyszkiewicz 1968a; Domonik 2003, 2008). Some joints have the form of discontinuities, albeit with a macroscopically invisible aperture. Their very subtle traces are sometimes emphasized by capillary water (see also Boretti-Onyszkiewicz 1968b). The barren open joints and the JRCV are the most advanced forms of jointing.

DOUBLE SHEAR ANGLE (2Θ) ANALYSIS

Following considerations presented earlier, the dihedral acute angle between conjugate D_R and D_L sets is the double shear angle (2Θ) (Lovering 1928; De Sitter 1964). Its value varies from 33 to 77° , but over 60% of measurements range from 55° to 69° (Text-fig. 14A). Such 2Θ values are typically observed in the Podhale flysch rocks and in other areas (e.g. Mastella and Mizerski 1977; Mastella and Ozimkowski 1979; Hancock 1985; Mastella and Zuchiewicz 2000; Mastella and Konon 2002; Lunina *et al.* 2005; Rubinkiewicz and Ludwiniak 2005). Similar results have also been obtained experimentally during triaxial compression tests (e.g. Paterson 1958; Handin *et al.* 1963; Gustkiewicz 1985; Łukaszewski 2005). In the study area, there is a regional tendency for 2Θ to increase from the axial zone towards the marginal parts of the Podhale synclinorium (Text-fig. 15 and 16).

Field observations and theoretical considerations suggest that the 2Θ value can depend, among other factors, on lithology, bed thickness, confining pressure and temperature conditions during rock deformation, pore fluid pressure and tectonic factors (i.e. Bradshaw and Hamilton 1967; Al Kadhi and Hancock 1980; Griggs *et al.* 1951; Paterson 1958; Handin and Hager 1957, 1958; Książkiewicz 1968). The influence of lithology on 2Θ has not been considered, because all data were collected from lithologically similar fine- and medium-grained calcareous sandstones. The effect of the pore fluid pressure was also neglected due to the lack of appropriate data.

The value of 2Θ changes only slightly (up to 5°) within a single outcrop, even in beds that differ in thickness. This is consistent with the observations by Tokarski (1975, 1977) and Aleksandrowski (1989), although contradictory results have also been published (Bradshaw and Hamilton 1967; Al Kadhi and Hancock 1980). Comparison made between beds that are more than twice as thick in adjacent outcrops in a similar tectonic position also suggest that in the study area the 2Θ value does not depend on bed thickness (Table 1).

It is well known that higher temperatures and confining pressures cause an increase in rock ductility (e.g. Griggs *et al.* 1951; Robertson 1955; Handin and Hager 1957, 1958; Donath 1961, 1970; Petley 1999; Łukaszewski 2008). The same rock may deform in a brittle or in a ductile manner at low or high confining pressure/temperature conditions respectively. Consequently, an increase in rock ductility can contribute to an increase in 2Θ values. The palaeotemperatures obtained from the studies of the illite/smectite diagenesis and the vitrinite reflectance vary from 50 to 140°C (Konon 1993; Marynowski and Gawęda 2005; Środoń *et al.* 2005, 2006). Generally, they increase downward, similarly as do the 2Θ values (Text-fig. 17). Thus, there is a coincidence between the increase in temperature and the increase in 2Θ values. However, laboratory studies of rocks from other areas (Griggs *et al.* 1951; Heard 1963; Rutter 1974) suggest that the influence of such temperatures on the mechanical properties of the ductility of the Podhale flysch sandstones was not significant. Despite that, the possibility that temperature could have acted synergically with other factors cannot be excluded.

Following the papers by Muehlberger (1961), De Sitter (1964), Muecke and Charlesworth (1966), Jaroszewski (1972), Tokarski (1977), Marín-Lechado *et al.* (2004) and Ismat (2008), the 2Θ value increases with increase in the confining pressure exerted by overburden. The validity of such dependence has been proven by laboratory tests (e.g. Hobbs 1960; Gustkiewicz 1985; Fakhimi 2004; Ramsey and Chester 2004; Łukaszewski 2005). Using confining pressure as a proxy for overburden loading, joints with high 2Θ angles in between are expected to form in the lower part of the flysch sequence. This agrees with current field observations: the highest 2Θ values occur generally in the Szaflary and the Lower Zakopane beds, whereas the lowest ones can be found in the younger lithostratigraphic members. The hybrid-fractures are present in the upper part of the flysch sequence: in the Upper Chochołów beds and in the upper part of the Lower Chochołów beds (Text-fig. 17A). The present-day thickness of the flysch is ca. 3500–4000m (Text-fig. 2), although the original thickness was greater. The present-day overburden exerts vertical pressure, which has been estimated to be ca. 93–106 MPa at the bottom of the flysch series (Text-fig. 17B)***. The pattern of diagonal system described above suggests that the tectonically-induced, NNE–SSW-oriented horizontal pressure (and consequently the σ_1 normal stress) must be greater than the vertical one. Therefore, it is very probable that non-uniform distribution of confining pressure in the

lithostratigraphic section of the Podhale Flysch is the reason for the variation in 2Θ angle.

The 2Θ values may also depend on tectonic factors; although often there are contradictory opinions about that (compare e.g. Sheldon 1912, Książkiewicz 1968 and Tokarski 1977). The 2Θ value is dependent on and increases with the value of differential stress ($\sigma_1 - \sigma_3$) (Muehlberger 1961; Muecke and Charlesworth 1966). The presence of transitions from single extension fractures ($0^\circ < 2\Theta < 10^\circ$ ****) to paired conjugate hybrid ones ($11^\circ < 2\Theta < 45^\circ$) and to paired conjugate shear fractures ($2\Theta > 45^\circ$) (Hancock 1985; Hancock *et al.* 1987) provides evidence that the variation in 2Θ values originates from the variation in differential stresses.

Aleksandrowski (1989) connected the change in the 2Θ angle with the local variation of differential stress during the development of fractures. Rubinkiewicz (1998), in turn, stated that 2Θ values increase upward in the flysch lithostratigraphic section of the Silesian Unit, which is related to the increasing compression from the Early Cretaceous until the Early Miocene. This opinion is in contradiction with Jaroszewski's (1972) statement that the 2Θ -angle in the Mesozoic rocks of the NE margin of the Holy Cross Mts decreases in an adjacent area of the marginal zone of the East European Platform which played "the role of a "transmitter" of the principal Laramide compression" (p. 208). In the study area, the highest 2Θ values have been found in the marginal zones of the Podhale synclinorium. These zones are more tectonically disturbed (especially the zone of contact with the PKB) than the axial one and they are located in the vicinity of the PKB and Tatra block, which could have played the role of a "transmitter" of compression. This influences 2Θ values in the same way as that caused by varying overburden thickness. Unfortunately, due to the type of the Podhale synclinorium structure, separation of these two overlapping phenomena is practically impossible. Thus, it is difficult to determine unambiguously to what extent, if at all, the observed 2Θ angles were dependent on the distance from the PKB and the Tatra block.

Generally, lower than average 2Θ values have been found in the northern part of the BDFZ (Text-fig. 15). Movements in the Mesozoic basement related to

the activity of large faults (Mastella 1975; Makowska and Jaroszewski 1987) caused flexure of the overlying flysch rocks. An extension which developed in the flysch during their flexuring (see e.g. results of modeling by Cosgrove and Ameen 2000) was approximately perpendicular to the strike of the BDFZ. It caused reduction in differential stress value and, consequently, development of conjugate joints with low 2Θ values.

DISCUSSION

The joints in the Podhale flysch rocks have been the main or secondary subject of several papers up to now (e.g. Boretti-Onyszkiewicz 1968a and b; Halicki 1963; Pokorski 1965; Mastella 1972a and b; Pepol 1972; Morawski 1972, 1973). The author's own observations have inspired him to comment on some theses included in these works.

1. The number and orientation of joint sets determined by the author differ from those determined by other researchers (Text-fig. 13). The reason for this is a different method of subdivision of the joint network into sets. The author has determined the joint sets on the basis of field observations, including not only joint orientation, but also joint style and their cross-cutting relationships found in individual outcrops, according to e.g. Hancock's (1994) suggestions. This method precludes the incorporation into a single set of joints that are similarly oriented but of different origin. The subdivision of the joint network performed by other authors (Pokorski 1965; Boretti-Onyszkiewicz 1968b) after fieldwork was mainly derived from a statistical procedure. They determined the numbers of joint sets and the ranges of their orientations from analysis of diagrams constructed for large areas. In particular, these authors included into one set joints having different morphological features and, consequently, different origins. This leads to discrepancy in the number and orientation of particular sets, especially for joints assigned by the author to the D_L and L' sets, and partly also to the D_R and T ones (cf. Text-fig. 7B, C and D).

None of the above-mentioned authors determined the T set joints. This is often invisible on diagrams made for large areas due to the small number of T set joints

** Studies of diagenetic history suggest that an estimated thickness of the removed overburden in the western Podhale would have been as much as almost 3-4 km (Środoń *et al.* 2006).

*** $<172-212$ MPa in the case of the flysch sequence enlarged with the missing section, respectively.

**** It is justified to assume that 2Θ values range from 0° to 10° for single extension joints (Hancock 1985; Hancock *et al.* 1987; see also Konon 2004). Due to the common strike dispersion within a single joint set, some joints can appear as apparently forming "conjugate" joint system with a very small dihedral angle.

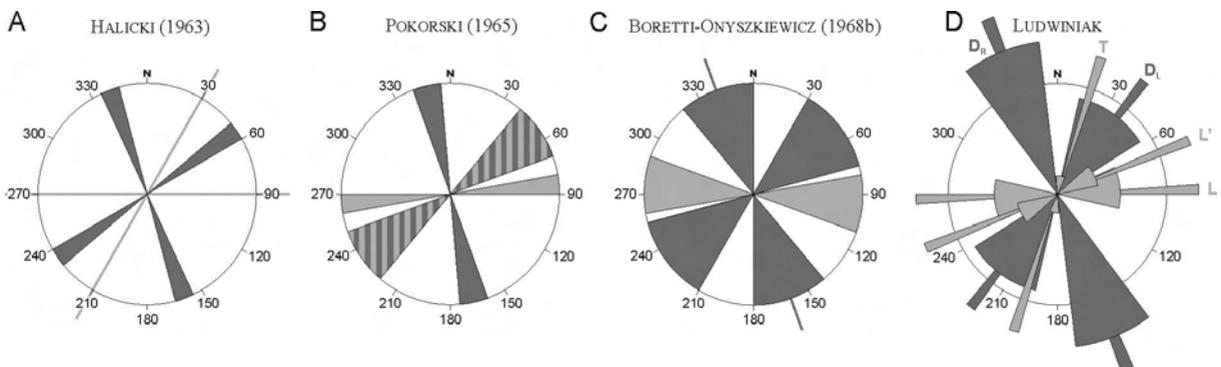
and, in some cases, their orientation close to those of the D_R or D_L set. Despite the lack of the T set joints on diagrams, the co-existence of the D_R , D_L and T sets has been recorded in dozens of outcrops.

The differences in joint orientation can be also due to the fact that Boretti-Onyszkiewicz (1968b) took the measurements from lithologically different layers, i.e. sandstones and shales, for which the deformational properties and values of internal friction angle (ϕ) are different (e.g. Pinińska 2003; Wines and Lilly 2003). She incorporated them into the same data set. According to her, there is no influence of lithology on the orientation of the joints (p. 121). However, this statement is in contradiction with the joint orientation diagrams presented by her (her fig. 6b and 6c – p. 112). They clearly show that the joint orientation within the same sub-area is more scattered in shales than in sandstones. Actually, this makes comparison of joint orientations in the whole area very difficult. The author has avoided this problem by taking all measurements from sandstone layers only.

2. The occurrence of mineral-filled joints (referred to here as JRCV) in the Podhale flysch rocks was mentioned vaguely by Halicki (1963, p. 357) and Mastella (1972b, pp. 79–80, 86–88). Since occurrence of the JRCV is connected only with some joint sets, Halicki ventured an opinion that at least two generations of joints exist. There is no doubt that within a single outcrop the JRCV are older (i.e. opened earlier) than the barren joints. Halicki did not specify, however, the orientation of the JRCV/barren joints composing each of these generations. Boretti-Onyszkiewicz (1966, pp. 118–121; 1968b, p. 121 and 124), who examined thin-sections of the JRCV, stated that calcite veins occur in

all sets, predominantly in those oriented approximately NNW–SSE. The author's observation is that the JRCV occur mainly within the D_R and D_L joint sets and less abundantly within the T set joints. Detailed observations of cross-cutting relationships throughout the study area showed that the joints of the D_R and D_L sets are older than those of the T set (see chapter *Description of joints morphology and cross-cutting relationships*). Boretti-Onyszkiewicz (1966, p. 119) analyzed thin sections cut perpendicular to bedding-surfaces, but parallel or perpendicular to the JRCV planes. In the first case, observations of such oriented thin sections make the recognition of the manner of JRCV opening practically impossible. In the second one, they do not give any answer to the question if there are any signs of shear displacement along the JRCV plane, parallel or near-parallel to the bed surface. The author's studies of thin sections (cut parallel to the bedding surfaces) along with macroscopic observations of markers in both wall-rocks indicate that there is no shear displacement along the JRCV/joint planes. Generally, they were opened as mode I fractures (Atkinson 1987). Only a few cases of JRCV opened obliquely to the wall-rocks have been found. The latter were opened in transtension with a negligible shear motion component.

3. Pokorski (1965) and Boretti-Onyszkiewicz (1968b) incorporated data collected from large areas into one data set without taking into account that particular outcrops might have been located in different tectonic positions. They did not restore the original orientations of the joints despite the fact that the recent orientation of some of the joints may differ from the original one as the result of deformation or rotational



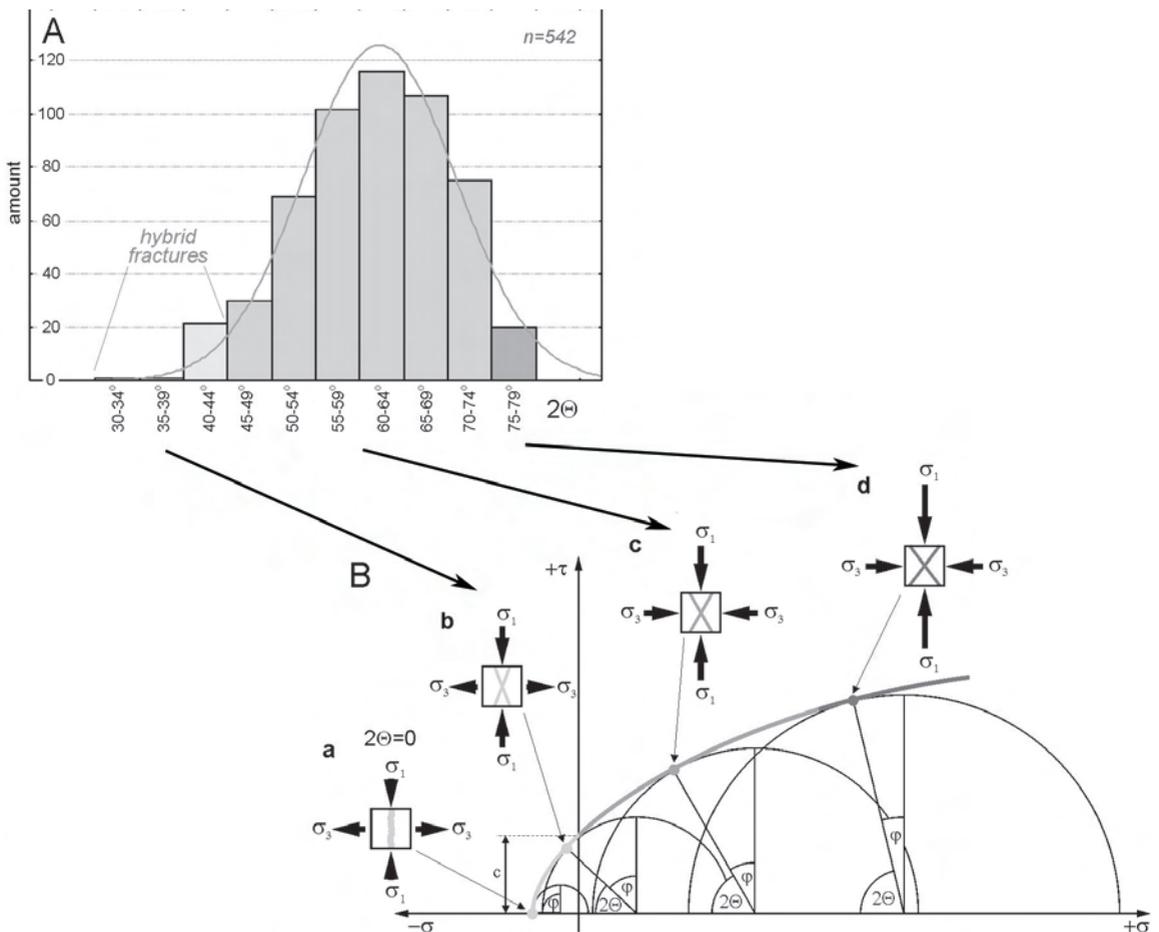
Text-fig. 13. Joint orientations in the Podhale synclinorium. The diagrams present the ranges of orientations for particular joint sets (wider circular sectors) obtained by the author (D) and by the other researchers (A–C). The narrower circular sectors and lines depict regionally dominant orientations of particular joint sets. D – the diagram has been constructed from the data collected from the whole study area, after the original orientation of joints had been restored. Differences in the radius of wider sectors reflect differences in the number of joints between particular sets. For other explanations see text

displacements of the beds after the formation of the joints. The results of tests conducted on data taken from both limbs of single mesoscopic folds point clearly to the unification of joint orientation after bedding correction (Text-fig. 5C, D) i.e. when the joint orientation depends on the later tectonic deformation. Thus, the back-tilting of strata is crucial for comparison of joint orientations in different outcrops throughout the study area which, in turn, is a necessary step for the restoration of the orientation of the palaeostress field. The omission of such a restoration by Pokorski and Boretti-Onyszkiewicz was one of the reasons why joint-set orientations obtained by them differ from those of the author. In the light of the above considerations, statements about the lack of influence of tectonic position on joint orientation (Boretti-Onyszkiewicz 1968b, p. 122; Pepol 1972, p. 594) do not seem to be justified.

4. The majority of authors stated that the joint network in the Podhale flysch rocks is older than the other

tectonic deformations and was formed before tilting of the beds. Boretti-Onyszkiewicz (1968b, p. 122) claimed that there is no evidence that particular sets are of different ages. The Pokorski's (1965, p. 618) suggestion, that the two joint sets of different age are placed in azimuth range from 160 to 175°, was only a loose remark. The author's analysis shows the different ages of particular joint-sets and the multi-stage development of the joint network. In fact, the majority of joints had formed before other tectonic deformations. There are some joints, however, that formed simultaneously or later than the other structures.

When discussing the age of joints, the researchers mentioned above omitted the aspect of two-stage development of joints belonging to particular sets, as follows from Price's theory (1959, 1966). Only Boretti-Onyszkiewicz (1968b, p. 131) stated, on the basis of the geometry and morphology of the joint surfaces, that two principal sets were initiated by shearing and opened in direction perpendicular to joint planes.

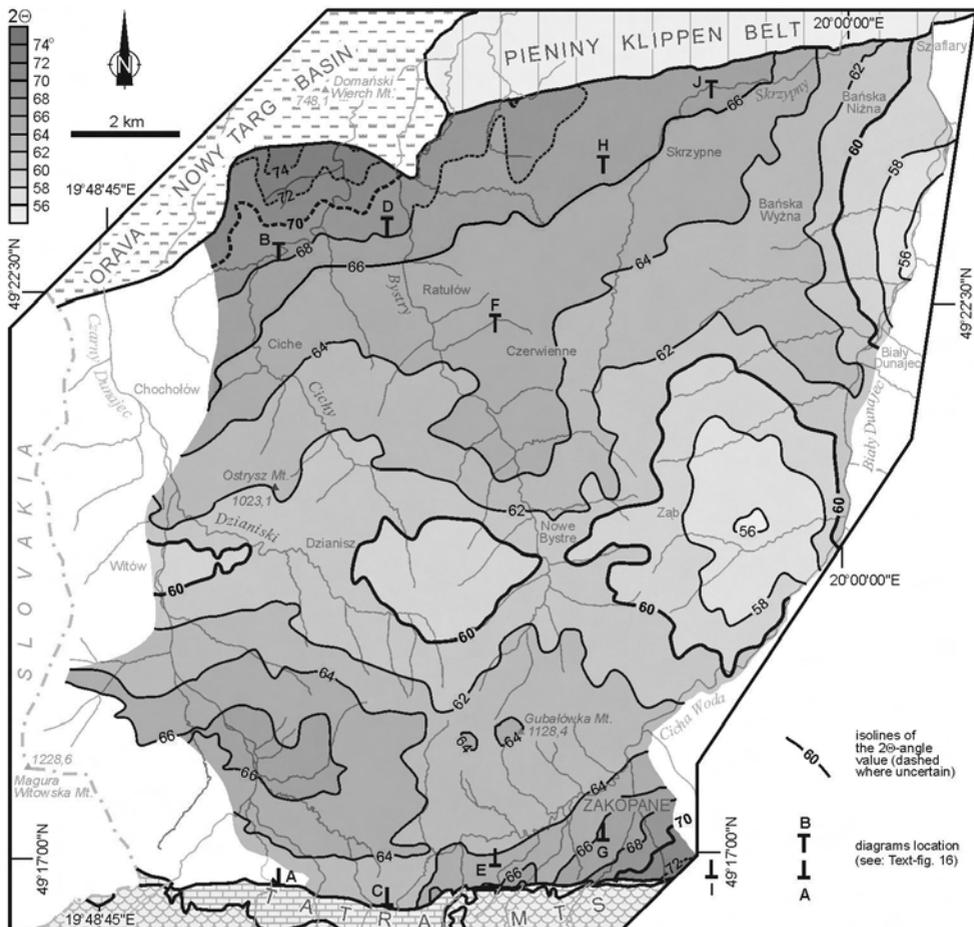


Text-fig. 14. A – Histogram of the double shear angle (2Θ) variability. B – Mohr's diagram showing stress conditions for the development of the diagonal joint system and the T set joints (after Price and Cosgrove 1990, Dadlez and Jaroszewski 1994, modified). ϕ – angle of internal friction, c – cohesion, τ – shear stress. For other explanations see text

She did not make, however, an unambiguous separation of the initiation/opening stages. It should be noticed, that the inception of joints as joint-anisotropy and the stage of opening of joints can be widely separated in time. Theoretically, in the case of the Podhale flysch rocks, these processes could have been separated by as much as ca. 30 Ma. Evidence of that has been presented here in the case of the diagonal system, where joint-anisotropy was formed during the incipient stage of the structural evolution of the Podhale synclinorium, whereas the actual opening of the joints took place during the uplift of the flysch, and also recently, as a result of exogenic processes.

5. Despite the obvious fact that observations of morphology are crucial for the determination of the origin of fractures (e.g. Hancock 1985; Price and Cosgrove 1990; Dunne and Hancock 1994; Twiss and Moores 2001) they were either omitted during previous studies or were very simplified. Small joint-related brittle structures have been observed only sporadically (Mastella and Mizerski 1977). For example, Halicki (1963 – p.537)

and Pokorski (1965 – p.617) stated that in some cases joint surfaces are very uneven in thick-bedded sandstones, but they did not assign them to any particular set. According to Boretti-Onyszkievicz (1968b – p.129), roughness of joint surfaces is connected with lithology, i.e. surfaces are rougher in conglomerates than in fine-grained sandstones. She also claimed that the morphology of joints of all sets is very similar (p.133). The first statement is unquestionable and obviously true, while the second one is inconsistent with the author's own observations (see chapter *Description of joints morphology and cross-cutting relationships* and Text-fig. 7A). The results of studies conducted in other areas (e.g. Lorenz *et al.* 1991; Rawnsley *et al.* 1998; Arlegui and Simón 2001; Zhang and Wang 2004), as well as research into the roughness of artificially produced fractures correlated with joint anisotropy (Domonik 2005) show that the morphologies of joints belonging to sets of different origin are different. Moreover, surfaces of artificial fractures become smoother in the transition from a single extensional to conjugate shear fractures (Ramsey and Chester 2004).



Text-fig. 15. Map of the 2θ angle based on the 1st order polynomial trend surface analysis

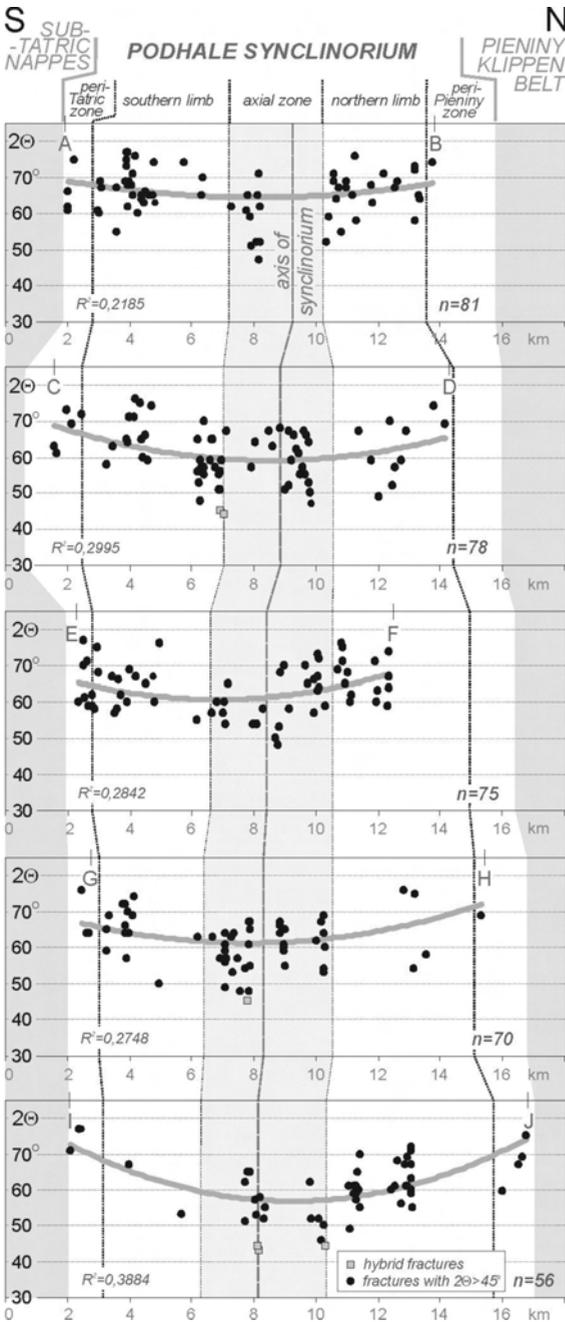
Pokorski (1965, p. 619) arbitrarily classified joints placed in an azimuth range from 40 to 70° as “shear joints”. Boretti-Onyszkiewicz (1968b, p. 134) allowed for the possibility that the sets oriented 140–180° and 30–75° were formed simultaneously “under

shear stresses induced by i.e. uniaxial compression”. In her opinion, these sets are conjugate and “shear” because they cross mutually and exhibit small mutual horizontal offsets of joint traces (up to several mm). The author agrees with these statements only to some extent. Conjugate shear discontinuities with the 2Θ angle $>45^\circ$ form under conditions of triaxial compression, as proved by the results of laboratory tests (e.g. Paterson 1958; Hobbs 1960; Donath 1961; Łukaszewski 2005). The mutual crossing of joints of both sets shows undoubtedly that they were opening synchronously. However, the horizontal offsets of joint traces within tilted beds could have been caused by any later strike-slip motion along them. Thus, they should be better classified as *faulted joints* (Cruikshank *et al.* 1991; Zhao and Johnson 1992; Wilkins *et al.* 2001). Moreover, this offset is often only apparent – it might be an effect of the mode I opening of a younger joint which crosses an older one obliquely. As follows from the considerations presented in this paper, the geometry of joint-related small brittle structures indicates only a tendency to strike-slip motion. Some of the several-millimetre long apparent “mode II offsets” observed in this work seem to have been formed recently as a consequence of small displacement of joint-bounded blocks caused by rock mass deterioration.

Thus, the classification of joints striking at 30–75° as “shear joints” by Boretti-Onyszkiewicz (and at 40–70° by Pokorski) does not seem to be well justified. As has been discussed here, joints having similar orientation but different in morphology and origin were incorrectly placed in this azimuth-range (i.e. D_L and L' sets).

6. The issue of the regional variability of 2Θ has not been considered up to now. As stated by some authors, the angle between joints which correspond to the D_R and D_L sets vary from 50 to 90° (Boretti-Onyszkiewicz 1968b, p. 119) or is even close to 90° (Morawski 1972, p. 581). Both the above authors suggest that these sets are conjugate and were formed by shearing, so the dihedral angle between them would have corresponded to the 2Θ angle. Pokorski (1965) claimed that a single set striking at 40–70° was formed by shearing. Assuming that the 2Θ value was equal to 45° ($2\Theta=90^\circ$) he determined the σ_1 direction. He did not, however, give any arguments that this set was indeed formed by shearing. Moreover, he also did not specify the orientation of the second conjugate set.

Such large values of 2Θ (close to 90°) are inconsistent with the field observations (e.g. Jaroszewski 1972; Aleksandrowski 1989; Hancock 1985; Dunne and Hancock 1994; Mastella and Zuchiewicz 2000;



Text-fig. 16. Diagrams showing distribution of the 2Θ angle values along N–S oriented scan lines (for location see Text-fig. 15). Each diagram has been plotted by the perpendicular projection of points of outcrop location on the N–S scan line located in the middle of a 2-km-wide zone.

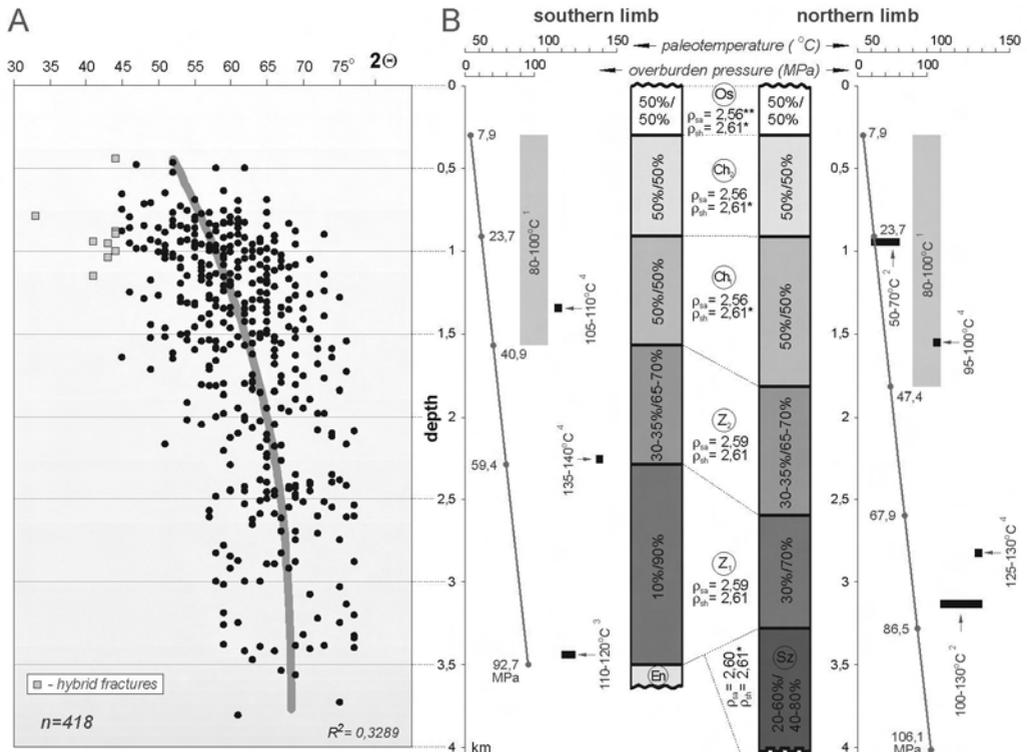
The 2nd order polynomial trend lines are marked on the diagrams

Rubinkiewicz and Ludwiniak 2005) and laboratory studies (e.g. Paterson 1958; Hobbs 1960; Handin *et al.* 1963; Gustkiewicz 1985; Ramsey and Chester 2004; Łukaszewski 2005, 2007) reported so far. It seems likely that 2Θ -values obtained by Pokorski, Boretti-Onyszkiewicz and Morawski are too large, because of the unjustified incorporation of some L' and D_L joints into a single set. Moreover, these authors did not take into account the aspect of the internal friction angle (ϕ) when considering the issue of 2Θ values. The occurrence of 2Θ values equal or close to 90° in the flysch sandstones would indicate that the ϕ -angle is equal or close to 0° . This might be possible only in the hypothetical case of non-existing material without any internal friction (De Sitter 1964; Jaroszewski 1980).

• Świerczewska *et al.* (2000, 2005) and Tokarski *et al.* (2006), based on the studies of joints and veins in the Magura Nappe, proposed another model for the development of joints and related veins. Accord-

ing to this model, all joint sets were formed originally as extension fractures and were reactivated by subsequent strike-slip displacement along their surfaces, due to a change in their attitude relative to the stress field, caused by the rotation of the latter one (see also: Wilkins *et al.* 2001; Eyal *et al.* 2006). For that case, the model proposed by them is convincing. However, it does not fit exactly the case presented in this paper.

• The architecture of the joint network described by Świerczewska and Tokarski differs from the one described here. According to them, the angular relationships between particular joint sets are more or less fixed. The joints of the D_1 and L sets, as well as those of the D_2 and L' sets respectively, are approximately perpendicular to each other. In the case of the western part of the Podhale synclinorium, the relationships described above are not achieved. One of the reasons is the variability of the angle between the D_R and D_L sets in the flysch lithostratigraphic section.



Text-fig. 17. A – 2Θ angle vs. overburden thickness of Podhale flysch rocks. Measurements have been collected from throughout the study area, excluding the Biały Dunajec fault zone. B – Diagram showing changes in overburden pressure and palaeotemperatures in lithostratigraphic columns of the Podhale Flysch (palaeotemperature values after Konon 1993¹, Marynowski and Gawęda 2005²; Śródoń *et al.* 2005³, 2006⁴). The percentage values point to the sandstones/shales summary thickness ratio within particular lithostratigraphic members. Bulk density of shales (ρ_{sh}) and sandstones (ρ_{sa}) from particular lithostratigraphic members is given after Bromowicz and Rowiński (1965) and Watycha (1968, 1976a). * ρ_{sh} values assumed on the basis of ρ_{sh} values from the Zakopane beds; ** ρ_{sa} values assumed on the basis of ρ_{sa} values from the Chochołów beds (for other explanations see Text-fig. 1C). The values of overburden pressure were calculated as a function of thickness of overlying rocks and their average bulk density. The calculations take into account the sandstones/shales summary thickness ratio

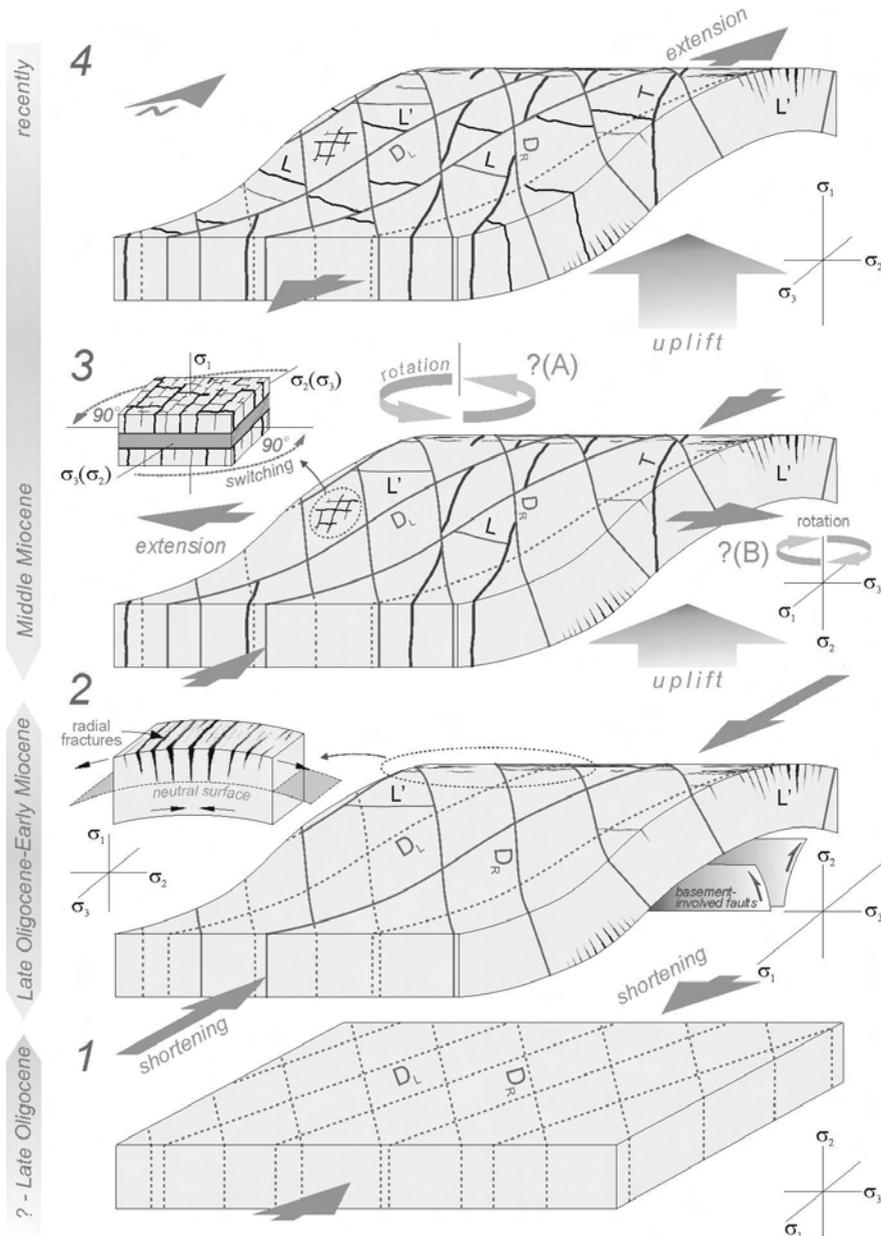
• The cross-cutting relationships, the occurrence of JRCV within certain joint sets and the attitude of joints/JRCV to other tectonic structures point to a distinctive sequence of set formation (see text).

• In the study area the majority of JRCV are *simple veins*. Thus, their architecture does not indicate their reactivation as a result of rotation of the palaeostress field. Only a few of the JRCV are asymmetrical *composite veins*. Unfortunately, most of the JRCV are composed of blocky grains and therefore are not good kinematic indicators.

Regional joint-network development stages

Detailed observations of the geometry, morphology, joint-related structures and veins, as well as the attitude of joints with respect to bedding and other tectonic structures clearly show that the joint network developed in several stages linked to the structural evolution of the Podhale synclinorium.

Stage 1. Development of the oldest diagonal system in the form of joint-anisotropy started when horizontal flysch strata became sufficiently lithified to enable the ac-



Text-fig. 18. Joint network development in the flysch of the Western Podhale area. Opened joints are indicated as solid lines and the "joints anisotropy" as dashed ones. The lengths of the σ -axes indicate the relative values of normal stresses. For other explanations see text

cumulation of stresses arising from NNE–SSW regional compression (Text-fig. 18-1) caused by the convergence of the European plate with the ALCAPA (e.g. Oszczytko and Ślącza 1989; Csontos *et al.* 1992; Plašienka *et al.* 1997; Fodor *et al.* 1999; Zoetemeijer *et al.* 1999). There is only limited field evidence to suggest that these joints started to develop within poorly lithified rocks.

Stage 2. The fundamental structural framework of the Podhale synclinorium was formed in the Late Oligocene–Early Miocene (Birkenmajer 1986) as parallel W–E-trending tectonic zones (Mastella 1975). It was caused by block movements related to rejuvenation of the faults in the Mesozoic basement (Mahel 1969; Soták and Janočko 2001). At this stage, some diagonal joints opened during the initial phase of folding (largely open and gentle in character), and joints genetically associated with the folding, i.e. the majority of the L' and a small portion of the L joints, were formed (Text-fig. 18-2).

Stage 3. In the Middle Miocene, the uplift of the Podhale synclinorium started (Sperner *et al.* 2002; at ~16–18 Ma ago – Środoń *et al.* 2006; Text-fig. 18-3) and the BDFZ and the Białka fault zone were formed (Mastella *et al.* 1996). During gradual uplift and under diminishing horizontal compression, the T set joints formed as an effect of WNW–ESE extension of the flysch rocks. The small angular difference (-8°) between the $\sigma_{1(D)}$ and the T set direction suggests a small clockwise rotation of the palaeostress field (i.e. σ_1 and σ_3 – Text-fig. 18-3 B) from that acting during the initiation of the diagonal system (*Stage 1*). However, a possibility of a small counterclockwise rotation of the basement of flysch rock masses cannot be excluded (Text-fig. 18-3 A). In this case, the rotation sense would be consistent with the results of palaeomagnetic investigations (Marton *et al.* 1999). Locally, a small portion of the L set joints formed simultaneously with the joints of the T set, creating a *grid-lock* pattern. This could have resulted from local, rapid, 90° alternate switching of σ_3 and σ_2 axes of nearly equal magnitude (σ_1 axis is parallel to the intersection line of both sets; Text-fig. 18-3) (Caputo 1995).

Stage 4. The lateral extension connected with the progressive uplift of the flysch rocks lasting up to recent times (Makowska and Jaroszewski 1987; Baumgart-Kotarba and Král 2002; Środoń *et al.* 2006) enabled relaxation of residual stresses within the flysch rock masses. As the result of the release of elastic strain energy accumulated during regional compression (Price 1959), the youngest extension joints were formed – the majority of the joints of the L set and a small portion of the L' set joints (Text-fig. 18-4). At the same time, the opening of joints belonging to older sets continued due to the weathering of the flysch rocks and protracted stress relaxation (Ostaficzuk 1973; see also Varnes and Lee 1972).

Acknowledgements

I am grateful to Leonard Mastella for all his help and long useful discussions. I would like to thank Andrzej Konon and Witold Zuchiewicz for detailed and very helpful reviews improving the original version of the paper. Sincere thanks are also due to Zbigniew Żytkiewicz and Chris J. Wood, for linguistic improvement. This paper was supported by internal grants: BW-1642/5 and BST-977/2 (Institute of Geology, University of Warsaw).

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Manuscript submitted: 15th March 2009

Revised version accepted: 15th April 2010