Conference Excursion 1: Structural Development of the Magura Nappe (Outer Carpathians): From Subduction to Collapse

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Introduction

The Outer Carpathians are a thrust-and-fold-belt, north-verging in the Polish segment (Fig. 1A). The belt, composed largely of Lower Cretaceous to Lower Miocene flysch strata, comprises several nappes. The innermost and largest of the nappes is the Magura Nappe. This nappe is subdivided by north-verging reverse faults into four slices which are named (from south to north) Krynica, Bystrica, Rača and Siary slices (Fig. 1B). To the north, the Outer Carpathian nappe pile is thrust over the Carpathian Foredeep, whereas to the south, the Magura Nappe contacts along steep faults with the Pieniny Klippen Belt (Fig. 1A). The Outer Carpathian nappe pile was formed due to Palaeogene-Neogene, southward-directed subduction of oceanic or sub-oceanic crust intervening between continental crust of the European plate and continental crust of the ALCAPA unit. The features produced during subduction-related shortening are overprinted by normal faults (Zuchiewicz et al. 2002 and references therein), which were formed during gravitational collapse. These faults bound intramontane basins filled with Neogene strata.

The map-scale structures of the Outer Carpathians have been recognised for a long time due to repeated mapping and numerous deep wells. On the other hand, still ten years ago, very little structural research have been done in the Outer Carpathian. During this excursion (Fig. 1C), we are going to show the results of our research of the past ten years.

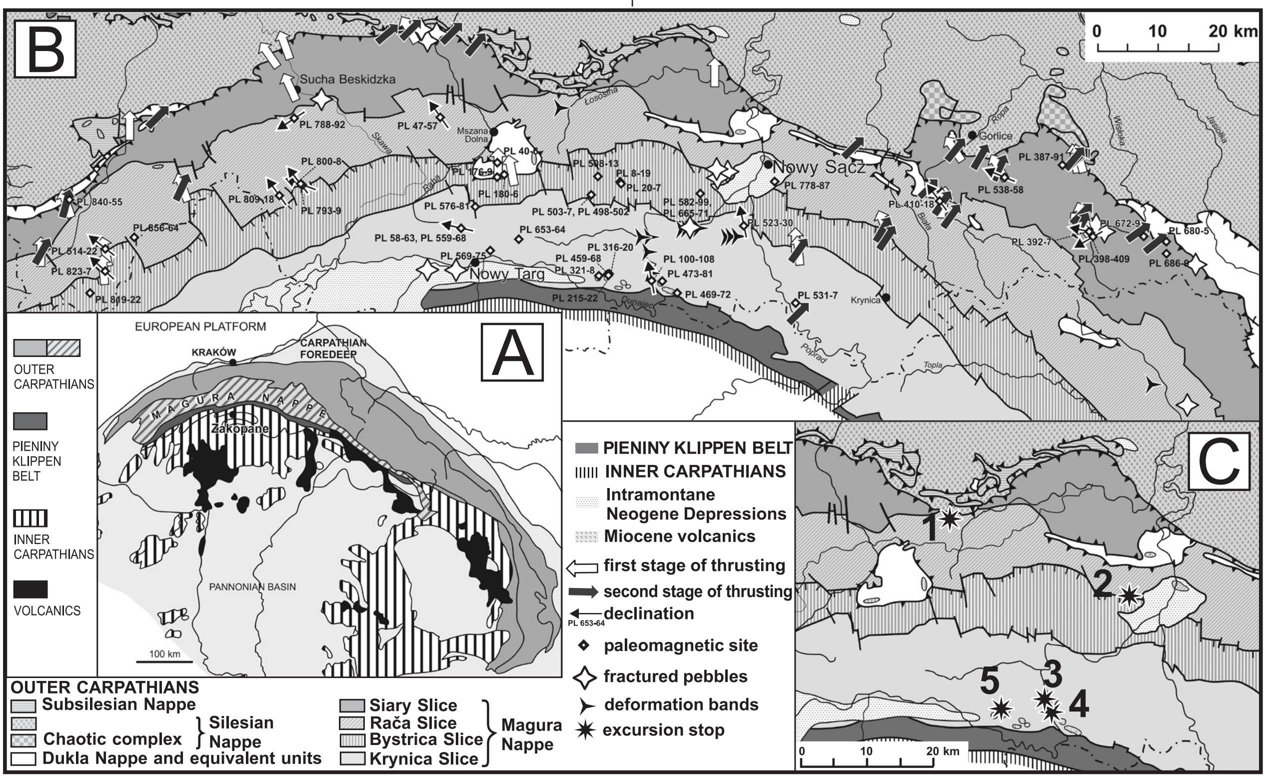


Fig. 1. A – structural scheme of the Carpathian arc; B – map showing: (1) results of kinematic analysis of the Magura Nappe (after Decker et al. 1999 and Galicia T. Group unpublished), (2) results of palaeomagnetic study in the Magura Nappe (after Márton et al. 1999, 2004 and Márton and Tokarski unpublished), (3) location of exposures of gravels and paraconglomerates bearing fractured clasts and (4) location of exposures in deformation bands were studied, C – itinerary of excursion.

Kinematic analysis

We have analysed small-scale structures at over 120 exposures located within all nappes. The results (Fig. 1B) show that the Outer Carpathian nappe pile was formed due to two successive stages of thrusting, the first one directed towards the NW, the second one to the NE. The inferred succession of thrusting confirmed earlier results of Aleksandrowski (1985) from the western part of the Magura Nappe. On the other hand, our results raised important question: whether the succession of differently verging thrusts results from clockwise far-field stress rotation or from anticlockwise body rotation of the Outer Carpathians. To answer this question, we have undertaken extensive palaeomagnetic study.

Palaeomagnetic study

From the Magura Nappe, fine-grained members of the flysch sequence, preferably marls, were collected for palaeomagnetic study at 34 localities (300 samples) distributed along the arc of the nappe (Fig. 1B). From the Mszana and Szczawa tectonic windows, additional four localities (24 samples) and from the Upper Badenian marls of the Nowy Sącz intramontane basin, two localities (21 samples) were sampled. As a result of standard palaeomagnetic measurements and evaluation, 15 localities from the Magura Nappe and one locality from the Nowy Sącz Basin yielded statistically good or acceptable palaeomagnetic directions (Table 1). They all indicate counterclockwise rotation, which is, however, characterizing post-folding/tilting tectonic movements, since the remnant magnetizations were acquired after folding/tilting (exception may be locality Pl 840-855). There seems to be no systematic change in declinations as we proceed from the west to the east, and although inclinations cluster around 60 °, there are a few localities with much shallower inclinations increasing the within-locality scatter.

Based on 14 localities (Pl 840-855 excluded) the palaeomagnetic mean direction of post folding/tilting age is Dec = 306°, Inc = 57 ° (k = 15, α_{95}° = 11), suggesting a general counterclockwise rotation of about 50 ° of the Magura Nappe, which could have taken place before the late Badenian, since declination for a single locality from the Nowy Sącz Basin does not show any deviation from the expected stable European declinations. For more details see **Stop 5**.

Timing of deformation

The termination of folding of the Magura Nappe is fairly well dated by: (1) the age of Miocene andesites which cut the already folded Magura sequence, and (2) the age of the unfolded Badenian strata filling Nowy Sącz intramontane basin. Conversely, the timing of the onset of folding has been not known; however, it had been traditionally accepted that it took place during the Early Miocene or Late Oligocene times (e. g. Roca et al. 1995). This opinion mostly comes from the apparent lack of angular unconformities within the flysch sequence, except for the local unconformities at the base of the Oligocene strata (Książkiewicz and Leško 1959), and within the Upper Eocene strata (Węcławik 1969); both unconformities occurring in the inner part of the Magura Nappe.

Locality	N (n)	D°	I°	k	α_{95}°	Dc°	Ic°	dip
Tenczyn Pl 47-57	7/11	321	+48	14	16	284	+70	168/30
Łąkcica Pl 100-108	5/9	149	−28	40	12	151	−5	352/25
Wołowiec Pl 392-397	6/6	282	+60	44	10	248	+23	220/50
Radocyna Pl 398-409	5/10	240	+64	26	15	224	+27	211/39
Łosie 410-418	8/9	323	+47	31	10	260	+42	205/56
Złatna Pl 514-522	5/9	128	−60	40	12	217	−86	120/30
Barcice Pl 523-530	6/9	168	−61	82	8	153	−59	266/9
Ropica Pl (538-558) 547-551	5/5 (21)	292	+57	37	13	264	+32	230/36
Klikuszowa Pl 559-568	4/11	283	+66	41	14	263	+37	245/32
Gołynia Pl 788-792	4/5	237	+59	25	19	183	+40	138/40
Zubrzyca I. Pl 793-799	5/7	290	+66	31	14	280	+69	160/5
Zubrzyca II. Pl 800-808	4/9	322	+31	28	18	302	+58	170/33
Zawoja Pl 809-818	7/10	324	+59	37	10	320	+36	315/23
Glinka Pl 823-827	4/5	313	+46	71	11	315	+47	345/5
Milówka Pl 840-855	7/16	84	+21	25	12	85	−20	150/85

■ **Tab. 1.** Palaeomagnetic results from the Magura Nappe.

The goal of our studies (Świerczewska and Tokarski 1998) was to demonstrate that folding in the Outer Carpathians started earlier than hitherto thought. We attempted this by dating the history of folding in relation to increasing induration of the strata involved. This was done by the analysis of deformation bands microstructures which indicate the degree of induration during a deformational event.

Deformation bands (DB) (Antonellini et al. 1994 and references therein) are roughly planar features which occur in porous granular materials. In sandstones, they form bands which are up to few millimeters thick and up to few hundred meters long. DB are accommodating offset like faults, but they do not contain planes of displacement discontinuity. DB are widespread in the strata of the Outer Carpathians. We have observed them within the Silesian and Magura nappes, in strata ranging from the latest Cretaceous through Late Oligocene in age.

We studied DB within strata of the Magura nappe (Fig. 1B) which is the innermost tectonic unit in the Polish segment of the Outer Carpathians. Assuming piggy-back style folding and thrusting in the Outer Carpathians (e.g. Decker and Peresson 1996 and references therein), it appears reasonable that folding first started in the Magura Nappe. Altogether, we have studied DB at 10 localities. For details and results see **Stop 1**.

PT conditions

Illite/smectite studies

X-ray diffraction studies of illite-smectite separated from about 270 claystone samples were use for a reconstruction of the burial and thermal history of the Magura Nappe (Świerczewska in press). Maximum palaeotemperatures were estimated based on the degree of smectite to illite transformation. The palaeotemperatures were calculated using the plot of Šucha et al. (1993) with a 10 °C correction, as suggested by Clauer et al. (1997).

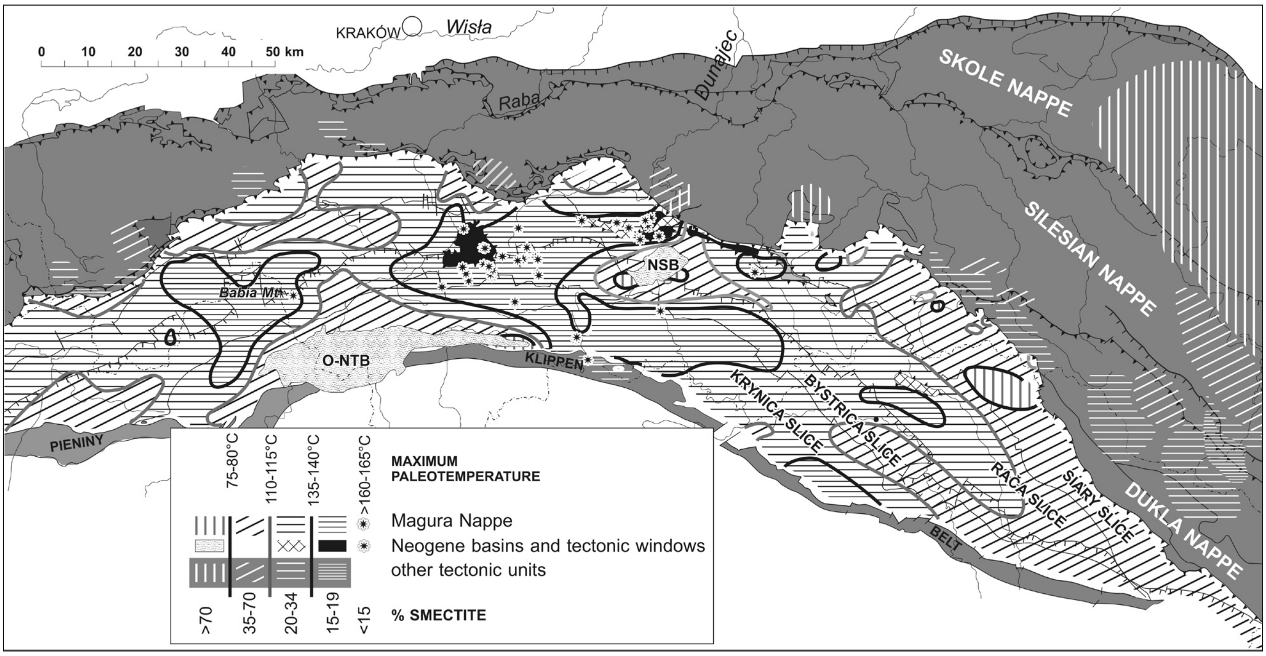
The Magura Nappe claystones contain from ≪5 to 90 % smectite in illite-smectite (I/S). These compositions indicate that the observed thermal alteration of the rocks on the present-day erosion surface of the nappe reflects temperatures that ranged from ≪75 °C to 200 °C (Fig. 2). These temperatures were related to tectonic burial at depths of between <4 and 11 km, assuming a mean value of 18 °C/km for the palaeo-geothermal gradient. For those exposures where this low palaeogradient was constrained by fluid inclusion data, burial depths calculated using I/S data agree with those based on fluid inclusions (cf. Hurai et al. 2004). This coincidence indicates that the influx of water-methane fluids and the maximum thermal alteration were coeval. The calculated values are a measure of the maximum thickness of eroded cover rocks.

Summary results of I/S studies for individual slices of the Magura Nappe and for the entire nappe show that for each slice, and for the entire nappe, there is a positive correlation between the age and degree of alteration of the sampled rocks. Only in small discrete portions of the nappe this correlation is not visible. Within the entire Magura Nappe, there is a clear general decrease in thermal alteration from the inner part outwards. A local and less distinct trend of decreasing thermal alteration inward, towards the Inner Carpathians, is apparent in the innermost part of the Magura Nappe. The maximum thermal overprint, corresponding to 160–165 °C and more, is registered in the central part of the nappe, especially in the region where tectonic windows occur. Rocks affected by the lowest palaeotemperatures (≪75 °C) occur only on the edge of the intramontane Nowy Sącz Basin, and in scattered localities in the eastern part of the Siary slice. In the central segment of the frontal overthrust, the Magura Nappe is thrust over significantly less altered rocks of the Silesian Nappe.

The observed thermal structure of the Magura Nappe was largely established during accretion, before emplacement of the Magura Nappe rocks in the present day tectonic setting. The outward-directed decrease in thermal alteration, positive correlation between rock age and thermal overprint in each of the slices and in the entire nappe, correlate with the growth of the accretionary wedge. The accretion-related thermal structure was greatly modified during the later thrusting of the Magura Nappe over its foreland. This event involved differential uplift and erosion. The uplift was significantly influenced by the morphology of Carpathian basement. The erosion was greatest above and inward from the regional basement slope distinguished by Ryłko and Tomaś (2001). This erosion resulted in exhumation of the most altered rocks. Two main stages of uplift and erosion can be distinguished. The first stage was related to the uplift of the accretionary wedge, when up to 4 km could have been removed by erosion. The second stage of uplift and erosion was related to thrusting of the Magura Nappe over its foreland and to coeval deformation within the latter.

Vein mineralization vs. structural development

Mineral fillings of fractures record conditions of structural evolution of a given tectonic object. Conversely, knowledge



■ **Fig. 2.** Results of illite/smectite studies, palaeoisotherms on the present day erosion surface, NSB – Nowy Sącz Basin, O-NTB – Orava-Nowy Targ Basin (after Świerczewska in press; geology from Żytko et al. 1989, Ryłko and Tomaś 2003).

of structural development enables for dating of mineralization phases. The results of our studies on the interrelationship between structural development and phases of calcite and calcite-quartz mineralization in the Magura Nappe (e.g. Świerczewska et al. 2000b) allowed us: (1) to date the phases of mineralization in relation to successive tectonic stages in a fragment of the Outer Carpathians and, (2) to reconstruct the PT conditions which took place during these stages.

In our reconstruction, the relationship between structural development and mineralization seems to be well proven for the last stage of the structural development (collapse). On the other hand, we were not quite satisfied of the relationship for the earlier stages of structural development. We believed that the suspected unfit could have resulted from the adopted model of jointing which comprised both shear and extensional joints (e.g. Dunne and Hancock 1994). Therefore, we decided to test our interpretation by adopting exclusively the extensional model of jointing (e.g. Dunne and Hancock 1994). For the object of this study we have chosen an exposure of Paleocene-Lower Eocene strata where the mineralization has been already intensively studied (Świerczewska et al. 2000a and b, 2001).

The exposure at Krościenko town (Fig. 1C, stop 4) is located in the innermost part of the Magura Nappe, the structural development of which is fairly well known (Świerczewska and Tokarski 1998). In that part of the Magura Nappe, mineral veins in Paleocene-Eocene sandstones are largely restricted to early cross-fault joints and faults. Vein textures show that mineralization occurred progressively with the evolving stress regime. The reconstructed process of mineralization fits well into the scheme of the Outer Carpathian structural development. Mineralization was most abundant during gravitational collapse, the last stage of structural development. The extensional model of jointing has been positively verified. For more details see **Stop 4**.

Fluid inclusions

Temperature of fluids penetrating rocks of the Magura Nappe was evaluated basing on fluid inclusion studies. These inclusions are hosted in calcite and calcite-quartz association filling fractures. Only quartz and calcite formed during collapse do contain immiscible aqueous_methane fluid inclusions. Trapping PT conditions for methane-bearing fluids are 160–210 °C and 0.75 to 2 kbar (Świerczewska et al. 2000a). Quartz –calcite veins were only found in the central segment of the Magura Nappe. They are more common in the basement of this nappe, namely in the Dukla Nappe and its equivalents exposed in tectonic windows. Fluid inclusions data collected from the basement show higher temperature (200–220 °C) and pressure (2.1–3.3 kbar) of trapping than those in the Magura Nappe.

Calcite formed in all stages of structural development contains mainly aqueous_inclusions. The latter homogenise between <50 and 145 °C (Świerczewska et. al. 2001). For most of the studied blocky calcites these temperatures are between 80 and 100 °C (A. Kozłowski unpublished), whereas for the columnar calcites they are more differentiated. Quartz overgrowths, which predate calcite mineralization, show homogenisation temperature below 50 °C.

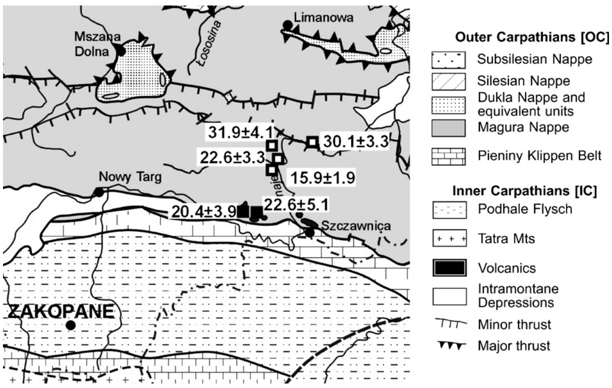
Volcanism

Andesites occur within the Pieniny Klippen Belt and the adjoining portion of the Magura Nappe (Fig. 1B). These rocks form numerous small intrusions (mostly dykes) exposed along the so-called Pieniny Andesite Line. There are two systems of intrusions representing two successive phases of volcanic activity (Birkenmajer and Pécskay 2000 and references therein).

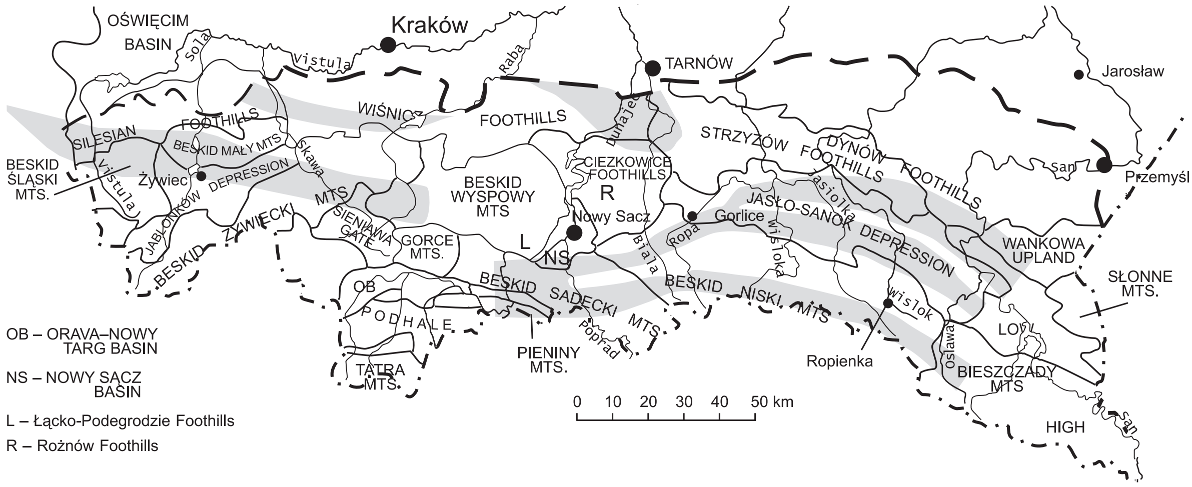
The Pieniny andesites represent basaltic andesites and andesites (according to TAS classification; LOI free basis). The SiO_2 content ranges between 51.5 and 61.5 wt%; LOI value varies from 1.3 to 3.8 wt% (Michalik et al. 2004). Numerous varieties of andesites were described according to phenocrysts assemblage composition (e.g. Małkowski 1958).

According to the results of K/Ar dating, the ages of andesites of both systems range from 13.5 to 11 Ma (Birkenmajer and Pécskay 2000). Both andesites and the host strata are hydrothermally altered (Szeliga and Michalik 2003 and references therein). Andesite samples bearing more altered mafic phenocrysts are rich in secondary chlorite, chlorite/smectite and Ba-enriched K-feldspars in matrix, whereas samples with less altered mafic phenocrysts containin illite/smectite (or vermiculite/smectite) (Michalik et al. 2006 submitted). The age of newly formed biotite in hydrothermally altered andesite is 11.35 Ma (Birkenmajer et al. 2004). This can suggest that either hydrothermal activity occurred directly after magma crystallization, or that dates of Birkenmajer and Pécskay (2000) correspond to the hydrothermal event.

New apatite fission track (AFT) analyses of two andesite samples from Mt. Wżar (Fig. 3) yielded ages of 20.4 ± 3.9 Ma and 22.6 ± 5.1 Ma. (Anczkiewicz unpublished). These ages are similar to the AFT ages obtained for the sandstones from the surrounding flysch of the Magura Nappe, which span an interval of 16–30 Ma. All AFT ages are younger than stratigraphic age of the strata involved and are clearly reset. Hence, the obtained ages both for andesites and for the surrounding flysch strata are interpreted as reflecting cooling of the Magura Nappe. Such interpretation implies that andesitic intrusions took place at least about 20 Ma ago. These results strongly contradict earlier K-Ar results. Clearly, further geo-



■ **Fig. 3.** AFT dating results from the Pieniny andesites (black squares) and Magura Nappe strata (white squares). For location see Fig. 1C (Stop 5).



■ **Fig. 4.** Physiographic units of the Polish segment of the Carpathians (based on Starkel 1991, modified), showing zones of Quaternary uplift marked by abnormally high river bed gradients (modified from Zuchiewicz 1998).

chronological studies, involving other geochronometers are needed in order to establish reliable timing of the Pieniny andesites.

Birkenmajer (2003) provided explanation of the origin of andesite magma. It can be a product of hybridisation of primary magma of mantle origin related to subduction of the European Plate. An alternative explanation (Birkenmajer 2003) suggests that the magma chamber was formed at the base of the accretionary wedge of the Outer Carpathians (possibly at a depth of 10 to 12 km). The continuation of the Pieniny Andesite Line matches the Odra Fault Zone (Lower Silesia), indicating a relationship between the Pieniny andesites and the Lower Silesian alkaline silica-undersaturated mafic rocks of the eastern termination of the Central European Volcanic Province (Birkenmajer and Pécskay 1999). According to Pin et al. (2004), the limited volumes of magmas were produced by partial melting of mafic rocks in the lower crust and/or enriched domains in the mantle, probably as a result of local decompression in a transpressive geodynamic setting. Jurewicz and Nejbert (2005) link the origin of andesitic magma with decompression during shear movements within a deep tectonic zone which can be the SE continuation of the Kraków-Lubliniec fault zone (tectonic contact of the Upper Silesia and Małopolska blocks).

Postscriptum (neotectonics)

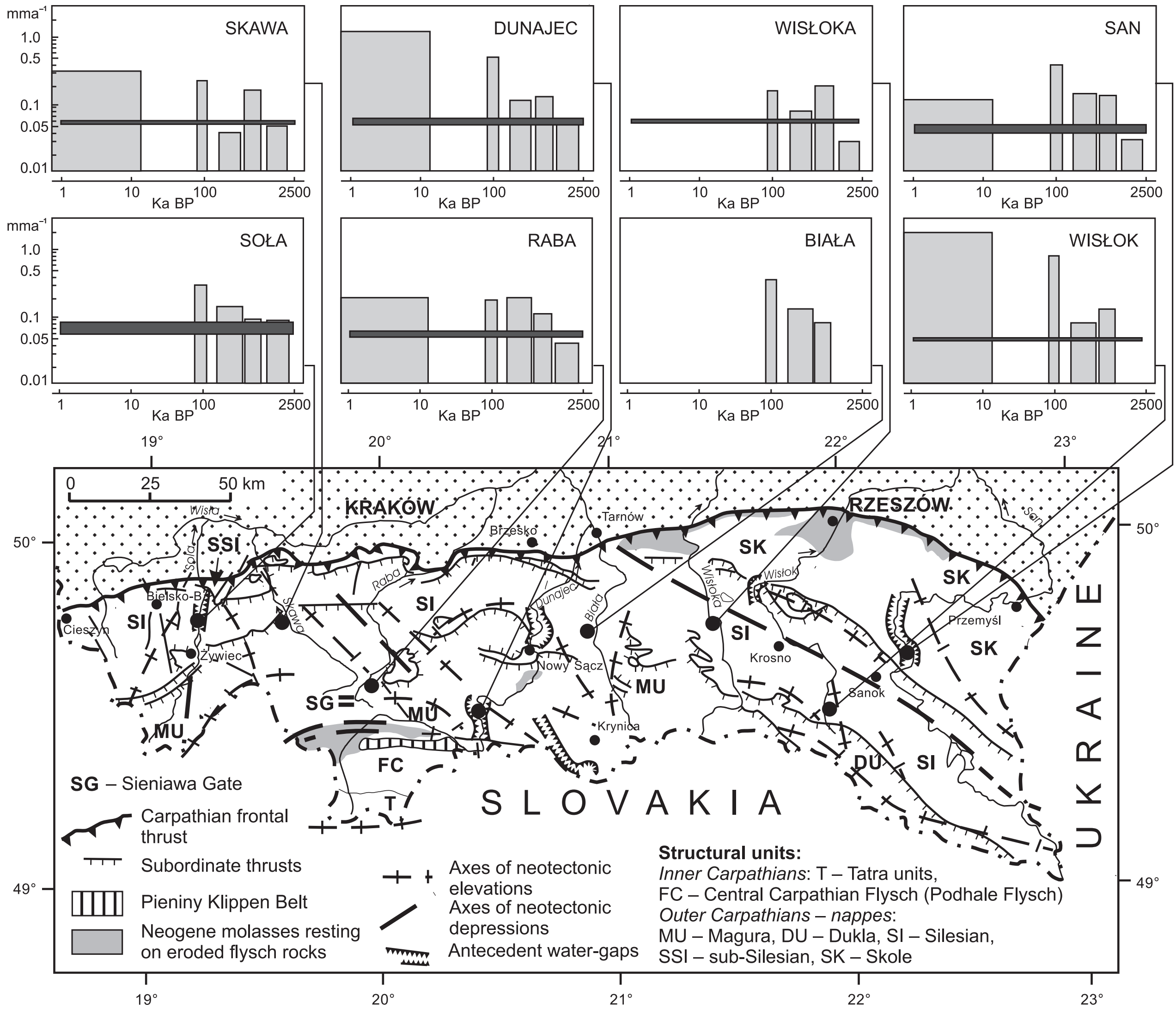
General features

The Outer Carpathian stack of nappes witnessed differential uplift in the Pliocene and Quaternary (Zuchiewicz et al. 2002 and references therein) (Figs 4, 5). The uplift in Pliocene times was probably of block-type in the western segment of the belt, whereas that in the medial and eastern segments was restricted to subparallel and relatively narrow upwarped zones. The total size of uplift, approximated by the amount of erosional dissection of the Plio-Pleistocene planation surfaces and straths of Quaternary terraces, varied from 150 m to ca. 900 m, averaging at around 300 m. The uplift could have been a result of both post-orogenic isostatic rebound related to erosional unroofing, particularly intensive in the western part of the area, and of the steepening of frontal parts of some nappes.

Episodes of intensified Quaternary uplift, restricted to relatively narrow zones oriented subparallel to the structural grain of the area and coinciding with frontal parts of overthrust nappes and larger slices, occurred during the Cromerian-Elsterian 1/2, Eemian-early Weichselian and Weichselian Late Glacial-Holocene times, at rates varying from 0.15 to 2.0 mm/yr (Fig. 5). These long (100–250 km) and narrow (15–25 km) zones of localized uplift appear to be a result of the Quaternary relaxation of horizontal stresses (cf. Zuchiewicz 1998 and references therein).

The available pieces of structural evidence imply that during the Late Neogene times structural development of the Polish segment of the Outer Carpathians was controlled by normal faulting. This interpretation is corroborated by geomorphic data indicative of *en block* uplift in the western part of the belt. However, there is no unequivocal evidence to decide whether the faulting was due to successive phases of alternating N-S and E-W extension or owing to one or more phases of multidirectional extension. Moreover, the geomorphic data from the medial and eastern parts of the belt suggest the occurrence of compressional stress regime during Pliocene times (cf. Zuchiewicz et al. 2002).

The data available for Quaternary times show an apparent contradiction. On one hand, different pieces of geomorphic evidence imply compressional stress arrangement, with the maximum compressive stress being oriented roughly perpendicular to the belt. This interpretation is compatible with the present-day orientation of the S_{Hmax} inferred from the breakout analysis and from focal solutions of the Krynica earthquakes (Fig. 6) (cf. Jarosiński 1998, Zuchiewicz et al. 2002). The zones showing tendencies to Recent uplift tend to be aligned subparallel to frontal thrusts of individual nappes and larger slices, suggesting the presence of Plio-Quaternary horizontal stresses in the flysch nappes. En echelon arrangement of theses zones, however, slightly different in the western and eastern parts of the study area, appears to indicate young sinistral motions along the Kraków-Lubliniec fault zone (Fig. 6) in the substratum of the overthrust nappes (Zuchiewicz 2001).



■ **Fig. 5.** Neotectonic sketch map of the Polish Carpathians (based on Zuchiewicz 1998, Zuchiewicz et al. 2002). Diagrams illustrate rates of Quaternary river dissection in those segments of river valleys which are located in neotectonically uplifted structures. Solid lines denote average rates of Quaternary dissection.

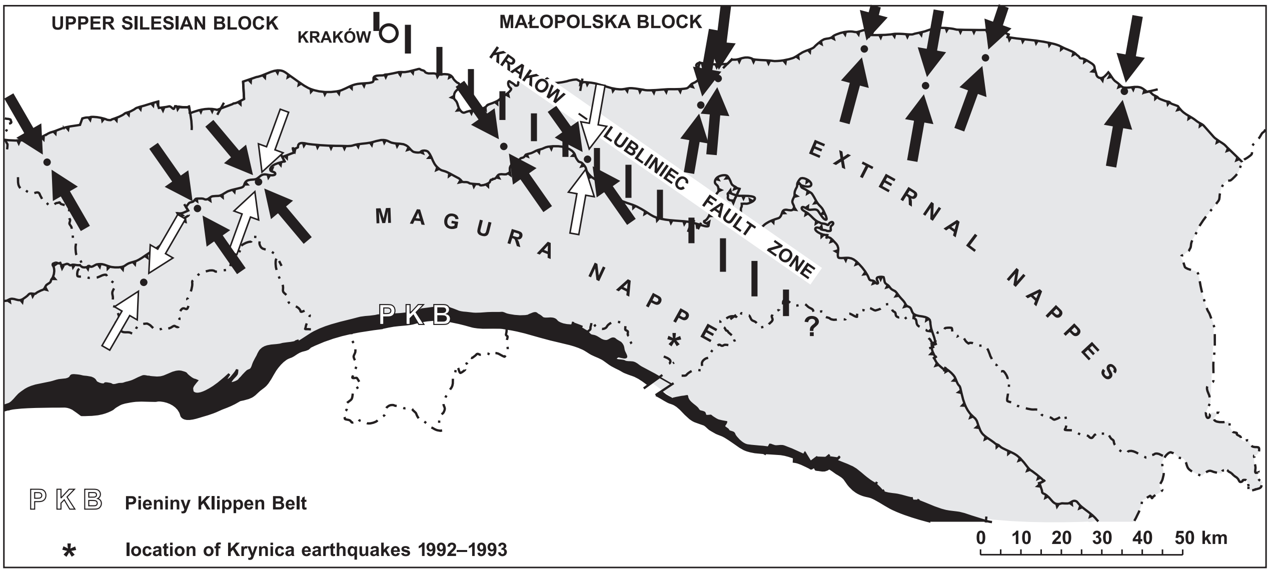
On the other hand, Quaternary normal faulting within the intramontane basins and in localized narrow zones of frontal parts of nappes and larger slices points to extensional stress arrangement. This contradiction can be explained by a concept of normal faulting restricted to certain zones affected by horizontal stresses. These processes were probably not uniform, as shown by differentiated rates of erosional dissection of Quaternary straths in individual geomorphic units within different Quaternary stages (Figs. 5, 7). Another, although not contradictory explanation, lies in the general isostatic post-orogenic uplift, being overprinted by coeval horizontal motions within the flysch cover.

The rates of recent vertical crustal motions in the Polish Outer Carpathians range between 0 mm/yr in the western and medial segment to ca. +1 mm/yr in the east (Wyrzykowski 1985), whereas those in the Pieniny Klippen Belt do not exceed 0.5 mm/yr (Ząbek et al. 1993, Czarnecki 2004). The results of recent GPS campaigns (Hefty 1998) point as well as borehole breakout analyses (Jarosiński 1998), in turn, to NNE-directed horizontal motions throughout the area (cf. Fig. 6).

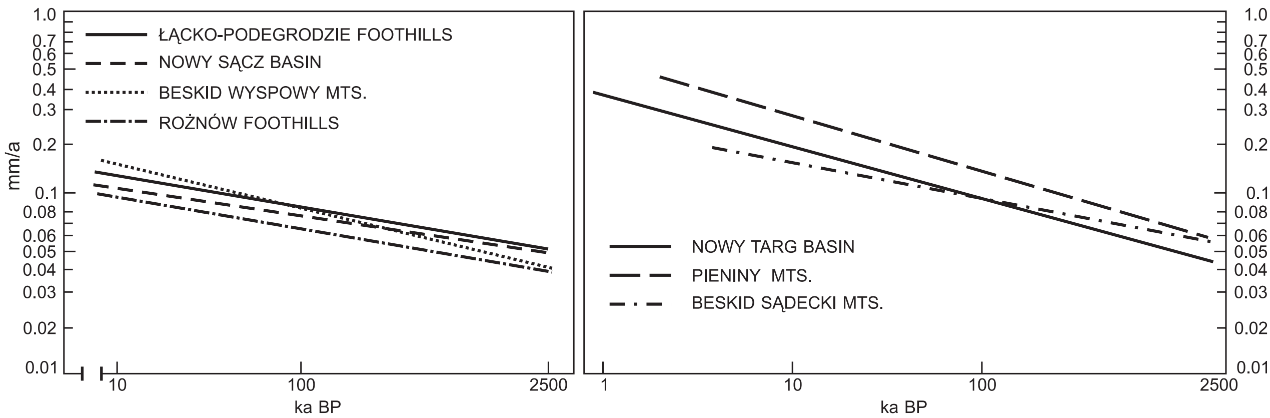
Magura Nappe

The apparent amount of Late Neogene-Quaternary uplift of the medial segment of the Magura Nappe has ranged from 150–430 m in the south and 360–900 m in the medial sector, to 180–310 m in the north. The size of purely Quaternary uplift of the southern part of the Magura Nappe has been estimated for some 150 m (cf. Zuchiewicz 1995).

A comparison between the pattern of elevated and subsided structures of the Magura floor thrust and the results of morphotectonic studies shows that in the western part of the Polish Outer Carpathians (Figs. 8, 9) the highest-elevated neotectonic structures (in the southern portion of that area) coincide with depressions of



■ **Fig. 6.** Mean SHmax directions obtained from breakouts in the autochthonous basement (black) and the Carpathian nappes (white) (based on Jarosiński's data in Zuchiewicz et al. 2002 simplified).



■ **Fig. 7.** Rates of erosional dissection of straths in different physiographic units of the medial portion of the Polish Carpathians (based on Zuchiewicz 1998).

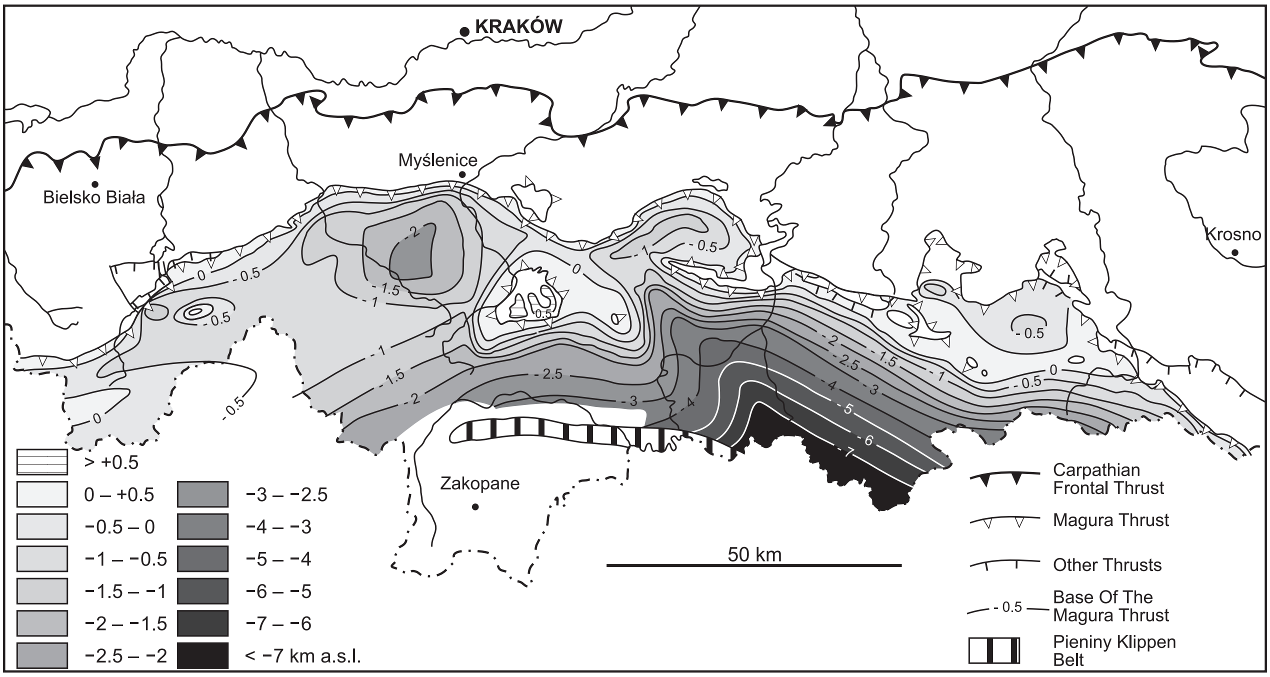
the Magura floor thrust, whereas farther north a reverse pattern becomes dominant: neotectonic elevations coincide either with exposures of the Magura frontal thrust or with elevations of its surface (Fig. 8). This is particularly true for an area comprised between 20° and 20° 30′E meridians. Moreover, the strongly uplifted region in this part of the Outer Carpathians is situated shortly south of the main elevation of the Magura floor thrust, represented by the Mszana Dolna tectonic window (Fig. 9).

The origin of such relationships is difficult to explain. We infer that one of possible factors could be Pliocene-Quaternary reactivation of faults cutting the Magura floor thrust, as well as the basal thrust of the Outer Carpathians, and particularly that which appears to separate the western-medial segment of the Outer Carpathians from the more eastern portion. A similar conclusion has been proposed by Zuchiewicz et al. (2002) when analysing morphometric parameters of uplifted neotectonic structures in the western and eastern portions of the Outer West Carpathians, whose en echelon arrangement is different on either side of the reactivated, probably sinistral, Kraków-Lubliniec deep fault zone (Fig. 6), located beneath the overthrust nappes. For more details see **Stop 3**.

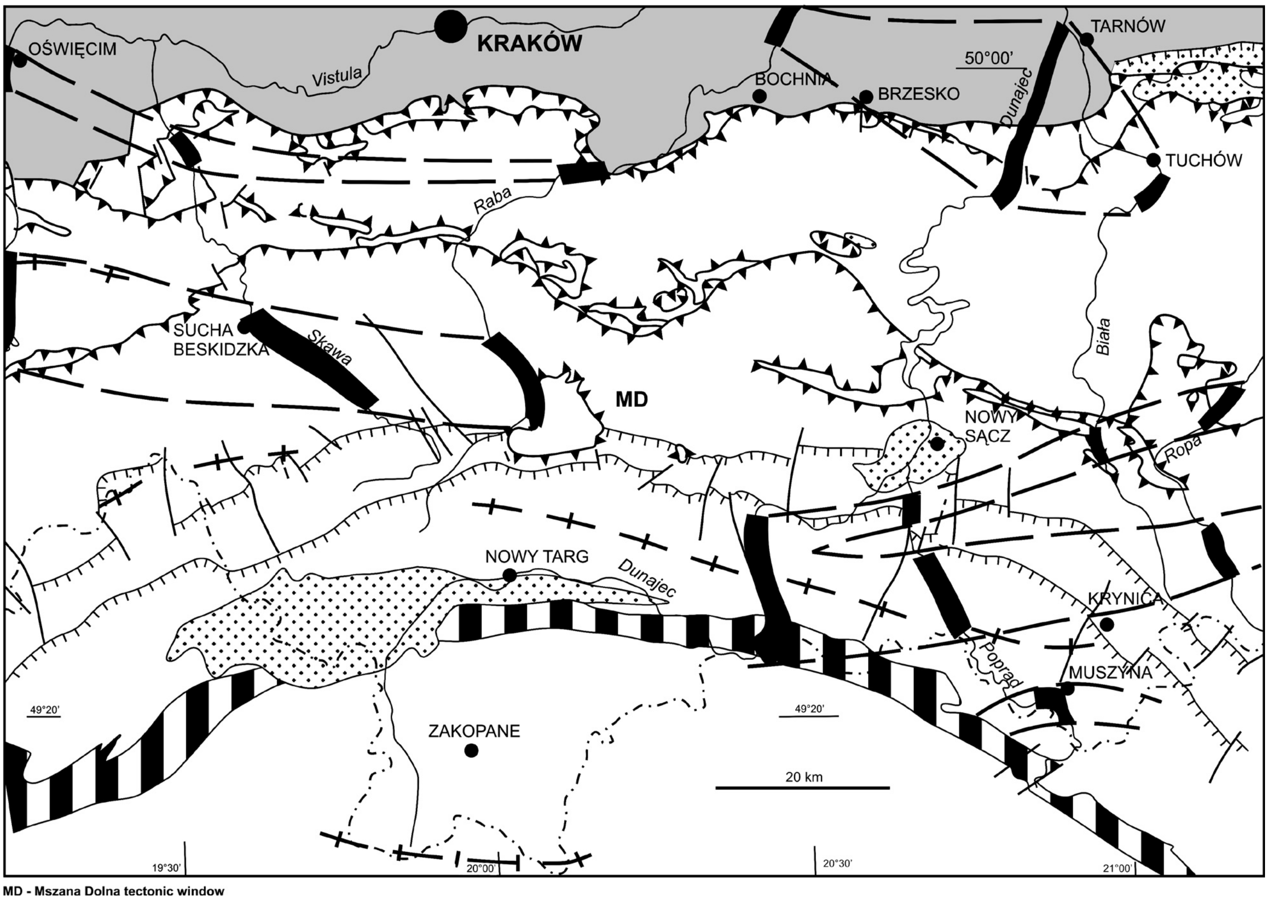
Fractured clasts

Analysis of fractured clasts in gravels and conglomerates has commonly been applied to palaeostress studies and dating of faulting during the past few tens of years (for review see: Tokarski and Świerczewska 2005). In the Polish segment of the Outer Carpathians, fractured clasts are fairly numerous in young Cenozoic (Miocene through Holocene) gravels and paraconglomerates which crop out close to regional overthrusts and faults (Fig. 1B), pointing to recent activity of these features. So far, the kinematics of this activity is poorly understood (e.g. Tokarski et al. submitted).

Fracture architecture in fractured clasts is either well-organized or chaotic. It is likely that only well-organized fracture architecture is suitable for kinematic analysis. The aim of our studies



■ **Fig. 8.** Topography of the Magura floor thrust (based on unpublished Oszczypko's data).



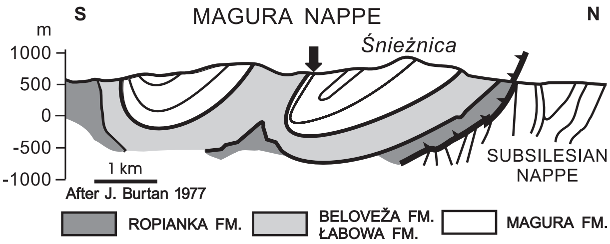
■ **Fig. 9.** Neotectonic sketch of the medial portion of the Polish Carpathians. Black vertical segments – Pieniny Klippen Belt, dots – Late Neogene molasses resting on eroded flysch units, shaded elongated segments – segments of river valleys showing abnormally high river-bed gradients, thick fence lines – neotectonically elevated areas.

has been to describe the influence of textural properties and lithology of gravels and paraconglomerates on fracture architecture, and to make an attempt at kinematic analysis of fractures. Up to now, we managed to document that: (a) the number of fractured clasts in a given clast population is positively correlated with the clast diameter and negatively correlated with the size of grains within clasts of detrital rocks; (b) the fractures include both those inherited after joints cutting host strata in the source areas, and those formed *in situ* in the studied rock series; (c) the inherited fractures, showing chaotic architecture, are mainly oriented at right angle or nearly perpendicular (80–90 °) to clast a-b surfaces; (d) neofractures formed *in situ*, showing well-organized architecture, are oriented both perpendicular (80–90 °) and at smaller angles (<80 °) versus a-b clast surfaces. For more details see **Stop 2**.

Stop 1. Deformation Bands in Eocene Sandstone, Gruszowiec

Description

Small, abandoned quarry in the Gruszowiec Village (Fig. 1C), located 300 m from the "Pod Cyckiem" (Under the Tit) bar. Thin- to thick-bedded sandstones intercalated with mudstone and claystone are exposed in the quarry. The strata belong to the Late Eocene Poprad Sandstone Member of the Magura Formation within the Rača slice (Fig. 1B). The exposure is situated in the overturned limb of the Śnieżnica syncline (Fig. 10). Four groups of planar minor structures have been observed (Fig. 11A): (1) water-escape sheets, (2) deformation bands (DB), (3) joints, and (4) minor faults. Water-escape sheets (1) and DB (2) were observed only within sandstones, whereas joints (3) and faults (4) also cut mudstones and claystones.



■ **Fig. 10.** Cross-section (after Burtan 1977) of the Śnieżnica syncline showing tectonic position of the discussed exposure. For location see Fig. 1C (stop 1).

(1) Water-escape sheets extend upwards from the bases of sand--stone beds. Some of the sheets root in wedge-shaped injections penetrating from the underlying mudstone. The water--escape sheets are oriented subvertical and cross-fold.

(2) Eighty-four DB have been studied in thin sections. The majority of DB are oriented fold-parallel, a minor number of bands shows of cross-fold orientation, and very few DB are oriented obliquely to the fold (Fig. 11B). Following the classification by Antonellini et al. (1994), three types of DB have been distinguished: (1) deformation bands with no cataclasis (NDB), (2) deformation bands with traces of feldspar cataclasis (TDB) and, (3) deformation bands with strong feldspar cataclasis (SDB). All types of DB contain traces of calcite cement, but we have not observed any calcite cataclasis. In the host rocks distribution of calcite cement is not uniform in relation to DB. Usually, DB separate areas showing different degree of cementation.

Inside the NDB and the TDB, mica flakes and elongated grains are oriented parallel to the band boundaries. The grain size, compared to the host rock, is only slightly reduced in the NDB, whereas it is reduced by factors between 5 and 10 in the SDB. The boundaries with the host rock are transitional for the NDB and clear cut for the SDB. The latter contain more matrix than the host strata whereas only local slight enrichment in matrix occurs in the former. The microstructures of the TDB are transitional between the NDB and SDB. The cross-fold DB are exclusively NDB. Occasionally, these bands cut fold-parallel NDB. Some of the cross-fold DB pass laterally into water-escape sheets. Some of the NDB pass laterally into the SDB. Numerous SDB cut NDB.

In all cases where the sense of shear can be determined, the observed offset along the fold-parallel DB is reverse. The fold-parallel DB display different dip angles ranging from subhorizontal to subvertical (Fig. 11C). Steep dips prevail for the NDB whereas the majority of TDB and SDB display shallow dip. Numerous DB form conjugate sets which intersect one another at 30–60 ° (Fig. 11A). The orientations of the bisecting planes of conjugate bands change from bedding-parallel to bedding-perpendicular.

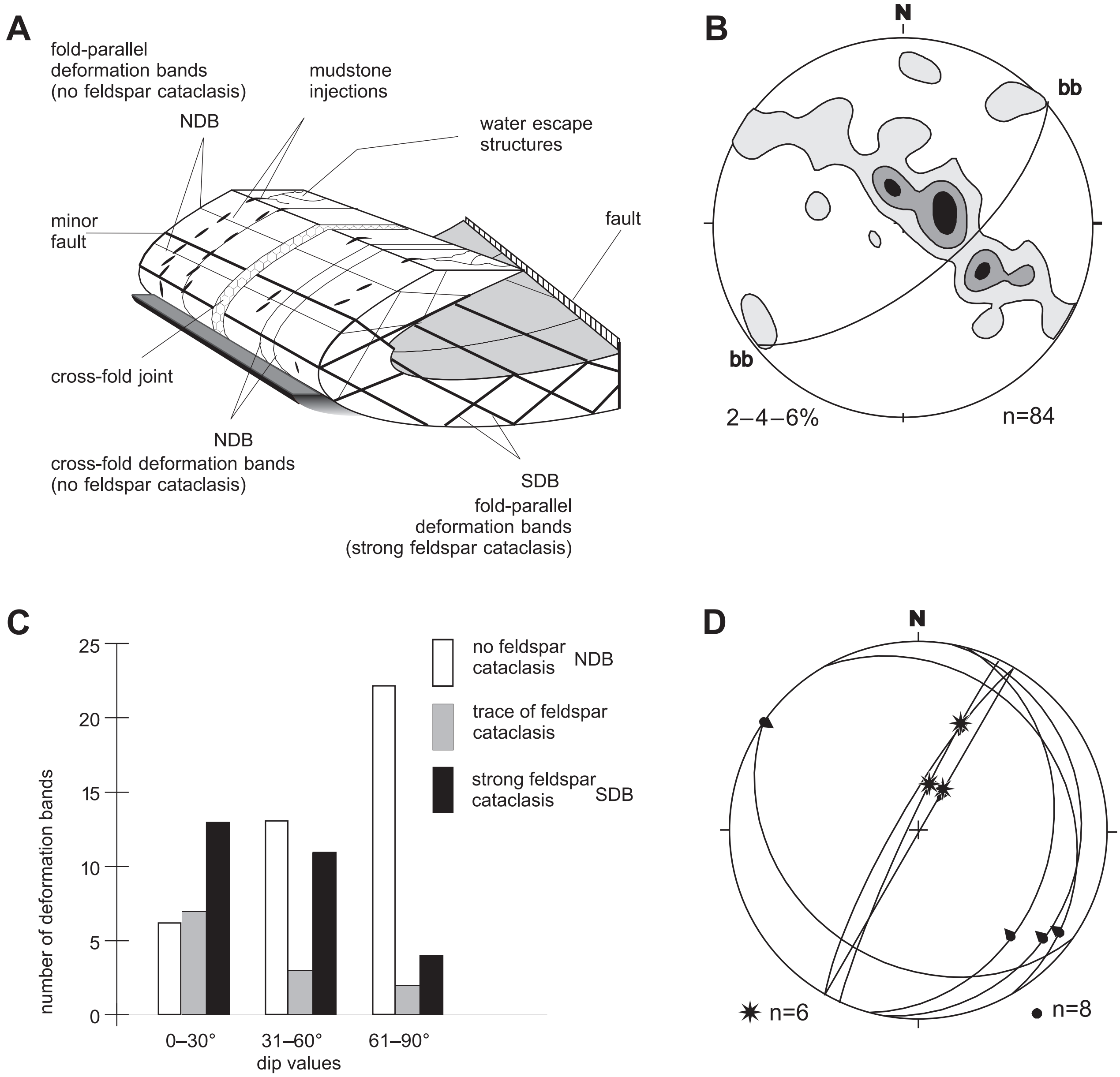
(3) Joints are sub-vertical and cross-fold (Fig. 11A). Some of the joints are filled by calcite veins up to 10 cm thick. In the host strata, close to the calcite-filled joints, the content of calcite cement diminishes abruptly away from joints.

(4) Minor faults are lined by breccia zones, up to 0.5 mm thick. The boundaries of these zones are clear-cut. In distinction to the DB, the breccia zones lining minor faults are devoid of mica flakes and display quartz and calcite cataclasis. The minor faults studied comprise shallow-dipping reverse faults, and subvertical, strike-subparallel, oblique faults (Fig. 11D).

Discussion

The orientation of water-escape sheets, joints and DB are either fold-parallel or cross-fold. This indicates that formation of these structures was related to regional folding.

Water-escape sheets are the features which form penecontemporaneously with sedimentation. This indicates that regional folding was already initiated during sedimentation of the strata involved (Fig. 11A). Some of the water-escape sheets pass laterally into cross-fold DB, whereas some of the latter cut some of



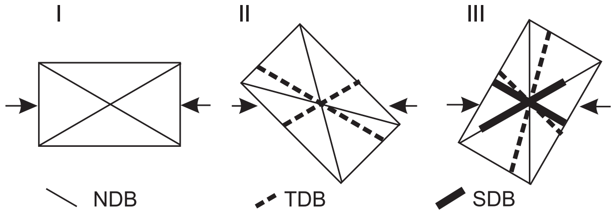
■ **Fig. 11.** Minor planar structures at the Gruszowiec exposure. A – model showing attitudes of structures in different structural positions; B – stereoplot of deformation bands; C – histogram of DB type versus DB dip; D – stereoplot of minor faults.

the fold-parallel DB. These relationships indicate that both cross-fold and fold-parallel DB also started to form during sedimentation. This conclusion is corroborated by the nature of the cross--fold DB which are exclusively NDB.

The fold-parallel DB present the whole spectrum of orientations, from bedding-parallel DB to bedding-perpendicular ones. They display reverse offset and at least some of them can be grouped into conjugate sets. We believe that all fold-parallel DB were formed as conjugate shear failures. In our interpretation (Fig. 12), these shear failures were occurring progressively during folding, becoming more and more steeply inclined to the bedding. Several fold-parallel NDB display shallow dips (Fig. 11C), indicating that they formed when the strata attained subvertical orientation close to the completing of folding. It follows that folding was completed when the host strata were still poorly indurated.

The number of fold-parallel TDB and SDB is negatively related to their dip angles. It follows that the majority of these bands formed when the strata were oriented subvertical, at least some of them as conjugate shear failures. The NDB are commonly cut by SDB, indicating that fold-parallel SDB which are steeply inclined with respect to bedding formed after completing of folding, when the host strata became more indurated than during the formation of the NDB.

We believe that those SDB which display steep dips also formed after completing of folding, resulting from reworking of the NDB. The observation that some NDB pass laterally into SDB corroborates this opinion. Not a single DB displays calcite cataclasis. This indicates that the calcite mineralization was introduced into host strata and into DB only after formation of SDB and, therefore, after completing of folding. Distribution of calcite cementation vs. DB indicates that DB, probably due to their low permeability, formed barriers to calcite-bearing fluids. It appears that this mineralization was introduced along the cross-fold joints (Fig. 11A). The studied minor faults display calcite cataclasis. It follows that the faulting occurred after introducing of calcite mineralization into the host rock and therefore after completing of folding.



■ **Fig. 12.** Cartoon showing the succession of appearance of particular DB types, I – before folding, II – during folding, III – after folding.

Conclusions

The above data provide good evidence for progressive pre-, syn- and postlithification of small-scale tectonic deformations. Microstructural studies indicate that deformation started in loose sediment as hydroplastic features and lasted until the host strata became fully indurated.

(1) The formation of DB was controlled by regional stress field in a compressional regime. Distinct types of DB formed progressively during and after regional foldind, during increasing induration of the host strata. Deformation bands with no cataclasis (NDB) started to form when the host strata were still in horizontal position and continued to evolve until the completion of regional folding. Deformation bands with traces of feldspar cataclasis (TDB) and bands with strong feldspar cataclasis (SDB) formed during and after regional folding.

(2) The folding occurred when the host strata were poorly indurated and possibly under low confining pressure. The folding took place during deposition of the host strata.

(3) At the completion of regional folding, the host strata were still poorly indurated. Introduction of calcite mineralization, which occurred along the joints, post-dated folding.

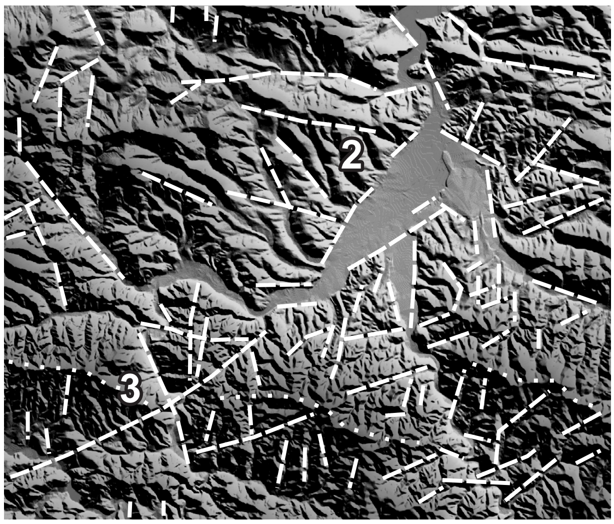
(4) Within the studied portion of the Magura Nappe, regional folding started not later than during the Eocene. Folding was completed when the host strata still were not fully indurated, most probably also during the Eocene.

Stop 2. Fracured Clasts in Miocene and Pleistocene Paraconglomerates, Brzezna

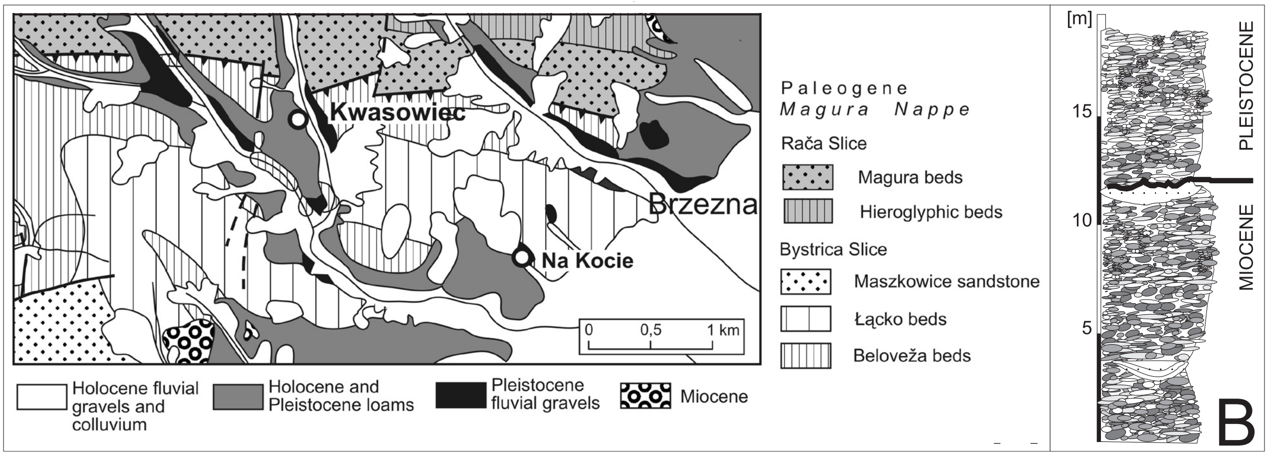
Natural exposure at the “Na Kocie” (Over the Cat) site is located in Brzezna village (Figs. 13, 14A). The choice of study object results from its interesting tectonic setting. This site is situated close to a regional thrust fault on the margin of the Nowy Sącz Basin, which is also tectonically controlled. In addition, the analysed strata bear a wealth of lithologically differentiated fractured clasts, and are disturbed by both small-scale (individual clasts) and larger-scale tectonic deformations. Finally, site “Na Kocie” is placed not far from an exposure of Pleistocene gravels at Kwasowiec, recently described by two of us (Tokarski and Świerczewska 2005). The latter gravels, showing poor roundness measures and monotonous lithology and bearing numerous fractured pebbles, provide good comparative material.

The studied exposure is situated in the hanging wall of the Bystrica slice thrust over the Rača slice, ca. 1,200 m away from the thrust surface. The present-day erosional margin of the Nowy Sącz Basin, coinciding with the extent of the Neogene infill, is placed 300 m from “Na Kocie” exposure.

The analysed rock series is exposed in a 19-m-high, undermined slope of a stream valley. The lower part, 11 m high (Fig. 14B), is composed of paraconglomerates bearing sandstone intercalations. This complex is of Late Badenian and/or Early Sarmatian age (Oszczypko et al. 1991 and references therein). The Miocene complex is unconformably overlain by 8-m-thick Pleistocene paraconglomerates dated to the Elsterian-2 (Butrym et al. 1989, Osz-



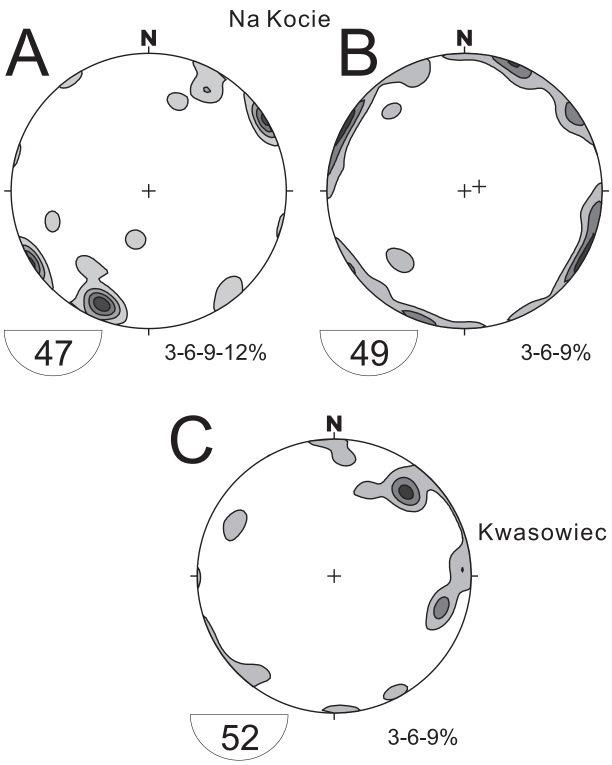
■ **Fig. 13.** Digital elevation model of the medial segment of the Polish Outer Carpathians, showing location of Brzezna (Na Kocie; NK) and Tylmanowa (TY) stops. Dashed lines refer to well-marked topolineaments, dotted line marks the axis of the most important neotectonic elevation of this region. Note a well-pronounced fault on the NW margin of the Nowy Sącz Basin and intersecting topolineaments close to Tylmanowa. For location see 1C (stops 2 and 3).



■ **Fig. 14.** A – study area showing the location of "Na Kocie" and "Kwasowiec" exposures (geology based on Oszczypko and Wójcik 1989); B – section of "Na Kocie" exposure; for location see Fig. 13.

czypko et al. 1992). Both complexes are separated by an erosional surface.

The clasts in the Miocene paracoglomerate are composed exclusively of rocks occurring in the Magura Nappe.The clasts are orderly arranged. Clast "a" axes plunge gently (<30°) to the NNW and SSE, less frequently to the NNE and SSW. The lower portion of the Miocene complex bears a lens-like intercalation of sandstones alternating with siltstones and claystones.

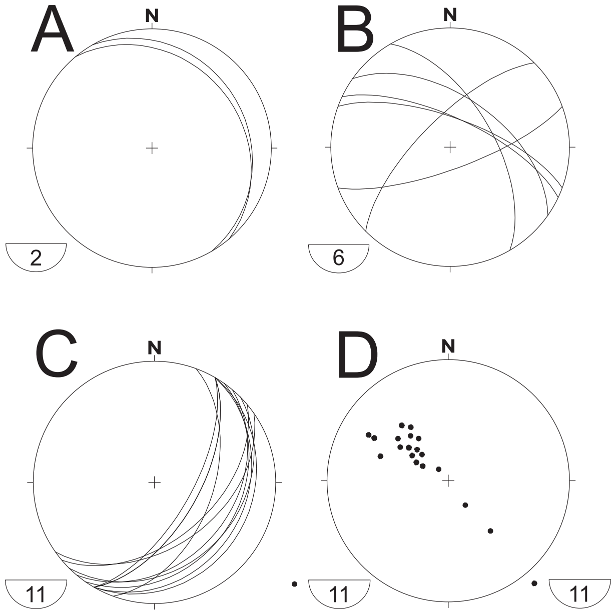


■ **Fig. 15.** Orientation of clast-cutting fractures: A – Miocene conglomerate at exposure "Na Kocie"; B – Pleistocene conglomerate at exposure "Na Kocie"; C – Pleistocene gravels at Kwasowiec.

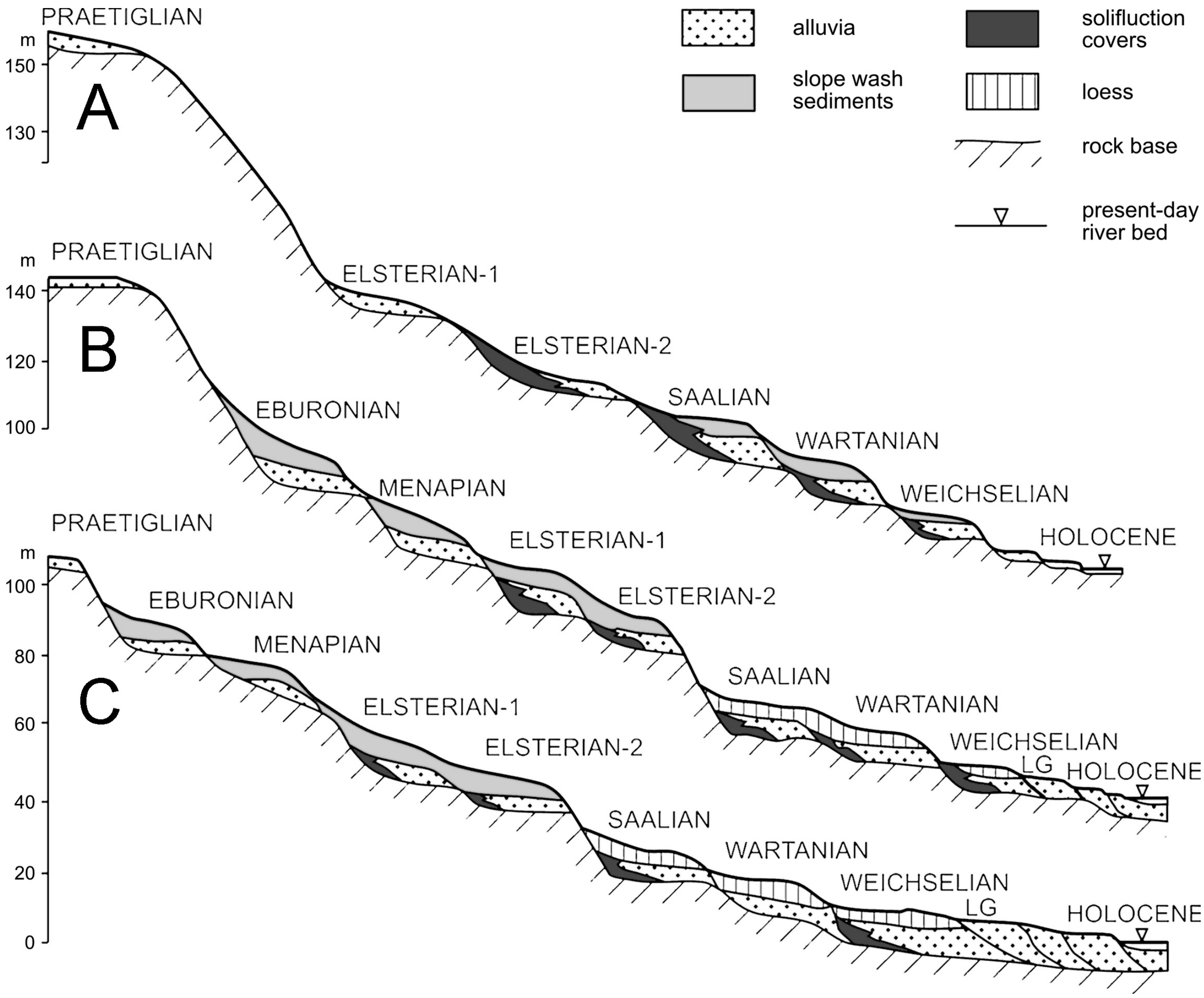
The clasts in the Pleistocene paraconglomerate are derived both from the Magura Nappe and from the Inner Carpathians (cf. Butrym et al. 1989, Oszczypko et al. 1992). The clasts are arranged orderly. The majority of clast "a" axes plunge gently (<30°) towards the west and south.

Fractured clasts

Paraconglomerates of both complexes bear numerous clasts cut by one fracture and some clasts cut by several fractures. These fractures are restricted to single clasts; the matrix is not fractured.



■ **Fig. 16.** A, B – normal faults cutting clasts within conglomerates of Miocene (A) and Pleistocene (B) age; C,D – structures comprised within sandstone intercalation in the lower part of the Miocene complex: A – reverse faults; B – poles to reverse faults (asterisks) and bedding surfaces (dots).



■ **Fig. 17.** Flights of Pleistocene and Holocene straths and complex-response terraces in three physiographic units dissected by the Dunajec River : A – Beskid Sądecki Mts. (site Tylmanowa), B – southern portion of the Beskid Wyspowy Mts. (Łącko – Podegrodzie Foothills), C – Nowy Sącz Basin (see Fig. 4 for location). LG – Weichselian Late Glacial.

In the Miocene conglomerate, a population of 100 large clasts (>2 cm) comprises 50 % of fractured clasts, and 50 % of unfractured clasts, while in the Pleistocene conglomerate, the respective share of these groups is 45 %, and 55 %. In the Miocene conglomerate, a population of 50 small clasts (<2 cm) contains 22 % of fractured clasts, and 78 % of unfractured clasts, and in the the Pleistocene conglomerate, these groups amount to 26 % and 74 %, respectively. Different lithologies represent different numbers of fractured clasts; clasts of detrital rocks tend to have more fractures compared to quartzites and magmatic rocks.

Clast architecture is well-organized and similar in both conglomerates (Fig. 15A, B). Most of the fractures are arranged subvertically, and tend to form two sets oriented NW and NE. The proportion of fractures situated at right angle and nearly perpendicular (80–90 °) to clast a-b surfaces amounts to 37 % and 50 % in the Miocene and Pleistocene conglomerates, respectively. In both conglomerates, infrequent clasts are cut by normal faults (Fig. 16A, B) which do not pass into the matrix.

Structures within sandstones

A sandstone intercalation exposed in the lower part of the Miocene complex is cut by numerous reverse faults of (Fig. 16C). Poles to both fault planes and bedding surfaces plot on a single great circle, whose axis is horizontal and oriented NE-SW (Fig. 16D). Joints cutting this sandstone body are oriented both perpendicular (NW) and parallel (NE) to the reconstructed axis of the great circle. Identically oriented fractures (NW and NE) cut a pebble situated in the lower portion of this body.

Interpretation

The architecture of clast-cutting fractures is well organized and comparable in both conglomerates (Fig. 15A, B), implying that fractures in the two complexes were formed *in situ* and are of the same origin.

We infer that the sandstone intercalation in the lower part of the Miocene complex originated during deposition of the latter as a channel fill which was later folded. This is testified to by the position of poles to bedding on a single great circle (Fig. 16D), the axis of which is horizontal and oriented NE-SW. The same great circle bears poles to reverse faults which cut the rock body in question. It means that both folding and reverse faulting took place in a compressional stress field of σ1 oriented NW-SE.

Most of clast-cutting fractures in the two conglomerates are subvertical and clustered into two sets oriented NW and NE. We conclude that these fractures represent joints formed in a compressional stress field of σ1 oriented NW. The NW-striking fractures are extensional features parallel to σ1/σ2 plane, whereas NE-striking fractures represent joint surfaces formed during stress relaxation (cf. Caputo 1995). In addition, fractures cutting clasts in the two conglomerates are parallel to joints within sandstone intercalation in the lower part of the Miocene complex. This implies that all these structures, except for normal faults, originated in the same stress field.

Discussion and conclusions

(1) In both complexes, the number of fractured clasts is positively correlated with the clast size, and negatively correlated with the diameter of grains in clasts of detrital rocks. The number of fractured clasts increases in clasts of detrital rocks, compared to those of quartzites and magmatic rocks. These data point to the role of clast size and lithology in controlling the number of fractured clasts, as already shown by Tokarski and Świerczewska (2005).

(2) Clast-cutting fractures are well organized (Fig. 15A, B), irrespectively of the orientation of clast "a" axes which are different for each complex. This means that such fractures were formed *in situ* and that clast orientation does not influence the bearing of fractures.

(3) The two conglomerates studied at "Na Kocie" exposure bear clasts cut by fractures oriented both at right angle or nearly perpendicular (80–90°) and at smaller angles (<80°) to the clast a-b surfaces. This observation supports our previous conclusion (Tokarski and Świerczewska 2005) that such an orientation of fractures can be diagnostic for fractures formed *in situ* in the studied gravels and conglomerates.

(4) Fractures cutting clasts within the conglomerates represent extensional joint fractures that were formed in a compressional stress field, of σ1 oriented NW-SE. The sandstone intercalation in the lower part of the Miocene complex was folded and reverse-faulted in the same stress field. It is probable that the origin of these deformations was related to neotectonic activity of the Bystrica thrust. Similar architecture of joints cutting clasts in the Pleistocene fluvial gravels at Kwasowiec (Fig. 15C) appears to suggest that this activity must have been of a wider extent. In contrast, the origin of normal faults cutting clasts of both complexes was probably different and associated with neotectonic activity of the margin of the Nowy Sącz Basin.

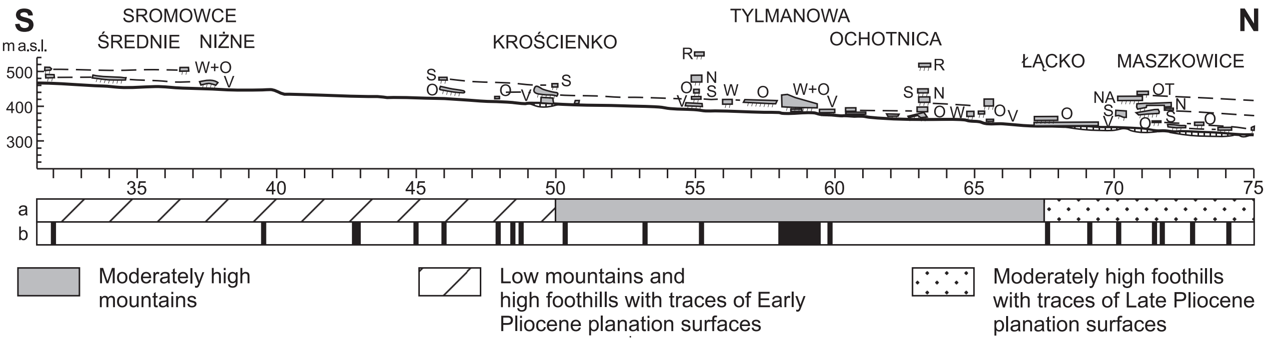
Stop 3. Uplifted Terraces in the Antecedent Dunajec River Gorge, Tylmanowa

Site Tylmanowa is situated in an antecedent gorge of the Dunajec River, dissecting the Beskid Sądecki Mts. Another gorge of this type cuts the Pieniny Mts., a few kilometres farther south. The "Beskid Sądecki" gorge segment includes two deeply cut meanders which are separated by a rectilinear valley, parallel to a fault line. This area is situated at a place of intersection of NNW, NE, and N-S striking topolineaments (Fig. 13).

The gorge is 15 km long and up to 700 m deep, its width changing from 75–100 m within the meanders to 450–500 m in other segments. The river-bed is cut into solid bedrock and its long profile is ungraded and of exceptionally high gradient, compared to the upstream and downstream valley reaches (cf. Fig. 9). The eastern valley sides are steep (50–66 %) and dissected by a network of short (up to 1.5 km) and high-gradient (200 %) minor tributary valleys and ravines. Outlets of tributary valleys are usually hanging over the present-day river bed, up to 10–15 m. Headwater parts of some of these tributaries represent hour-glass valleys. The surrounding ridges bear traces of four planated surfaces that rise 900 m, 770–830 m, 500–590 m, and 450–500 m a.s.l.

The eastern valley sides are mantled by weathering debris and loams, while the western ones are dominated by a flight of straths and complex-response terraces (Fig. 17), whose alluvial covers were deposited during the Pleistocene glacial stages: Praetiglian (150–155 m to 154–161 m), Elsterian-1 (75–84 m to 78–96 m), Elsterian-2 (51–55 m to 52–65 m), Saalian (26–41 m to 29–41 m), Wartanian (17–24 m to 20–31 m), and Weichselian (10–11 m to 16–18 m), as well as in the Holocene (6–10 m, 4–5 m, 2–3 m). The thickness of terrace alluvium is between 3–4 m and 10–14 m (cf. Zuchiewicz 1995 and references therein). These covers are composed of poorly rounded and poorly sorted, both Outer Carpathian flysch (sandstones, siltstones, rare conglomerates) and Tatra-derived (granites, quartzites) gravels and cobbles. Limestones shed from the Pieniny Klippen Belt can only be found within the youngest, i.e. Weichselian and Holocene alluvium; limestone clasts of older fluvial series became completely dissolved. The Early and Middle Pleistocene covers include a large proportion of angular clasts, pointing to the role of intensive solifluction within glacial stages. All Pleistocene fluvial covers interfinger with solifluction tongues, those dated to the last and penultimate glacial stages being also overlain by slopewash and/or solifluction-slopewash covers, which are 3–8 m thick. Such interfingering enables for relative dating of the preserved terrace covers to individual glacial ages.

Long-profiles of Pleistocene straths (Fig. 18) clearly show increased relative heights of the latter within meander loops. These



■ **Fig. 18.** Longitudinal profile of Pleistocene terraces of the Dunajec River valley between the Nowy Targ Basin and the southern portion of the Beskid Wyspowy Mts., showing location of site Tylmanowa. Symbols: a – landscape types, b – fault zones within bedrock. Grey patches on the profile refer to alluvial covers, barbed lines mark straths of individual terraces. Pleistocene glacial stages: R – Różce (Praetiglian), OT – Otwock (Eburonian?), NA – Narew (Menapian?), N – Nida (Elsterian-1), S – San (Elsterian-2), O – Odra (Saalian), W – Warta (Wartanian), V – Wisła (Vistulian, Weichselian). Numbers below the profile refer to distance (in kilometers) from Nowy Targ.

heights are greatest within the entire Polish segment of the Outer Carpathians; straths of equivalent age within the remaining Dunajec River valley reaches, and those of other Carpathian river valleys are lower, even by 30 m in case of the oldest Pleistocene straths (cf. Zuchiewicz 2001 and references therein). Deformations of Pleistocene straths combined with intense erosional downcutting appear to indicate Pleistocene uplift of the axial part of the Beskid Sądecki Mts., part of the most strongly elevated neotectonic structure in the Polish Outer Carpathians (cf. Starkel 1972, Zuchiewicz 1995, cf. also Figs. 4, 9, 13). The amount of fluvial accumulation throughout the Middle and Late Pleistocene in the Dunajec River gorge can be estimated at 555×10^6 cubic metres, that removed by erosion attaining ca. 652×10^6 cubic meters. The lack of fluvial covers from the Eburonian and Menapian stages points to intense, tectonically-controlled, erosion before the Elsterian; the equivalent-age terrace covers are to be found immediately north of the gorge (Fig. 18).

Stop 4. Joints and Mineral Veins in Paleocene Sandstone, Krościenko

Description

Natural exposure of the Paleocene-Lower Eocene Szczawnica Formation on the Dunajec River bank at Krościenko town (Fig. 1C) is the site where the model of jointing has been tested.

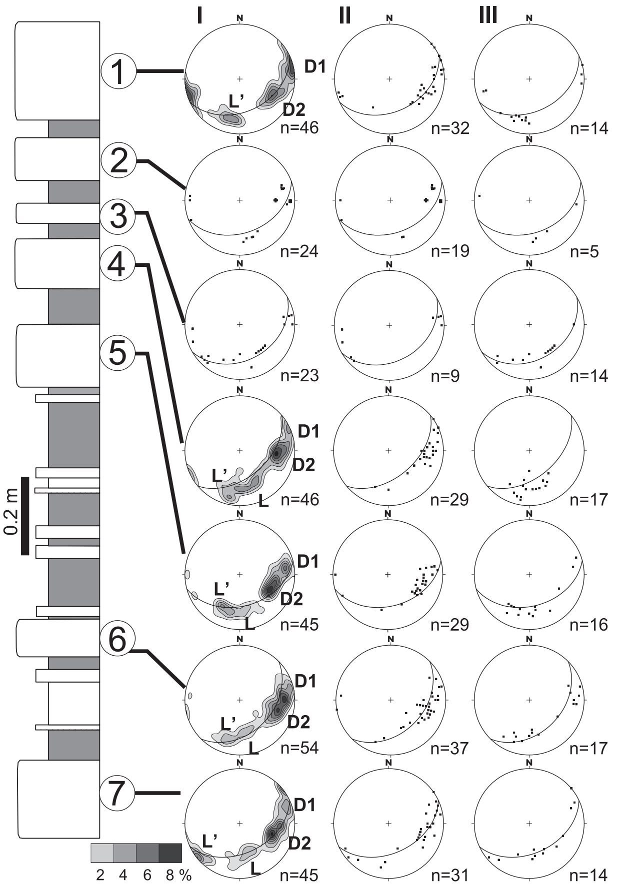
Our study focused on mineralized joints and small-scale faults that cut sandstone beds in a sandstone-mudstone-claystone sequence, 2 m thick (Fig. 19). Altogether, attitudes of 283 joints and 15 faults were measured, and all joints, mineral veins and small-scale faults cutting 7 sandstone beds were examined in detail.

The architecture of joints and faults was reconstructed. Macroscopic observations of mineral veins, especially their cross-cutting relationships, were carried out in the field. The results of our already published microscopic and geochemical studies were also used in the interpretation (Świerczewska et al. 2000b, 2001).

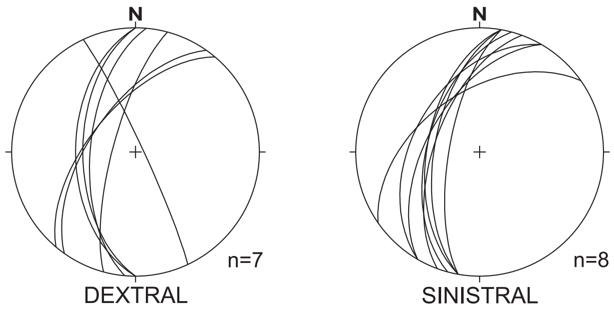
Joints. Attitudes of 283 joints were measured. Four sets of joints (D1, D2, L, and L') can be distinguished in most of the beds. The joints of sets D1 and L, and joints of sets D2 and L' are oriented sub-perpendicular (81–107°) to each other. The joints of sets D1 and D2, and those of sets L and L' are oriented under small angles (17–42°) to each other. The architecture of mineralized joints is distinctly different to that of the barren ones. Most joints of sets D1 and D2 are filled by mineral veins, whereas joints of sets L and L' are mostly barren.

Small-scale faults. Attitudes of 15 minor faults were measured (Fig. 20). Only strike-slip faults, both dextral and sinistral ones, were observed. The faults display strikes from N25W to N50E. The dips of the faults are similar to those of the joints which strike parallel to the faults. Whenever striae occur on fault surfaces, the striae are parralel to the intersection between fault plane and bedding plane. All studied faults are filled by mineral veins.

Textures of mineral veins - microscopic observations. Most of the D1 and D2 joints are healed by columnar (CC), blocky (BC) and drusy (DC) calcite, as well as by blocky calcite and blocky to drusy quartz (CQ). Simple veins filling these joints are composed of CC, BC or DC. In composite veins, the following successions are observed: (i) from CC to BC to DC, (ii) from CC to BC, (iii) from CC to QC to DC. The L and L' joints are filled exclusively by CC. Crystals oriented either perpendicular (CCP) or oblique (CCO) to the walls of veins were observed in



■ **Fig. 19.** Lithological log and attitudes of joints cutting particular beds (1–7): I – all joints, II – mineralized joints, III – barren joints (after Świerczewska et al. 2005).



■ **Fig. 20.** Attitudes of small-scale faults.

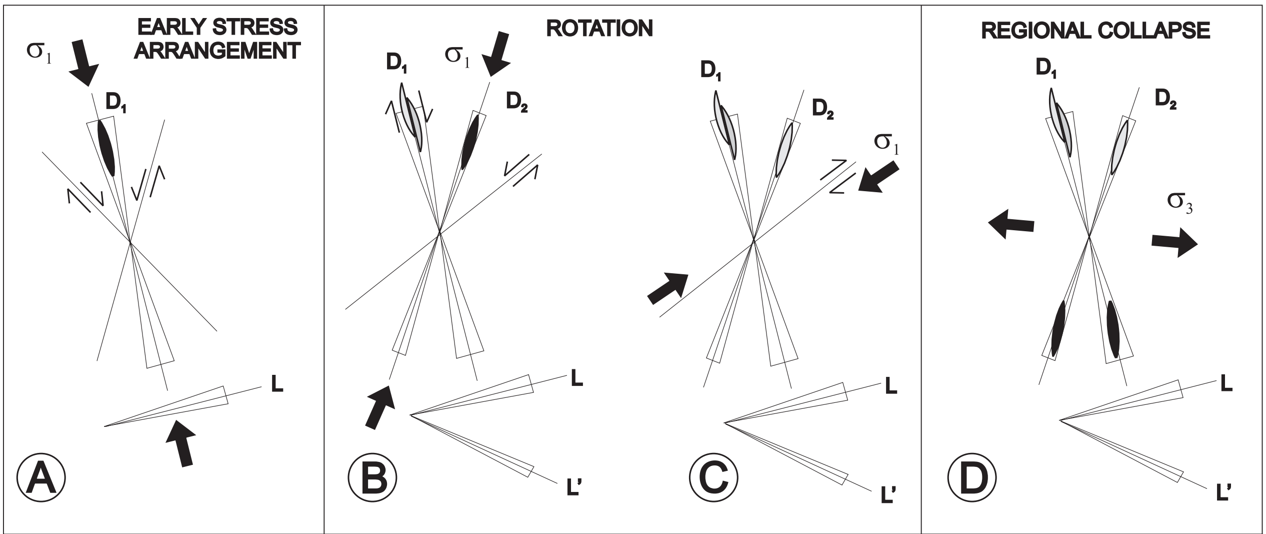
joints	I	II	III	IV
D_1	CCP	CCO	CCO	CCP CCO BC-CQ-DC
D_2		CCP	CCO	CCP CCO BC-CQ-DC
L	CCP	CCO	CCO	
L'		CCP	CCO	

time →

■ **Fig. 21.** Observed mineral successions in mineral veins filling D_1, D_2, L and L' joints.

joints of all sets. CCO indicate either sinistral or dextral sense of strike-slip movement along microfaults (cf. Fig. 20). Synkinematic fibrous calcite was found on surfaces of some small-scale strike-slip faults. In some faults CQ post-date this fibrous calcite.

Microscopic observations were supplemented by macroscopic examination of vein intersections. Results of studies showing succession in filling of the D1, D2, L and L' joints are summarized in Fig. 21. Cross-cutting relationships between composite veins yielded most information about mineral sequences. Basing



	A	B	C	D
D_1	CCP-*BC-DC*	CCO	CCO	CCP CCO BC-CQ-DC
D_2		CCP-*BC-DC*	CCO	CCP CCO BC-CQ-DC
L	CCP	CCO	CCO	
L'		CCP	CCO	
FAULTS	FC CCO			

■ **Fig. 22.** Cartoon (drawn in horizontal plane) illustrating progressive formation of extensional joints marked by ellipses (black when active, grey when inactive) and active small-scale faults (lines with half-arrows); hypothetic sequences of textures in mineral veins related to continuous mineralization are presented in the bottom part of the figure (parts of sequences which were not observed are marked in italics); see text for further explanation.

on these cross-cutting relationships, as well as on cross-cutting of composite and single veins it was possible to recognize CC formed during different stages of infilling.

Discussion and conclusions

The described architecture of joints fits very well the extensional model of jointing (e.g. Engelder and Geiser 1980). Results of kinematic analysis of small-scale faults show that the faults and all joints were formed progressively before and during anticlockwise rotation of the Carpathians (Fig. 22A-C). The amount of the rotation was at least 60°. Collapse-related deformation (Fig. 22D) was due to a WNW-ESE oriented extension.

Hypothetic sequences of textures in mineral veins related to progressive mineralization in a changing stress arrangement are presented in the bottom part of Fig. 22. Except for the *BC-DC* sequence (stages A and B), all sequences can be observed in veins (compare Fig. 21).

Cross-cutting relationships between mineral veins filling joints and small-scale faults appear to be complicated. However, these relationships fit in the hypothetic sequences of textures in mineral veins. Therefore, we believe that the adopted extensional model of jointing is valid in the discussed case. Moreover, the observed cross-cutting relationship of veins shows that the bulk of mineralization filling joints and microfaults was introduced during collapse. This confirms our earlier conclusion (Świerczewska et al. 2000b). Furthermore, there occur two temperature ranges of homogenisation (<50 °C and 100–145 °C) obtained from fluid inclusions trapped in CCP (Świerczewska et al. 2001). This appears to confirm our conclusion that CCP was formed during three stages of structural evolution (A, B, D – Fig. 22).

Stop 5. Pieniny Andesites at Wżar Hill – Mineralogy, Petrology and Palaeomagnetism

Two exposures will be presented: (1) old quarry close to the Snozka pass (road Krościenko-Nowy Targ) – two generations of andesite dykes; petrographic varieties, contact metamorphism, palaeomagnetic results, and (2) exposure near lower station of the ski-lift in Kluszkowce village - two petrographic varieties of andesite; high degree of hydrothermal alterations.

In the old quarry two generations of andesite dykes are well visible. The older generation is usually rusty and significantly altered. The younger one (which was quarried here) is "fresh-looking", dark grey in colour, with porphyritic texture. Abundant amphibole and pyroxenes phenocrysts are typical of this variety. Feldspar phenocrysts are less abundant (feldspar is common in rock matrix). Xenoliths of sedimentary rocks and dark enclaves (composed of agglomerations of mafic minerals), and mafic megacrysts are relatively common in both varieties (Kardymowicz 1957, Bakun-Czubarow and Białowolska 2004).

The older generation of dykes belongs to the first phase intrusions (e.g. Birkenmajer and Pécskay 1999). These are oriented subparallel to the northern boundary fault of the Pieniny Klippen belt. The dykes of younger generation (second phase intrusions) are related to transversal SSE-NNW faults (e.g. Birkenmajer and Pécskay 1999).

Andesites at Wżar hill are emplaced in Palaeogene flysch rocks of the Magura Nappe. Both andesite and sedimentary rocks are altered near contacts. Contact alterations are well visible in the old quarry (changes in colour, hardening of sedimentary rocks) (e.g. Birkenmajer 1958). Contact metamorphism is expressed in formation of garnet and pyroxene in sandstone with abundant carbonate cement (Pyrgies and Michalik 1998). Two chemical varieties of both garnet and pyroxene could be related to two phases of andesite emplacement. Andesite at contact is significantly enriched in calcite. Hydrothermal activity–related minerals overlap those related to contact metamorphism (Pyrgies and Michalik 1998).

In the exposure near lower station of the ski-lift also two varieties of andesite are exposed. Feldspar phenocryst-rich rock, characterized by numerous voids filled with secondary minerals, contacts with a variety dominated by mafic minerals. Hydrothermal alterations zones and hydrothermal veins exposed in this outcrop have not been studied in detail. (e.g. Gajda 1958 a, b, Szeliga and Michalik 2003).

Mineralogy. Plagioclase phenocryst composition varies from oligoclase to bytovnite. Most plagioclase phenocrysts are zoned. In some samples, cores of plagioclase phenocrysts rich in glass inclusions are rimmed by euhedral and inclusion-free zones (Michalik et al. 2004). Plagioclases are partly replaced by calcite and chlorite (especially in feldspar phenocryst- rich variety from the outcrop at Kluszkowce village). Amphibole phenocrysts represent magnesian hastingsitic hornblende and magnesian hastingsite and ferroan pargasite, ferroan pargasitic hornblende, edenitic hornblende, according to Leake's (1978) classification. Amfiboles are often replaced completely or partly by chlorite, calcite, Fe-Ti oxides, and titanite. Coronas of complex composition (fine-crystalline aggregates of plagioclases, pyroxene, Fe-Ti oxide minerals) are common around amphiboles. Pyroxene phenocrysts are represented by diopside (Morimoto's 1988 classification). Alterations of pyroxene result in formation of chlorite and calcite. Fe-Ti oxides phenocrysts belong to ulvöspinelo-magnestite series. Fibrous illite-like mineral is commonly observed using SEM. Mn-Fe oxides of corn-flake morphology aggregates are developed around mafic phenocryts (Michalik et al. 2004, 2005).

XRD study of < 2 μm fraction separated from andesite samples indicates that it is composed of illite/smectite (and vermiculite/smectite?) and cristobalite. Less than 2 μm fraction separated from andesite samples and studied using XRD method are composed of illite/smectite (and vermiculite/smectite?) and cristobalite (samples from Wżar) (Michalik et al. 2005). Abundance of secondary minerals related to hydrothermal activity and presence of newly formed K-containing minerals (illite) are the reason for difficulties in K-Ar dates interpretations.

Palaeomagnetism. On Wżar Hill both the older and younger Miocene andesites crop out in abandoned quarries. The contact zone between the two phases of intrusions is well exposed, but heavily weathered. Palaeomagnetic sampling of the andesites required special care, for the intrusions themselves are also quite altered. Thus, we drilled long cores and found that weathering indeed fully remagnetized the upper 10 cm or so of the andesites. Useful palaeomagnetic signal was only obtained from the deepest part of the about 20-cm-long cores. Both phases of intrusions possess composite natural remnant magnetization. The mostly dominant component was reversed polarity and exhibits no rotation with respect to stable European reference declination. The other, more stable, component shows moderate counterclockwise rotation (of about 30 ° in average). This component was not found during previous palaeomagnetic investigations, most probably because the samples were demagnetized by alternating field, a method which was not suitable to separate components.

The Wżar Mts. andesites are of key importance for the tectonic interpretation of palaeomagnetic data obtained also from other Pieniny andesite outcrops. They suggest that the area intruded by the andesites rotated counterclockwise after the second intrusion phase.

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