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Tectonics of the Beskid Wyspowy Mountains (Outer Carpathians, Poland)

Andrzej KONON

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The structure of the Magura Nappe, within the zone of maximal bending of the Western Outer Carpathians, is described; investigations were concentrated mainly in the Beskid Wyspowy Subunit. A zone, bounded to the north and south by duplexes, comprises large characteristic synclines (Śnieżnica, Lubogoszcz, Szczebel, Klimas, Łopień, Ćwilin, Luboń Wielki). These appear on maps as isolated "island mountains", in the Beskid Wyspowy Subunit. These synclines contrast strongly with the belt-like distribution of regional folds to the west and east of the area. The synclinal massifs developed gradually. Fold belts several hundred metres long developed in the first phase of overthrusting of the Magura Nappe, with horizontal N–S compression dominant.Thrusts separating the individual subunits developed when the face of the overthrusting Magura Nappe stopped and the stress continued to push its southern parts forwards. The next phase, with continuing horizontal N–S stress included the development of strike-slip faults and the bending of the Carpathian Arc, resulting in extension of this part of the orogen. The syncline zone within the Beskid Wyspowy Subunit underwent disintegration and particular blocks became independent. Rotation of blocks with individual synclines took place along fault zones. In the part of the Polish Outer Carpathians investigated this stage is also characterised by a change of compression from N–S to NNE–SSW. Due to the uplift of this part of the Carpathians, strike-slip faults changed into dip-slip faults in the terminal part of this phase.

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Key words: Western Outer Carpathians, Magura Nappe, brachysynclines, mesostructures, block rotation, paleostresses.

INTRODUCTION

This paper is a study of the complex tectonics of the Magura Nappe in the central part of the Beskid Wyspowy Mts. and parts of their foreland, in the zone of maximal bending of the Western Outer Carpathians (Figs. 1 and 2). The area is characterised by synclines occurring prominently on maps as isolated "island mountains". The synclines strongly contrast with the belt-like distribution of regional folds, occurring to the west and east of the area. Investigations were focussed on the formation of these synclines in relation to the evolution of Western Outer Carpathians.

GEOLOGICAL SETTING

According to Gołąb (1947), the first mentions of the geology of this region were by Paul (1886), Szajnocha (1895) and Nowak (1927). Structural maps of the Beskid Wyspowy were published by Świderski (1953*a*, *b*), Burtan (1974), and Burtan and Skoczylas-Ciszewska (1966).

LITHOSTRATIGRAPHY

Two lithostratigraphic successions — the Magura and Silesian (Burtan and Skoczylas-Ciszewska, 1966; Burtan, 1974, 1978) (Fig. 3) correspond to two first-order tectonic units: the Magura and Silesian Nappe (Fig. 2). The age and thickness of particular beds are after Burtan (1974, 1978), Burtan *et al.*, (1992) and partly after Oszczypko (1992) (Fig. 3).

Two main lithofacies occur within the Magura succession (Węcławik, 1969) — the Rača and Siary units, the latter documented only in the northern margin of the area. Tectonic investigations were concentrated on the area within the Rača Unit.

The Silesian Unit is documented only in a small part of the area (Fig. 2) with deposits ranging from Early Cretaceous to Late Oligocene in age.

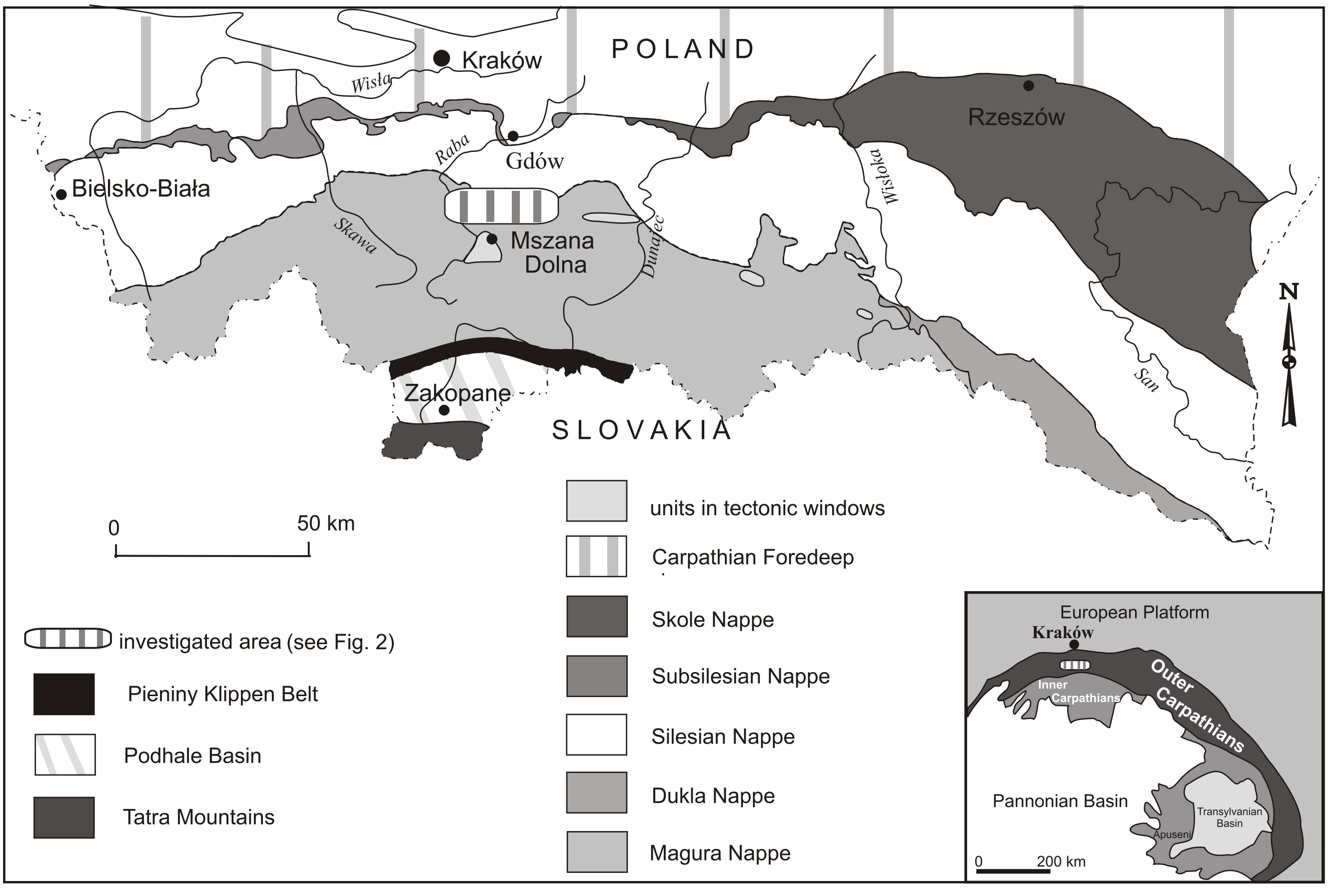

Fig. 1. Location of the Beskid Wyspowy Mts.

TECTONICS

The area investigated comprises mainly the Magura Nappe, which to the north adjoins a fragment of the Silesian Nappe (Figs. 1 and 2), and to the south adjoins the Mszana Dolna tectonic window within the Magura Nappe (e.g. Świderski, 1953*a*; Burtan, 1974; Mastella, 1988).

Magura Nappe. Two tectonic subunits can be distinguished within the Magura Nappe (Fig. 2) (Świderski, 1952; Mastella, 1988); the Beskid Wyspowy Subunit (Góry Wyspowe of Świderski, 1953*b*; Beskid Wyspowy of Mastella, 1988) and to the north, the Kiczora Subunit (Świderski, 1953*b*). The Beskid Wyspowy Subunit is built of deposits included within the Rača lithostratigraphic zone (Fig. 3), which forms a part of the Zawoja-Jordanów synclinorial zone (Książkiewicz, 1972).

Large synclines — Klimas, Szczebel, Luboń Wielki, Lubogoszcz, Śnieżnica, Ćwilin and Łopień — dominate here (Figs. 2 and 4). They form isolated "island mountains" (Świderski, 1953*b*) (Fig. 2), a result of thick-bedded Magura Sandstones in their cores, which are more resistant than the older beds (Inoceramian Beds, Variegated Shales or Hieroglyphic Beds; Nowak, 1927; Świderski, 1952, 1953*b*).

Silesian Nappe. Investigations within this unit were focussed only on a small fragment of 12 km^2, in the foreland of the Magura Nappe, (the "Skrzydlna bay"; Fig. 4), where at least 9 slices are present in this area. They are composed of the Cieszyn Beds, Verovice Beds, Menilite Beds and Krosno Beds (Burtan, 1974; Burtan *et al*., 1992) (Fig. 2).

METHODS OF INVESTIGATION

Investigations were based on a 1:10 000 geological map (Konon, 1999), supplemented by interpretation of air-photos on a 1:25 000 scale and of radar images on a 1:100 000 scale, together with structural analysis of tectonic mesostructures, tectonic microstructures in fault rocks and palaeostress analysis. Slickensides were investigated by methods introduced by Angelier (1979) and Ratschbacher *et al*. (1994).

Following Jaroszewski (1972) statistical analysis was applied within homogeneous domains (areas with homogeneous features, which can be analysed as a whole). Plane orientations and directions were noted respectively as dip direction/dip angle and as azimuths (0–360°).

Diagrams were prepared as lower hemisphere projections, normals to planes being used. Strike-slip faults were projected after (1979, 1994), and dip-slip faults after Hoeppener (1955).

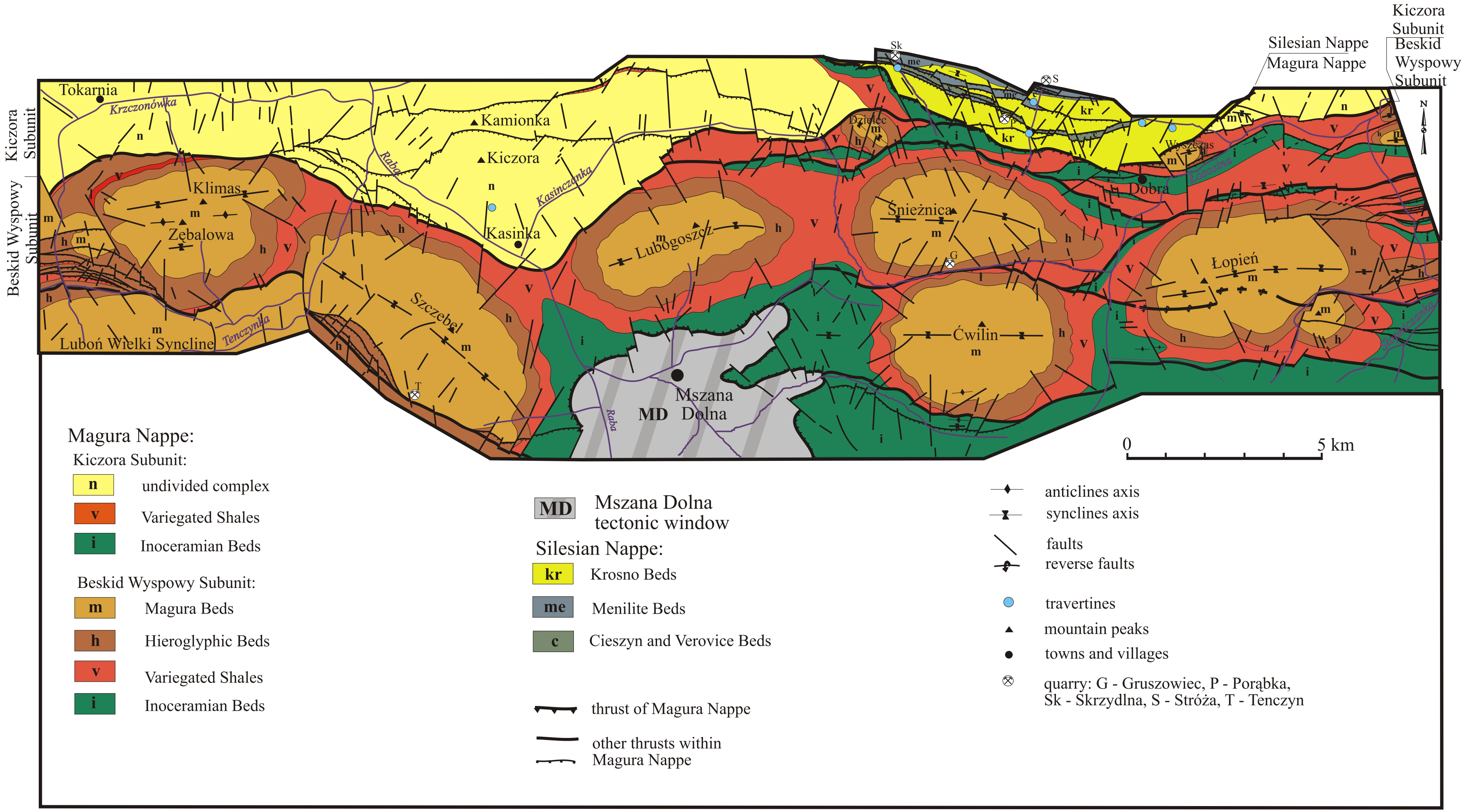

Fig. 2. Tectonic map of the central part of the Magura Nappe in the Beskid Wyspowy region (Konon, 1999)

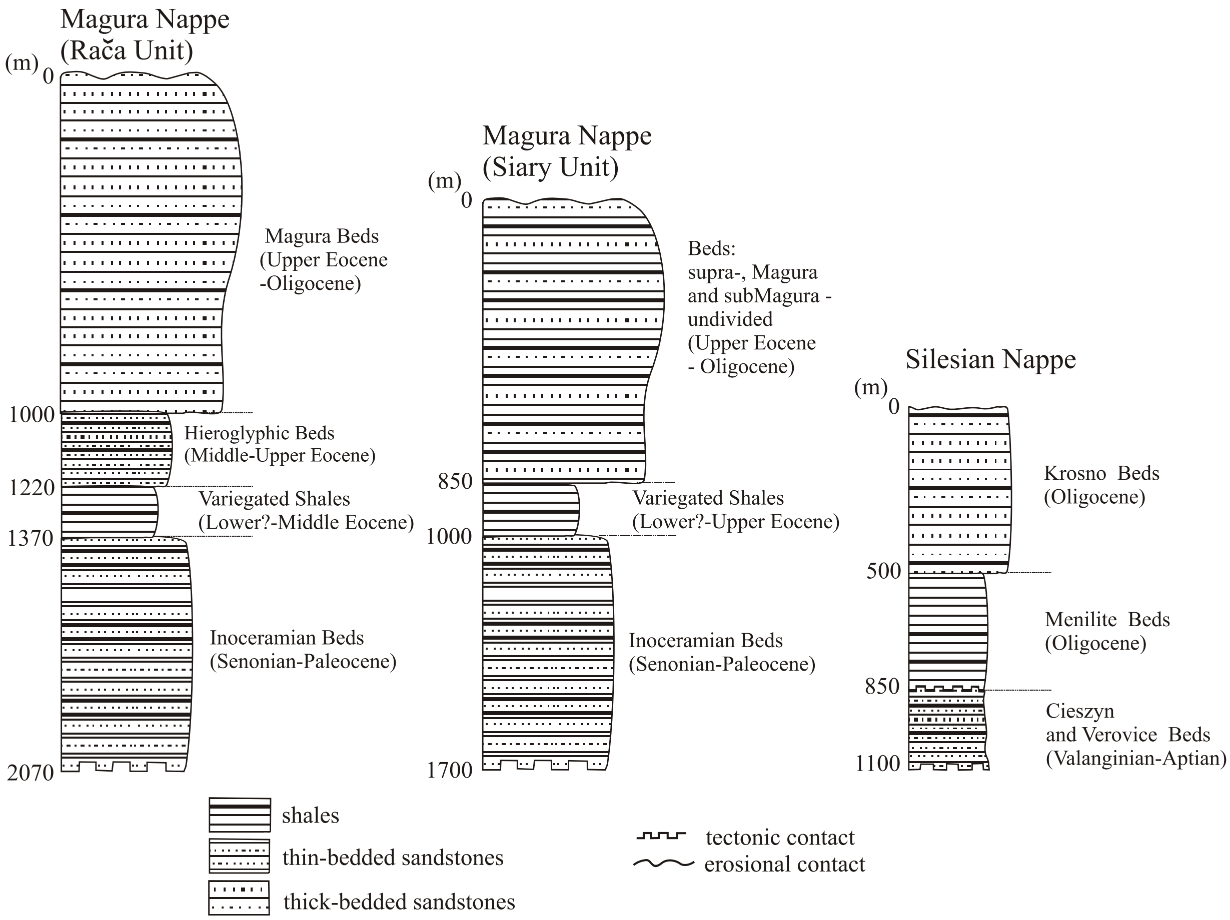

Fig. 3. Lithostratigraphic columns of the Magura and Silesian successions in investigated area — based on Burtan (1974, 1978), Oszczypko (1992) and investigations presented in this paper

These methods showed the tectonic evolution of the Beskid Wyspowy Mts. in relation to the evolution of the entire Western Outer Carpathians. Also, previously unrecognised tectonic zones, were distinguished in the Magura Nappe:

— in the Beskid Wyspowy Subunit — zones of synclines and slices (northern slices and southern slices);

— in the Kiczora Subunit — three larger slices.

New mesostructures, e.g. three cleavage systems, kink folds and small antiformal stack structures were noted for the first time in this part of the Outer Carpathians, and various microstructures and shear lenses were observed in the thrust zones.

Analysis of joints, folds and strike-slip faults has yielded palaeostress patterns for these particular tectonic zones.

NEW TECTONIC DIVISIONS

MAGURA NAPPE

THE BESKID WYSPOWY SUBUNIT

The subunit comprises a synclinal zone, bordered to the north and south by two slice zones — northern and southern, (Figs. 2 and 4).

Synclinal zone. The synclines form two belts — the northern (Śnieżnica, Lubogoszcz, Szczebel and Klimas synclines) and southern (Łopień, Ćwilin and Luboń Wielki synclines) (Figs. 2 and 4).

The Klimas, Szczebel, Lubogoszcz, Śnieżnica, Ćwilin and Łopień are, according to Jaroszewski (1994), brachysynclines passing into troughs. These synclines represent concentric folds, whereas the anticlines, occurring in a much-reduced form, represent similar folds (Figs. 2, 5, 6). Using a geometric classification (Elliott, 1965; Ramsay, 1967), the synclines can be referred to as parallel folds. The Klimas, Lubogoszcz, Ćwilin and Łopień synclines represent upright folds, and according to the criteria of Ramsay (1974), they are tectonically poorly developed (Figs. 5–7). The Śnieżnica and Szczebel synclines are overturned folds in their southern parts near the thrusts, and advanced as regards tectonic development (Figs. 5–7).

The upper parts of the synclines, built of thick-bedded sandstones of the Magura Beds, resemble concentric folds and are indistinctly deformed. The lower parts, in turn composed of shales and thick- to medium-bedded sandstones, are secondarily folded and sliced (Figs. 5 and 6).

The occurrence of tectonic striae concordant with the dip direction suggest that the synclines formed during flexure, with flexural slip.

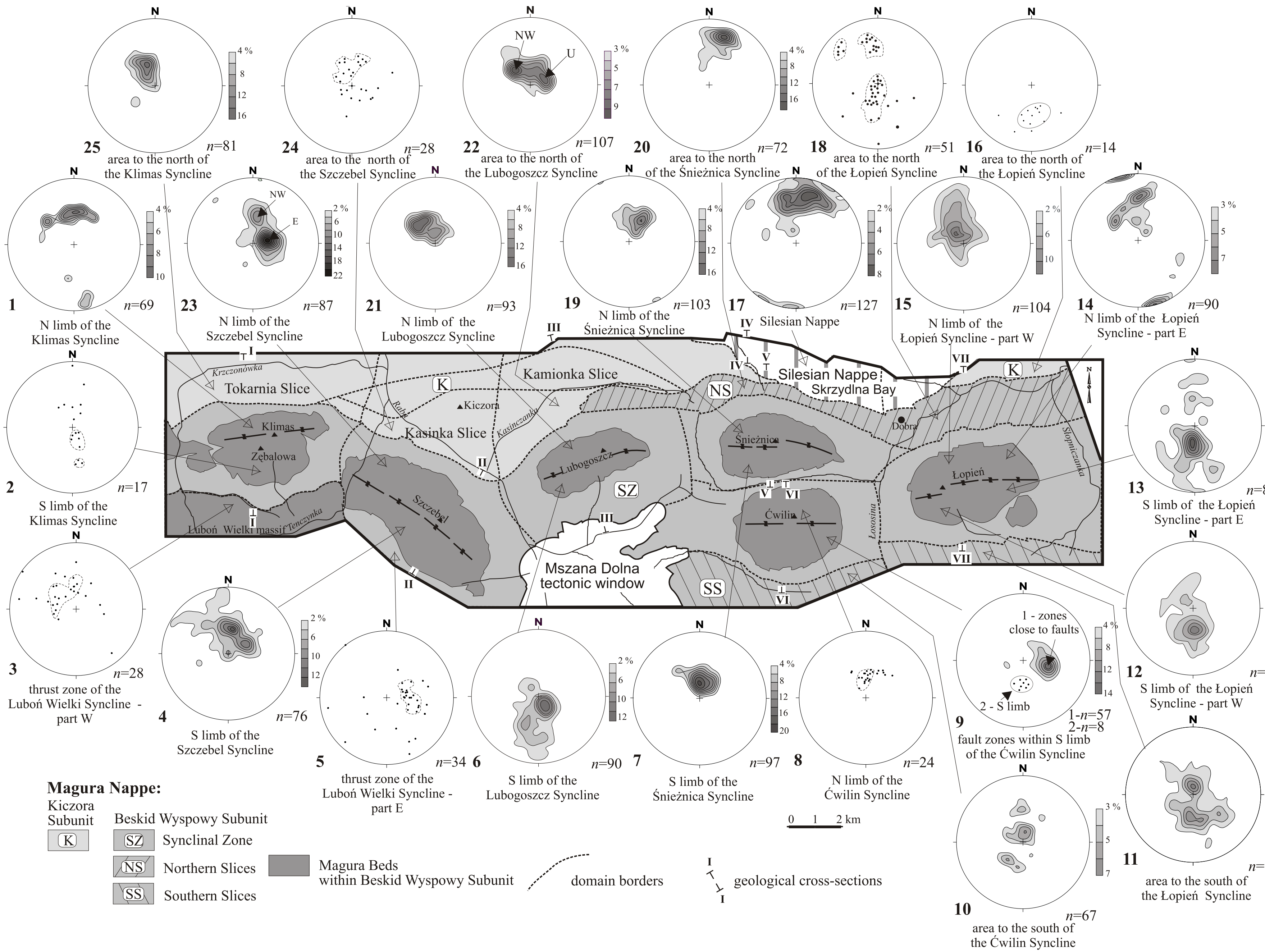

Fig. 4. Subdivision of the investigated area into structurally homogeneous domains with diagrams of bedding (1–25)

Other explanations see Fig. 2

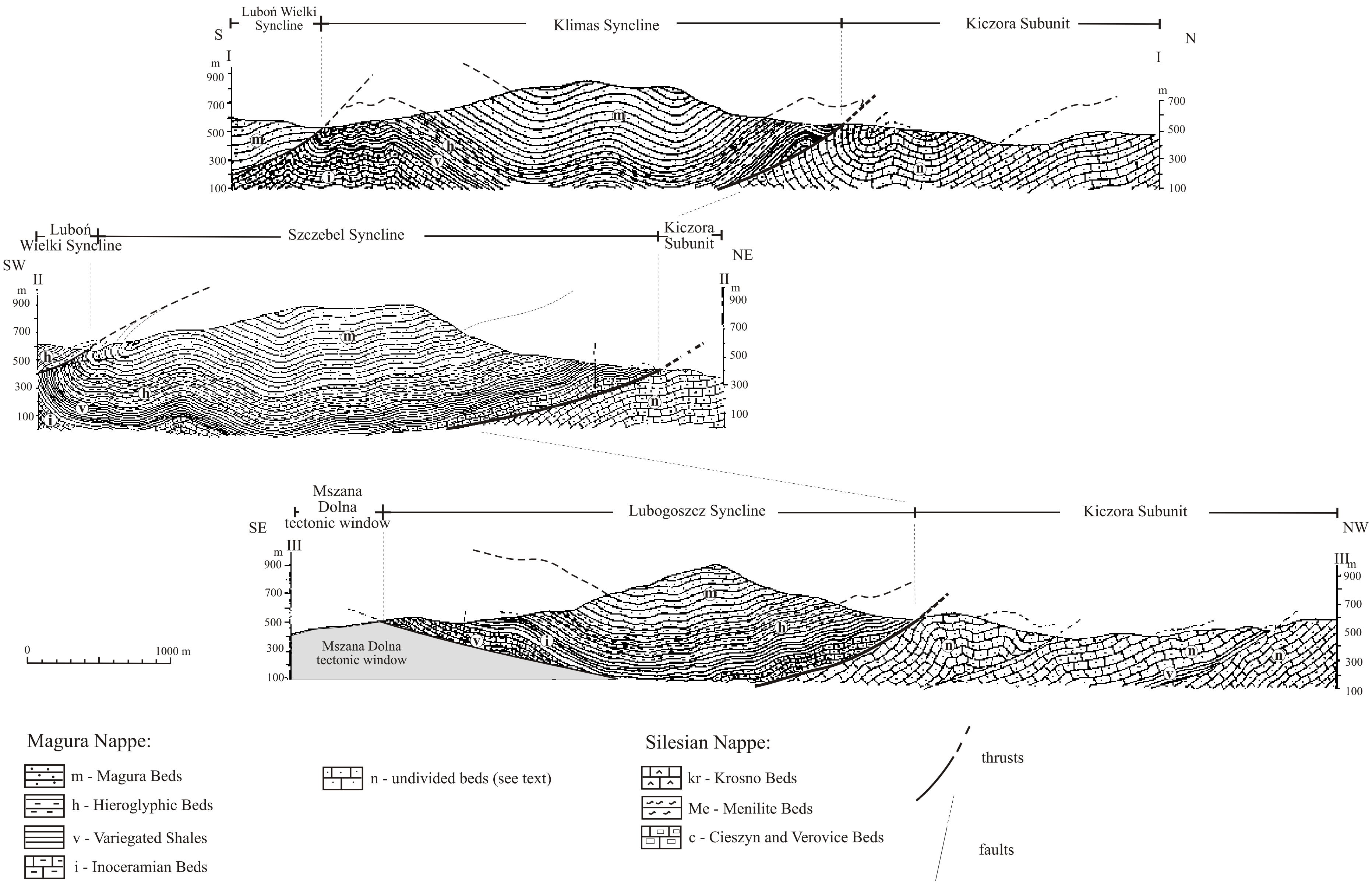

Fig. 5. Geological cross-sections I–I, II–II, III–III

For location see Fig.4

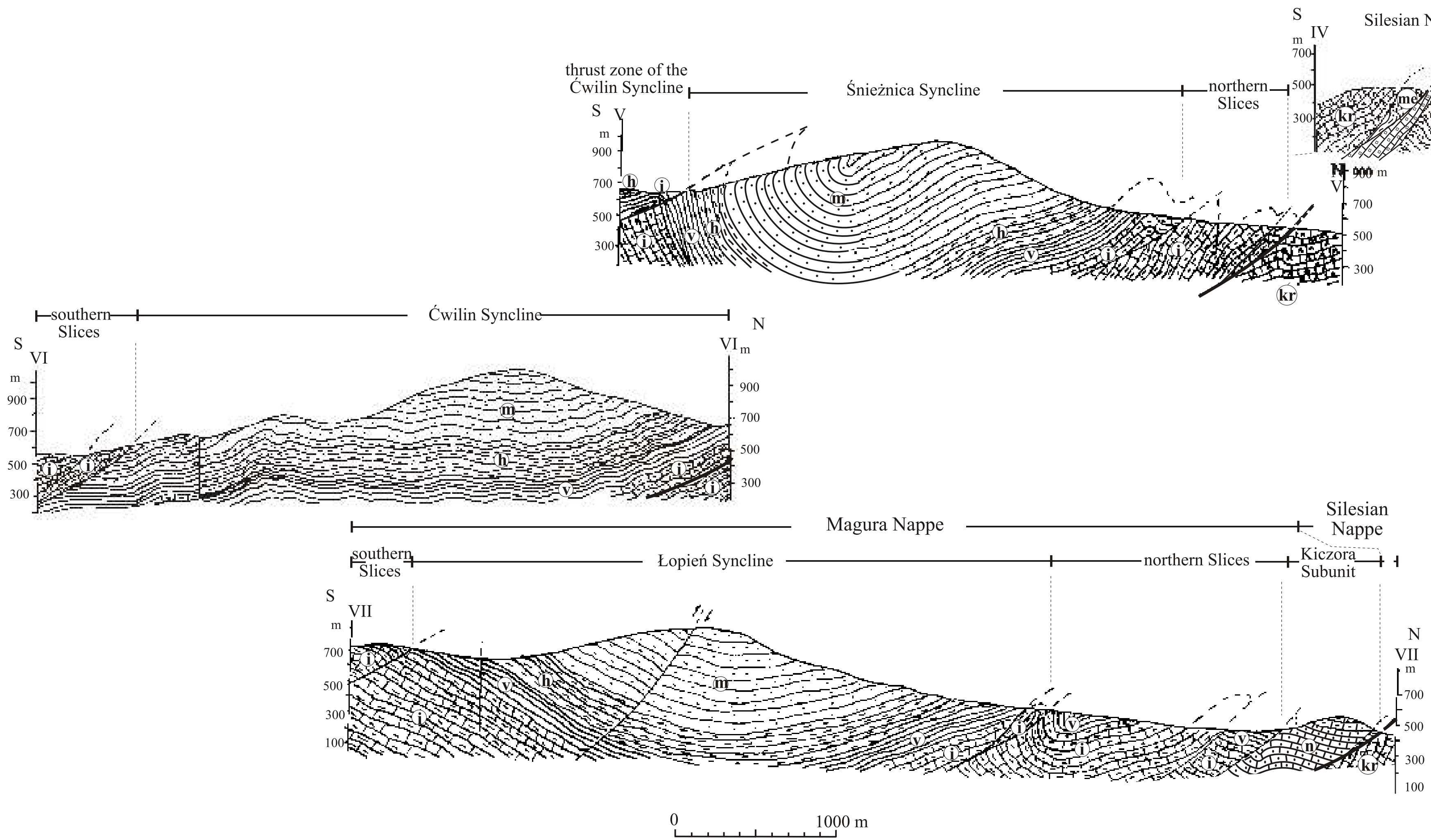

Fig. 6. Geological cross-sections IV–IV, V–V, VI–VI, VII–VII

For location see Fig. 4; for explanation see Fig. 5

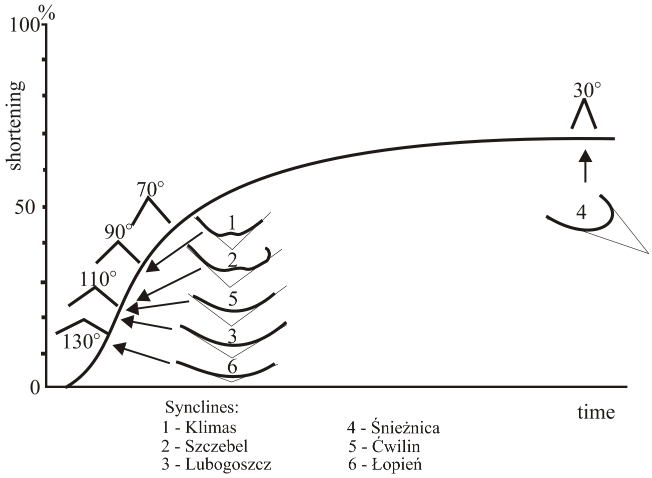

Fig. 7. Interlimb angles of synclines in the Beskid Wyspowy Subunit in view of the relation between the interlimb angles and fold development (modified after Ramsay, 1974)

The axes of individual synclines within this zone are variably deflected from the regional trend of this part of the Western Carpathians Arc (Fig. 8); deflections range from clockwise (Szczebel Syncline — of about 35–40°; Śnieżnica Syncline — of about 10°), to counter-clockwise (Lubogoszcz Syncline — of about 20°, Klimas Syncline — of about 15°, and Łopień Syncline — of about 10°) (Figs. 2 and 8).

Slice zones. The slice zones border the synclinal zone to the north and south, respectively.

The northern slice zone occurs between the synclinal zone and the Silesian Nappe in the central part, and between the synclinal zone and the Kiczora Subunit in the eastern and western part of the area (Figs. 2 and 4). This zone is characterised by the presence of at least 5 imbricated, overthrusted slices. In the region lying directly in the foreland of the Śnieżnica Syncline, the slices are covered by the synclinal zone thrust (Fig. 2).

The northern slices are dominated by 197°/65° beds (in the Ś Syncline foreland), and 10°/35°–60° and 165°–175°/40°–65° beds (in the Łopień Syncline foreland) (Fig. 4).

The beds within the slices are strongly folded (Fig. 6). Similar folds prevail, whereas concentric folds are much rarer, a result of the dominant lithology. The anticlines are characterised by a strong reduction of the lower limbs , and in some synclines the S limbs are overturned (Fig. 6). Parallel slices dominate, concordant with the direction of the structures. The Dzielec Slice, trending 125°, and the Wyszczas Slice, trending 65°, are deflected from this direction (Fig. 2).

The southern slices occur between the synclinal zone and the Mogielnica Subunit thrust, which lies beyond the area investigated. Two slices occur in the area within this zone (Fig. 2). The western part is characterised by beds dipping 193°/23°, and the E part by beds dipping 180°/20° and 10°/37° (Fig. 4). Similar folds prevail within the slice, with anticlines characterized by reduced lower limbs (Fig. 6). The southern slice zone probably continues laterally towards the SW as a zone of strong folding to the south of the Luboń Wielki Syncline (Świderski, 1953*a*).

The occurrence of the northern slices between the thrusts of the synclinal zone and Kiczora Subunit, partly above the Magura Nappe thrust, and of the southern slices between the thrust of the southern slice zone on the synclinal zone and the thrust of the Mogielnica Subunit (Świderski, 1953*a*), indicates that these zones represent a typical duplex, according with the model of Boyer and Elliott (1982).

The Kiczora Subunit. The Kiczora Subunit occurs to the north of the Beskid Wyspowy Subunit (Fig. 4). Investigations were carried out only in its southern part, around its contact with the Beskid Wyspowy Subunit (Figs. 2 and 4). Three larger slices: the Kasinka, Kamionka and Tokarnia slices, which are internally sliced, were distinguished here (Figs. 2 and 4). These approximately parallel slices form imbricate structures transported over each other along S dipping thrusts. The slices comprise the Inoceramian Beds, Variegated Shales and an overlying unit of sub-, supra- and Magura Beds, not separately distinguished here.

TECTONIC STRUCTURES

TECTONIC MESOSTRUCTURES IN THE MAGURA NAPPE

The area of the Magura Nappe investigated is characterised by different tectonic mesostructures, genetically linked with the folding and overthrusting. Attention was focussed on folds, faults, slickensides and cleavage.

Folds. The folds occur in packets of shales and thin- to medium-bedded sandstones. Drag folds, associated with flexural slip, prevail. They are developed within in folds of higher orders (e.g. Fig. 9) or, more rarely, along gently dipping faults and shears.

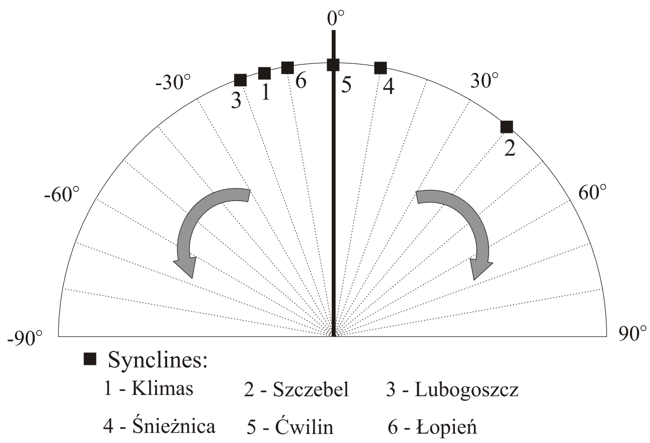

Fig. 8. Deflection of syncline axes in the Beskid Wyspowy Subunit from the regional trend of tectonic structures in the investigated part of Outer Carpathians

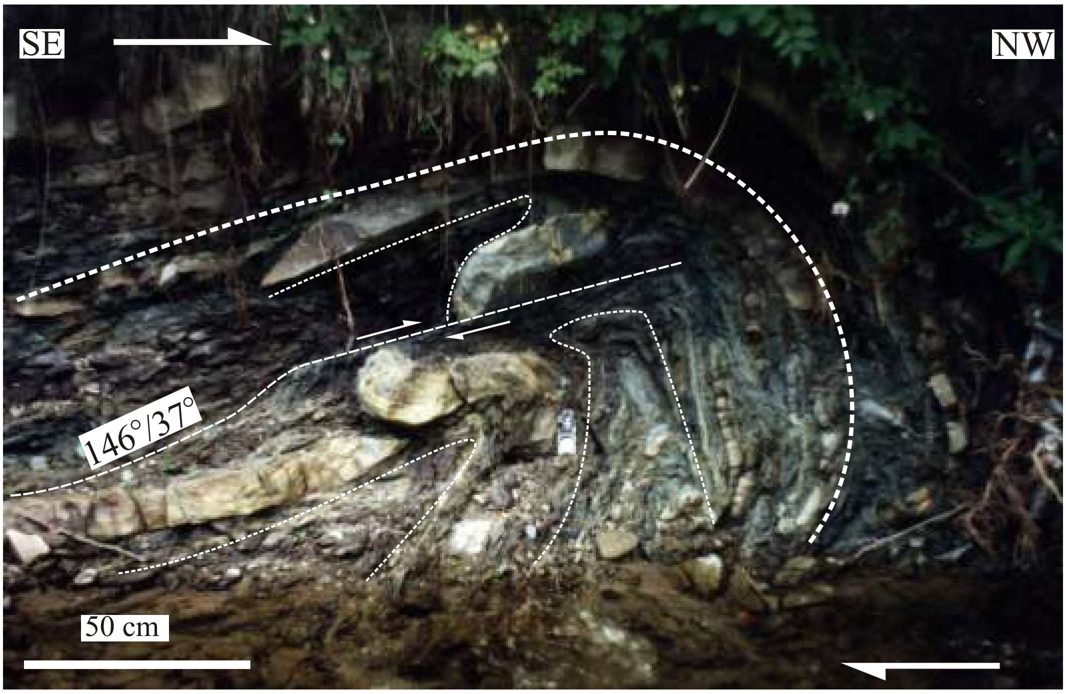

Fig. 9. Mesofold in the Variegated Shales in the Kasinczanka creek (about 8.5 km away from the Raba River)

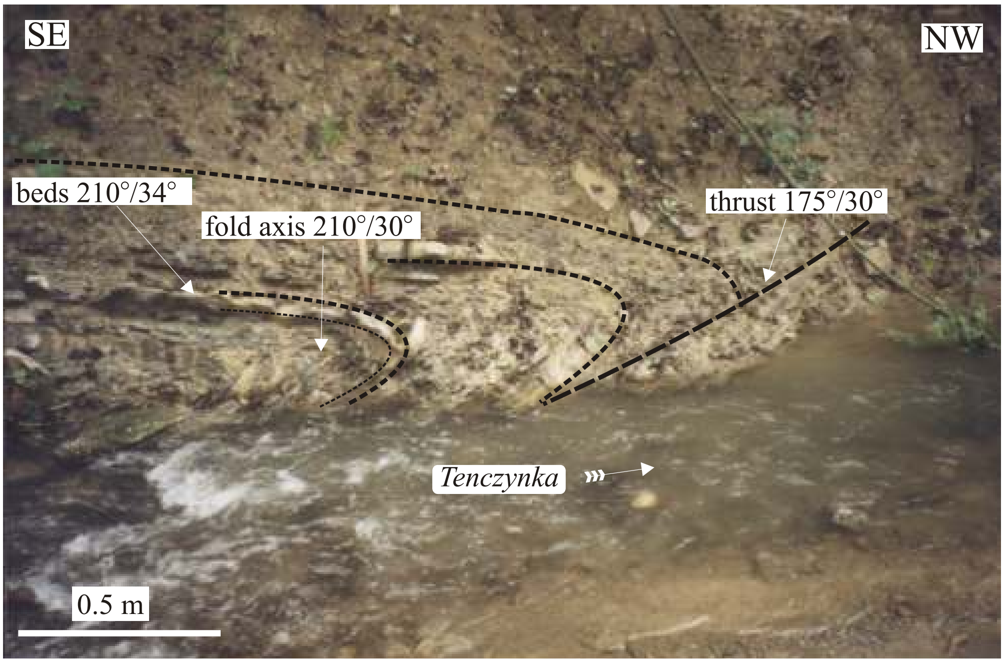

Fig. 10. Overturned fold in Hieroglyphic Beds in the Tenczynka creek (about 4 km from Tenczyn quarry)

Drag folds occurring in the limbs of regional folds are characterised by a simple limb structure and strong deformation of the cores (Fig. 9). Axes of drag folds are similar to the axis of the fold, in the limb of which they are developed.

Drag folds underwent gradual deformation during the steepening of the fold limbs. In the initial phase they formed simultaneously with the folds, within which they developed as typical drag folds. When the limbs became steeper in the superior folds, the pair of forces acted more strongly, deforming the cores of drag folds. In the terminal phase of folding, dispersed reverse faults were formed.

The gently dipping parts of regional folds are characterised by synclines and anticlines up to several decimetres high. The main displacement of the deformed beds took place along shear planes passing into bedding planes.

Folds are characterised by a strong asymmetry of limbs and axes similar to the axes of folds, in the limbs of which they are developed.

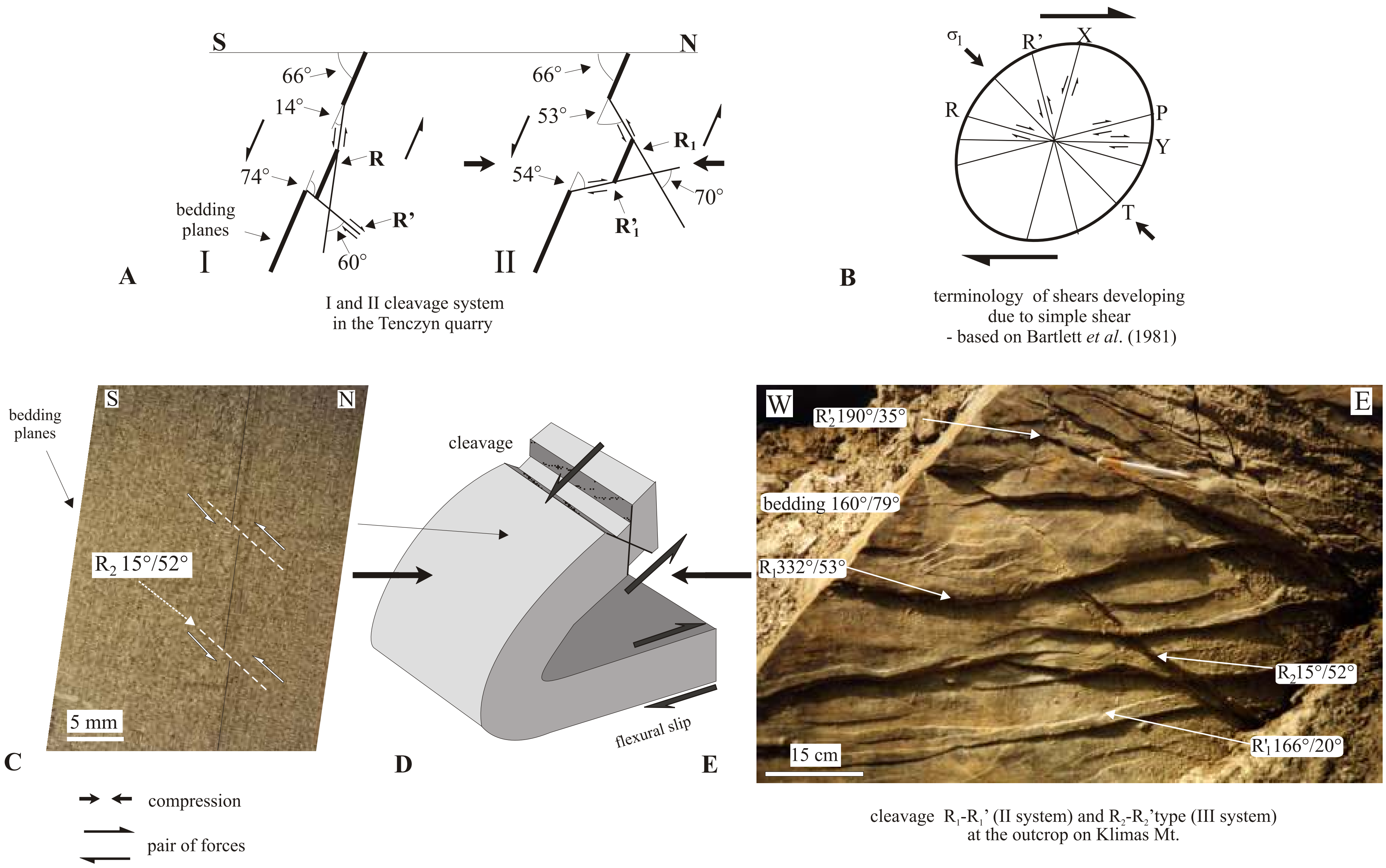

Fig. 11. Cleavage systems in the Magura Nappe

Part explanations for forces see Fig. 13

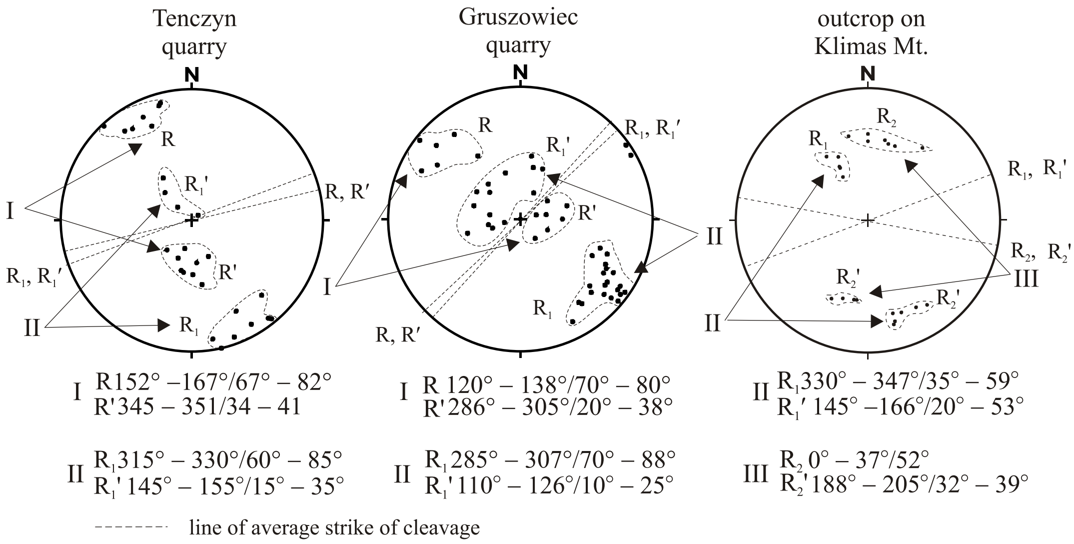

Fig. 12. Diagrams of cleavage planes in the Magura Nappe

Such folds largely form as a result of the differences in the competence of the folded deposits, where more competent beds — in this case sandstones — can slip more easily on less competent beds (in this case shales), linked with the tendency to form small ramps. Their formation, according to Knipe (1985), may be initiated by a slip plane attempting to by pass an object restricting free movement. Einstadt and DePaor (1987) have alternatively suggested that ramps may form as a result of their development within brittle rocks, in this case in sandstones, associated with horizontal slip planes within the shales. Both models suggest the formation of drag folds during simple shear.

In the Magura Nappe, folds have also been recognised close to faults and overthrusts. Their amplitude ranges from a few to over a dozen metres (Fig. 10). The folds verge to the north and occur on the upthrown sides of reverse faults dipping approximately 10–50°S. They form where substantial friction took place along the fault plane (Mitra, 1992). Various authors (e.g. Suppe, 1983; Chester and Chester, 1990; Mitra, 1990, 1992), refer to these folds to as fault-propagation folds, commonly described from the Outer Carpathians (e.g. Mastella, 1988; Aleksandrowski, 1989).

Cleavage. Closely spaced shear fractures (Fig. 11) were commonly recognised in sandstone and siltstone beds within the upper limbs of overturned synclines. Series of parallel steps of variable size (a few millimetres wide, in some cases up to 1–2 m long and at intervals of up to 10 cm) were formed on bed surfaces during displacement along the shears.

These features indicate, following Jaroszewski (1972, 1980, 1994) and Boyer (1984), that these fractures represent cleavage. Earlier, such fractures were described by Aleksandrowski (1980, 1989) as step/shear lineation, by Misiuwianiec (1992) and Jaźwiński (1994) as cleavage and by Tokarski *et al.* (1995) and Marko *et al.* (2000) as deformation bands.

Individual cleavage systems differ in strike, in their angles to bedding and in their senses of movement along these surfaces.

The first system (I) comprises parallel sets of fractures marked as R and R' (e.g. Fig. 12 — Tenczyn and Gruszowiec quarries). These fractures cross each other at acute angles of about 60–70°.

The second system (II) comprises R_1 and R_1' sets (e.g. Fig. 12 — Tenczyn and Gruszowiec quarries, Klimas Mt. exposure). The R_1 set occurs at a larger angle to the bedding plane than set R (Fig. 11), whereas set R_1' lies at similar angles to the bedding plane as set R', but has the geometry of a reverse fault, and not a normal fault as in the case of set R'. The fractures of the second system cross each other at an acute angle of about 70°.

The third system (III) comprises sets referred to as R_2 and R_2' (Figs. 11 and 12). They have similar angles to bedding and, as in the sets of system II, they cross each other at an acute angle of about 70°.

The parallel strikes of the cleavage sets, constant angles formed by the sets to bedding and sense of movement along their surfaces indicate that each of these cleavage systems comprises conjugate shear sets, similar to those described by Muff (1909, *bona vide* Price and Cosgrove, 1990) in the Craignish phyllites in the United Kingdom. Some cleavage systems from the Magura Nappe are parallel to the fold axes and some are not. The first group comprises systems I and II, and thus they can be linked with the flexural slip; whereas the second group comprises system III and is linked with the overthrust of the upper parts of the Magura Nappe.

System I represents classic Riedel shears (Riedel, 1929). This cleavage, following Jaroszewski (1972), Vialon (1979) and Mastella (1988), developed first, when the rock could increase its volume. Therefore, it could have originated as a result of flexural slip in beds with low dips. System I cleavage planes in later deformation locally pass into small faults, as observed also by Świerczewska and Tokarski (1996, 1998).

System II is "rotated" in relation to the bedding planes (Jaroszewski, 1972) by comparison with typical Riedel shears, demonstrated by experiments and field investigations (e.g. Riedel, 1929; Morganstern and Tchalenko, 1967; Tchalenko, 1970; Harris and Cobbold, 1985; Naylor *et al*., 1986; Woodward *et al*., 1988). The shears probably developed from a pair of forces acting in a plane perpendicular to bedding, as well as

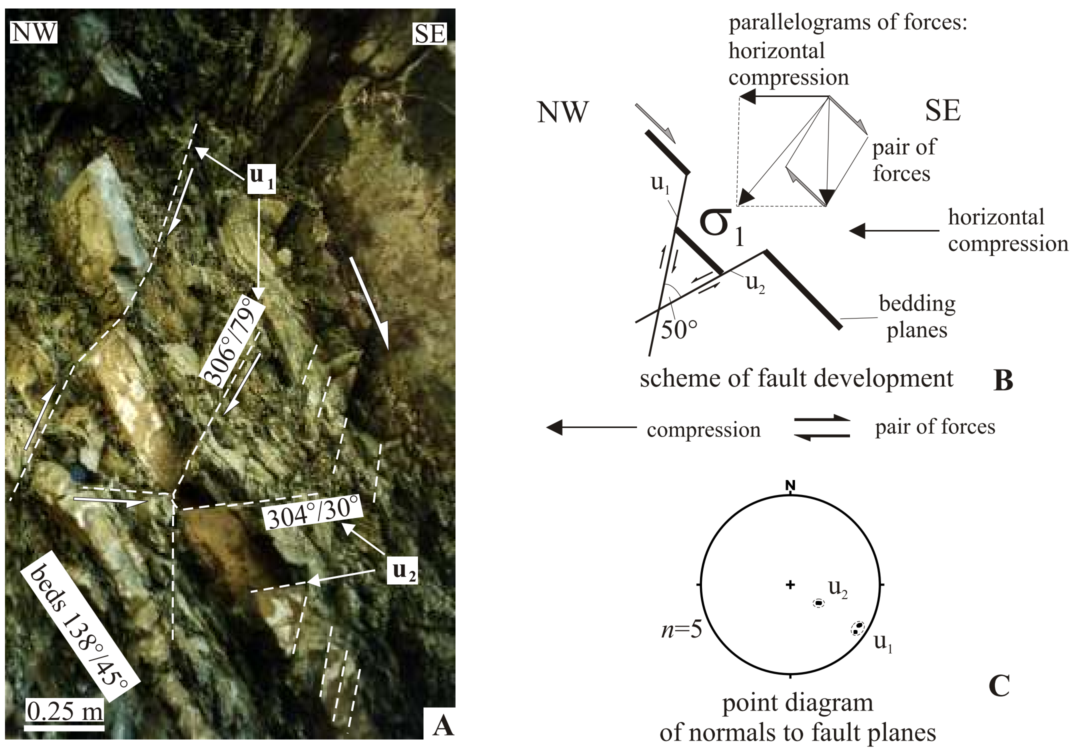

Fig. 13. Dip-slip faults in the Gruszowiec quarry

For location see Fig. 2; other explanation in the text

by horizontal compression caused by the overthrust of the upper elements of the Magura Nappe, which means that it developed later than the cleavage of system I.

System III, the strikes of which (around 100°) are not parallel to the fold axes, appeared last, also due to a pair of forces and horizontal compression. Therefore, the system formed due to the overthrust of the upper parts of the Magura Nappe towards 10°.

Dip-slip faults. Dip-slip faults were commonly observed in the exposures. They include both normal and reverse faults, with a throw reaching several metres, and fissures about 1m wide, filled with breccia, fault gouge or cataclasites. Their inclination in relation to the beds change with lithology (e.g. Fig. 13A), consistent with the observations of Peacock and Sanderson (1992). In more ductile beds, the angles are about 60°, whereas in less ductile beds, they are about 40°. Such faults are common in the steep, reversed limbs of the synclines. Two sets of faults have been distinguished there: u_1 — comprising reverse faults, with a dominant trend of at 215°/78° in the Szczebel Syncline and 306°/79° in the Śnieżnica Syncline; and u_2 — comprising normal faults with a trend of 304°/30° in the Śnieżnica Syncline (Fig. 13A). The acute angle between the u_1 and u_2 sets is about 50°.

As in the case of the cleavage, the angular relations between the surfaces of the faults and the bedding, the senses of movement along the surfaces and the parallelism of the strikes of the planes of the u_1 and u_2 sets, all indicate that these faults originated by the action of a pair of forces in a plane perpendicular to bedding and by horizontal compression (Fig. 13B, C).

This suggests development by increased flexural slip related to the overturning of the southern limbs of the synclines.

Such faults are characteristic for steep, reversed fold limbs (Price, 1967; Perry, 1978), and the angles between the faults and the bedding indicate development by the imposition of a pair of forces and horizontal compression (Fig. 13B), taking place due to the shortening of this part of the Magura Nappe.

The development of cleavage systems II and III, as well as of the dip-slip faults of sets u_1 and u_2, has a direct relation with the steep, reversed setting of the beds (Konon, 1998*a*), associated with increased flexural slip. According to Tanner (1989), in pure elastic bending the differences between the values of stress in normal and reversed limbs depend on the values of flexural slip, which increases in the steep limbs where the dips exceed values of 60–70°.

The considerable differences in the values of stress between the normal and reverse limbs of folds were noted by Cloos (1947), observing the degree of deformation of ooids; by Mitra (1978), studying quartz grains; and by Coward (1984), who analysed folds.

Thus cleavage systems I and II as well as the u_1 and u_2 faults appeared as folds gradually developed, whereas cleavage system III appeared during overthrusting of the upper parts of the Magura Nappe.

INTERNAL STRUCTURE OF THE SLICES IN SILESIAN NAPPE

The slices are dominated by 192°/70° — striking beds (Fig. 4). Concentric folds occur in thick-bedded sandstone

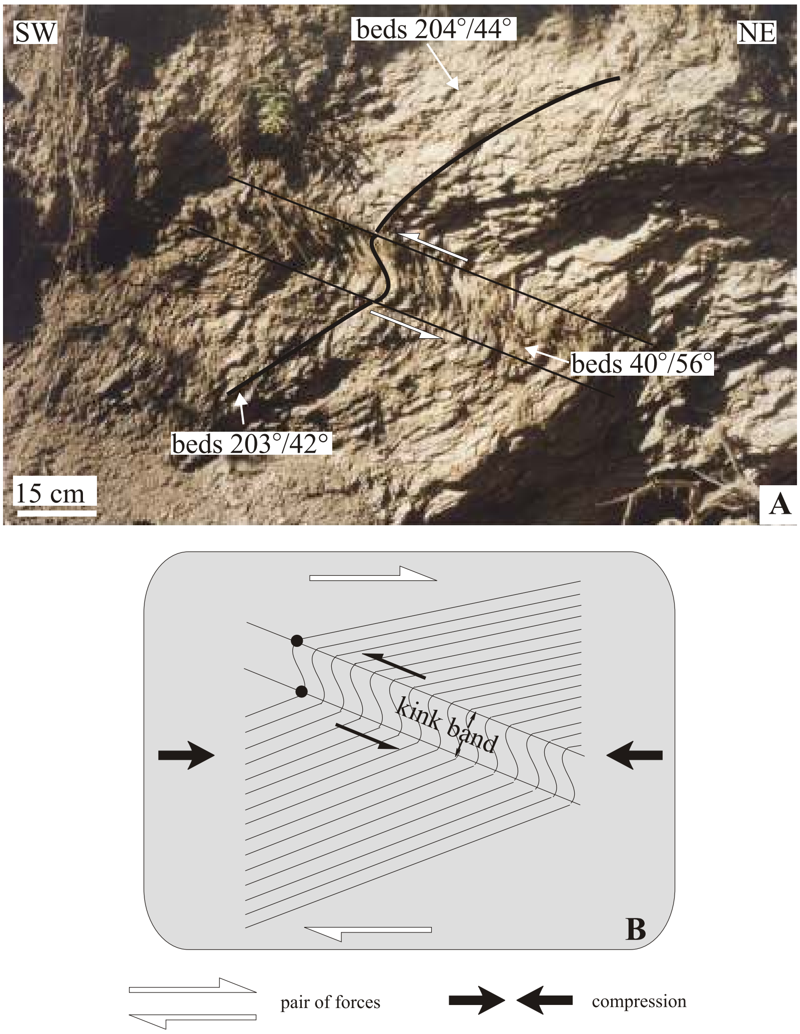

Fig. 14. Kink-folds in the Skrzydlna quarry

For location see Fig. 2

units, whereas similar folds are observed within thin-bedded sandstones and shales.

Slickensides were noted on the bedding planes, with tectonic striae approximately parallel to the dip, thus indicating flexural slip, which suggest that folds originated during flexure. The fold axes are concordant with the elongation of slices, and have a northern vergence.

Strikes of the thrusts separating the slices vary between 110–120° in the west, through parallel in the central part, to 80° in the east (Fig. 2). This pattern of slices may indicate an elevation in the basement, around which the overthrusting Silesian Nappe "flew". The presence of such elevation is suggested by the structural map of the Miocene basement of the Carpathian Foreland (Nowotarski *et al.*, 1994).

Folds up to several metres in amplitude, with acute hinges and arranged in bands (Fig. 14) were recognised within the Menilite Beds in the Skrzydlna quarry (Fig. 2). The axes of these folds lie between 124°–149°/9°–39°, with a concentration at about 130°/25°, whereas the dominant position of the band set is at 35°/30°. The bands of square-like deformed shales indicate (e.g. Paterson and Weiss, 1966) tendencies to shearing.

The fold geometry, numerous bedding planes of the Menilite Beds acting as anisotropy planes, and the occurrence of bands where shearing took place, indicates that the folds represent kink-folds, as described by e.g. Paterson and Weiss (1966), Verbeek (1978), or Williams and Price (1990).

Typically, (e.g. Paterson and Weiss, 1966), kink-folds originated during contraction, approximately parallel to the anisotropy plane, in a high surrounding pressure (e.g. Anderson, 1974; Gay and Weiss, 1974; Williams and Price, 1990), which impeded slip between the beds.

In experiments (Paterson and Weiss, 1966), two conjugate kink bands appeared, whereas only one kink-fold band appeared in this case. If the kink-folds described originated in typical conditions (Paterson and Weiss, 1966; Ramsay, 1967) of contraction near-parallel to bedding, they formed at about 30° to the axis of the greatest stress σ_1, which is rarer, according to e.g. Dewey (1966), than the most frequently accepted value of 45°–60° (e.g. Paterson and Weiss, 1966; Żelaźniewicz, 1976; Peacock, 1993). This indicates that the second band of kink-folds did not develop as a result of stress relief along the bedding within the Menilite Shales (part of the Menilite Beds),

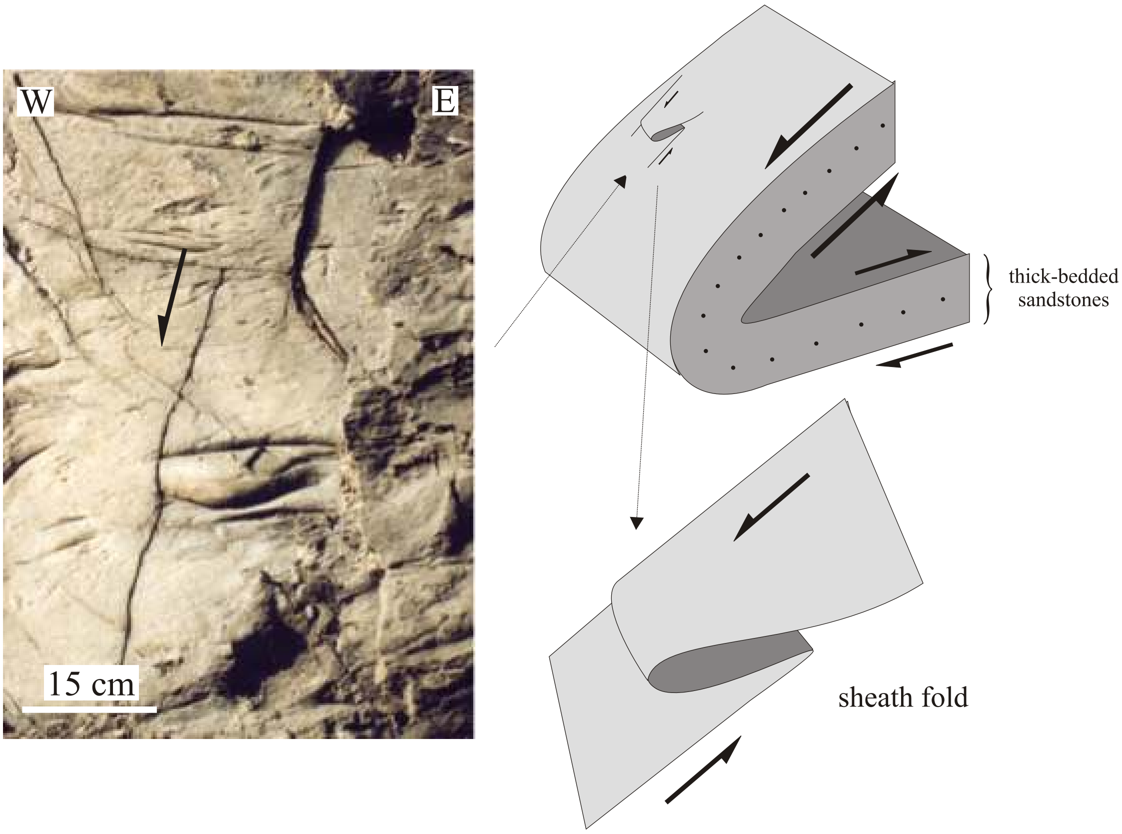

Fig. 15. Sheath folds in the Porąbka quarry

For location see Fig. 2

resembling cleavage sets described by Jaroszewski (1972) from the Holy Cross Mountains.

The folds most probably appeared in the terminal phase of deformation, as is common (e.g. Ramsay, 1962; Babaie and Speed, 1990), and as was described also from the nearby Mszana Dolna tectonic window (Mastella, 1988). Their vergence indicates formation during simple shear in a vertical plane of general direction SW–NE.

Other asymmetrical small folds, up to a dozen centimetres long and tending to deport from a cylindrical trend, characterised by a strongly curved axis, approximately parallel to the bedding (Fig. 15) were observed in the Porąbka quarry (Fig. 2). Such geometry suggests that they represent sheath folds (e.g. Cobbold and Quinquis, 1980; Ghosh and Sengupta, 1987; Price and Cosgrove, 1990). Their vergence is concordant with the flexural slip.

Cleavage is present within the steeply inclined thick-bedded sandstones of the Krosno Beds. The most common comprise the R_1 and R_1' sets, which occur in the southern limb (about 175°/70°) of the overturned syncline, exposed in the Stróża quarry (Fig. 2). They possess the geometry of reverse faults (Fig. 16). They correspond to system II of the Magura Nappe, and, like the latter, originated during flexural slip connected with horizontal compression due to overthrusting of the Silesian Nappe lying farther south, or even of parts of the Magura Nappe.

Dip-slip faults within the part of the Silesian Nappe investigated are represented by normal and reverse faults, with strikes of 60–85° (e.g. Fig. 16), independent of bed position. Two systems of faults have been distinguished, based on the dips and sense of movement along the dip planes: system I, which developed without the overloading of the Magura Nappe and which is linked with the overthrusting of the Silesian Nappe; and system II, which developed during overthrusting of the Magura Nappe and the resulting increase in loading.

The first fault system (I) comprises normal faults R — characterised by shallow dips of 25–33°N; and reverse faults R' — characterised by steep dips of 60–80°N (e.g. Fig. 16). R faults have fault zones up to several centimetres thick, with numerous

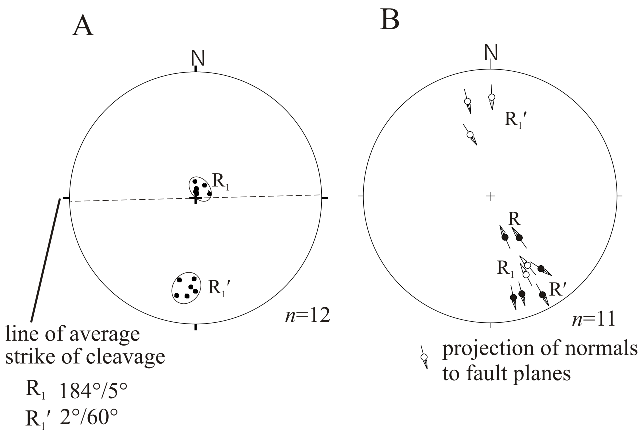

Fig. 16. **A** — diagrams of cleavage planes from the Stróża quarry, **B** — dip-slip faults from the Porąbka quarry

For location see Fig.2

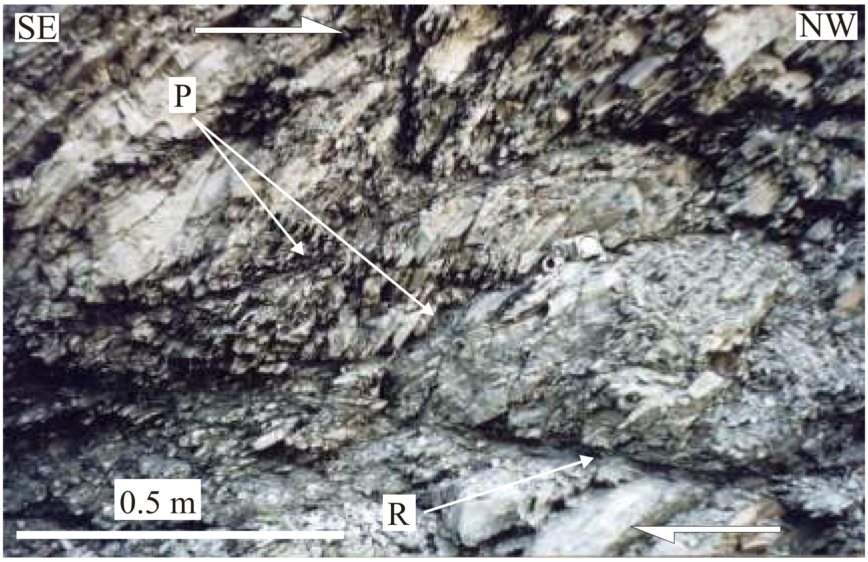

Fig. 17. Shear lenses in the Kasinczanka creek outcrop (about 8.7 km away from the Raba River)

slickensides, whereas R’ faults have fault zones up to 1 m thick. The acute angle between these sets is about 50–60°. This angle, senses of movement along the R and R’ planes, and the parallel strikes of their fault planes indicate that they represent faults corresponding to Riedel shears (Riedel, 1929; Barlett *et al*., 1981).

The R and R’ faults developed by simple shear in a vertical plane, consistent with the model of Vialon (1979) and Mastella (1988), under low loading, allowing widening towards the top. The development of the high-angle R’ set, with wider breccia zones, suggests opening of these faults as loading increased.

The second fault system (II) comprises normal faults R_1 of dips about 60°N and R_1’ of dips 55–70°S (e.g. Fig. 16). The acute angle between these sets is about 60°. The setting of these faults, senses of movement of their walls and the angle between them indicate that they appeared by the action of a pair of forces under the overloading of the Magura Nappe. The model of Vialon (1979) and Mastella (1988), suggests origination due to simple shear in a vertical plane, but in conditions of increased loading, most probably caused by the overthrust of the Magura Nappe on to this fragment of the Silesian Nappe.

The I and II fault systems in the Silesian Nappe appeared during overthrusting of the Magura Nappe, coevally with the formation of cleavages I and II in the Magura Nappe.

THRUSTS

Numerous shallow dislocations of different range occur in the area investigated, which following Elliott and Norris (1981) are referred to as thrusts. The most important, with a regional extent, is the Magura Nappe thrust. Lower-range overthrusts include the Beskid Wyspowy Subunit thrust, the northern and southern slices, the syncline zone thrust and the thrusts separating individual slices.

Magura Nappe thrust. This thrust can be traced over a distance of about 10.5 km in the region of the “Skrzydlna bay”. The oblique contact of the slice overthrusts with the Magura Nappe overthrust (Fig. 2) suggests that the northern border of the Magura Nappe thrust is erosional, consistent with the conclusions of Świderski (1953*b*) and Książkiewicz (1972).

The thrust zone is poorly exposed. In the field it can be traced by the presence of breccia and cataclasites locally over 12 m thick. Around the “Skrzydlna bay” it is recognised by lithological differences, forming morphological steps.

The Magura Nappe thrust is cut by numerous faults, passing through the Magura as well as the Silesian Nappe (Fig. 2). Analysis of intersection lines shows that the thrust plane is characterised by variable dip angles. In the vicinity of the “Skrzydlna bay”, the values reach about 30–40°S in its central part, whereas in the eastern and western part they reach about 20°S. This indicates that the Magura Nappe was thrust over an uneven basement, as also suggested by Połtowicz (1985) and Mastella (1988). The general direction of overthrusting of the Magura Nappe was approximately, from S to N, consistent with Książkiewicz (1972) and Mastella (1988).

LOWER-RANGE THRUSTS

Lower-range thrusts include those of mappable size, with displacements of up to several kilometres (the Beskid Wyspowy Subunit, northern and southern slices and syncline zone thrusts) or with small displacements, e.g. thrusts separating individual slices (Fig. 2). Analysis of minor structures, suggest overthrusting generally from S to N, with many local deflections from this direction, probably resulting from uneven basement.

The **Beskid Wyspowy Subunit thrust** occurs in the eastern part of the area investigated along the northern slices, where the “Skrzydlna bay” overlaps the Magura Nappe thrust, and in the western part of the area where it overlaps the syncline zone thrust (Fig. 2). The dip of this thrust is 20–40°S and the strike is about 90° (Fig. 2), as in the northern slices thrust, in the eastern parts of the area in the vicinity of the “Skrzydlna bay”. The general direction of displacement along these overthrusts was S to N.

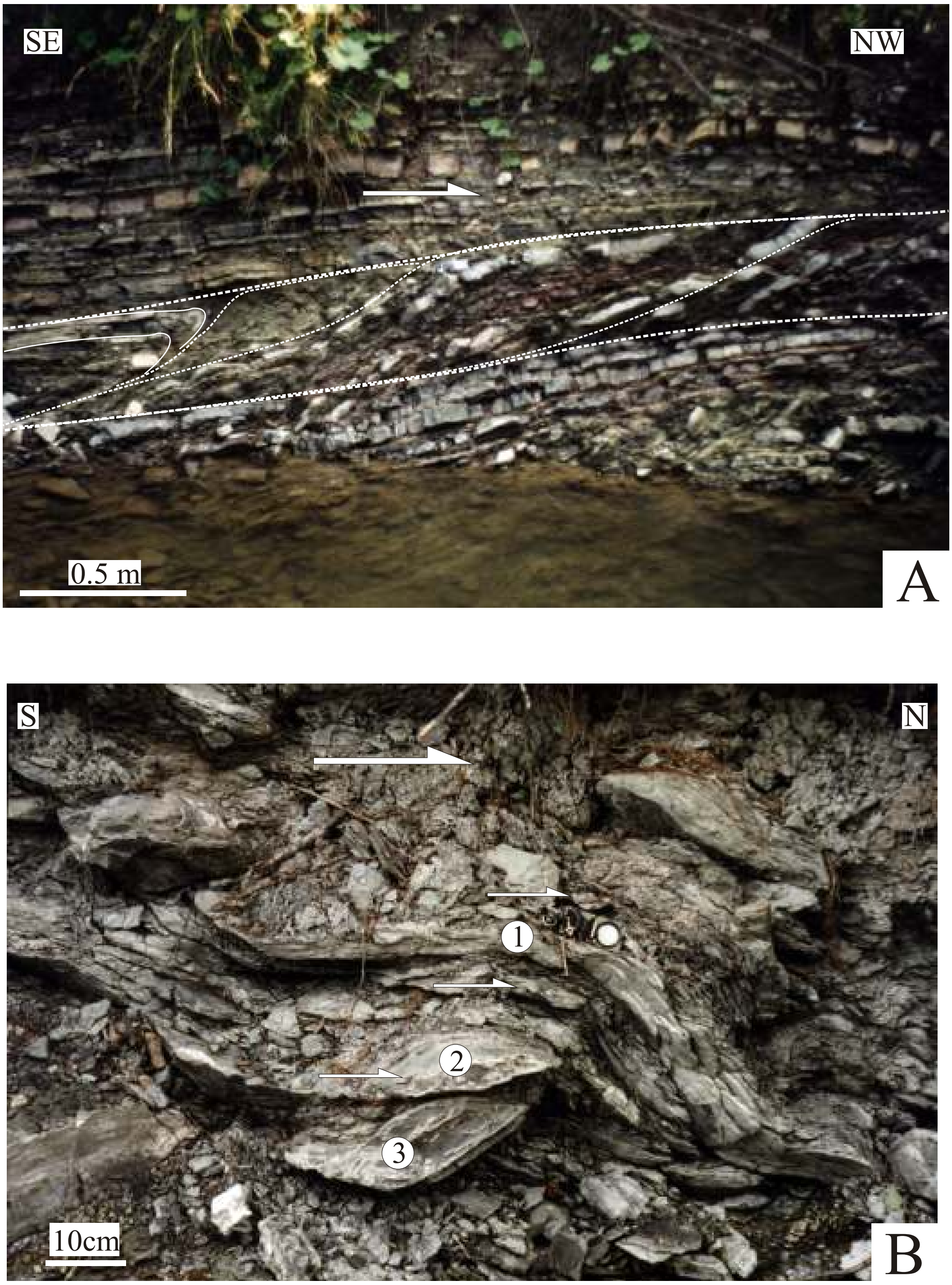

Fig. 18. **A** — duplex structure in the Krzczonówka creek (W of Klimas Syncline), **B** — antiformal-stack structure in the Słopniczanka River (1, 2, 3 — sequence of slices development)

Northwards from the Lubogoszcz, Śnieżnica and Łopień synclines, the thrusts of individual synclines pass smoothly into the thrust of the entire syncline zone (Fig. 2). The thrusts are distinctly marked in the southern synclines belt (Fig. 2). The thrust of the Ćwilin Syncline on the Śnieżnica Syncline, the thrusts of the Łopień Syncline on the northern slice zone and of the Luboń Wielki Syncline on the Klimas Syncline (Świderski, 1953*a*), passing probably into the Ćwilin thrust, are easily distinguishable (Fig. 2).

All the thrusts are characterised by dips of 15–40°S and mainly parallel strikes.

The **southern slices thrust**, easily distinguishable south of the Ćwilin and Łopień synclines (Fig. 2) dips at 25°S and its strike is also approximately parallel to the general trend (Fig. 2 and 6). Its continuation to the west is questionable, where it disappears at the margin of the Mszana Dolna tectonic window (Mastella, 1988) and probably continues beyond the area investigated south of the Luboń Wielki Syncline.

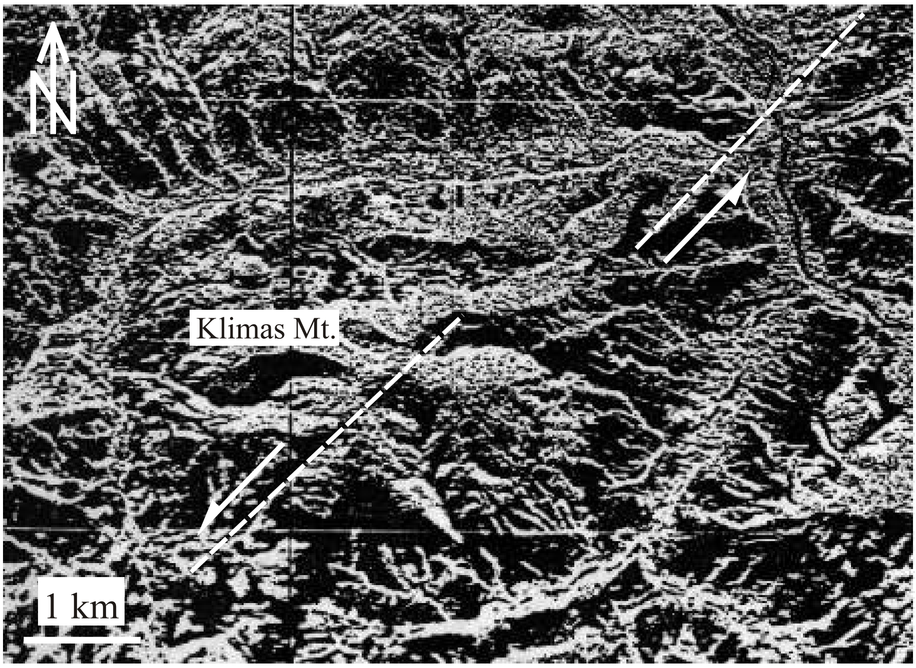

Fig. 19. Example of a sinistral strike-slip fault (radar image of the Klimas Mt.)

Apart from these thrusts, a series of much smaller displacements, with variable dips up to 50°S, can be distinguished. However, they generally flatten out. The strikes of these thrusts are very variable, from generally parallel to 110–140°, e.g. NW and SW of the Szczebel Syncline (Fig. 2). The general direction of tectonic transport along these displacements was SW to NE. The strikes of thrusts differ also from generally parallel in the eastern part of the "Skrzydlna bay", where their azimuths reach 70–80° (Fig. 2). Analysis of mesostructures along the thrusts indicates SSE to NNW thrusting.

Thrusts, the planes of which pass into bedding planes (e.g. Fig. 5 — section II–II), as is typical (Niño *et al.*, 1998) for the transfer of part of the stresses during thrusting movements, also occur sporadically in the area investigated. The presence of such thrusts is indicated by differences in setting directly above and below the thrust. Such blind thrusts (Thompson, 1979) are common, as confirmed by many authors, e.g. Thompson (1979, 1981) from the Rocky Mountains, and Berger and Johnson (1982) in the Appalachians.

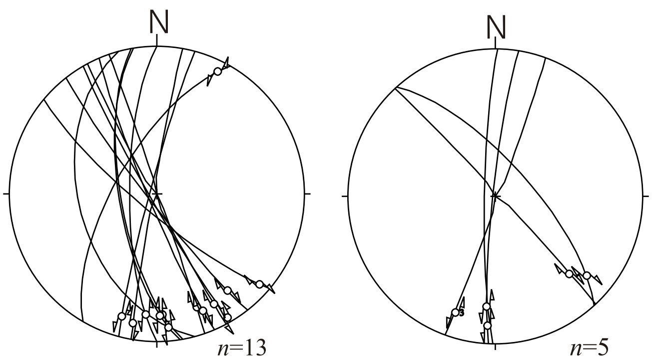

Fig. 20. Examples of strike-slip fault diagrams (Krzczonówka creek — N from the Klimas Syncline)

SETTING OF THE THRUST ZONES

The thrust zones, from about a dozen centimetres to several metres thick, comprise rock, cut by faults and fractures.

Mesostructures. The faults and fractures form occasionally, within less brecciated competent rocks, a characteristic lens-like arrangement. Such lens-shaped structures are best developed in the Beskid Wyspowy Subunit overthrusts, northwards from the Lubogoszcz Syncline (Fig. 2). The lenses are built of shales, and are from several tens of metres to about 2 m thick. Individual bodies are bordered by shears, striking 75–90°, represented by normal faults, dipping about 15°NW and with shears represented by reverse faults, dipping about 25°SE (Fig. 17). According to the terminology of Barlett *et al.* (1981), these represent R and P type shears. The P shears connecting the R shears occur in long series, forming shear lenses (Skempton, 1966 *bona vide* Groshong, 1988; Naylor *et al.*, 1986). The occurrence of R shears (Vialon, 1979; Mastella, 1988) and the presence of connected series of R–P shears indicates (Davison, 1994) a tendency to concentrate shearing in a narrow zone, that is, in the case of the structures described, under the loading of the upper parts of the Magura Nappe (Mastella, 1988).

Microstructures. No structures indicating the directions of tectonic transport were distinguished macroscopically in cataclasites, most commonly occurring as fault rocks comprising mainly clay minerals and quartz. SEM and optical microscope analysis of oriented, undisturbed samples, as in the investigations of Lash (1989) and Maltman (1988), indicate a high degree of arrangement of platy clay minerals along the shears (Konon, 2000). The observed shears in clayey cataclasites suggest the action of a pair of forces, along with vertical compression, caused by the influence of the overburden loading linked with the transport of rock masses of the upper parts of the Magura Nappe in a generally S to N direction (Konon, 2000).

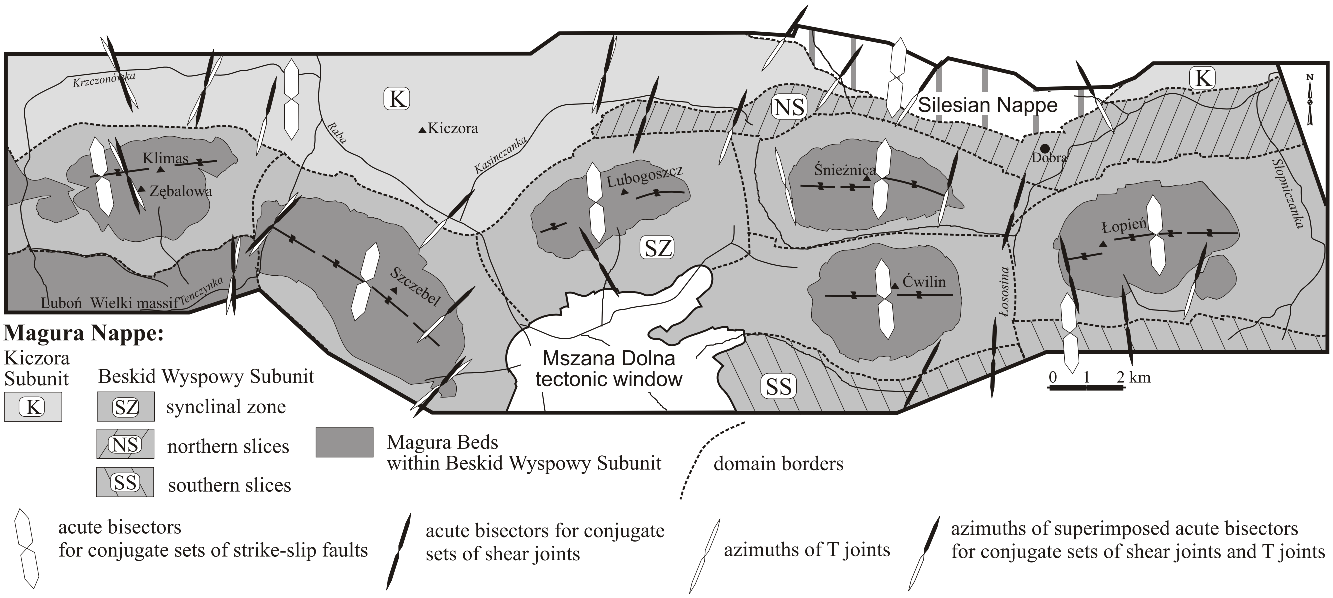

Fig. 21. Azimuths of acute bisectors for conjugate sets of strike-slip faults and for conjugate sets of shear joints and T joints

Other explanations see Fig. 4

Fault rocks. Fault rocks composed mainly of clay minerals, sporadically bearing larger fragments, up to 12 cm across, of less deformed rocks, have been observed in the thrust zones. According to Sibson (1977) and Wise *et al*. (1984), partly supplemented in relation to the internal direction by Groshong (1988), the fault rocks in the investigated area represent breccia and cataclasites formed during brittle deformation (Konon, 1998*b*, 2000).

Duplexes. A series of imbricated slices were recognised within the Magura and Silesian Nappe. In the Magura Nappe these represent larger zones, up to 2–3 km wide and several tens of kilometres long: the northern slices zone (with at least 5 slices) and the southern slices zone (with at least 2 slices), whereas in the Silesian Nappe they represent a slice zone (with at least 9 slices) within the “Skrzydlna bay”.

Apart from these, smaller slices, several kilometres long and wide, were distinguished in the SW part of the southern limb of the Klimas Syncline, to the south of the Szczebel Syncline and in the NE part of the northern limb of the Łopień Syncline (Fig. 2). The slices from these zones have a sigmoidal shape (Fig. 18A) in N–S cross-section and a dominant inclination of 30–70°S.

Much secondary deformation took place within the slices (Figs. 5 and 6), mainly in those places, where the formation of duplexes encompassed the Inocermian Beds, Variegated Shales and the Hieroglyphic Beds, comprising mainly shales and fine- to medium-bedded sandstones, where many slip zones were activated. Fragments of unfolded rocks occur only sporadically. The overlap of back-slices on the preceding slices was also observed in some cases (Fig. 18B), a result of the differences in transport velocity of individual rock packets.

Faults determined between slices represent reverse faults. Moreover, the slices occur under larger thrusts, e.g. the northern slices zone occurs underneath the synclines zone overthrust (Fig. 2). These features indicate that sets of such slices, as described earlier from the Outer Carpathians by Teisseyre (1921), can be at present referred to as duplexes (in the sense of Dahlstrom, 1970) (Mastella, 1988; Aleksandrowski, 1992; Konon, 1996, 1997; Mastella and Rubinkiewicz, 1998).

Sets of horses were developed consistent with the most common model (Mitra, 1986; Mitra and Boyer, 1986; Bowler, 1987; Averbuch and Mansy, 1998) of the kinematic development of a duplex (Boyer and Elliott, 1982). The duplexes thus formed in contraction conditions, as is typical (Woodcock and Fisher, 1986) for contraction duplexes, where an imbricate stack develops beneath the roof sequence along the main initially developed thrust (Boyer and Elliott, 1982; Averbuch and Mansy, 1998).

The formation of a series of duplexes was probably induced by impeded slip in the displaced rock packets. This resulted in the formation of flat slices, and, in cases of strong resistance, to their steepening. In consequence, according to the model of Boyer and Elliott (1982), earlier slices overlapped the later ones (Fig. 18B), thus a more advanced duplex structure — an antiformal stack (e.g. Boyer and Elliott, 1982; Morley, 1986) was formed.

The impedance to slip might have been partly caused by an increase in internal friction, or by the occurrence of steps in the basement of the overthrusting nappes, the presence of which in the area investigated has been suggested by e.g. Kozikowski (1958) and Mastella (1988), and accords with Knipe (1985) as regards more intense flaking of beds where the nappe passed over steps in the basement.

STRIKE-SLIP FAULTS

The Magura and Silesian nappes are cut by numerous strike-slip faults. The sizes of these faults range from map-scale

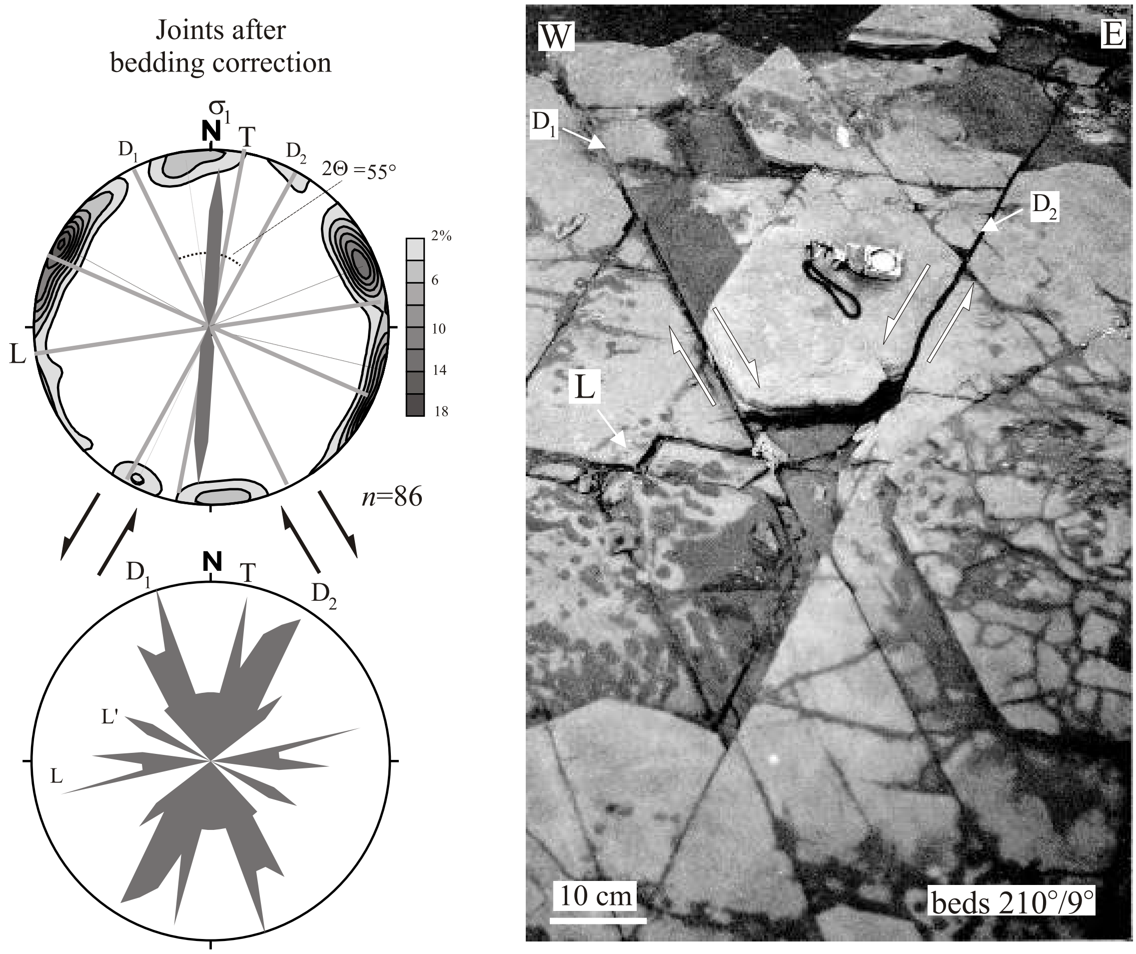

Fig. 22. Joints in sandstone beds in the Łososina River — south slices (SE from the Ćwilin Syncline)

to those distinguishable only in individual exposures. In exposures these faults are marked as zones up to several metres wide, filled with fault rocks — fault gouge, cataclasites or tectonic breccia. These faults were observed on air-photos and radar images (e.g. Fig. 19) and form narrow steep-sided ravines.

Strike-slip movement along the faults was inferred from the presence of slickensides with horizontal tectonic striae (e.g. Fig. 20), the en echelon arrangement of the faults (Fig. 19), sigmoidal bending of beds, vertical axes of folds occurring on both sides of the faults, and rotation of fragments of sandstone beds in the fault zones. Younger tectonic striae, characterised by steep attitudes, have been noted on some of the strike-slip faults planes, thus indicating the reactivation of the faults in a later phase as dip-slip faults.

Recent travertine has been observed around the faults, mainly within the Kiczora Subunit and the Silesian Nappe (Fig. 2). This might indicate, as suggested by Gruszczyński and Mastella (1986) from the Mszana Dolna tectonic window, continued activity along the fault zones. Attitudes of some of these faults agree with the dislocations already marked on maps (Mastella, 1988) or discovered by geophysical investigations (Doktór *et al.*, 1987; Nowotarski *et al.*, 1994).

Strike-slip faults cut folds and in some cases even several slices, and some of them cut the thrust of the Magura Nappe on the Silesian Nappe. Therefore, most of these faults represent post-folding and post-thrust faults (Fig. 2), as indicated by other investigations (e.g. Tokarski, 1975; Mastella, 1988; Aleksandrowski, 1989; Mastella and Szynkaruk, 1998).

The strike-slip faults form a dense, regular net comprising sets of sinistral faults, with trends between 5 and 45°, and dextral faults, with trends between 150° and N–S (e.g. Fig. 20). Strike-slip faults of both sets cut one another at an angle of about 50°. Following Freund (1974), the senses of movement along their planes and the angle between them indicate that both sets form a conjugate sets of faults, formed during pure shear.

The directions of horizontal compression calculated from the two sets generally reach values between 350–20° (Fig. 21).

The strike-slip faults here originated due to horizontal compression, in an apparently stable N–S direction (Książkiewicz, 1972). The variability of compression directions, calculated for domains encompassing individual synclines within the syncline zone (Fig. 4) indicates slight rotation. This suggests that, from the moment of fault formation, the Klimas Syncline underwent counter-clockwise rotation of about 10°, whereas the Lubogoszcz, Ćwilin and Łopień synclines and part of the zones from the Kiczora Subunit and the southern slices underwent a counter-clockwise rotation of about 5°. The Szczebel Syncline probably underwent a clockwise rotation about 15°, and the Śnieżnica Syncline about 5°.

Tectonic structures	Synclines					
	Klimas	Szczebel	Lubogoszcz	Śnieżnica	Łopień	? Ćwilin
Joints	17°	42°	28°	13°	17°	?
Fold axis	15°	35-40°	20°	10°	10°	0°
Strike-slip faults	10°	15°	5°	5°	5°	? 5°

direction of domain rotation; present direction of maximal stress determined from structural data

Fig. 23. Direction of domain (syncline massifs) rotation (determined from analysis of joints, fold axes and strike-slip faults)

JOINTS

Numerous fractures cutting singular beds, perpendicular to them and at distances more or less similar to the bed thickness have been observed within the Magura and Silesian Nappe (Mastella, 1972). These fractures form a regular regional network. They show little or no displacement and are joints in the sense of Price (1959), Hancock (1964) and Jaroszewski (1972).

The joint pattern was investigated using the methodology of Jaroszewski (1972), Mastella *et al.* (1997), Zuchiewicz (1997, 1998), Mastella and Zuchiewicz (2000).

The initial, pre-folding, position of joints was reconstructed (Price, 1959; Jaroszewski, 1972). Thus joint-bearing beds were rotated to a horizontal position, as in Hancock and Al-Kadhi (1982) and Mastella *et al.* (1997).

Joint sets were classified according to their geometric relation to present-day fold axes. As in Mastella *et al.* (1997) and Mastella and Zuchniewicz (2000), two sets of cross-fold joints were investigated: diagonal and approximately perpendicular to map-scale fold axes (e.g. Fig. 22).

Diagonal joints. Diagonal joints occur in two sets: D_1 — dextral and D_2 — sinistral. The diagonal joint is characterised by generally smooth fracture planes. Occasionally, the fractures are filled with calcite. Fringe structures rarely occur, which pass on bedding planes into *en echelon* fractures. The sets cut one another at 50–60° (Fig. 22) and in some cases pass into one another in form of arcs, indicating that the jointing has a shear and conjugate character (Jaroszewski, 1972). This allows calculation of the axis of maximal stress σ_1 of the stress field, in which they formed (Fig. 21).

Perpendicular joints. Perpendicular joints (T) are much less common and are mostly uneven. Feather fractures are often present on the planes, similar to those observed by Bahat and Engelder (1984) in the Appalachians. The fissures are often open and filled with calcite. Planes of perpendicular joints cut and thus post-date the diagonal joint sets. The fractures originated in a horizontal position of the σ_1 axis and negative, horizontal σ_3 axis (Price, 1959).

Analysis of palaeostress. The azimuths of the bisector of the angle between the diagonal joints D_1 and D_2 as well as the azimuths of the joints T reveal small deflections, ranging from 2 to 10 degrees in the same exposures (Fig. 21). Although the directions of the σ_1 axis, calculated from analysis of the diagonal and perpendicular joints, show a strong scatter in different domains, the differences are insignificant within individual domains (Fig. 21).

The joints originated in a regional stress field, the diagonal being first (preceding the folding), followed by the perpendicular joints (Rubinkiewicz, 1998; Mastella and Zuchiewicz, 2000). Assuming that the direction of horizontal compression for this part of the Outer Carpathians was constant (N–S) (Książkiewicz, 1972), deflection from this direction within individual domains, encompassing entire synclines within the syncline zone in the Beskid Wyspowy Subunit, occurred as the joint sets formed, more or less coevally with the formation of folds and strike-slip faults (Fig. 23). Therefore movement of entire fragments of the Beskid Wyspowy is postulated.

TECTONIC EVOLUTION OF THE CENTRAL PART OF BESKID WYSPOWY

This study suggests that the geological evolution of the Beskid Wyspowy is more complicated than previously postulated (Świderski, 1953*a*, *b*; Kozikowski, 1953, 1972).

The Łopień, Śnieżnica, Ćwilin, Lubogoszcz, Szczebel and Klimas synclines represent characteristic individual synclinal massifs, strongly contrasting with the belt-like arrangement of regional folds, occurring to the west (Burtan *et al.*, 1981) and east of the latter (Książkiewicz, 1971; Golonka *et al.*, 1979).

This structural and cartographic analysis as well as earlier papers (Świderski, 1953*a*, *b*; Kozikowski, 1953, 1972; Burtan, 1974, 1978) indicate that the formation of these massifs was controlled by lithological differences. The thick-bedded Magura Sandstones filling the syncline axes contrast with the underlying successions comprising shales and fine- to medium-bedded sandstones of the Hieroglyphic Beds, Variegated Shales and partly also of the Inoceramian Beds. Differences in the competence of rocks facilitated the separation of the syncline belt into smaller fragments, brachysynclinal in character, in later phases of development.

1. Laterally elongated fold belts several tens of kilometres long comprising wide synclines and narrower anticlines, appeared in the first phase of the thrusting of the Magura Nappe

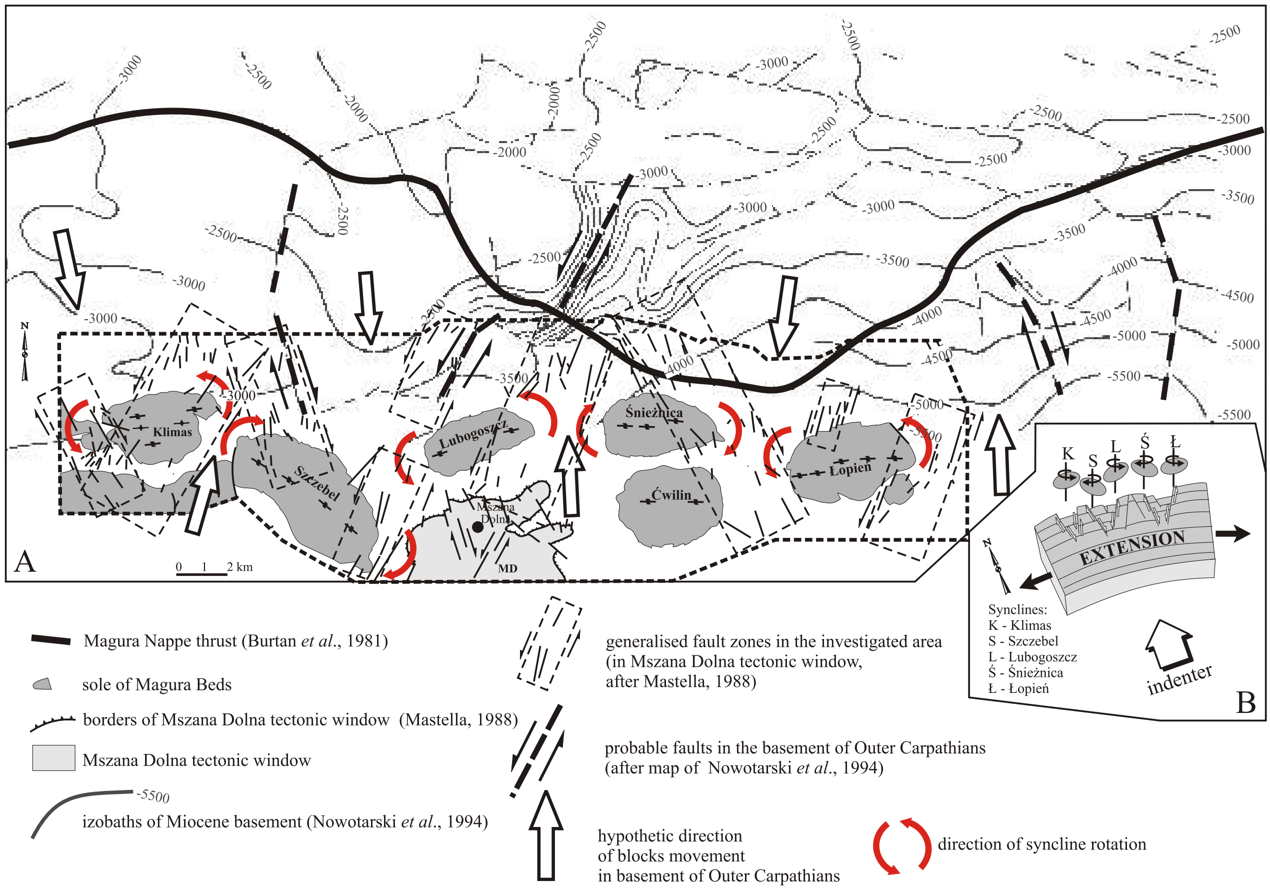

Fig. 24. Syncline rotation in the Beskid Wyspowy Subunit

under dominant N–S horizontal compression, (Książkiewicz, 1972). The Beskid Wyspowy Subunit represented at that time one of these wide, continuous belts of synclines.

2. During the next phase of development, after the transformation of horizontal compression into a pair of forces acting in a vertical plane, the folds developed a northern vergence. With stopping of the face of the overthrusting Magura Nappe and continuous stress of its overthrusting southern parts, thrusts separating the particular subunits and duplexes were formed, i.e. the northern and southern slices zone in the Beskid Wyspowy Subunit. Formation of duplexes took place also during the thrust of the Magura Nappe over the Silesian Nappe. In effect, the southern limbs in some synclines underwent overturning.

3. In the following phase of development, in an approximately N–S horizontal stress field, strike-slip faults originated. In effect, the syncline belt in the Beskid Wyspowy Subunit was torn apart. The individual blocks became independent, as in the case of boudinage formation suggested by Kibitlewski (1989) for rock massifs of the Czorsztyn succession or by Gates (1996) for megaboudines in the Hudson Highlands Unit.

4. As a result of the displacement along these zones of strike-slip faults, correlated with the dislocations interpreted (Fig. 24A) from maps of the Miocene basement of the Carpathians (Nowotarski *et al*., 1994), the rotation of blocks encompassing individual synclinal massifs took place.

The succession of the structures (joints–folds–slices–strike-slip and dip-slip faults) and the directions of stress calculated on their basis (Fig. 23), indicate that the rotation of blocks had a continuous character.

EVOLUTION OF THE MAGURA NAPPE IN THE BESKID WYSPOWY REGION IN A CONTEXT OF THE TECTOGENESIS OF THE WESTERN OUTER CARPATHIANS

The collision of the African and Eurasiatic Plates, between which several microplates were present (Royden *et al*., 1982), caused the formation of the Outer Carpathians (Burchfiel and Royden, 1982; Royden *et al*., 1982; Royden, 1988).

1. The displacement of one of these microplates (a fragment of Pannonia — Royden, 1988; N Pannonian — Alcapa, Csontos *et al*., 1992) within the western Outer Carpathians, in the central part of the Polish Carpathians, caused approximately N–S compression from the Oligocene onward (e.g.

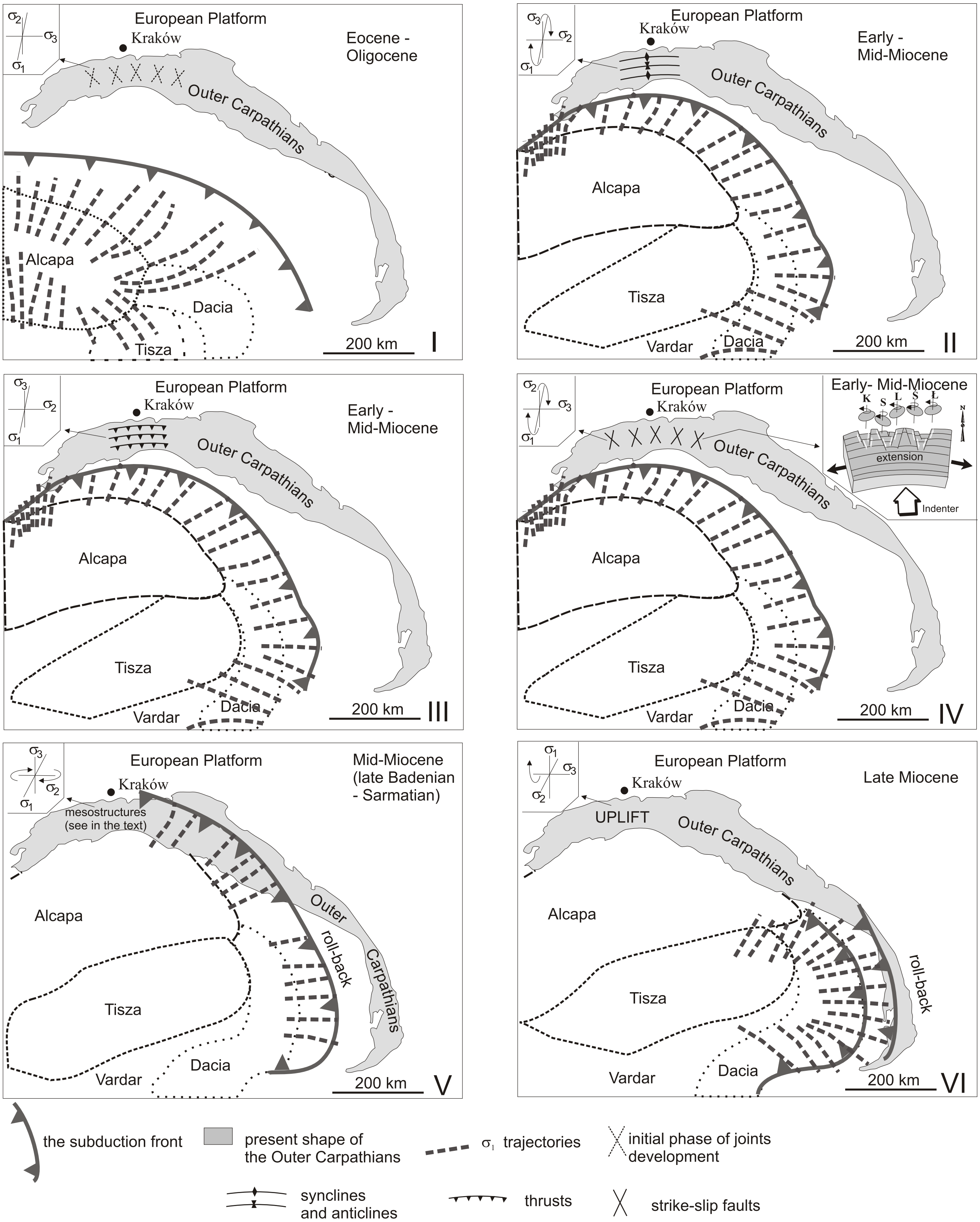

Fig. 25. Evolution of the Beskid Wyspowy Mountains in the view of the tectogenesis of the Carpathian region

Sketch of evolution major blocks and σ_1 trajectories in the Carpathian region after Fodor *et al*. (1999); synclines within the Beskid Wyspowy Mountains: K — Klimas, S — Szczebel, L — Lubogoszcz, Ś — Śnieżnica, Ł — Łopień; principal stress — $\sigma_1 > \sigma_2 > \sigma_3$; other explanation see in the text

Książkiewicz, 1972; Birkenmajer, 1985; Cieszkowski *et al*., 1992), and this is linked with the initial phase of joint development in the Beskid Wyspowy region (Fig. 25I).

2. The further movement of this microplate to the north (Royden, 1988) increased compression in the Early Miocene, which resulted in the overthrusting of the Magura Nappe and initial folding (Fig. 25II).

3. The displacement of the microplate was accompanied by advance of the Eurasiatic Plate (Tapponnier, 1977; Tokarski, 1978; Royden, 1988). In the Miocene, this caused the transformation of horizontal compression into the action of a pair of forces in a vertical plane, with the sense of the main component S to N, and this resulted in the formation of lower-range thrusts, duplexes and the overturning of some fold limbs (Fig. 25III).

4. Collision with the Eurasiatic Plate resulted in indentation within the earlier formed accretionary margin, thus probably causing bending of the deformed Carpathian orogen. As a result, a fan-like concentric regional arrangement of stress trajectories, propagating radially, with uneven marginal trajectories (Aleksandrowski, 1989; Nemčok, 1993; Nemčok and Nemčok, 1994; Mastella *et al*., 1997; Szczęsny, 1998) appeared, due to which a fan-like network of strike-slip faults, similar to those recognised by Laubscher (1972) in the Alpine foreland in the Jura Mountains, Tapponnier and Molnar (1976) in eastern Asia, or Angelier *et al*. (1986) and Huchon *et al.* (1986) on Taiwan, developed in the Outer Carpathians.

This regional arrangement of stress trajectories, formed in conditions making possible a large range of a uniform stress field (Jaroszewski, 1994; Jarosiński, 1998, 1999), caused rotation of blocks in the bended orogenic belts; counter-clockwise in the left part of the deformation belt, and clockwise in the right part of the deformation belt (Doglioni, 1995).

Such movements of blocks in the nappes basement point to tectonic bending of the Carpathian Arc due to the indentation of a continental microplate from the south within the accretion margin, most probably by its convex margin, as suggested by experiments of Zweigel (1998). In consequence, strong extension took place in the zone in which the brachysynclines of the Beskid Wyspowy formed (Fig. 24). Such extension is common in many collisional orogenic belts and in their forelands (Hancock and Bevan, 1987). This mechanism was noted, e.g. in the Cantabrian Nappes belt (Julivert and Arboleya, 1984), in the Aquitaine basin — in the foreland of the thrusted Pyrenees, where extension is sub-parallel to the margin of the Pyrenees (Hancock and Bevan, 1987), within the Western Alps (Dietrich, 1989) and in the Apennines (Doglioni, 1995).

The presence of extensional stresses in the outer zone of the Carpathian Arc, causing increase in its length, suggests the occurrence of a bending mechanism during the formation of the Carpathians, as confirmed by Nemčok *et al.* (1998*a*).

This mechanism of bending of the Carpathian Arc during its formation may be inferred from analysis of the tectonic data of e.g., Birkenmajer (1979, 1985), Kibitlewski (1989), Jurewicz (1997), and Nemčok and Nemčok (1994), from the Pieniny Klippen Belt; Marko (1993) in the Muráň fault zone, Nemčok (1993) and Nemčok *et al*. (1998*a*, *b*) on areas directly to the north and south of the Pieniny Klippen Belt, as well as by Mastella and Szynkaruk (1998) and Mastella and Zuchiewicz (2000) in selected areas of the Outer Carpathians.

The tectonic bending of the Carpathians caused the disintegration of the syncline belt in the northern zone and the formation of isolated synclinal massifs in the Beskid Wyspowy Subunit (Fig. 25IV).

5. A change of compression from N–S to approximately SSW–NNE (documented by the origin of the III system of cleavage and kink-folds) started to take place in the next phase, probably linked with the lateral movement of part of the Eastern Alps (Ratschbacher *et al.*, 1989, 1991) as well as the Carpathians (Nemčok, 1993) towards the east, taking place from the Badenian onwards. This was probably directly related to with the relocation of the subduction zone towards the Eastern Carpathians by a roll-back mechanism (Burchfiel and Royden, 1982; Nemčok *et al*., 1998*a*, *b*; Fodor *et al.*, 1999) (Fig. 25V).

6. In the terminal part of this phase, from the Sarmatian, the faults changed their character from strike-slip to dip-slip faults as a result of uplift of this part of the Carpathians (Książkiewicz, 1972; Mastella, 1988) (Fig. 25VI).

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