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STUCTURAL EVOLUTION OF THE GNIEŹDZISKA SYNCLINE – REGIONAL IMPLICATIONS FOR THE SW MESOZOIC MARGIN OF THE HOLY CROSS MOUNTAINS (CENTRAL POLAND)

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Abstract: Basing on analysis of tectonic mesostructures, the structural evolution stages of the Gnieździska Syncline have been determined. The structure represents a typical example of folds occurring in the SW margin of the Holy Cross Mountains. The well-exposed syncline displays a wide variety of structures, including: shear and extension joints, stylolites, cleavage, strike-slip and dip-slip faults as well as master joints. Tectonic structures resulting from flexural slip indicate that the Gnieździska Syncline developed as a flexural-slip fold as a result of horizontal NE-SW compression. The subsequent deformation phase included mesostructures pointing to the increasing activity of a nearby Gnieździska–Wola Morawiecka dextral strike-slip fault of regional extent. In the terminal phase of the post-kinematic uplift of the Holy Cross Mountains, T joints and master joints appeared.

Key words: mechanisms of fold deformation, palaeostress reconstruction, Holy Cross Mountains, Poland.

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INTRODUCTION

This paper focuses on the determination of the geometry, as well as mechanisms and stages leading to the formation of the Gnieździska Syncline, as well as on the conclusions arising for the tectonic pattern of the south-western Mesozoic margin of the Holy Cross Mountains.

The Gnieździska Syncline is one of the synclines occurring in a belt along the northern border of the SW Margin of the Holy Cross Mountains (Stupnicka, 1972) (Fig. 1). In the eastern part, the synclines include the Bolmin Syncline, Góra Leśna Syncline and Ostrów Syncline. Westwards, the Gnieździska Syncline, referred to also as the Krasocin Syncline (Filonowicz & Lindner, 1987), passes into the Mnin Syncline (Lewiński, 1912) (Fig. 1B).

The area of the Gnieździska Syncline was subject to many detailed cartographic works on the scale 1:10 000 (Sadkowska, 2000) and on the scale 1:25 000 (the Piekoszów C sheet – Ozimkowski *et al.*, 1999). The surveys were also supplemented by photointerpretation of aeral photographs on the scale 1:18 000 and radar images on the scale 1:100 000. The surveys became the starting point for the structural analysis of the entire area. In point and contour diagrams, the orientations of structural planes (bed attitude, fault planes etc.) are marked as dip directions/dip projected on the lower hemisphere of the Schmidt net. In the analysis of strike-slip faults, the P-b-t axes were calculated for each fault plane according to Turner (1953) and Wallbrecher (1986).

LITHOSTRATIGRAPHY

The rocks occurring in the investigated area (Fig. 2) represent only a fragment of the stratigraphic column of the western Mesozoic margin of the HCM in this region (Filonowicz & Lindner, 1986). The oldest strata are represented by clays, claystones and siltstones of Rhaetian age. Due to their brown-reddish colour they are marked well in the debris. They are overlain, with an erosional gap, by black clays of Batonian (Filonowicz & Lindner, 1986) and Lowermost Callovian age (Barski, 1999). The strata occur in form of two belts in the south-western and north-eastern part of the arȩa (Fig. 3). In the eastern part of the latter they are reduced by a fault (Fig. 3). They crop out only within a quarry on Dybkowa Hill, squeezed out through fractures in those places were the exploitation reached the lowermost part of the Callovian gaizes. Rocks of Callovian age are represented by calcareous gaizes with cherts, yellow in colour and locally with pink or green-greyish coatings, with belemnites and borings. They are intercalated by rare thin-bedded marly shales (Fig. 2) (Matyja *et al.*, 1996). Their up to 24-

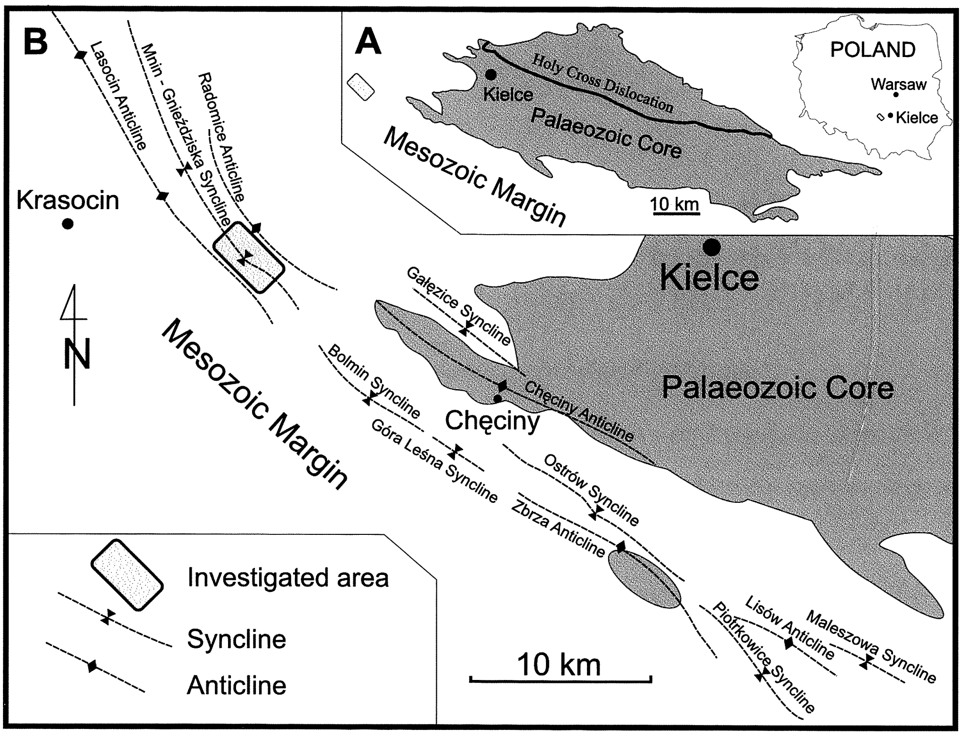


Fig. 1. Location map

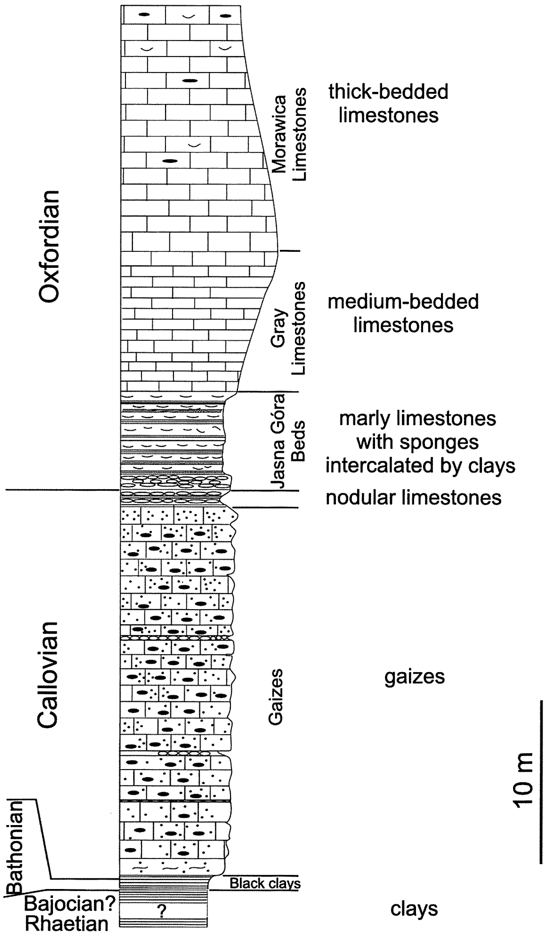


Fig. 2. Lithostratigraphic column of the Mesozoic beds in investigated area – based on Matyja *et al.* (1996)

m-thick complete section is exposed in a quarry on Dybkowa Hill (Fig. 3).

The Oxfordian (Matyja *et al.*, 1996) begins with thin-bedded, white marly limestones, passing into light-grey, bedded platty limestones, intercalated by marly shales (Jasna Góra beds) (Fig. 2), overlain by bedded, grey medium- and thick-bedded limestones (Grey Limestones) passing continuously into thick- and medium-bedded spotty limestones with numerous sponges (Morawica Limestones). Locally, massive limestones occur within them, exposed on the Lipia and Dębowa hills (Fig. 3).

TECTONICS

The resistant to erosion Oxfordian deposits along, with the underlying Callovian gaizes play the main role in the architecture of the Gnieździska syncline. They compose two belts of hills representing the north-eastern and south-western limbs of the syncline (Fig. 3). The north-eastern limb is dominated by 240/18 beds (Figs 4, 5), although in the eastern part of the limb the dips increase to ca. 30/S along the fault cutting the limb from the north (Figs 3, 5). In turn, bed strikes in the fault zones are rotated and reach ca. 155° (Figs 3, 5). The south-western limb is distinctly steeper. 46/56 beds dominate, with a small subdominant of 28/42, characteristic for beds in the eastern part of this limb.

Within the 5 km long analysed part of the axis zone, the width of the syncline measured between the Callovian/Oxfordian boundary beds in both limbs varies from 600 to 700 m (Fig. 5). The syncline axis in the analysed part is NW–SE with a tendency to plunging towards the NW (Fig. 3). The position of the axial surface varies slightly from ca. 220/70

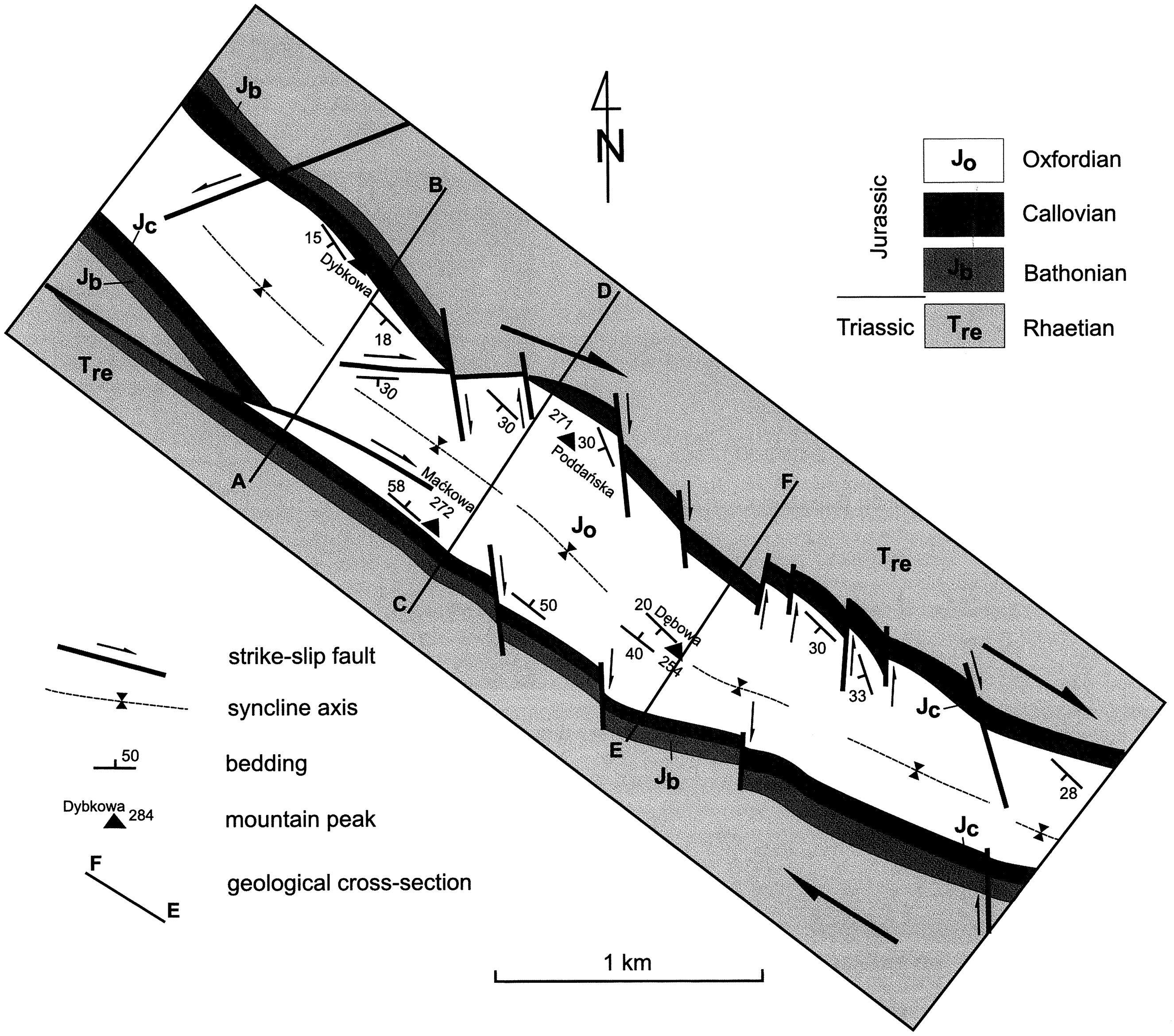


Fig. 3. Tectonic map of the investigated area

in the NW to ca. 220/85 in the SE, and the interlimb angle is 110° (Fig. 5). The structure represents therefore, an inclined fold, poorly advanced in tectonic development according to the criteria of Ramsay (1974).

MESOSTRUCTURES

A wide spectrum of tectonic mesostructures occurs in both limbs of the Gnieździska Syncline. Among them only joints, stylolites, cleavage, slickensides and master joints are frequent enough to enable structural analysis. Joints and stylolites were particularly useful due to their penetrativeness on the local scale.

Joints

Systematic fractures meeting the definition (Jaroszewski, 1972; Dune & Hancock, 1994; Mastella & Zuchiewicz, 2000) of joints (Fig. 6) occur in Callovian and Oxfordian strata. The joint network comprises four sets: two diagonal (S_R and S_L), and approximately sub-perpendicular (T) and sub-parallell (L) to the Gnieździska Syncline axis (Figs 6, 7). Such a pattern is typical for folded complexes of brittle rocks (Cosgrove & Ameen, 2000), and in regional scale also for the Holy Cross Mountains (Jaroszewski, 1972; Stupnicka, 1972).

Following Price (1959, 1966) and Jaroszewski (1972), the gravitational-relaxational, at incipient stage prefolding origin of joints is adapted here. Therefore, the orientation of joint sets, similarly as in Murray (1967) and Mastella & Zuchiewicz (2000), was determined after bedding correction.

S_R joints cluster at N10–20°E, with the dominant N15°E, whereas S_L joints cluster at N65-85°E with the dominant of N75°E (Figs 6, 7). The acute angle at which they intersect varies at particular outcrops within a narrow range of 50–65°. The acute bisector orientation, concordant with the orientation of σ_1, is NE–SW. Joint surfaces of both sets are smooth, and traces of their intersection with the bed-

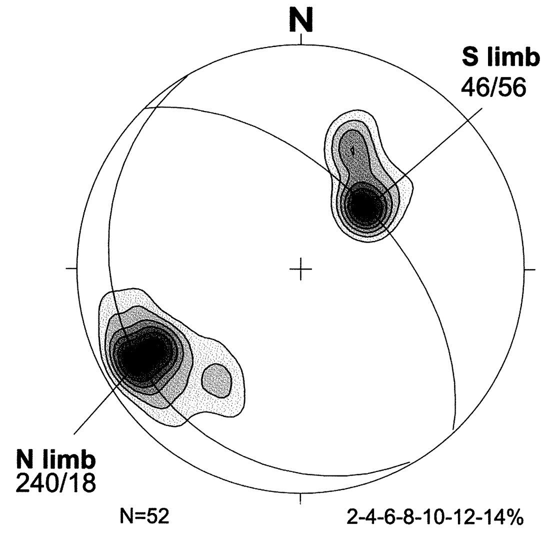


Fig. 4. Diagram of bedding in the Gnieździska Syncline

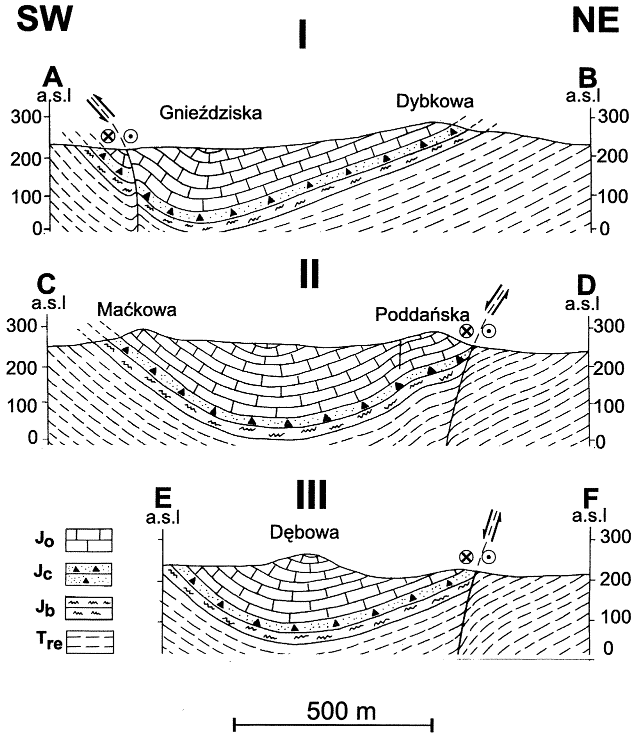


Fig. 5. Geological cross-sections

ding are rectilinear (Fig. 6). Sporadically, fringe structures and feather fractures occur. Their pattern indicates that at the incipient phase (Price, 1959) they were formed as shear joints. S_R joints retained a tendency to dextral, whereas S_L joints – to sinistral displacements (Fig. 6).

The sense of displacement and intersection at angles of ca. $2\Theta = 60°$ indicate that they represent a typical (Jaroszewski, 1972; Mandl, 1988; Mastella & Zuchiewicz, 2000) conjugate shear system. After Price (1959, 1966), it can be concluded that as in the case of the NE Mesozoic margin of the Holy Cross Mountains (Jaroszewski, 1972), this system was initiated in a shear triaxial stress field ($\sigma_1>\sigma_2>\sigma_3$) with horizontal σ_1 and σ_3, whereas its opening began already in extension conditions with accompanying residual stresses.

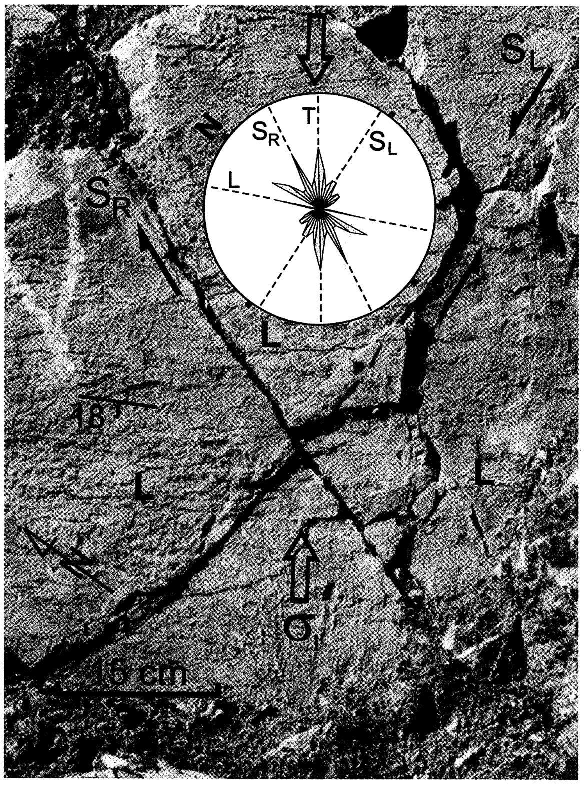


Fig. 6. Joints in limestones in the quarry in the Dybkowa Mt.

T joints are the most common joints in both limbs of the syncline (Fig. 7). Their orientation clusters at azimuths N35–55°E, with dominants of N45°E in the NE limb, and 40° in the SW limb.

On the contrary, L joints are rather rare, in the NE limb very poorly developed (Fig. 6) and showing variable orientations. In the SW limb, in turn, they cluster within a narrow range of N130–140°E (Fig. 7).

The opening of T joints is larger than in the remaining sets of joints and reaches up to several centimetres. Surfaces of a single T or L joint typically behave as form-cast coupling, with variable fracture structures e.g plumose structures indicating (Bankwitz, 1965; Engelder, 1985) extensional mode of opening. Traces of their intersection with the bedding are usually curvilinear.

According to Jaroszewski (1972), T joints appeared in a $\sigma_1>\sigma_2>-\sigma_3$ stress field with horizontal σ_1 and $-\sigma_3$ axes, at horizontal compression (σ_1), decreasing during the uplift of the Holy Cross Mountains. In turn, the residual development of L joints in the shallow NE limb of the syncline, stronger development in the steep SE limb (Fig. 7), as well as immediate disappearance of joints inwards point to their relation with buckling of the syncline. In this case, L joints are considered of the same origin as radial fractures (Jaroszewski, 1980).

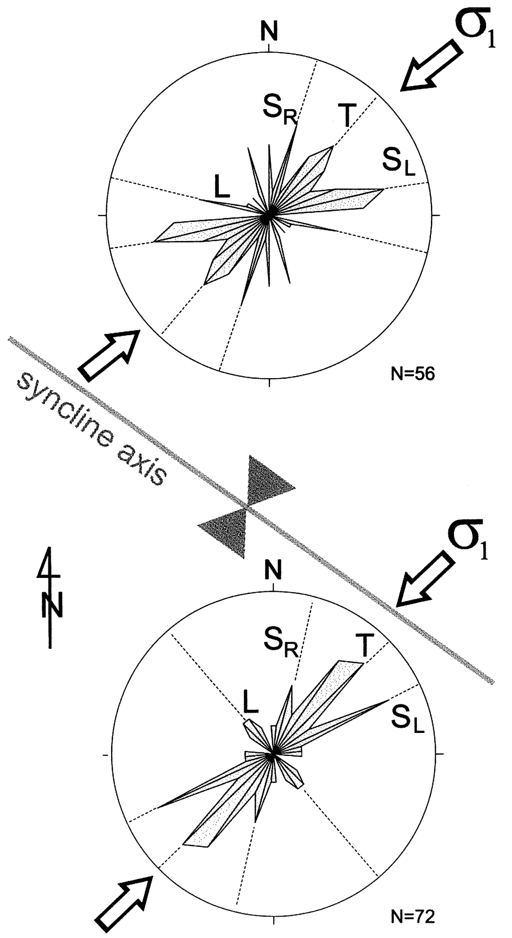


Fig. 7. Rose diagrams of joints in the N and S limbs of the Gnieździska Syncline

Master joints

A series of vertical fractures was recognised in both limbs of the Gnieździska Syncline, which cut entire bed complexes up to several meters (Fig. 8), have large lateral range, devoid of features pointing to displacement and have fracture fissure 10–30 cm wide. The cut the remaining tectonic structures. They form one set clustering at 30–63/90 (Fig. 8).

The common occurrence of vertical fractures in both limbs of the Gnieździska Syncline and the fact that they cut beds, faults and other tectonic structures indicates that they can be referred to as master joints. These extension fractures developed, according to the classical interpretation (e.g. Jaroszewski, 1972; Dunne & Hancock, 1994), in course of extension, at horizontal σ_1 clustering at N50–60°E, normal to σ_3, at vertical orientation of the axis σ_2, in one of the last stages of tectonic deformation in the analysed area.

Stylolites

Stylolites represent the most common mesostructures occurring within both limbs of the Gnieździska Syncline. They occur on surfaces typically normal, rarely parallel to bedding (Fig. 9). The stylolites are most distinct, as stylolitic seams in cross sections transverse to the stylolite surface (Fig. 9). Within limestone beds, several sets of stylolites have been determined. In most cases the stylolite teeth are aligned normal to the solution surfaces. According to the classification of Guzzeta (1984), the stylolite profiles are of sharp-peak type.

To identify the interval in which the stylolite sets were formed, a comparison of plane orientation for all measured stylolitic seams and columns in the northern and southern limb of the Gnieździska syncline, was carried out (Fig. 9). This comparison and rotation of particular sets in relation to rotating of beds to horizontal position, allows us to distinguish three stylolite sets.

Set I, occurring rather rarely, is characterised by bedding-parallel stylolites and sub-vertical columns (Fig. 9). The columns have lengths from 0.3 to 1.5 cm, and their density is rather small.

The next sets, II a and II b, are characterised, similarly as in e.g. Arthaud & Mattauer (1969) and Helmstaedt & Greggs (1980), by bed-normal stylolites (Fig. 9). Set II a,

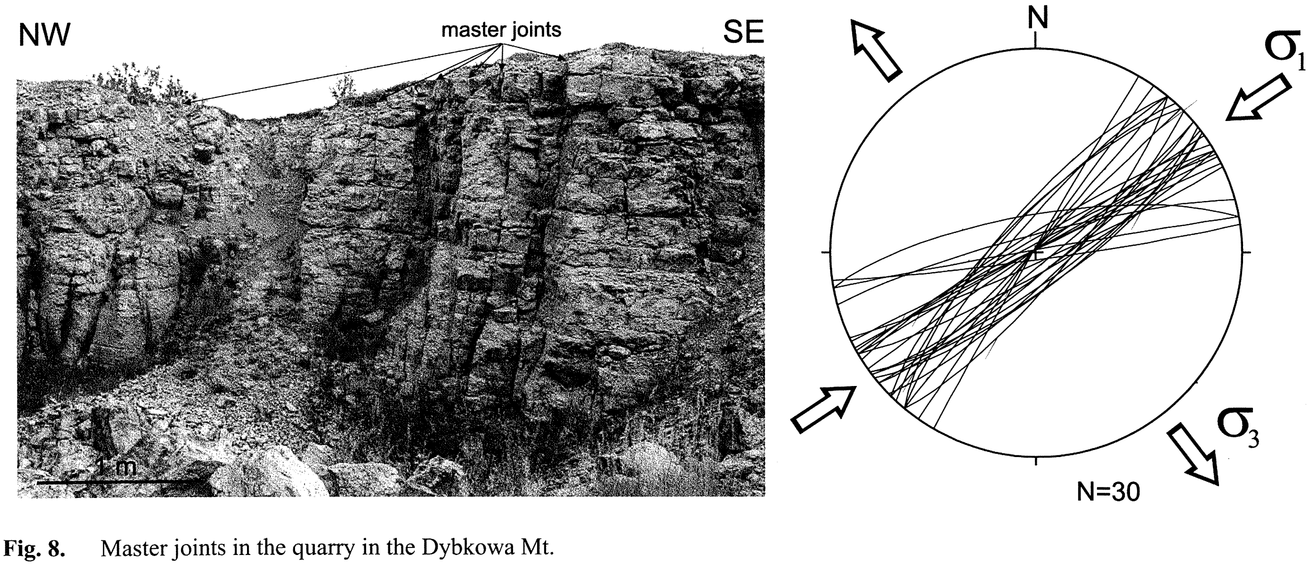


Fig. 8. Master joints in the quarry in the Dybkowa Mt.

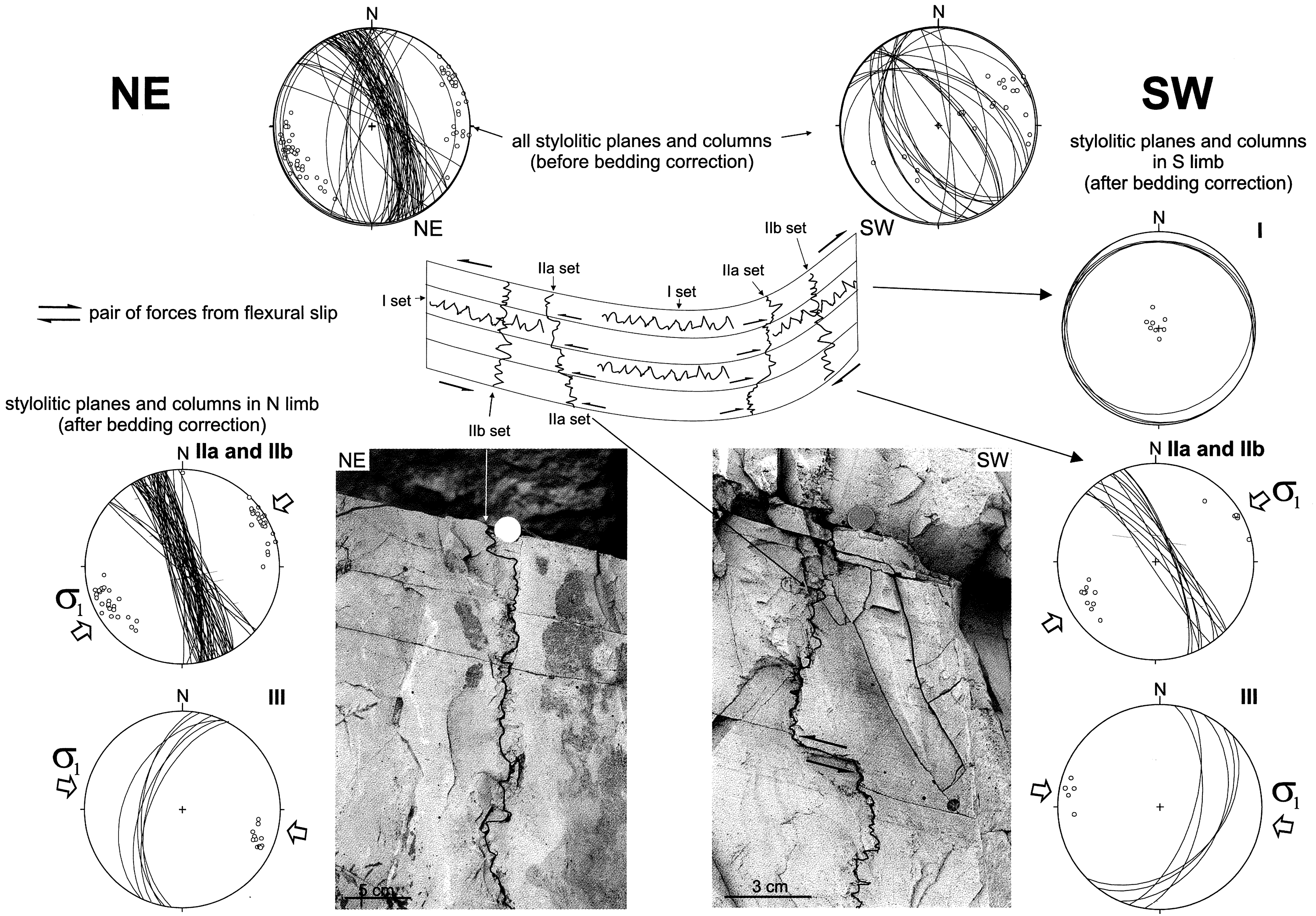


Fig. 9. Stylolite systems in the Gnieździska Syncline

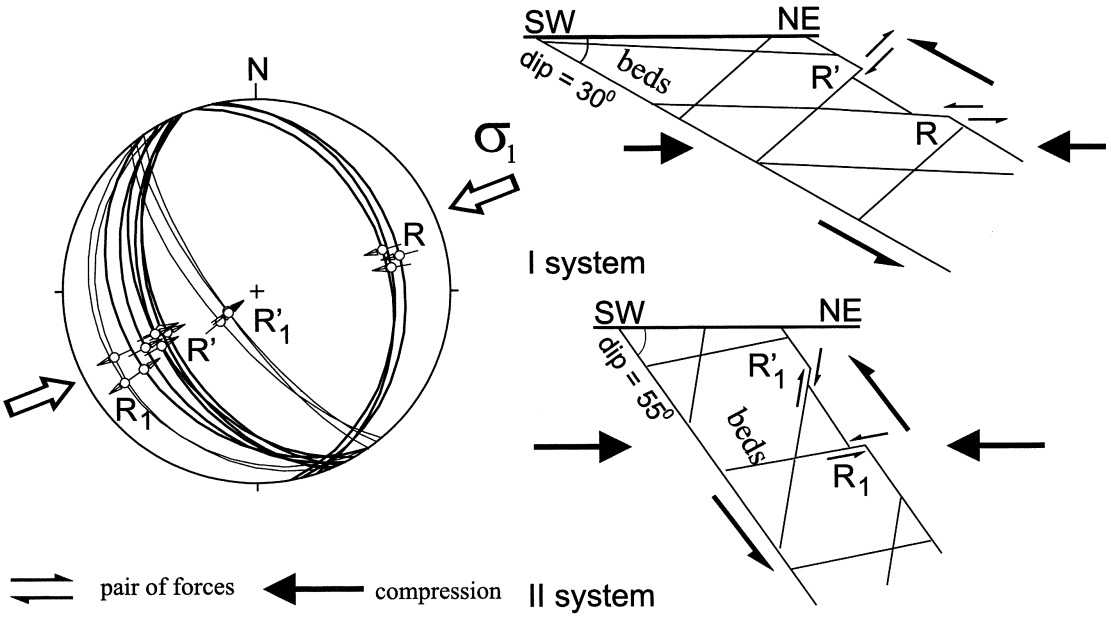


Fig. 10. Scheme of development of I and II cleavage systems

contrary to set II b, is distinctly displaced along the bedding for 3–8 cm (Fig. 9). The length of columns for both sets is from several millimetres to ca. 4.5 cm. Approximately 2–5 stylolitic seams per 1 meter occur both in sets II a and II b.

The stylolitic seams cluster around 60–80/80–90, whereas columns normal to these planes trend around 55–70/0–21 and 230–250/2–20 (Fig. 9).

Set III, similarly as set II b, lacks displacements along the bedding, and is characterised by oblique-bed stylolites. The orientation of stylolitic seams varies distinctly from other stylolites – 120–140/50–65 and 275/55, whereas of columns sub-normal to these planes 270–285/5–20 and 100–110/10–20, respectively.

The position of stylolitic seams and columns in relation to bedding indicates that set I formed under loading of cover.

Sets II a, II b and III, contrary to set I, are typical tectonic stylolites.

Sets II a and II b, the orientation of which is similar to some stylolite sets recognized by Wartołowska (1972) and Świdrowska (1980), developed in horizontal beds, what is testified to by bed-normal stylolite seams, identical orientation of columns in both syncline limbs, as well as by stylolitic seams parallel to fold axis. The stylolites developed in course of layer-parallel shortening. Set II b, younger than set II a, developed in gently dipping beds, however during distinct activation of flexural slip. Azimuths of columns (e.g. Blake & Roy,1949; Rispoli, 1981) of sets II a and II b indicate the general axis of shortening of N50–60°E.

Set III developed in dipping beds, most probably due to stresses caused by the nearby faults, what is testified to by a distinct deflection of stylolitic seams from the Gnieździska syncline axis.

Measurements of amplitudes for stylolite seams basing on sets II a and II b, from 7 outcrops in the NE limb of the Gnieździska syncline and 4 in the SW limb according to the Stockdale (1926) method, indicate a similar mean shortening for both limbs reaching 6.5–8%.

Cleavage

A series of densely-spaced fractures with strikes parallel to bedding and cutting beds at distances up to 5 cm (in thick-bedded limestones), bearing features of cleavage, were determined within the limestone beds. According to the classification of Powell (1979), rhythmic, cutting cleavage prevails, considered by Jaroszewski (1972) as fracture cleavage.

Four fracture sets forming 2 cleavage systems were determined on the basis of angles between cleavage planes and the bedding and the sense of movement along these planes (Fig. 10). The acute angle between the sets is always orientated at the roof-floor line of the bedding.

The first system comprises sets of fractures determined as R and R', lying at angles of 25° and 85° to the bedding, respectively. The dihedral angle between the sets is ca. 70°. Sets R and R', characterised by dips of 28/N and 46/S, are always consequent to the sense of flexural slip (Fig. 10). Fractures of this system are rare.

The second system comprises set R_1, orientated at 65–75 to the bedding and consequent, as well as set R_1' lying at 60° to the bedding, obsequent in relation to the sense of flexural slip. The dihedral angle between these sets is ca. 55°.

The general features of cleavage sets – strikes parallel to the syncline axis, sense of movement along these planes and orientation of the acute angle between sets of particular systems indicate (Bartlett *et al.*, 1981; Mastella, 1988; Konon, 2001) that each of the described two cleavage systems comprises conjugate shear sets, which developed due to the action of a pair of forces linked with flexural slip.

System I (R, R') can be most probably considered as classical Riedel shears (Riedel, 1929), developed parallel to

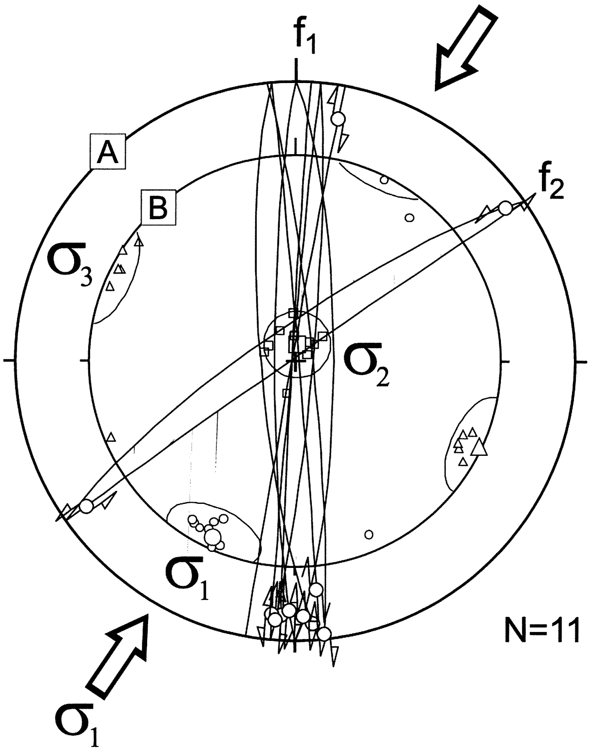


Fig. 11. Secondary strike-slip faults in the N limb of the Gnieździska Syncline (**A**); P-b-T axis (**B**)

the fold axis due to the action of a pair of forces from flexural slip, at a rather low influence of horizontal compression when bed dips were small.

System II (R_1, R_1') appeared later than system I, also parallel to the fold axis. The angles of sets of this system with the bedding are different from classical positions (e.g. Riedel, 1929, Morgenstern & Tchalenko, 1967; Bartlett *et al.*, 1981; Naylor *et al.*, 1986). This indicates that, besides simple shear resulting from the action of a pair of forces from flexural slip, compression normal to bedding was also responsible for the formation of this system. The compression was a result of the increased normal component due to regional horizontal compression, when dips exceeded 45° (e.g. Tanner, 1989).

This suggests that system I developed earlier, at small dips and low flexural slip. Larger flexural slip and compression normal to the bedding were necessary for the formation of cleavage of system II. Such conditions could have taken place in the late phases of folding (e.g. Mitra, 1978; Tanner, 1989), what is testified by later appearance of cleavage system II.

FAULTS

Strike-slip faults

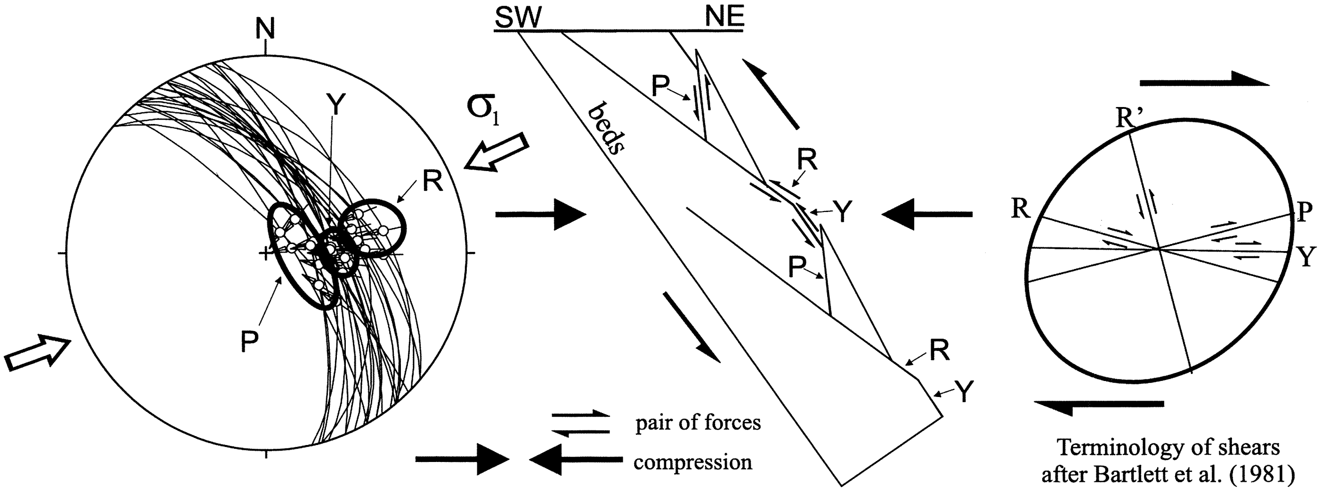
The described Gnieździska syncline is cut by numerous faults with variable strike-slip component. The largest fault cuts the NE limb of the syncline, east of Dybkowa Hill (Fig. 3). Its orientation is approximately consequent with the orientation of the Oxfordian strata in this limb (Fig. 3). This longitudinal fault is a continuation of a large fault zone, distinguished already by Czarnocki (1938) and Stupnicka (1972). Eastwards, this zone lies near Morawica and Wola Morawicka (op.cit.). Westwards the fault zone passes into the fault between the Dybkowa and Poddańska hills (Sadkowska, 2000) (Fig. 3). As a result, Rhaetian claystones contact with Callovian gaizes in the NE wall of this fault (Figs 3, 5 – III), or even with Oxfordian limestones of the SW wall, as in western part of the Poddańska Hill (Figs 3, 5 – II). In the field this fault is marked by breccia zones comprising Callovian gaizes and Oxfordian limestones. The occurrence of older lithostratigraphic members in the NE wall of the fault, as well as steepening of beds from ca. 16/S to ca. 30/S in the zone close to fault (Fig. 3), indicate the presence of a dip-slip component of this fault. The downthrown side of the fault is the SW part, and its throw in the vicinity of Poddańska Hill reaches at least 70 m.

The second fault of similar orientation cuts the SW limb of the syncline, disappearing just near Maćkowa Hill (Fig. 3). A large fault oblique to the syncline axis was determined by Sadkowska (2000) as a oblique-slip fault in the NW part of the area (Fig. 3).

Both limbs of the syncline are cut by a series of small steep faults (Figs 3, 11). In the NE limb these faults cooccur with the discussed longitudinal fault. In the SW limb this series occurs on the elongation of a similar longitudinal fault (Fig. 3). The faults, due to distinct lithological contrast between resistant Oxfordian limestones and yellow weathering Callovian gaizes, and forming depressions and red weathering Rhaetian deposits, are easily determined in the field as well as on air photographs. Fault zones, up to 0.5 m wide are filled with breccia, cataclasites and fault gouge. As determined from map analysis (Fig. 3), slicolites and slickensides, generally dextral strike-slip faults with azimuths of 340–20° occur, with sporadic sinistral strike-slip faults of azimuths N40–60°E (Fig. 11). The sets intersect one another at 60° and, according to Freund's (1974) criteria, can be considered as conjugate faults. Sinistral faults with azimuths of N15°E, cutting this zone NW of Dębowa Hill, are the exception (Fig. 3).

Series of *en echelon* strike-slip faults can be considered a typical (e.g. Wilcox *et al.*, 1973, Schreurs & Coletta, 1998) initial form of strike-slip faults, and in the further stage of their development are incorporated within their fault zone (e.g. Tchalenko, 1970; Mollema & Antonellini, 1999). Their pattern in relation to the fault zone and the sense of the slip component, in turn, indicates that they were developed due to the action of a dextral pair of forces within a horizontal plane (Fig. 3). In this case, the oblique dextral strike-slip faults should be identified as R-type Riedel shears (Fig. 3), developed due to the lack of possibilities of lateral movement, or even the narrowing of the fault zone (Vialon, 1979; Gamond & Giraud, 1982). On the other hand, sinistral faults with azimuths of 15° NW from Dębowa Hill would be R'-type Riedel shears (Fig. 3). This might indicate (op. cit.) that in this part (Fig. 3) the discussed fault zone could widen laterally.

Thus, the longitudinal fault in the NE limb of the Gnieździska syncline would be an oblique-slip transpressive fault



Terminology of shears after Bartlett et al. (1981)

Fig. 12. Dip-slip faults in the S limb of the Gnieździska Syncline

with the NE down thrown side. Only on a small part NE from Dębowa Hill it developed without regional compression normal to the fault (Fig. 3). A similar transpressive and oblique-slip character can be determined in a longitudinal fault west of Maćkowa Hill, whereas the belt of oblique faults occurring in its eastern continuation represents its incipient form.

Dip-slip faults

Most planes of dip-slip faults, noted in form of slickensides, were observed in the southern limb of the Gnieździska Syncline. In the northern limb they occur sporadically.

Dip-slip faults in the southern limb of the syncline occur in series arranged at angles low to bedding or directly of bedding planes. Three sets of reverse faults – R, Y and P were distinguished. Their strikes are parallel to the fold axis.

R faults include faults with fault planes clustering at 70/40, lying at ca. 15–20° to the bedding planes (Fig. 12).

Y faults include faults with planes parallel to the bedding (Fig.12).

P faults are characterised by almost normal position of fault planes, clustering at 70–80/70–80 and lying at ca. 15–25° to bedding (Fig. 12).

The R, Y and P faults occurring in the south-western part of the syncline, according to the terminology of Bartlett *et al.* (1981), correspond to R, Y and P shears. The R and P faults are arranged in R-P arrays. After the development of slips parallel to bedding, representing Y faults, R faults appeared, followed by P shears (Tchalenko, 1970; Bartlett *et al.*, 1981). The occurrence of faults representing R shears (Vialon, 1979; Mastella, 1988) and the presence of R-P arrays indicates (Davison, 1994) a tendency to concentrate shearing in a narrower zone, thus in the case of the analysed structures, in course of increasing flexural slip due to increasing dip.

CONCLUSIONS

The above structural analysis points to several stages of the structural evolution of the Gnieździska Syncline.

In the earliest deformation phase, as indicated by set I stylolites, the rocks underwent only influence of loading.

In the next stage in still horizontal beds, NE-SW horizontal compression started to act. The shear joints (S_R, S_L) were formed as "potential shear surfaces"and stylolites of set IIa appeared, being followed by stylolites of set IIb when the fold limbs, as a result of progressive folding, were slightly inclined. The continuous process of bed buckling caused the development of flexural slip, larger in the steeper, southern limb of the syncline. In consequence, cleavage systems I and II, as well as R, Y and P dip-slip faults were developed, pointing to the increasing influence of horizontal compression.

In the syncline buckling phase, L joints parallel to the syncline axis were also formed.

In the next phase, in which the limbs of the Gnieździska Syncline retained positions close to the present-day, one secondary strike-slip faults developed due to the influence of the Gnieździska–Wola Morawicka fault.

In the terminal deformation phase the post-kinematic uplift began, which lead to the formation of T joints and, due to further uplift, master joints of the same orientation. The formation of T joints and master joints indicates directly the phase of extension sub-parallel to the fold axis, at decreasing horizontal compression (σ_1).

The deformations of the Gnieździska Syncline indicate that it was formed at the dominating horizontal compression of N50–60°E orientation. The observable differences in the development of the northern and southern limbs are caused by the influence of a large regional strike-slip fault Gnieździska–Wola Morawicka.

The development of the Gnieździska–Mnin Syncline is probably linked mainly with the tectonic inversion of the Mid-Polish Swell, between the Maastrichtian and the Palaeocene (Kutek & Głazek, 1972; Lamarche, 1999).

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REFERENCES

Arthaud, F. & Mattauer, M., 1969. Exemples de stylolites d'origine tectonique dans le Languedoc, leurs ralations avec la tectonique cassante. *Compte Rendu Sommaire des Seances de la Societe Geologique de France*, 8: 311 pp.

Bankwitz P., 1965. Über Klüfte, II – Die Bildung der Kluftfläche une eine Systematik ihrer Strukturen, *Geologie*, 15: 896–941.

Barski M., 1999. Dinocyst stratigraphy of the Jurassic black clays from Holy Cross Mts area (Central Poland). (In Polish, English summary). *Przegląd Geologiczny*, 47: 718–722.

Bartlett, W. L., Friedman, M., Logan, J. M., 1981. Experimental folding and faulting of rocks in limestone layers. *Tectonophysics*, 79: 255–277.

Blake, D. B. & Roy, Ch. J., 1949. Unusual stylolites. *American Journal of Science*, 247: 779–790.

Cosgrove, J. W. & Ameen, M. S., 2000. A comparison of the geometry, spatial organization and fracture patterns associated with forced folds and buckle folds. In: Cosgrove, J. W. & Ameen, M. S., (eds.), Forced Folds and Fractures. *Geological Society Special Publication,* London, 169: 7–21.

Czarnocki, J., 1938. Carte géologique générale de la Pologne, feuille 4, Kielce. Echelle 1: 100,000. Edition du Service Géologique de Pologne.

Davison, I., 1994. Linked fault system; extensional, strike-slip and contractional. In: Hancock, P. L. (ed.), *Continental Deformation*. Pergamon Press, Oxford, pp. 121–142.

Dunne, W. M. & Hancock, P. L., 1994. Palaeostress analysis of small-scale brittle structures. In: Hancock P.L. (ed.), *Continental Deformation*. Pergamon Press, Oxford, pp. 101–120.

Engelder, T., 1985. Loading paths to joint propagation during a tectonic cycle: an example from the Appalachian Plateau, U.S.A. *Journal of Structural Geology*, 7: 459–476.

Filonowicz, P. & Lindner, L., 1986. Szczegółowa Mapa Geologiczna Polski w skali 1 : 50 000. Arkusz Piekoszów. (In Polish only). Wydawnictwa Geologiczne. Warszawa.

Filonowicz, P. & Lindner L., 1987. Objaśnienia do Szczegółowej Mapy Geologicznej Polski w skali 1 : 50 000. Arkusz Piekoszów. (In Polish only). Wydawnictwa Geologiczne, Warszawa.

Freund, R., 1974. Kinematics of transform and trancurrent faults. *Tectonophysics*, 21: 93–134.

Gamond, J. F. & Giraud, A., 1982. Identification des zones de faille á l'aide des associations de fractures de second ordre. *Bulletin de la Societe Géologique de France*, 24: 755–762.

Guzzeta, G., 1984. Kinematics of stylolite formation and physics of pressure solution process. *Tectonophysics*, 101: 383–394.

Helmstaedt, H. & Greggs, R. G., 1980. Stylolitic cleavage and cleavage refraction in lower Palaeozoic carbonate rocks of the Great Valley, Maryland. *Tectonophysics*, 66: 99–114.

Jaroszewski, W., 1972. Mesoscopic structural criteria of tectonics of non-orogenic areas: an example from the north-eastern mesozoic margin of Świętokrzyskie Mountains. *Studia Geologica Polonica*, 38: 1–215.

Jaroszewski, W., 1980. *Tektonika uskoków i fałdów*. Wydanie II. (In Polish only). Wydawnictwa Geologiczne, Warszawa: 360 pp.

Kutek, J. & Głazek J., 1972. The Holy Cross area, Central Poland, in the Alpine cycle. *Acta Geologica Polonica*, 22: 603–653.

Konon, A., 2001. Tectonics of the Beskid Wyspowy mountains (Outer Carpathians, Poland). *Geological Quarterly*, 45: 179–204.

Lamarche, J., Mansy, J. L., Bergerat, F., Averbuch, O., Hakenberg, M., Lewandowski, M., Stupnicka, E., Świdrowska, J., Wajsprych, B. & Wieczorek, J., 1999. Variscan tectonics in the Holy Cross Mountains (Poland) and the role of structural inheritance during Alpine tectonics. *Tectonophysics*, 313: 171- 186.

Lewiński, J., 1912. Utwory jurajskie na zachodnim zboczu Gór Świętokrzyskich. (In Polish only). *Sprawozdania z posiedzeń Towarzystwa Warszawskiego*, 5. Warszawa.

Mandl, G., 1988. *Mechanics of tectonic faulting; models and basic concepts*. In the collection: Developments in structural geology. Elsevier Science Publications, Amsterdam, 407 pp.

Mastella, L., 1988. Structure and evolution of Mszana Dolna Tectonic Window, Outer Carpathians, Poland. *Annales Societatis Geologorum Poloniae*, 58: 53–173.

Mastella, L. & Zuchiewicz, W., 2000. Jointing in the Dukla nappe (Outer Carpathians, Poland): an attempt at palaeostress reconstruction. *Geological Quarterly*, 44: 377–390.

Matyja, B. A., Wierzbowski, A. & Drewniak A., 1996. Węglanowe osady basenu późno-jurajskiego zachodniego obrzeżenia Gór Świętokrzyskich. (In Polish only). In: Karnkowski, P.H. (ed.), Analiza basenów sedymentacyjnych a nowoczesna sedymentologia. *Materiały Konferencyjne V Krajowego Spotkania Sedymentologicznego*, A1–16. Warszawa.

Mitra, S., 1978. Microscopic deformation mechanisms and flow laws in quartzites within the South Mountain Anticline. *Journal of Geology*, 86: 129–152.

Mollema, P. N. & Antonellini, M., 1999. Development of strike-slip faults in the dolomites of the Sella Group, Northern Italy. *Journal of Structural Geology*, 21: 273–292.

Morgenstern, N. R. & Tchalenko, J. S., 1967. Microstructural observations on shear zones from slips in natural clays. *Geotechnical Conference, Oslo, Proceedings*, 1: 147–152.

Murray, F. N., 1967. Jointing in sedimentary rocks along the Grand Hogback Monocline, Colorado. *Journal of Geology*, 75: 340–350.

Naylor, M. A., Mandl, G. & Sijpesteijn, C. H. K., 1986. Fault geometries in basement-induced wrench faulting under different initial stress states. *Journal of Structural Geology*, 8: 737–752.

Ozimkowski, W., Dzierżek, J., Konon, A., Mastella, L., Rubinkiewicz, J., Szczęsny, R. & Szumański, A., 1999. Szczegółowa Mapa Geologiczna Regionu Świętokrzyskiego w skali 1 : 25 000. Arkusz Piekoszów C. (In Polish only). Państwowy Instytut Geologiczny, Warszawa.

Price, N. J., 1959. Mechanics of jointing in rocks. *Geological Magazine*, 96: 149–167.

Price, N. J., 1966. *Fault and Joint Development in Brittle and semi-brittle Rock.* Pergamon Press, Oxford, 176 pp.

Powell Mc, A. C., 1979. A morphological classification of rock cleavage. *Tectonophysics*, 58: 21–34.

Ramsay, J. G., 1974. Development of chevron folds. *Geological Society of America Bulletin*, 85: 1741–1754.

Riedel, W., 1929. Zur Mechanik geologischer Brucherscheinungen. *Zentralblatt für Mineralogie, Geologie und Paläontologie*, Abt. B, 354 pp.

Rispoli, R., 1981. Stress fields about strike-slip faults inferred from stylolites and tension gashes. *Tectonophysics*, 75: T29–T36.

Sadkowska, A. 2000. Tektonika obrzeżenia mezozoicznego Gór Świętokrzyskich w rejonie Gnieździsk (In Polish only). Unpubl. MSci. Thesis, University of Warsaw, 67 pp.

Schreurs, G. & Colletta, B., 1998. Analogue modelling of faulting in zones of continental tanspression and transtension. In: Holdsworth, R. E., Strachan, R. A. & Dewey J. F. (eds.), Continental transpressional and transtensional tectonics. *Geological Society Special Publication*, London, 135: 59–79.

Stockdale, P. B., 1926. The stratigraphic significance of solution in rocks. *Journal of* Geology, 34: 399–414.

Stupnicka, E., 1972. Tektonika południowo-zachodniego obrzeżenia Gór Świętokrzyskich. (In Polish only). *Biuletyn Geologiczny Uniwersytetu Warszawskiego*, 14: 21–114.

Świdrowska, J., 1980. Stylolity tektoniczne jako wskaźnik tektogenetyczny na obszarze południowo-zachodniego obrzeżenia Gór Świętokrzyskich. (In Polish only). *Przegląd Geologiczny*, 3: 159–164.

Tanner, P. W. G., 1989. The flexural-slip mechanism. *Journal of Structural Geology*, 11: 635–655.

Tchalenko, J. S., 1970. Similarities between shear zones of different magnitudes. *Geological Society of America Bulletin*, 81: 1625–1639.

Turner, F. J., 1953. Nature and dynamic interpretation of deformation lamellae in calcite of three marbles. *American Journal of Science*, 251: 276–298.

Vialon, P. P., 1979. Les déformation continues-discontues des roches anisotropes. *Eclogae Geologicae Helvetiae*, 72: 531–549.

Wallbrecher, E., 1986. *Tektonische und gefügeanalytische Arbeitsweisen*. Enke-Verlag, Stuttgart, 244 pp.

Wartołowska, J., 1972. An example of the processes of tectonic stylolitization. *Bulletin de L'Academie Polonaise des Sciences. Série des Sciences de la Terre*, 20: 197–204.

Wilcox, R. E., Harding, T. P. & Seely, D. R., 1973. Basic wrench tectonics. *American Association of Petroleum Geologists Bulletin*, 57: 74–96.

Streszczenie

EWOLUCJA STRUKTURALNA SYNKLINY GNIEŹDZISK – REGIONALNE IMPLIKACJE DLA POŁUDNIOWO-ZACHODNIEGO MEZOZOICZNEGO OBRZEŻENIA GÓR ŚWIĘTOKRZYSKICH

Andrzej Konon & Leonard Mastella

W oparciu o analizę mezostruktur tektonicznych określone zostały etapy ewolucji strukturalnej synkliny Gnieździsk, stanowiącej typowy przykład fałdów występujących pasem w południowo-zachodnim mezozoicznym obrzeżeniu Gór Świętokrzyskich (Fig. 1). Występujące na badanym terenie skały (Fig. 2) obejmują niewielki – od retyku po kelowej – fragment profilu litologiczno-stratygraficznego mezozoicznego obrzeżenia Gór Świętokrzyskich. W badaniach oparto się o analizę mapy tektonicznej (Fig. 3), położeń warstw (Fig. 4), przekrojów geologicznych (Fig. 5), ciosu ścięciowego i ekstensyjnego (Fig. 6, 7), ciosu przewodniego (Fig. 8), stylolitów (Fig. 9), kliważu (Fig. 10), oraz uskoków przesuwczych i zrzutowych (Fig. 11, 12).

Przeprowadzona analiza strukturalna wskazuje na wieloetapowość ewolucji synkliny Gnieździsk.

W najwcześniejszym etapie deformacji, na co wskazują stylolity I zespołu, skały były poddane jedynie oddziaływaniu ciężaru nadkładu (Fig. 9).

W następnym etapie w ciągle płaskich warstwach, zaznaczyło się oddziaływanie kompresji horyzontalnej o kierunku NE-SW. Spowodowało to założenie ścięciowego systemu ciosu (Fig. 6, 7) oraz powstanie stylolitów – najpierw IIa, a następnie, gdy skrzydła fałdu w wyniku postępującego procesu fałdowania uległy niewielkiemu wychyleniu powstanie zespołu IIb. Ciągły proces zginania warstw spowodował powstanie posuwu fałdowego, większego w stromszym, południowym skrzydle synkliny. Doprowadziło to do powstania systemów kliważu I i II (Fig. 10) oraz uskoków zrzutowych R, Y i P, wskazujących ponadto na zwiększający się udział kompresji horyzontalnej (Fig. 12). W etapie wyginania synkliny powstał również, równolegle do osi synkliny cios L.

W kolejnym etapie, w którym skrzydła synkliny Gnieździsk uzyskały położenia zbliżone do dzisiejszych (Fig. 4), powstały w związku z oddziaływaniem uskoku Gnieździska–Wola Morawicka drugorzędne uskoki przesuwcze (Fig. 11).

W końcowym etapie deformacji rozpoczęło się wypiętrzanie, co doprowadziło do powstania cios typu T oraz w dalszym etapie wypiętrzania, na tych samych kierunkach – ciosu przewodniego (Fig. 8). Powstawanie ciosu T i ciosu przewodniego wskazują bezpośrednio na etap ekstensji subrównoległej do osi fałdu, przy zmniejszającej się horyzontalnej kompresji (σ_1).

Cechy deformacji synkliny Gnieździsk wskazują, że powstała ona przy dominującej przez cały okres kompresji poziomej o kierunku 50–60°. Zauważalne różnice w rozwoju północnego i południowego skrzydła, są spowodowane oddziaływaniem dużego regionalnego uskoku przesuwczego Gnieździska–Wola Morawicka.