Fault geometry and evidence of depocentre migration within a transtensional intra-basinal high – a case study from the Łączna Anticline (Intrasudetic Synclinorium, SW Poland)

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INTRODUCTION

Intra-basinal highs, also known as central basin horsts (Sims et al., 1999; Dooley and Schreurs, 2012), are very common structural elements observed within extensional basins, particularly in a transtensional setting (Anders and Schlische, 1994; Hözsel et al., 2008; Wu et al., 2009). Their evolution and geometry are linked with cross-basin fault zones (CBFZs; Dooley et al., 2004) that led to the formation of uplifted, horst-like structures in the central part of the basin. Similarly to intra-basinal “pop-up structures” (McClay and Bonora, 2001) formed in a transpressional regime, transtensional intra-basinal highs are topographic elevations with older rocks exposed in the core (Sylvester, 1988; Sugan et al., 2014). They have strongly influenced the sedimentation and migration of depocentres within pull-apart basins (Wu et al., 2009). Several morphological, intra-basinal elevations (Ulîčny, 2001; Wojewoda, 2007, 2009) have been recognized and described within the Intrasudetic Synclinorium (IS) – a distinct intramontane depression situated in the northeastern termination of the Bohemian Massif (Fig. 1). The IS is the largest geological unit (70 km long and 35 km wide) of the Sudetes and is built of a weakly deformed sedimentary succession of the Intrasudetic Basin. The basin infill comprises Mississippian to Lower Permian volcanioclastic rocks, which are unconformably overlain by Triassic continental and Cretaceous marine sediments. Development of the Intrasudetic Basin as a narrow intramontane trough (Wojewoda and Mastalerz, 1989) was initiated probably in the Middle or Early to Late Viséan times (Nemec et al., 1982; Turnau et al., 2002), especially in an extensional strike-slip setting (Porębski, 1980; Wojewoda, 1997; Aleksandrowski, 1998). The Intrasudetic Basin was filled with clastic, mainly continental deposits during Mississippian to Early Permian times. The basin infilling process was interrupted by sub-volcanic intrusions and tectonic deformations before the end of the Pennsylvanian (the first stage of basin inversion; according to Żelaźniewicz et al., 2011). The youngest, Late Cretaceous (Ziegler, 1987; Mazur et al., 2006) and Neogene (Wojewoda, 2004) inversions of the basin resulted in the formation of numerous structural depressions (“sub-basins”; Ulîčny, 1999) bounded by NW–SE-trending normal and strike-slip faults (Wojewoda, 1997; Ulîčny, 2001; Grygar and Jelinek, 2003; Wojewoda, 2007). The sub-basins within the IS are interpreted as individual pull-apart basins (Ulîčny, 2001; Wojewoda, 2007) separated by transversely oriented elevations (Wojewoda and Mastalerz, 1989). This paper presents the results of geological mapping and structural analysis of the Łączna Anticline area (LA; Ger.
The presented investigations suggest that the Łączna Anticline constitutes an intra-basinal high that separates two minor, NW–SE-trending depressions within the IS: the Krzeszów and Police brachysynclines (KB and PB sub-basins, respectively; Jerzykiewicz, 1969; Táslér et al., 1979).

**GEOLOGICAL SETTING**

The study area is located at the Czech-Polish border, within the northwestern part of the Intrasudetic Synclinorium, between the towns of Lubawka and Mieroszów (Fig. 1). According to regional physiographical divisions (Kondracki, 2002), the region is...
assigned to the Zawory Range (Cz. Závora) that constitutes the northernmost part of the Stołowe Mountains in the Central Sudetes. Sedimentary rocks exposed in the area represent Permian (Rotliegend), Triassic (Buntsandstein) and Cretaceous (Upper Cenomanian) deposits (Fig. 2). In the peripheral parts of the basin occur the oldest, Lower Permian sediments, included in the Radków Formation in the Polish part of the IS (Nemec et al., 1982) or in the Trutnov Formation in the Czech part of the IS (Tásler, 1964). Lower Permian sediments consist mostly of coarse-grained conglomerates (“fanglomerates”; Dziedzic, 1961) and conglomeratic sandstones interpreted as alluvial fan and braided river deposits (Aleksandrowski et al., 1986; Wojewoda, 2008). Sediments of the Radków (Trutnov) Formation pass upward into sandy conglomerates with calcareous intercalations and dolomites of the Chelmsko Śląskie Beds (Dziedzic, 1961; Śliwiński, 1984) or of the Bohuslavice Formation in the Czech part of the IS (Tásler et al., 1964). These deposits are classified as late Saxonian (Śliwiński, 1984) or Thuringian (Holub, 1972) in age, respectively. They are interpreted as fluvial and lacustrine deposits (Tásler, 1979) with calcite (caliche) and travertine horizons (Śliwiński, 1981, 1984). Lower Triassic arkosic sandstones of the Bohdašín Formation (Tásler, 1964; Prouza et al., 1985) nearly concordantly overlie Permian deposits and are considered as typical braided river and alluvial fan deposits (Mroczkowski, 1977; Prouza et al., 1985). The topmost part of the Bohdašín Formation consists of strongly kaolinized, weakly lithified sandstones of undefined age (Middle Triassic?), which occur only locally in the IS (Wojewoda et al., 2016). They were first distinguished in the Czech part of the IS as the Deřek from or the Barchoviny Member sandstones (Holub, 1966; 1972). Kaolinitic sandstones were interpreted as shallow marine (Holub, 1966; Wojewoda et al., 2016), lacustrine (Prouza et al., 1985) or even aeolian sediments (Mikulaš et al., 1991; Uličný, 2004). In the study area, the sandstones of the Bohdašín Formation crop out mainly in the vicinity of Łączna (Czarna Struga Valley) and are discordantly covered by Cretaceous strata. The boundary between the Triassic and Cretaceous sediments corresponds to the bottom of the “basal conglomerates” (Skoček and Valečka, 1983) that mark a nearly horizontal transgressive surface. Cretaceous marine sediments occur in the central parts of the brachysynclines and build the hills surrounding them. Between the Krzeszów and Polič brachysynclines, the sandy facies of these strata (glauconitic and calcareous sandstones) are assigned to the Upper Cenomanian (Peruc-Koryčany Formation; Dvořák, 1968; Jerzykiewicz, 1971). The sandstones pass upward into glauconitic and siliceous mudstones (gaizes and spongiolites) assigned to the Upper Cenomanian (Jerzykiewicz, 1971) and Lower Turonian (Biłă Hora Formation; Dvořák, 1968). Mudstones build the highest elevations of the region. Glauconitic mudstones (“glaoonite horizon”; Berg, 1909) thin out to the south within the siliceous mudstones (Berg, 1909; Jerzykiewicz, 1971; Ziółkowska, 1990). The fine-grained deposits pass upward into Middle and Upper Turonian calcareous mudstones and sandstone outcropping in the central parts of the KB and PB. Cretaceous rocks, which crop out within the IS, constitute the western part of a larger regional geological unit – the Bohemian Cretaceous Basin (Skoček and Valečka, 1983), located in the central part of the Bohemian Massif. It is worth mentioning that the Triassic and Cretaceous strata of the Krzeszów Basin constitute also the northernmost part of the existing Mesozoic sedimentary cover of the IS.

The tectonic setting of the Łączna area is poorly recognized. According to Jerzykiewicz (1969) and Don et al. (1981), the northern part of the Zawory Range (a mountain belt comprising the Rög, Drogosz and Chochol hills; Puc and Traczyk, 2006),

<table>
<thead>
<tr>
<th>CHRONOSTRATIGRAPHY</th>
<th>LITHOSTRATIGRAPHY</th>
<th>[m]</th>
<th>LITHOLOGY</th>
</tr>
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<tr>
<td>Upper Cretaceous</td>
<td></td>
<td>250</td>
<td>calcareous mudstones</td>
</tr>
<tr>
<td></td>
<td></td>
<td>200</td>
<td>siliceous mudstones with sandstone intercalations</td>
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<tr>
<td>Lower Cretaceous</td>
<td></td>
<td></td>
<td></td>
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<tr>
<td></td>
<td></td>
<td>150</td>
<td>kaolinitic sandstones</td>
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<tr>
<td></td>
<td></td>
<td>100</td>
<td>arkosic sandstones and conglomerates</td>
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<tr>
<td></td>
<td></td>
<td>50</td>
<td>sandstones mudstones</td>
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<td></td>
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<tr>
<td>Buntsandstein</td>
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<tr>
<td>Bohdašín Formation</td>
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<td></td>
<td></td>
</tr>
<tr>
<td>Lower Permian</td>
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<td></td>
<td></td>
</tr>
<tr>
<td>Saxonian</td>
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<tr>
<td>Radków Formation</td>
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</table>

Fig. 2. Stratigraphy and lithology of the Łączna Anticline area
The Adršpach, Police and Batorów was subdivided by some authors into three minor tectonic units: strata, especially within jointed glauconitic sandstones. The KB and PB are deformed by vertical or sub-vertical sets of Quader sandstein; although the Krzeszów Brachysyncline was always considered as a sin-

gle brachyfold (Jerzykiewicz, 1971; Don et al., 1979), especially on the west-

of the LF strikes obliquely to the main fault intersection trace. The eastern segment of NNW–SSE-trending faults. The eastern segment (loc. 3). The north-west-

ern branch of this fault, especially on the northern slopes of Drogosz Hill, was interpreted earlier as the main fault zone of the LF (Berg and Dathe, 1905/1906) and as the “Southern Marginal Flexure Fault” (Jerzykiewicz, 1971). Detailed mapping of this area has shown that the northern, fault-affected slopes of Drogosz Hill are covered by a landslide colluvium (loc. 4; Kowalski, 2017). The surface of rupture of this landslide probably coincides with the upper part of a listric-shaped fault plane (Kowalski, 2017). Moreover, the area between the synthetic fault splays represents a relay ramp structure (loc. 3; e.g., Pea-

METHODS

Geological and structural mapping of the Łączna region was conducted in numerous localities including abandoned and periodically active quarries, road crosscuts and natural exposures. Representative samples of the main lithological types were collected. Textural and structural features of selected Triassic and Cretaceous sandstones were analysed during fieldwork and supplemented with petrographic microscopic observations. Structural mapping involved measurements and description of joints, fault planes, and preserved fault scarp s. Numerous mesostructural kinematic indicators (Pettit, 1987), such as striated ridges, grooves, slickensides, low-angle shears, hackles and en echelon cracks, were described and measured (Table 1). The measurements were plotted using a lower hemisphere stereographic projection.

During fieldwork, much attention was paid to accurate localization and type of lithological boundaries (tectonic and sedimentary contacts). In many cases, basic geomorphological mapping and observations were carried out. Structural features were identified and precisely localized in the field with application of the Nomad Trimble GPS and Pentagram PathFinder Logger P3106. The effect of the investigations—a geological map at the scale of 1:10,000, was superimposed on a shaded LIDAR Digital Terrain Model (DTM) of ~1 m resolution (Fig. 3). Contours generated from the DTM had a 0.5 m interval and were applied to construct regional stratigraphic and structural surfaces with the use of Microdem Software v. 2015.8. (developed by Peter Guth) and Global Mapper v. 15.0. This allowed examination of the accuracy of the current geological maps. LIDAR-based geomorphometric analysis was also made.

TECTONICS OF THE ŁĄCZNA REGION

LIPIELICA–ŁĄCZNA FAULT (LF)

Almost the entire KB area is segmented by NW–SE-ori-


tined and minor ENE–WSW-trending faults. The Cretaceous rocks of the KB, Triassic outcrops of the Czarna Struga Valley, and the surrounding hills (Miłna and Dziób) are cut by the LF that continues also to the NW of the KB within the Permian volcanic rocks of the IS (Krucze Mts.). The vicinity of this fault is not well-exposed, thus especially cartographic and geomorphological evidences of faulting can be observed. The LF dips steeply to the north-east and the maximum postulated throw on the fault reaches 30 m (Jerzykiewicz, 1971; Don et al., 1981). In the northern part of the study area, the tectonic contact between siliceous mudstones and older sandstones of Cenomanian age indicates amplitude of ~23 m within this fault segment (locality no. 1).

Several geomorphic evidences of tectonic activity along the LF were identified, especially during field investigations and geomorphometric analysis based on LiDAR data. To the north, they include, e.g., splitting valleys (loc. 2) formed above two segments of NNW–SSE-trending faults. The eastern segment of the LF strikes obliquely to the main fault intersection trace. Exposures of glauconitic mudstones constitute in this case part of a northeasterward tilted, wedge-shaped block bounded by fault splays. On the southeastern slope of Drogosz Hill, the LF diverges into two linkage fault segments (loc. 3). The northwestern branch of this fault, especially on the northern slopes of Drogosz Hill, was interpreted earlier as the main fault zone of the LF (Berg and Dathe, 1905/1906) and as the “Southern Marginal Flexure Fault” (Jerzykiewicz, 1971). Detailed mapping of this area has shown that the northern, fault-affected slopes of Drogosz Hill are covered by a landslide colluvium (loc. 4; Kowalski, 2017). The surface of rupture of this landslide probably coincides with the upper part of a listric-shaped fault plane (Kowalski, 2017). Moreover, the area between the synthetic fault splays represents a relay ramp structure (loc. 3; e.g., Pea-
Fig. 3. Detailed geological map superimposed on a LiDAR DTM showing the main structural features of the Łączna area (author: A. Kowalski)

SF – “Southern Marginal Flexure”; other explanation as in Figure 1; contours generated from LiDAR data (interval 10 m)
and steps, R’ – Riedel shears; S – slick en sides; U – soft-sed i ment de for ma tion dense frac tures; E –
Quader out crops across the cuesta ridge (loc. 6; Fig. 3).

Stratig ra phy and li thol ogy: UC – Up per Cre ta ceous; MT – Mid dle Tri as tic; LT – Lower Tri as tic. GS – glau conitic sand stones; G&SM –
glauconitic and si li ceous mudstones; KS – kaolinitic sand stones. Ki ne matic in di ca tors and other di ag nos tic struc tures: B – brec cia; D –
NW–SE-trending dextral strike-slip move ments. This is also
confirmed by ob serv a tions of steep exten sion frac tures with
hackl es (Herman, 2009) and oblique pinnate joints (Hancock,
1985; Engelder, 1989), which occur in artifi cial ex posures of the
Lower Quader on the nor thern slopes of Mi elna Hill (loc. 7).

RÓG FLOWER STRUCTURE

NW- and NNW-trending, both strike-slip and dip-slip faults,
were also recog nized in the south west ern ter mi na tion of the
KB. The great est con cen tra tion of these faults is ob served in ar-
ificial ex posures of the Lower Quader around Róg Hill – the
most pro min ent table hill (712.8 m a.s.l.) within the Zawory
Ridge. Accord ing to Jerzykiewicz (1969, 1971), the south ern
part of the Róg platea u is re lated to the axis of the Łączna
Anticline.

Fault planes crop out main ly in ex posures along for est
roads tra vers ing the nor thene astern and nor ther n slopes of Róg
Hill (hereinafter: N Róg faults). The first dam age zone (loc. 8;
Fig. 4A, B) com prises nu mer ous NW–SE-trending, both dextral
strike-slip and nor mal faults. Fault sur faces are ver ti cal or
steeply in clined at 50–90° and strike ob li quely to the main
lithological bound aries (Fig. 3). Low-angle R-shears, hackl es
and slickensides ob served on the frac ture sur faces in di cate
that these fac ults are re lated to the axis of the Łączna
Anticline.

Moreover, a wa ter shed was formed within the south-
east ern part of the LF be tween Mi elna and Dziób hills (loc. 6).
The oc cu rence of hor setail and low-angle Riedel shear frac-
tures in ex posures within this sec tion of the fault indi cates
NW–SE-trending dextral strike-slip move ments. This is also
con ferred by ob serv a tions of steep ex tend sion frac tures with
hackl es (Herman, 2009) and oblique pinnate joints (Hancock,
1985; Engelder, 1989), which occur in artifi cial ex posures of the
Lower Quader on the nor thern slopes of Mi elna Hill (loc. 7).

Table 1

<table>
<thead>
<tr>
<th>Locality (no.)</th>
<th>Stratigraphy and lithology</th>
<th>Coordinates</th>
<th>Fault geometry</th>
<th>Fault plane orientation</th>
<th>Kinematic indicators and other diagnostic structures</th>
</tr>
</thead>
<tbody>
<tr>
<td>N Chochol (2)</td>
<td>UC; G&amp;SM</td>
<td>50°41’01.65” N 16°08’02.65” E</td>
<td>strike-slip, normal fault</td>
<td>100/70</td>
<td>D, S, R’</td>
</tr>
<tr>
<td>N Mi elna (7)</td>
<td>UC, GS</td>
<td>50°39’56.62” N 16°08’05.77” E</td>
<td>strike-slip, normal fault</td>
<td>280/85</td>
<td>H, R’</td>
</tr>
<tr>
<td>NW Róg (8)</td>
<td>UC, MT; KS, GS</td>
<td>50°40’31.81” N 16°05’43.14” E</td>
<td>strike-slip, normal faults</td>
<td>40/80</td>
<td>D, H, R’, S</td>
</tr>
<tr>
<td>N Róg (9)</td>
<td>UC, MT; KS, GS</td>
<td>50°40’35.04” N 16°05’55.14” E</td>
<td>strike-slip, normal faults</td>
<td>240/80</td>
<td>B, D, P, R</td>
</tr>
<tr>
<td>S Róg (11)</td>
<td>UC, GS</td>
<td>50°39’51.80” N 16°07’01.43” E</td>
<td>strike-slip, normal fault</td>
<td>240/85</td>
<td>D, H, P, R, R’</td>
</tr>
<tr>
<td>SW Róg (12)</td>
<td>UC, GS</td>
<td>50°39’53.73” N 16°06’52.52” E</td>
<td>strike-slip, normal faults</td>
<td>230/85</td>
<td>D, E, R</td>
</tr>
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<td>E Łączna (13)</td>
<td>LT, KS</td>
<td>50°40’28.44” N 16°08’56.21” E</td>
<td>strike-slip; normal faults</td>
<td>282/85</td>
<td>H, R’, S, D, Fe</td>
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<td>Samotna (14)</td>
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<td>strike-slip faults</td>
<td>90/55</td>
<td>E, S</td>
</tr>
<tr>
<td>N Dziób (15)</td>
<td>UC, GS</td>
<td>50°40’11.10” N 16°08’48.80” E</td>
<td>strike-slip fault</td>
<td>100/85</td>
<td>R, S</td>
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<tr>
<td>Zdoñov 1 (16)</td>
<td>LT, KS</td>
<td>50°39’19.98” N 16°08’57.85” E</td>
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<td>170/50</td>
<td>D, Fe</td>
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<td>Zdoñov 2 (16)</td>
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<td>50°39’21.86” N 16°09’02.70” E</td>
<td>normal fault</td>
<td>186/30</td>
<td>E, U</td>
</tr>
</tbody>
</table>

Stratig ra phy and lithol ogy: UC – Upper Cre ta ceous; MT – Mid dle Tri as tic; LT – Lower Tri as tic. GS – glauconitic sand stones; G&SM –
glauconitic and siliceous mudstones; KS – kaolinitic sandstones. Kinematic indicators and other diagnostic structures: B – breccia; D –
dense frac tures; E – en ech e lon cracks; Fe – iron com pounds within frac tures; G – gouge; H – hack les; P – polished fault sur faces; R – ridges
and steps, R’ – Riedel shears; S – slick ensides; U – soft-sed i ment de for ma tion

cock and Sanderson, 1991) that affects the local drainage pat-
tern of the nor ther n slopes of Drogosz Hill. A simi lar case is
present within the south ern escarpment of the Zawory Ridge,
where the eastern fault block is tilted to the NE and dis placed
approximately 500 m tow ards the SW (loc. 5). A tilted block is
associ ated with a relay ramp be tween two over step ping nor-
mal faults with a small strike-slip com ponent (e.g., Pea cock and
San der son, 1991) that af fects the lo cal drain age pat-

Fault planes crop out main ly in ex posures along for est
roads tra vers ing the nor thene astern and nor ther n slopes of Róg
Hill (hereinafter: N Róg faults). The first dam age zone (loc. 8;
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hackl es (Herman, 2009) and oblique pinnate joints (Hancock,
1985; Engelder, 1989), which occur in artifi cial ex posures of the
Lower Quader on the nor thern slopes of Mi elna Hill (loc. 7).
ing the Róg plateau are cut by narrow, rectilinear, N–S and NNW–SSE-trending deep valleys (loc. 10), whose formation was probably induced by a large concentration of fractures within brittle shear/fault zones.

Similarly to the N Róg faults, brittle deformations, which crop out in the southern part of the Róg plateau (Fig. 3), show a NNW–SSE trend (320 ± 340°), with fault dips ranging within 80–90° (loc. 11). Vertical fault planes were observed in several abandoned quarries of the Lower Quader. Characteristics of the fault planes and their orientation similar to that of the regional system of extensional fractures indicate that the fault zones are strictly related to joints (Fig. 5A, C, G). Some of the vertical NNW–SSE-trending joints with a maximum spacing of 30 cm consist of packages of intensely sheared and deformed sandstones (Fig. 5E). They should be referred to as tensile cracks reactivated as cataclastic shear zones (e.g., Pluim and Marshack, 2004). Fracture surfaces are visibly affected by low-angle, short R-shear fractures (Fig. 5F), hackles and en echelon cracks (Fig. 5B), which indicate dextral movements with a normal component. Most of these faults led to a change in the orientation of Cretaceous strata (cf. Fig. 3).

Fault surfaces, contrary to the N Róg strike-slip and normal faults, are usually striated (Fig. 5A) and display the presence of asymmetric steps, ridges (Fig. 5D) and slickensides (Fig. 5A). In addition, plumeose structures can be observed on adjacent joint surfaces (Fig. 5G). Nearly vertical strike-slip faults are strictly associated with listric or nearly planar normal faults and their splays (Fig. 5D) that form conjugate fault sets. Listric faults trend mainly parallel, less oblique to the strike-slip fault planes (Fig. 5F). In a small abandoned quarry (loc. 12) located to the west of the previously described exposures, en echelon fault-arrays were recognized. They are related to left-step, strike-slip faults determined also by cartographic methods. In this quarry, the sandstones are strongly fractured, which was probably the main reason of its abandonment.

The fault pattern and their kinematics indicate that several splay and en echelon faults formed a negative flower structure that runs in a NNW–SSE direction across the Róg table mountain (Fig. 6).

**Fig. 4. Structural features of the N Róg faults**

A – dextral fault (sf on the diagram) with low-angle R-shears and the perpendicular joint surface (j on the diagram) affected by normal fault (nf on the diagram) – western limb of the Róg Flower Structure (loc. 8); B – set of normal faults with a strike-slip component – western limb of the Róg Flower Structure (loc. 8); C – breccia zone within the western limb of the Róg Flower Structure (loc. 9), the symbols ⊗ and ⊘ show motion between blocks bounded by strike-slip faults: away and toward an observer, respectively; D – contractual faulted hybrid joints within the breccia zone (loc. 9).

**OTHER EVIDENCES OF FAULTING IN THE ŁĄCZNA REGION**

Apart from the major faults described above, several minor faults and fault zones were identified in the Łącżna area. They are manifested in the morphology of the valleys and ridges and exposed in numerous quarries and natural exposures surrounding the highest hills of the region. Distinct evidence of faulting was found within the Buntsandstein exposures in the western termination of the Czarna Struga Valley, between Dziób and Chochol hills. In a small quarry located to the north of the road connecting the villages of Różana and Łącżna, nearly vertical NNW–SSE-trending joint sets and extension fractures linked with normal faults can be observed. Strongly fractured sandstones occur in the eastern part of the quarry (loc. 13;
Fractures are related to the N–S and NNW–SSE-oriented sinistral strike-slip faults that cause a change in orientation of arkosic sandstones. The orientation of these fractures strongly coincides with the elongation of well-preserved tors (Samotna, Bliźniaczki and Czartowskie Skąły groups) on the convex slopes of Dżób and Chochol hills (Fig. 3; loc. 14). On the eastern wall of the Samotna tor (Fig. 7B) occur left-stepping, en echelon cracks and low-angle R-shears. Low-angle fractures and slickensides were recognized also on the steep walls of the Czartowskie Skąły group. Tectonically affected tors form N–S-oriented ridges (Dżób and Rogal hills), which are perpendicular to the axis of the Czarna Struga Valley and emphasize the structurally controlled local watershed zone between the Nysa Kłodzka and Metuja River basins.

Fig. 7A). Fractures are related to the N–S and NNW–SSE-oriented sinistral strike-slip faults that cause a change in orientation of arkosic sandstones. The orientation of these fractures strongly coincides with the elongation of well-preserved tors (Samotna, Bliźniaczki and Czartowskie Skąły groups) on the convex slopes of Dżób and Chochol hills (Fig. 3; loc. 14). On the eastern wall of the Samotna tor (Fig. 7B) occur left-stepping, en echelon cracks and low-angle R-shears. Low-angle fractures and slickensides were recognized also on the steep walls of the Czartowskie Skąły group. Tectonically affected tors form N–S-oriented ridges (Dżób and Rogal hills), which are perpendicular to the axis of the Czarna Struga Valley and emphasize the structurally controlled local watershed zone between the Nysa Kłodzka and Metuja River basins.

Fig. 5. Structural features of the S Róg faults (Lower Quader exposures, loc. 11)

A – nearly vertical planes of strike-slip faults (fp, sf on the diagrams) with visible slickensides (s) and ridges (r); note the SW-dipping normal fault (dotted line) linked with strike-slip deformation; B – en echelon cracks associated with normal fault; C – set of normal- and dextral strike-slip faults associated with joints (j); low-angle R-shears (R) are visible and indicate dextral sense of movements; D – fault Splays (= secondary faults; marked with arrows) associated with strike-slip fault (slickensides marked with dotted lines), perpendicular joint surface (j) affected also by strike-slip fault with ridges (r); E – cataclastic shear zone between vertical joint surfaces; F – dextral strike-slip fault with associated fractures and oblique normal, listric-shaped fault (fp); G – part of a plumose structure (pl) on nearly vertical plane (fp) of strike-slip fault; other explanations as in Figure 4.
Fault geometry and evidence of depocentre migration within a transtensional intra-basinal high... 787

Fig. 6. Block diagram of the northern part of the Róg Flower Structure (vertical exaggeration 1.5)

Geology and localities according to Figure 3; explanations as in Figure 4

Fig. 7. Deformations within arkosic sandstones (Buntsandstein)

A – sinistral strike-slip faults in an abandoned quarry (E Łączna, loc. 13); B – left-stepping en echelon cracks and R-shears on the tor surface (Samotna – loc. 14); C – low-angle listric fault (fp); D – synsedimentary microfaults and cracks (marked with arrows and dotted lines) within coarse-grained sandstones; lower part of layer is undeformed (loc. 16); other explanations as in Figure 4
The N–S-trending ridge of Dziób Hill, untypical for this area, is bounded by steep escarpments (40°) and constitutes probably the prolongation of this displacement zone (Fig. 3). It is also confirmed by poorly exposed cataclasites and fault breccias that occur within the Lower Quader exposures of the northern slopes of Dziób hill (loc. 15).

Several meso-scale deformations were also recognized in the Buntsandstein exposures near Zdoòov (loc. 16). They include, for example, low-angle listric faults with a throw up to ~0.5 m (Fig. 7C) and synsedimentary microfaults with a few centimetres throw (Fig. 7D).

Several mesofault planes and deformation structures were also observed in abandoned quarries of the Lower Quader in the Czech part of the study area, especially in the Libná region. In periodically active pits ("Levobřežní lomy u Libné"; loc. 17) situated on the southern slopes of an unnamed hill (659 m a.s.l.; Fig. 3), distinct fault mesostructures were observed. Similarly as in the N Róg faults, the fault planes are mainly vertical, N–S-trending and strictly coinciding with orthogonal joints (Fig. 8). Damage zones related to normal- and strike-slip faults contain clay gouges that developed within brittle shear zones. Features of fault planes (stratified ridges, slickensides and grooves) indicate dextral movements between sandstone blocks.

**KAOLINITIC SANDSTONES – DISTINCT HORIZON WITHIN THE INTRASUDETIC SYNCLINORIUM**

Kaolinitic sandstones occur only locally in the topmost part of the Bohdašín Formation. In the study area, they were recognized in exposures on the western slopes of Róg Hill, on the southern slopes of Drogosz Hill and on the western slopes of Strážný Vrch and Milena hills (Fig. 3). In the Krzeszów Brachysyncline, kaolinitic sandstones were also documented in abandoned quarries located near Jawiszów and Krzeszów (Góra Świętej Anny locality). Despite the fact that artificial exposures of kaolinitic sandstones are small and limited only to road crosscuts in the study area, narrow exposures of these sediments are highlighted by the occurrence of distinct kaolinitic soils with dispersed pebbles. Results of geological mapping suggest that the preserved thickness of the kaolinitic sandstones is between 10 and 15 m (Fig. 3).

Kaolinitic sandstones (Ger. *Weißer Kaolinsandstein*) were marked as a separate lithological unit on the first detailed geological map of the Łączna region by Berg and Dathe (1905/1906), and assigned to the Middle Buntsandstein. Mroczkowski (1977) assumed that they might constitute a relic of an ancient weathering cover developed at the top of the Buntsandstein deposits from the Early Triassic to the Late Cenomanian. Don et al. (1981) marked these sediments also on top of Buntsandstein arkosic sandstones in the Róg area and speculated that they might represent Lower Triassic sediments that were reworked during the Late Cretaceous marine transgression. According to Prouza et al. (1985), the kaolinitic sandstones that crop out in the Zawory Ridge are lithologically similar to the Devět křížů sandstones (Holub, 1966) described in detail by Uličný (2004) and Wojewoda et al. (2016). The Devět křížů sandstones are exposed in the close vicinity of the southwestern boundary of the Intrasudetic Synclinorium (the Hronov–Poříč Fault Zone HPFZ, cf. Fig. 1), within the morphological elevation between the Náchod and Trutnov basins (Wojewoda et al., 2016). Both the origin and age of these deposits remain controversial (see: Geological Setting), although based on regional stratigraphic subdivisions and the lithological similarity to the kaolinitic sandstones from the adjacent North Sudetic Synclinorium area, they are probably late Early or early Middle Triassic in age (Chrzastek, 2002).

In the lower part of the succession, kaolinitic sandstones are represented by coarse-grained, weakly lithified and strongly kaolinitised arkosic wackes. Towards the top of the profile, sandstones are whitish to light grey and reveal distinct platy parting, which is usually bedding-parallel. They represent moderately to well-sorted, medium- and coarse-grained, sub-lithic and sub-arkosic arenites. Most of framework grains range from 0.5 to 0.8 mm, rarely 1 mm in size (Fig. 9). Based on the classification of Pettijohn (1975), the grains are usually sub-angular, intermittently sub-rounded and rounded. Monocrystalline quartz is the major mineral component of these deposits. Its content varies between 70 and 80%. Strongly fractured and kaolinitised feldspar grains constitute up to 10% of the framework grains. Lithic fragments are represented by metamorphic (mostly gneisses and quartzites) and igneous rocks (granites, granodiorites, rarely rhyolites), which are also affected by kaolinisation processes. Most of the lithic grains, probably remnants of mica schists, are strongly sericitised and kaolinitised. The sandstone matrix is composed of kaolinite-group minerals with a small admixture of illite.
DISCUSSION AND SUMMARY

The geometry and kinematics of major deformation structures cutting the study area indicate mainly a normal and dextral strike-slip component of the movements. The oldest deformations of the area include pre-lithification (synsedimentary) faults and fissures, which were formed due to NE–SW tension and coincide with the orientation of the youngest brittle deformation structures: joints, normal faults and strike-slip faults. NW–SSE-oriented joints developed presumably as opening mode (mode I) fractures (Pollard and Aydin, 1988) that are parallel to the σ1 axis and perpendicular to the maximum tensile stress (Ramsay, 1980). The problem of joints formation within the IS has not been examined in detail so far and their compressional origin cannot be excluded. However, many authors consider that joints formed in an extensional setting (Jerzykiewicz, 1968; Stejskal et al., 2012; Wojewoda, 2012). Well-exposed plumose structures and hackles occurring on the joint surfaces confirm this view and their tensile origin (cf. Mierzejewski, 2015). Significant changes in the joint orientation are caused by shearing along joint surfaces (Wilkins et al., 2001; Myers and Aydin, 2004), which produced fractures and fault planes oblique to the pre-existing joints. Hence, most of the observed fault planes should be classified as reactivated joints (“faulted joints”; Wilkins et al., 2001). Shearing of joints is confirmed by numerous NW–SSE-trending vertical strike-slip and normal faults, which are parallel and sub-parallel to the joint surfaces. ESE–WW–oriented systematic sets of joints, perpendicular to the joints described above, were also reactivated with a small normal component. Short low-angle R-shears and en echelon fractures, recognized in the study area, have been formed obliquely to the strike-slip faults and suggest permanent dextral movements occurring in a transtensional setting. Sinistral displacements occurring in exposures to the east of Łączna represent high-angle R-shears, antithetic to the main, NW–SE-trending dextral strike-slip faults. Contractual structures comprise only faulted hybrid joints (Marin-Lechado et al., 2004), formed due to local compression within the negative flower structures.
Structural analysis has shown that the Łączna Anticline area displays distinctive structural, morphological and lithological features typical of intrabasinal elevated areas bounded by extensional strike-slip and normal faults. The southern branch of the LF and the oblique faults constitute the principal displacement zone (PDZ; Christie-Blick and Biddle, 1985) that subdivides this part of the IS into two rhomb-shaped, half-graben units with distinct subsidence centres (Fig. 11). The PDZ is associated with negative flower structures (Róg Flower Structure) to the west and normal- and strike-slip faults to the east. The LF affected the development and geometry of the entire KB area. This is confirmed by differences in total vertical displacements on this fault observed within the KB and LA, which are ~40–50 m within the KB and 20–25 m within the LA.

Kaolinitic sandstones constitute a distinct horizon in the northern part of the IS. Their distribution is strictly related to tectonic zones that occur within the southern limb of the KB and is also limited to a morphological elevation (Zawory Range). Moreover, kaolinitic sandstones do not occur below the Cretaceous deposits in adjacent, present-day basin structures, which is confirmed by borehole data (Wojtkowiak et al., 2011). During the Early Triassic, the KB and PB constituted a sedimentary basin with a single, central depocentre formed above the strike-slip fault (Fig. 11A). This is confirmed by Early Triassic and even Permian palaeogeographic reconstructions offered by Mroczkowski (1977) and Siwiński (1981). Permanent transtension of the basin basement resulted in the formation of dual-depocentre geometries (Dooley and Schreurs, 2012) within the interior of a pull-apart basin (the KB in this case, Wojewoda, 2007). During late Early and probably Middle Triassic times, the southwestern and northeastern terminations of the KB constituted narrow basin depocentres separated by an elongated, NW–SE-oriented elevation (Fig. 11B). This process is well-documented by analogue modelling of transtensional basins (Wu et al., 2009). Probably in the late Early Triassic, strongly kaolinitic, regolith-type weathering covers developed on the sandstones of the Bohdašín Formation. The regolith succession in the topmost part of the sandstones developed without disintegration of their primary structure. It is documented by sedimentary structures observed in the lowermost part of the kaolinitic sandstones succession, similar to the Buntsandstein deposits. Strong kaolinisation of the uppermost part of the Buntsandstein deposits was claimed by many authors (Jerzykiewicz, 1971; Mroczkowski, 1977; Don et al., 1981). During the Early to Middle Triassic, the KB region was a flat plain area periodically flooded by a shallow sea (Wojewoda et al., 2016). A conglomerate bed in the bottom of the “slab sandstones” is interpreted herein as a transgressive deposit. Small-scale, wave-ripple lamination observed on the bedding surfaces of kaolinitic sandstones shows that the depth of individual, probably isolated water bodies, did not exceed a few decimetres. It is therefore quite probable that the local depressions were occupied by shallow ephemeral salt lakes. The playa-like depressed area was occasionally cut by narrow river
channels. This is documented by shallow erosional scours within the kaolinitic sandstones. Strongly bioturbated coaly mudstones that occur within the kaolinitic sandstones in the vicinity of Krzeszów (Jerzykiewicz, 1971) represent salt marsh environment deposits (Kowalski, 2016) linked with a shallow sea transgression. After regression and a long period (~140 My) of denudation, in the Late Cretaceous (Late Cenomanian), the study area was flooded by a shallow epicontinental sea. At that time, the KB area was a flat, low-lying penepaled land built of older sedimentary rocks (Wojewoda, 1997; Uličný, 2004). Preservation of kaolinitic sandstones below the transgressive Upper Cenomanian clastic deposits was possible only within local structural depressions near Łączna and Krzeszów.

Regional palaeogeographic reconstructions and structural analysis conducted in the KB and LA areas reflect general consistency and permanent geodynamic development of this area in a transtensional strike-slip setting. Presented results of the structural and sedimentological studies do not coincide with palaeo- and recent stress reconstructions proposed by Nováková (2014) and Prouza et al. (2015) for the western border of the Police Brachysyncline (the HPFZ). According to these authors, the HPFZ and adjacent fault zones consist of inverse faults and even gently dipping thrusts, associated with...
rocks are caused only by brittle faulting (Fig. 12) linked with the deformation. Changes in dip orientations of the sedimentary Cretaceous sediments do not show any evidence of continuous al., 2016). It should be also emphasized that the Triassic and IS development was termed earlier as the “Saxonian Ce nozoic (Mio cene?) stage of basin in ver sion. This stage of the IS, pro posed by other authors (Wojewoda, 2007; Wojewoda et Nováková, 2014), strike-slip (Wojewoda and Kowalski, 2016) and even flexural (Valenta et al., 2008) tectonics. The total absence of inverse faults and pop-up structures in the study area, especially in the vicinity of the LF and other recognized faults, significantly contradicts the opinion about strong subhorizontal compression (Prouza et al., 2015) produced by the “Alpine orogeny” (Täsl er et al., 1979; Kozdrój and Cymerman, 2003) within investigated part of the IS. Described structures confirm the thesis about the permanent extensional development of the IS, proposed by other authors (Wojewoda, 2007; Wojewoda et al., 2016). It should be also emphasized that the Triassic and Cretaceous sediments do not show any evidence of continuous deformations. Changes in dip orientations of the sedimentary rocks are caused only by brittle faulting (Fig. 12) linked with the Cenozoic (Miocene?) stage of basin inversion. This stage of the IS development was termed earlier as the “Saxonian tectonogenesis” (Jeryzkiewicz, 1971; Tásler et al., 1979; Stejskal et al., 2012). The terms brachyanticline or brachysyncline are usually applied to folds whose amplitude decreases to zero in both directions (Park, 2005). There is no evidence of the presence of brachyfolds or folds in the study area. Hence, the term “Lączna Elevation” is adequate and correct in the case of the study area and reflects its morphological and structural position.

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