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# Tectonic evolution of the Carpathian Foredeep and its influence on Miocene sedimentation

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Structural analysis of the Miocene deposits of the northern marginal part of the Carpathian Foredeep, from the Nida Trough in the west to Tarnobrzeg in the east shows that three tectonic phases affected Neogene deposits of the Carpathian Foredeep: (1) an Early Styrian phase — Early Badenian synsedimentary movements, (2) a Moldavian phase (Late-Styrian) — Early/Late Badenian, after deposition of gypsum and during deposition of the lower part of the *Pecten* Beds, (3) an Attican phase — Sarmatian-Pliocene(?), after deposition of the Krakowiec Beds. The two Badenian phases involved reactivation of NW–SE basement faults Early-Styrian phase activity along these faults resulted in facies changes and thickness variability in the Baranów Beds. Reactivation of these faults during the Moldavian phase resulted in a considerable increase of their throws. Strata in the footwalls of faults were antithetically rotated to form structural traps for subsequent hydrocarbon and sulphur deposits. During the Attican phase, a horizontal compression stress field (compression direction — 30–50°) produced transverse and oblique faults (NE–SW and ENE–WSW) with strike-slip and oblique-slip movements. These faults were responsible for "pumping" hydrocarbons into the earlier formed traps. Minor deformations within gypsum deposits and seismotectonic features of the Witów Gravels are also discussed.

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Key words: Carpathian Foredeep, Neogene, neotectonics, mesostructural analysis, teledetection, synsedimentary faults, strike-slip, faults, tectonic breccias.

## INTRODUCTION

This paper discusses Miocene outcrops in the northern marginal zone of the Foredeep (Fig. 1). In the west, the Nida Trough extends south as far as the Vistula river (Fig. 2), while in the east the Osiek–Baranów (Fig. 4) and Machów–Jeziórko–Jamnica (Fig. 3) sulphur deposits occur. These Miocene rocks were studied using:

— mesostructural analysis,

— remote sensing,

— geological cross-sections based on boreholes and geophysical data.

Mesostructural analysis of the Nida Trough involved: Miocene outcrops between Gacki and Czarkowy along the left bank escarp of the Nida river (Fig. 2) (Nida Horst slope — Fig. 1); outcrops in the Proszowice region (Gniazdowice, Koniusza — Fig. 2; Słomniki Horst — Fig. 1); and four exposures in the central part of the Nida Trough (Trzonów, Działoszyce, Witów, Majkowice — Fig. 2; Działoszyce Depression — Fig. 1). In the east, the Machów sulphur open-pit near Tarnobrzeg was the field of study (Figs. 1, 3). Strike and dip of beds, joints, faults and fault throws were measured (analysis of joint patterns will be presented in a separate paper).

Remote sensing comprised satellite images and radar photographs interpreted by  and Graniczny (1983) and supplemented by Piątkowska (Figs. 2–4), and interpretation of aerial photographs by the author (Figs. 2, 4).

Data from 55 boreholes drilled through the Miocene sequence and reaching underlying rocks were used for the construction of geological cross-sections (Figs. 5, 7). Variations in the dip of the Chemical Beds comes from shallow seismic sections, by the Geological Research Firm in Cracow 1977 (Fig. 5, sections I–I – III–III) and/or from field observations and borehole data correlations.

Other studies collated included recent interpretation or reprocessing of older seismic sections (Jawor, 1983*a, b*, 1997; Pietsch and Krzywiec, 1994; Dziadzio and Jachowicz, 1996; Krzywiec and Pietsch, 1996; Pietsch *et al.*, 1997; Krzywiec, 1997*a, b* and others) showing the fault zones which transect

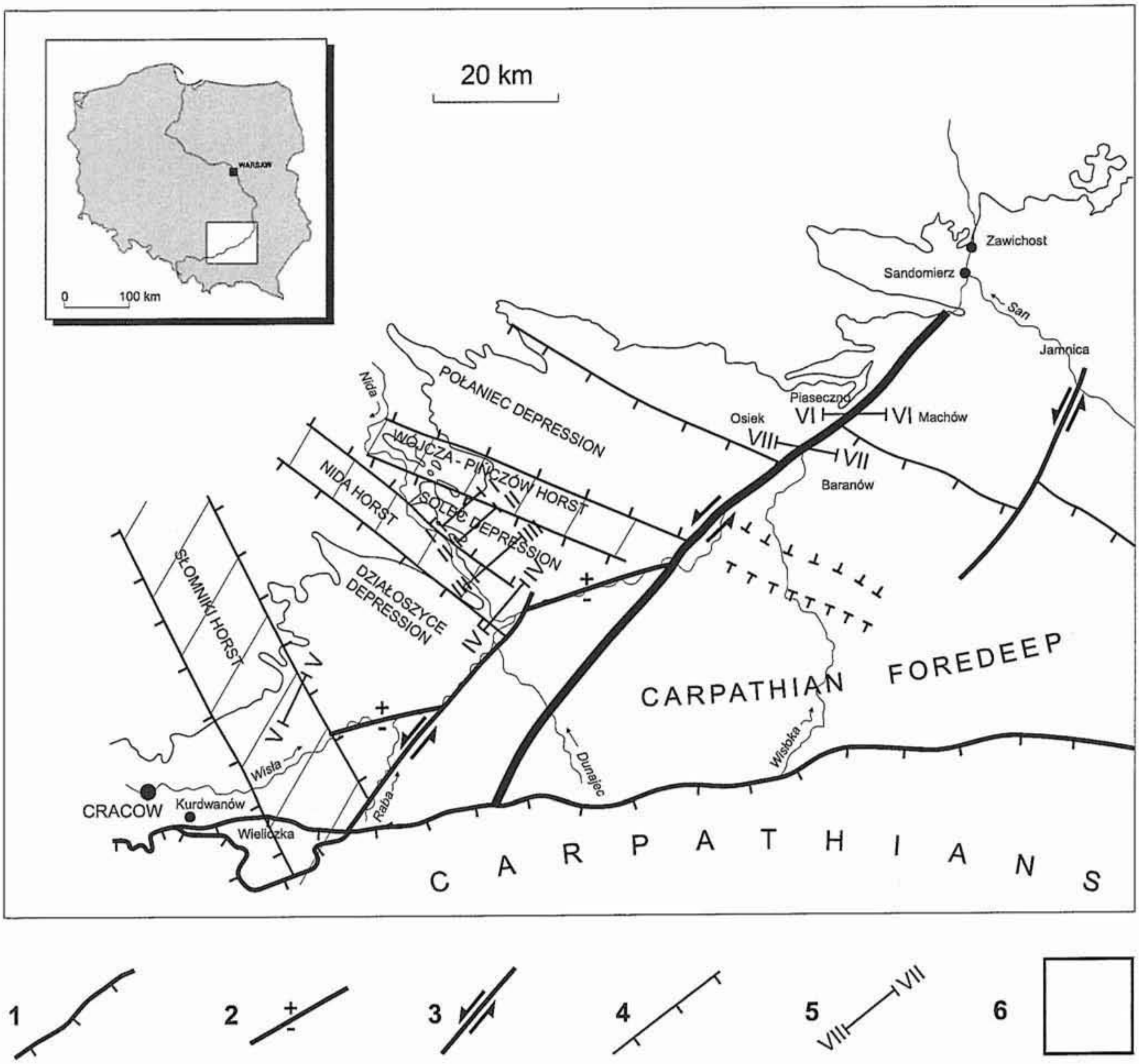


Fig. 1. Structural sketch of Carpathian Foredeep (central part) and location of the study area

1 — northern Carpathians margin, 2 — normal faults, 3 — strike-slip faults, 4 — normal faults bordering the horsts, 5 — geological section lines, 6 — study area

Miocene deposits, especially in the central and eastern parts of the Carpathian Foredeep (south of the Vistula river).

The northern part of the study area is less suitable for geophysical investigations, as the Miocene deposits are thinner. Seismic boundaries visible here in seismic sections reflect velocity contrasts within the Mesozoic and basement, and all the Miocene sequence is recorded within the limit of image separation.

A new plate tectonics-based interpretation of the structural evolution of the Carpathian Foredeep during the Miocene has been given by Jarosiński (1998).

Of relevance here is the possible influence of compaction, rather than tectonics, on the formation of normal faults within the Miocene deposits (particularly in the central part of the Carpathian Foredeep) (Poprawa, 1999; Poprawa and Krzywiec, 1999). Preliminary results of subsidence analysis for the whole Carpathian Foredeep have been given by Oszczypko (1996, 1997, 1999).

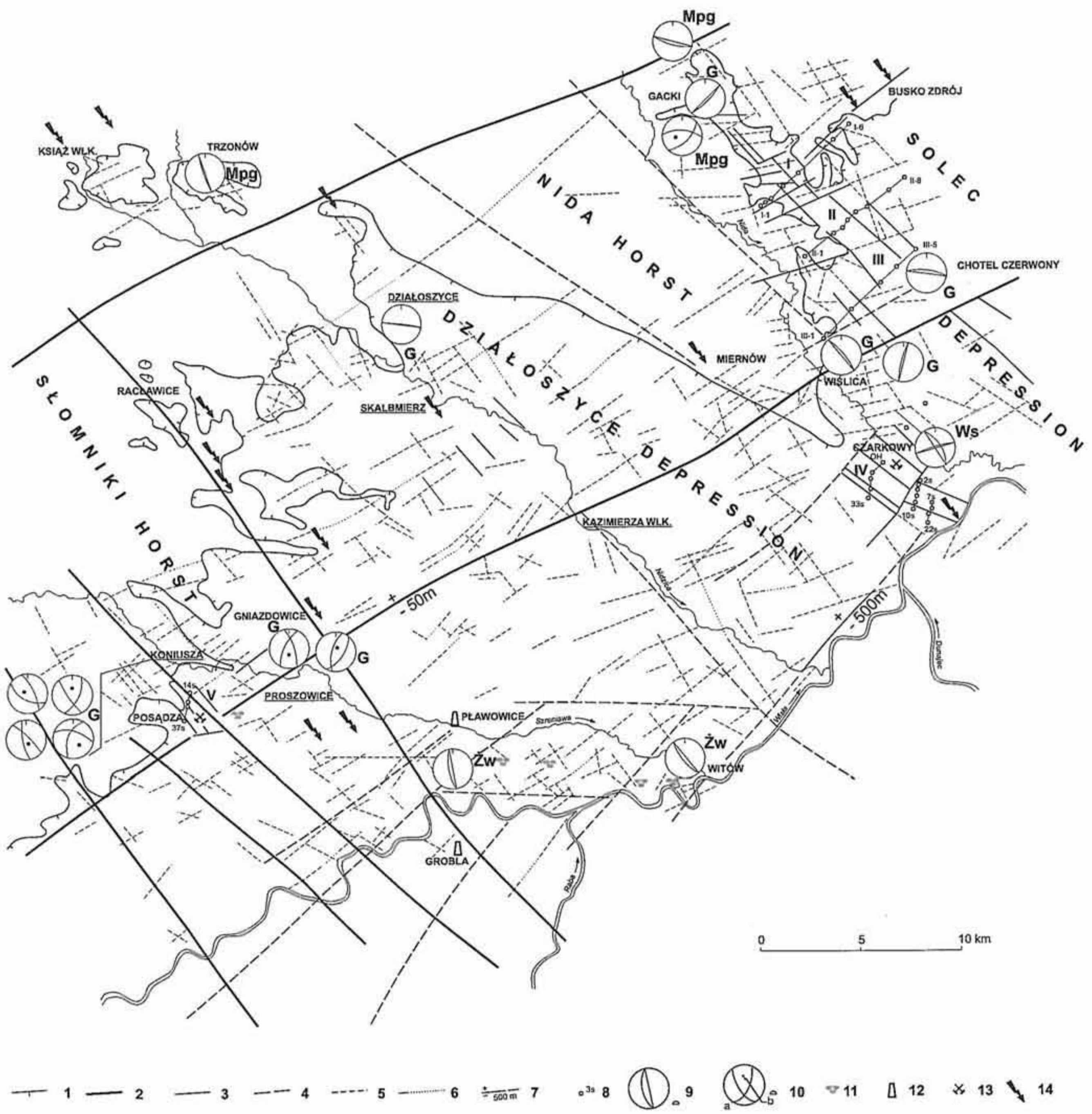


Fig. 2. Tectonic map of the southern part of the Nida Trough

1 — northern extent of Miocene deposits (after Czarnocki, 1950), 2 — important regional dislocations, 3 — dislocations documented from borehole and geophysical data and visible on the air-photos, 4 — satellite image lineaments, 5 — air-photo lineaments, 6 — obscured segments of lineaments, 7 — fault throw estimated from borehole data for the bottom of the anhydrite horizon, 8 — boreholes, 9 — fault orientations, 10: a — orientations of conjugate complementary faults, b — orientations of principal stress $\sigma_1$, 11 — outcrops of the Witów Gravels, 12 — Miocene natural gas field, 13 — abandoned sulphur-mine, 14 — seismic shock effects; Żw — Early Quaternary (Preglacial), Witów Gravels, Miocene, Ws — limestones with native sulphur, G — gypsum, Mpg — sub-gypsum beds; I–III — cross-sections

## FAULT PATTERN

Until recent the main source of information on the location of fault zones within the Miocene of the Carpathian Foredeep was the structural map of the basement (Karnkowski and Łapinkiewicz, 1965; Kruczek, 1968; Jawor, 1970; Oszczypko *et al.*, 1989). The fault zones could not then be properly elucidated owing to the lack of good quality seismic data. At that time it was believed that the faults cutting the Miocene deposits

Fig. 3. Lineament map of the sulphur deposit Machów–Jeziórko–Jamnica

were a simple propagation of basement faults (Cisek and Czernicki, 1965; Jawor, 1970; Łyczewska, 1971, 1975; Osmólski, 1972; Pawłowski *et al.*,1985; Kubica, 1992, 1996, and others). Structural maps of the Carpathian Foredeep Miocene, derived from closely spaced seismic sections, are now in preparation (Krzywiec, 1999) or already exist for certain gas field areas (Baran *et al.*, 1996; Fik, 1996). Therefore, the author has used remote sensing methods, on the assumption that the pattern of lineaments should mostly reflect the fault pattern, since Miocene deposits are at outcrop or underneath thin Quaternary deposits. It is significant that recent tectonic activity has been recorded in this area (see Fig. 2).

The lineament pattern of the area studied is composed of 2 major sets (Figs. 2–4):

1. NW–SE to WNW–ESE, or longitudinal.
2. NE–SW to NNE–SSW, or transverse.

Most of the lineaments are transverse, with an azimuth of 50–80°. The longitudinal probably correspond to older basement faults, which have not cut through the Upper Miocene argillaceous deposits.

## IMAGE AND DEPTH EXTENT OF FAULTS

Transverse faults have strike-slip or oblique-slip (Fig. 6) and less frequently dip-slip motions (Pl. I, Figs. 1–3; Tab. 1). This group of faults is represented by the Kurdwanów–Zawichost Zone (Figs. 1, 7) (*cf.* Osmólski *et al.,* 1978; Jarosiński, 1992; Kotański, 1997), the northern boundary of the foredeep which is visible as a distinct morphostructure. Large throws on these faults occur south of the Proszowice–Kazimierza Wielka–Wiślica line, where they reach about 30–100 m and increase up to about 500 m near the Vistula river (Fig. 2) in the Nida Trough (Kurdwanów–Zawichost Zone). North of this line, vertical slips on transverse faults usually do not exceed 10 m. In the Tarnobrzeg sulphur deposits area, throws on transverse faults belonging to the Kurdwanów–Zawichost Zone do not exceed 40 m (Fig. 7). Between Kraków and Nowy Korczyn, the Vistula river valley twice turns from E–W to NE–SW. These changes correspond with two lineament sets which, according to Osmólski *et al.* (1978), represent strike-slip (NE–SW) and dip-slip (E–W) faults. The maximum stress direction axis $\sigma_1$, obtained using conjugate disjunctions and complementary shears in Miocene deposits within the Kurdwanów–Zawichost Zone (Krysiak, 1986, fig. 2) indicates compression at an azimuth of 30–50°, and this is responsible for the NE–SW strike-slip activity of the transverse faults. The existence of horizontal stresses from the SW in this area was already assumed by Osmólski (1972).

Połtowicz (1978) considered that the Kurdwanów–Zawichost Zone as a whole is an oblique-slip structure, and its horizontal displacement in Jurassic deposits attains 8 km.

According to Trafas (1975), the bends in the Vistula river valley between Kraków and the Raba river mouth resulted from southward displacement (between Kraków and Niepołomice)

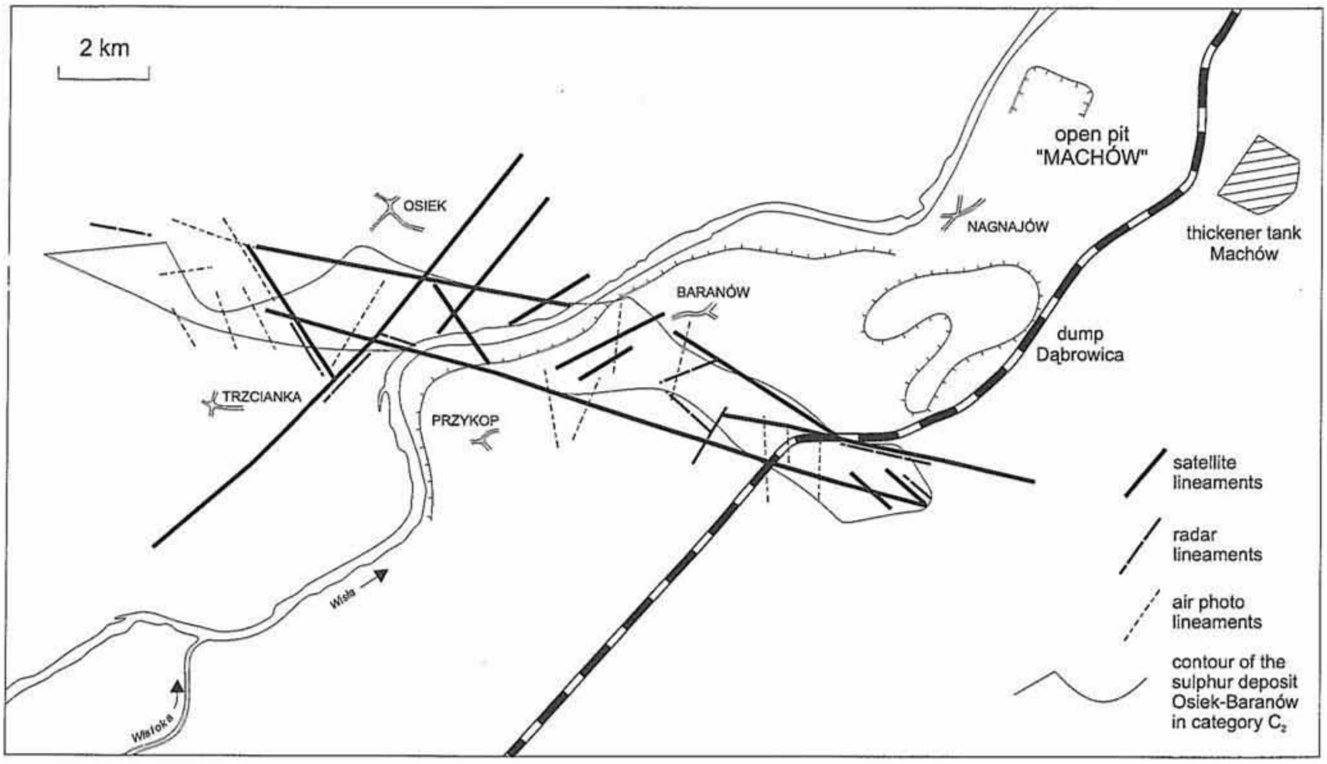


Fig. 4. Lineament map of the sulphur deposit Osiek–Baranów

and northward displacement (near the Raba river mouth) by alluvial fans of the left and right bank tributaries of the Vistula river. However, these river bends and tributary configurations are fault-influenced (Osmólski *et al.*,1978). I consider that this indicates neotectonic fault activity affecting erosion and accumulation in the river system.

Longitudinal faults affecting the Miocene deposits are dip-slip faults. They form a row of horsts and troughs and wider tectonic depressions (Figs. 1, 5–7). Longitudinal faults are steep or inclined towards their footwalls (with dips of 40–70°). In general, the strata in the footwall are normally rotated, i.e. towards the throw direction, although sometimes, near the horsts, reverse drag of beds is observed (Fig. 5, section V–V).

Maximum throw on the longitudinal faults in the study area is 20–40 m (Fig. 5, sections I–I – III–III, V–V; 7), sporadically reaching 100 m (Fig. 5, section IV–IV). Kubica (1996) recorded throws reaching 120 m along faults bordering the Osiek–Baranów sulphur deposit, but this value was calculated by correlating the base of the Chemical Beds across considerable distances (about 20 km). The great thickness variability of the Miocene deposits in this area suggests that this might be an over-estimate (*cf.* Krysiak, 1986*b*). Elsewhere, throws are of up to twenty metres or, along the strike-slip faults, there may be no vertical displacement. These small throws incurred doubt, formerly, about faulting in the Miocene deposits of the Carpathian Foredeep (in particular Jucha, 1974, 1985, 1995).

Only in the central and eastern part of the foredeep (e.g. Wielkie Oczy Trough), do fault throws attain large values.

## SYNSEDIMENTARY FAULTING DURING SEDIMENTATION OF THE BARANÓW BEDS

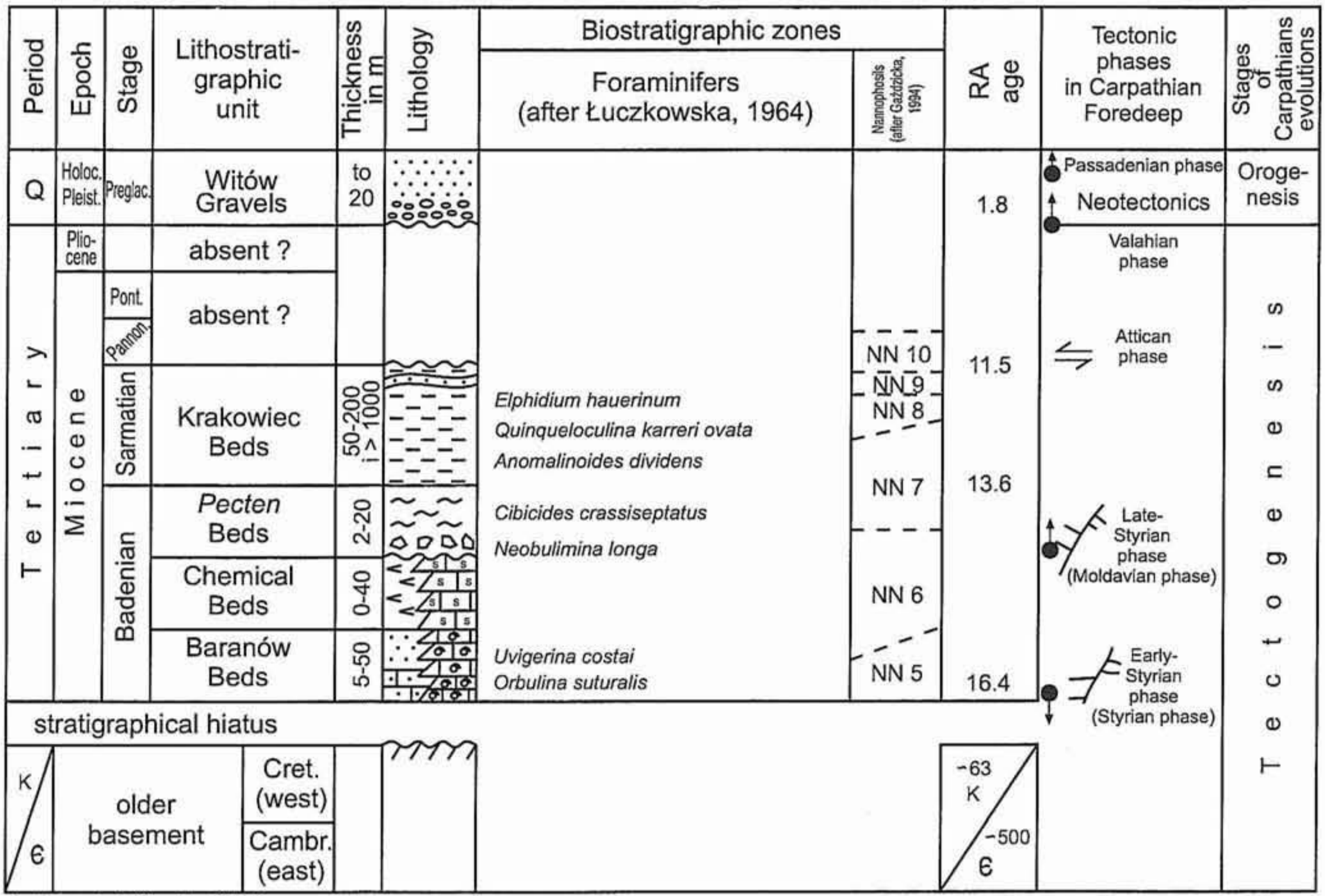
The nature of the Miocene marine transgression on to the Carpathian Foreland, and the deposition of the Lower Badenian deposits underlying the evaporitic series raise the question of what relief existed during the transgression. There are two different opinions on the subject. Supporters of significant pre-Miocene basement relief (Dżułyński, 1953; Radwański, 1968, 1969, 1973; Oszczypko and Tomaś, 1976; Konior, 1981; Szymanko and Wójcik, 1981; Cisek, 1983; Jawor, 1983*a*; Peryt and Kasprzyk, 1992; Połtowicz, 1994, 1997) note the varying thicknesses and facies of Badenian sub-evaporitic deposits, and changes in microfaunal assemblages accompanying the lithofacies variations (Alexandrowicz, 1979). The varied relief of the Miocene sea floor was attributed to pre-Miocene tectonic movements and to fluvial erosion (Dżułyński, 1953; Radwański, 1973; Konior, 1981), or exclusively to fluvial activity (Oszczypko and Tomaś, 1976; Oszczypko, 1996; Cisek, 1983; Jawor, 1983*a*; Połtowicz, 1994). A different opinion is represented by Gradziński (1962, 1963), Bogacz (1967) and Ney *et al.* (1974) who considered that the Miocene sea transgressed on a flat area with relatively low relief, what precludes major Paleogene tectonic activity. Differences in the base of the Miocene deposits were attributed to later tectonic movements. My present studies show evidence for both arguments.

The following has been ascertained:

1. Senonian marls underlying Miocene deposits in the Nida Trough often have high dips (30–40°), and Badenian deposits

Table 1

**Stratigraphic-lithologic-tectonic table**



| Period | Epoch | Stage | Lithostratigraphic unit | Thickness in m | Lithology | Biostratigraphic zones: Foraminifers (after Łuczkowska, 1964) | Biostratigraphic zones: Nannophosis (after Gaździcka, 1994) | RA age | Tectonic phases in Carpathian Foredeep | Stages of Carpathians evolutions |
|---|---|---|---|---|---|---|---|---|---|---|
| Q | Holoc.<br>Pleist. | Preglac. | Witów Gravels | to 20 | | | | 1.8 | Passadenian phase<br>Neotectonics | Orogenesis |
| Tertiary | Pliocene | | absent ? | | | | | | Valahian phase | Tectogenesis |
| | Miocene | Pont.<br>Pannon. | absent ? | | | | NN 10 | 11.5 | Attican phase | |
| | | Sarmatian | Krakowiec Beds | 50-200<br>i > 1000 | | *Elphidium hauerinum*<br>*Quinqueloculina karreri ovata*<br>*Anomalinoides dividens* | NN 9<br>NN 8<br>NN 7 | 13.6 | | |
| | | Badenian | *Pecten* Beds | 2-20 | | *Cibicides crassiseptatus*<br>*Neobulimina longa* | NN 6 | | Late-Styrian phase (Moldavian phase) | |
| | | | Chemical Beds | 0-40 | | | | | | |
| | | | Baranów Beds | 5-50 | | *Uvigerina costai*<br>*Orbulina suturalis* | NN 5 | 16.4 | Early-Styrian phase (Styrian phase) | |
| stratigraphical hiatus | | | | | | | | | | |
| K<br>Є | older basement | | Cret. (west)<br>Cambr. (east) | | | | | -63 K<br>-500 Є | | |

Absolute age after Kováč and Zlinská (1998)

overlie them unconformably (Osmólski, 1972; Łyczewska, 1975). This unconformity can be traced in many boreholes across the Nida Horst zone, particularly around Czarkowy, and also at the boundary of the Nida Horst with the Solec Depression in exposures along the Nida river escarpment.

2. Breccias composed of fragments of Cretaceous marls cemented with glauconitic clays (Fig. 8) commonly occur at the base of the Miocene succession: on slopes, at the margins of horsts and in tectonic troughs.

3. Sub-gypsum deposits in the Solec Depression and the Nida Horst near Czarkowy show structurally controlled facies variability. In the horsts, Lower Badenian white marls, similar to the Cretaceous ones, predominate (Fig. 8, sections I–I, IVa–IVa, IVb–IVb), or the horsts are composed exclusively of Cretaceous deposits (Fig. 8, sections III–III, IV–IV, IVb–IVb, V–V). Grey clays and glauconitic marls occur in tectonic troughs (Fig. 8, sections I–I – IVa–IVa). Good examples of such facies changes are seen in exposures along the Nida river escarpment. In the hanging wall of the fault exposed in the Gacki quarry (Pl. I, Fig. 1), selenite gypsum deposits overlie the Badenian light cream-coloured marls, which resemble the underlying, slightly paler Senonian marls. In the footwall the gypsum deposits overlie dark grey glauconitic clays of the Baranów facies.

In the Działoszyce Depression, the lithological profile of the Sub-gypsum Beds is bipartite (Fig. 8, section V–V): marls occur at the bottom and clays in the upper part. Therefore, only the thickness variations of the Sub-evaporitic Beds and occasionally the extent of basal breccias indicate that the transgression was syntectonic here. Naturally, the character of sedimentation was influenced not only by water depth in turn determined by structural position, but also by basement lithology. Hence, while in the Działoszyce and Solec Depressions, marls and clays were deposited upon Senonian marls, in the area of the Wójcza–Pińczów Horst was lithothamnium reefs developing on hard Jurassic rocks, and the sandy Baranów Beds were deposited on quartzites and shales (uppermost Precambrian-Cambrian) in the Tarnobrzeg area.

4. Thicknesses of the Badenian Sub-gypsum Beds vary widely depending on structural position: from 0–3 m on the horsts to 10–50 m in the tectonic troughs (Fig. 8; *cf.* Osmólski, 1972; Kubica, 1996). Likewise, in other parts of the Carpathian Foredeep, the thickness of the Baranów Beds varies between 0 m up to a few tens of metres (Ołtuszyk, 1968; Cisek, 1983; Jawor, 1983*a*); thicknesses in boreholes exceed 150 m only in

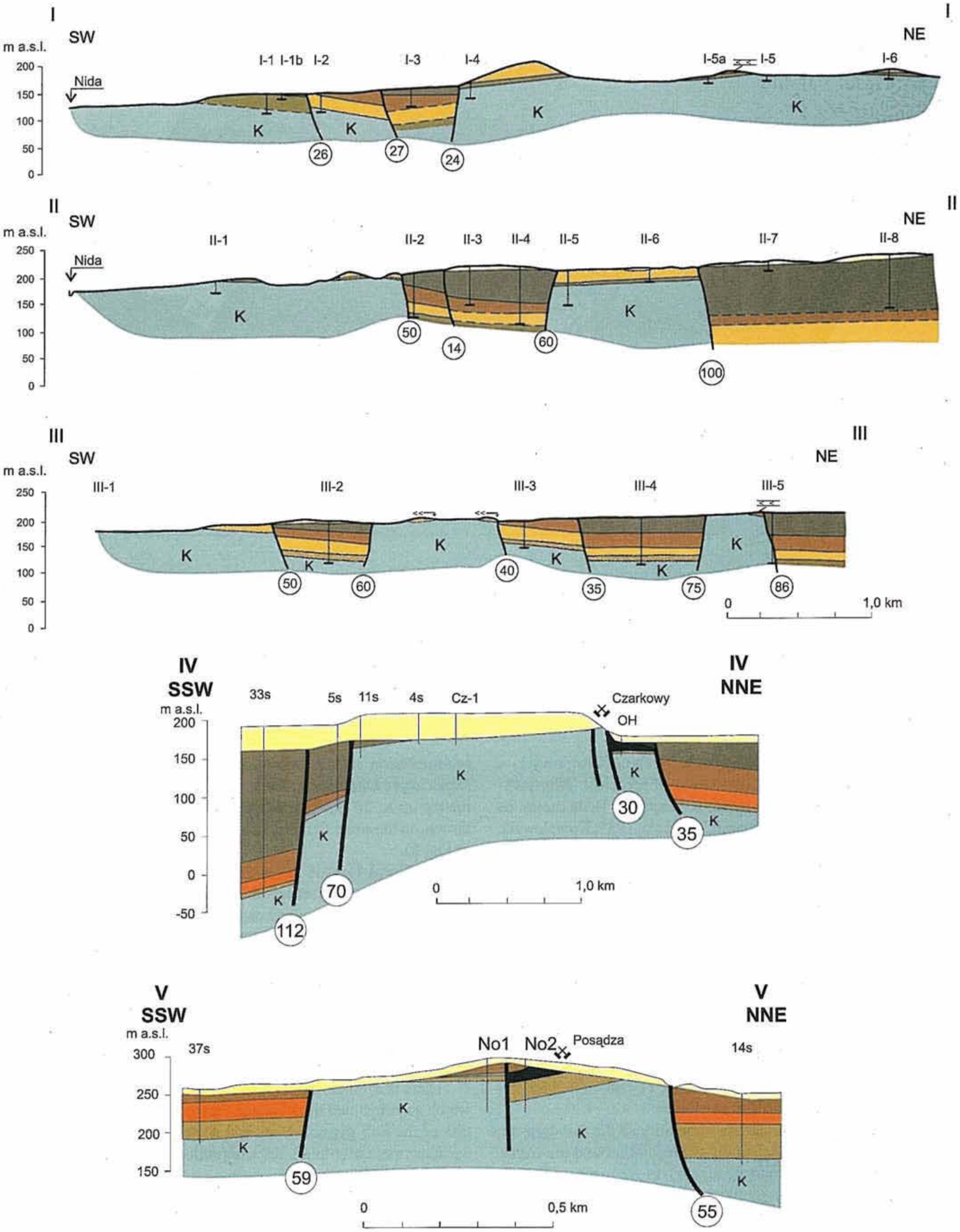


Fig. 5. Geological cross-sections: I–I, II–II, III–III — Solec Depression, IV–IV — Nida Horst, V–V — Słomniki Horst; for location see Fig. 1

For explanations see Figs. 6 and 7

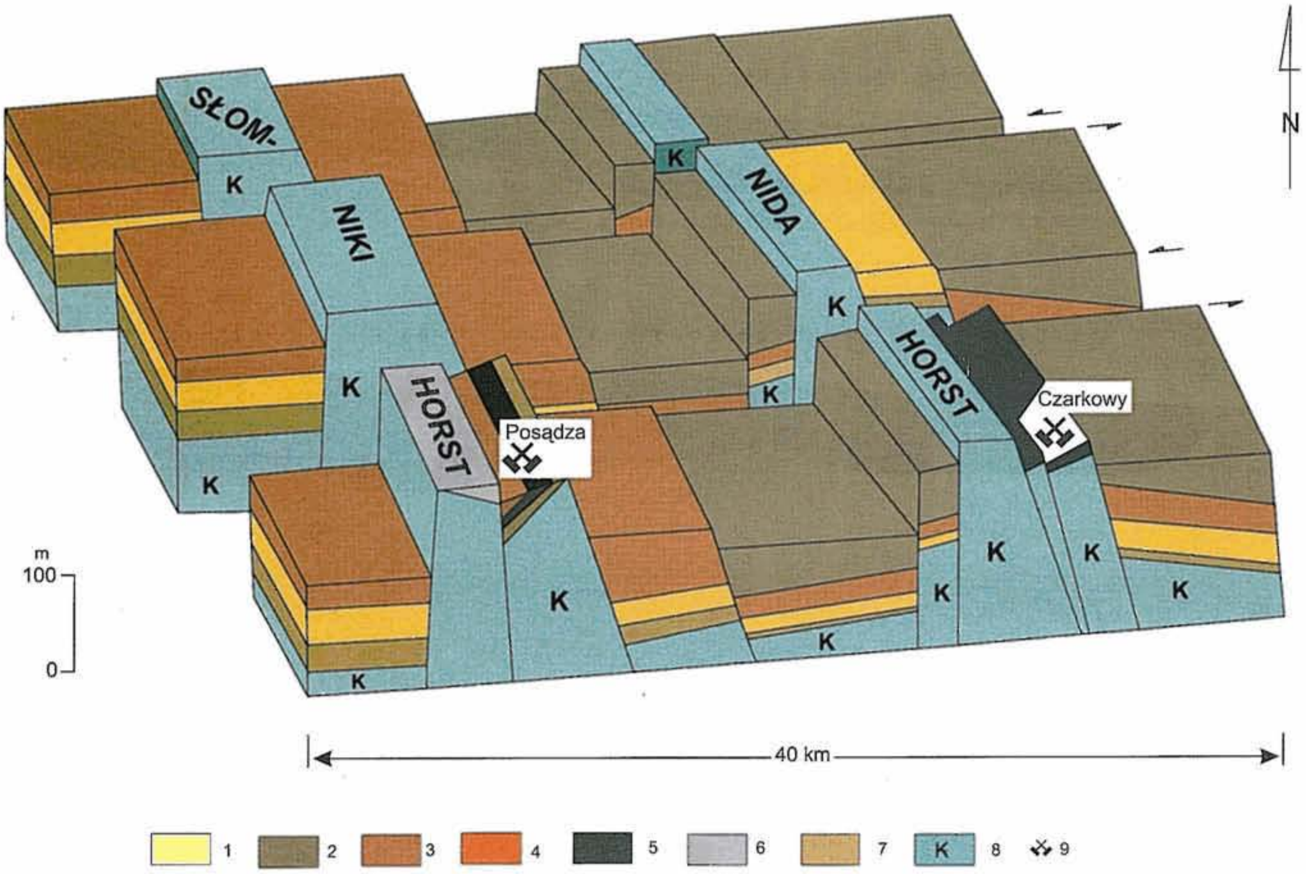


Fig. 6. Block diagram — structural model of the Nida Trough

1 — Quaternary, Tertiary (Miocene), 2 — Krakowiec Beds (on Fig. 7, section VI–VI — undivided clay cover on gypsum), 3 — *Pecten* Beds, 4 — gypsum, 5 — limestones with native sulphur, 6 — barren limestones, 7 — Baranów Beds, 8 — Cretaceous, 9 — abandoned sulphur-mine

vicinity of the Carpathians (e.g. in the borehole Kowalowy 1 — Jawor [1983] including at least 180 m of breccias). This thickness variability in the Sub-gypsum (Baranów) Beds cannot be related to later erosion, because (Czarnocki, 1935; Kowalewski, 1966; Garlicki, 1979), there is sedimentary continuity between the Sub-gypsum Beds and the gypsum.

5. There are areas where gypsum deposits immediately overlie the Palaeozoic or Mesozoic basement without the Baranów Beds occurrence (see Czarnocki, 1947; *cf.* Fig. 5, sections I–I, III–III, IV–IV), or only the Krakowiec Clays overlie the basement (Osmólski, 1972; *cf.* Fig. 7, section VI–VI).

These facts suggest that:

1. The Nida Trough area, composed of poorly weathering-resistant Cretaceous marls, must have been affected by complete peneplanation during the Paleogene break in sedimentation (which lasted 46 My — Tab. 1), obliterating any relief caused by post-Cretaceous tectonics.

2. The facies and thickness variability of the Sub-gypsum Beds, as well as the distribution of breccias around the Cretaceous horsts therefore indicate tectonic movements just before the transgression and/or simultaneous with sedimentation.

3. Because the thickness variability of the Sub-gypsum Beds does not result from erosion, it reflects the magnitude of tectonic movements which accompanied the sedimentation. We can therefore estimate throw values for particular faults during sedimentation of the Sub-gypsum Beds. In the Solec and Działoszyce Depressions the values of throws amount to a few metres up to 20 m, sporadically to 40 m (Fig. 8). Nowadays throws on the same faults are 2 to 30 times greater (*cf.* Fig. 5, as in Fig. 8).

4. A local lack of facies and thickness variability corresponds to areas of lesser tectonic activity.

5. The distribution of breccias, facies changes, and thickness variability of the Baranów Beds (Fig. 8) shows that the only active fault zones at that time were those aligned NW–SE. This pattern seems to hold also for other parts of the Carpathian Foredeep (*cf.* "Thickness map of the Sub-anhydrite Beds" — Karnkowski and Ołtuszyk, 1968), where isopachs of these deposits are arranged along NW–SE trends, and transverse faults do not apparently disturb them (also *cf.* Kubica, 1996).

There is a remaining problem: why the tectonic movements which accompanied the Miocene transgression and sedimentation of the Sub-gypsum Beds had such a small spatial extent, constant orientation and small amplitude. It seems that the direct reason for such movements was the influence of Miocene marine transgression upon strongly fractured Cretaceous marls (in the west) or brittle and weathered Old Palaeozoic rocks (in the east). This might have triggered the sinking of second order troughs on the peneplanated surface of the Paleogene land by a hydrotectonic mechanism. An analogous trigger operates

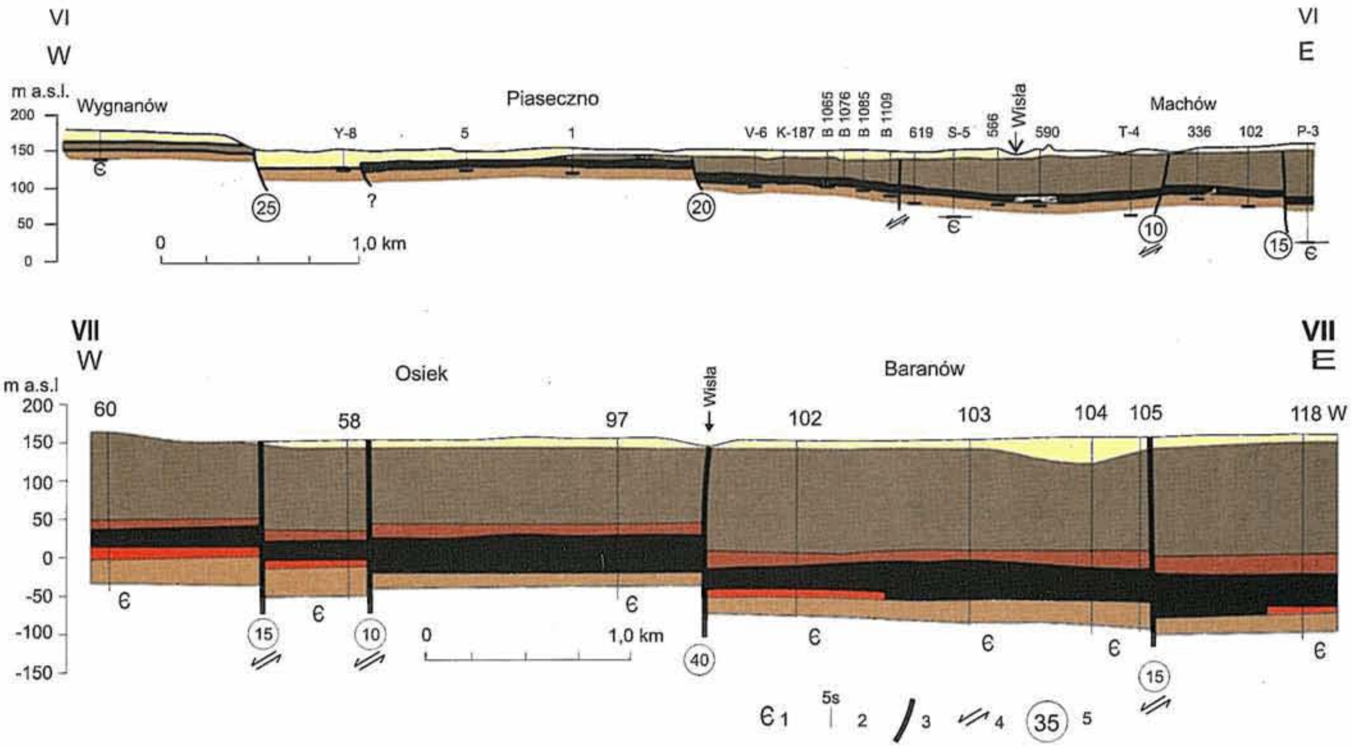


Fig. 7. Geological cross-sections through Kurdwanów–Zawichost Zone: VI–VI — Piaseczno–Machów (Tarnobrzeg sulphur deposits), VII–VII — Osiek–Baranów (Tarnobrzeg sulphur deposits)

1 — Cambrian, 2 — boreholes, 3 — normal faults, 4 — strike-slip faults, 5 — fault throws; other explanations see Fig. 6

during the filling of dam basins with water (Sherard *et al.*, 1974; Gupta and Rastogi, 1979). The main fault zones, stretching NW–SE, were the zones of weakness, and this explains why the Lower-Badenian synsedimentary structures inherited this direction. Talus of Cretaceous marl debris formed accumulations at the feet of horsts (Fig. 8), whereas sedimentation of clays and glauconitic marls continued in the troughs. The highest horsts remained deposit-free (Fig. 8, sections III–III, IV–IV, IVb–IVb, V–V) and white marls were deposited on shallowly submerged horsts (Fig. 8, sections I–I, IVa–IVa, IVb–IVb). In the eastern part of the studied area (Machów–Jeziórko — Fig. 7, section VI–VI and Osiek–Baranów — Fig. 7, section VII–VII) — the Upper Precambrian to Lower Cambrian sandstones and quartzites supplied material to the Baranów sands and sandstones, which are the dominant facies in the central part of the Carpathian Foredeep. Locally, on higher elevations and shoals, a lithothamnium limestone facies appeared (e.g. Jamnica sulphur deposit). The faults in the Baranów Beds thus reproduce the basement structural pattern.

This explanation of the lack of distinct relief in the peneplaned pre-Miocene surface is contrary to most previous interpretations (e.g. Pawłowski, 1965; Ney *et al.*, 1974; Cisek, 1983; Jawor, 1983*a*; Pawłowski *et al.*, 1985) who considered that sedimentation of the Badenian sub-gypsum members flattened the basin floor "palaeorelief". Evidence cited included sedimentary continuity with the Sub-evaporitic Beds and the constant thickness of gypsum deposits (16–20 m) where these had not been affected by later erosion (*cf.* Karnkowski and Otłuszyk, 1968). I consider this argument unconvincing, and discuss it below.

This synsedimentary tectonics correspond with the Early-Styrian phase (Tab. 1).

## SMALL-SCALE SYNSEDIMENTARY DEFORMATION OF GYPSUM DEPOSITS

I consider the time of gypsum sedimentation as, in general, tectonic quiescent. The opposite point of view is presented by Peryt and Kasprzyk (1992) and Peryt (1996) who relate variability of the gypsum deposits to sea-level changes controlled by strong tectonic movements. Distinct tectonic movements seem in fact to have taken place as late as at the end of sulphate evaporation (see Tab. 1). Within the gypsum, the numerous sedimentary-diagenetic deformations (Kwiatkowski, 1972; Bąbel, 1991, 1996, 1999; Kasprzyk, 1995, 1999*a, b*) are accompanied by deformations which may *sensu lato* be assigned to tectonic structures related to tectonic compaction of the deposits (pressure-solution structures — e.g. small folds Pl. II; Figs. 1, 2) and to seismic shocks (e.g. breccia vein in a fault fissure — Pl. II, Fig. 4). These structures occur within the laminated gypsum deposits, and include synsedimentary flexures of different scales and geometries, as well as brittle deformations: breccia veins and breccia complexes of considerable thicknesses.

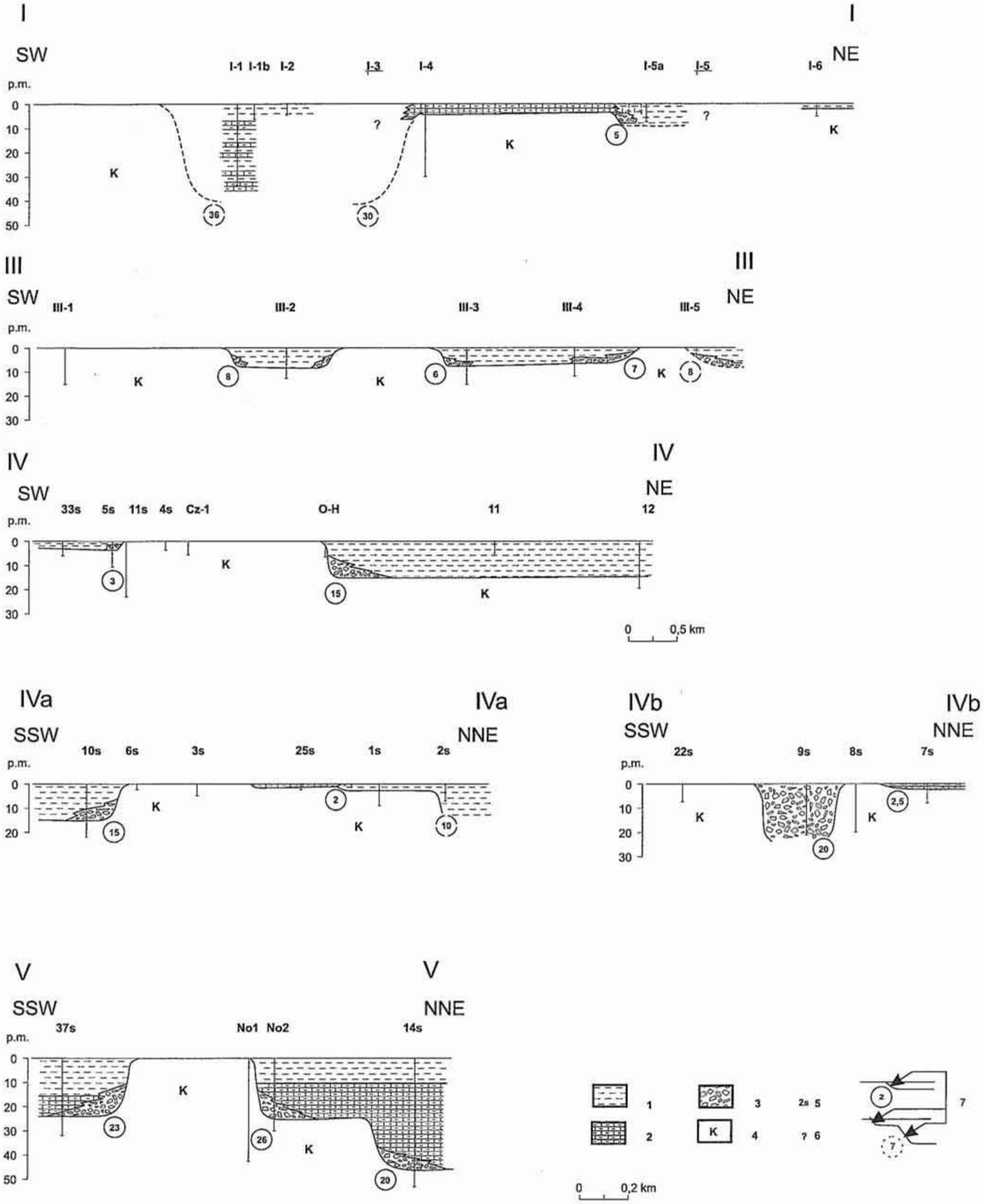


Fig. 8. Tectono-facies cross-sections for time of sub-gypsum beds sedimentation: I–I, III–III — Solec Depression, IV–IV, IVa–IVa, IVb–IVb — Nida Horst, V–V — Słomniki Horst

**Tertiary**: Lower Badenian, sub-gypsum beds: 1 — glauconitic clays, 2 — marls, 3 — breccias; **Cretaceous**: 4 — marls and gaizes; 5 — boreholes, 6 — undrilled sub-gypsum beds, 7 — faulted margins of synsedimentary depressions, throw — magnitude, where known, in metres (broken circle — estimated value)

## PLASTIC DEFORMATIONS

In the Gacki quarry, broad folds in the laminated gypsum beds (10–20 m wide, 1–3 m high) occur immediately above crystalline gypsum deposits with an uneven top surface. The laminated gypsum strata are a little thicker in fold limbs than hinges, suggesting a synsedimentary or early diagenetic age of these structures, which thus form some sort of drape folds. Folds of this type are nearly E–W, though some folds of N–S strike have also been observed (Fig. 9). In the Działoszyce Depression, these structures are 10–15 m wide, and 1–5 m high. A synsedimentary trough (Fig. 10) whose amplitude was about 5 m and was bordered along one side by a flexure in a laminated gypsum bed, was visible in the middle part of the eastern wall of Gacki quarry. The thickness of the gypsum bed varies from 1 m in the hanging wall and 0.2 m in the hinge, up to 2 m in the trough. Within the trough, above the laminated gypsum strata, a breccia fills the trough and is composed of fragments of these deposits. The structure is overlain by disturbed laminated gypsum. The basal parts of the structure were not visible, but coarsely crystalline gypsum ("selenite") of the lower unit was exposed a few metres below. This suggests that the trough visible above is a sinking of layer effect, developed on an uneven top surface. Some flexures of 1–2 m amplitude, occurring within laminated gypsum deposits, have also been found in Gacki quarry. Their strike, like that of the trough described, is E–W. A characteristic feature of all these forms are compensatory variations in thickness of strata (Fig. 11). This suggests a synsedimentary or early diagenetic age, when the sediment was not completely lithified.

I consider small plastic deformations such as: kink folds (Pl. II, Fig. 1), recumbent folds (Pl. II, Fig. 2) disharmonic in relation to the underlying and overlying beds, not as tectonic *sensu stricto*, but as diagenetic-synsedimentary in origin (*cf.* Kwiatkowski, 1972; Bąbel, 1991). Peryt and Kasprzyk (1992) suspect that these structures formed as a result of seismic shocks. However, their small size, limited spatial extent, isolated nature and disharmony suggest otherwise. A comparison of these structures (*op. cit.*) with syn-rift seismites from the Red Sea (*cf.* Plaziat *et al.*, 1990) seems not relevant because the latter represent larger-scale phenomena.

## BRECCIA VEINS AND BRECCIA COMPLEXES OF CONSIDERABLE THICKNESS

Breccia veins were observed in a few exposures in the Solec Depression (Gacki quarry — Pl. II, Fig. 4), Chotel Czerwony, Górki, Leszcze, Wola Zagójska (Fig. 2). In the laminated gypsum deposits these are only 0.1–0.5 m wide. In one case, a breccia vein occurs within a fault zone of a *ca.* 1 m throw. Typically, these veins have a sharp boundary with undeformed gypsum deposits. The veins are infilled with closely packed fragments, up to twenty centimetres in diameter, of laminated gypsum. Neither continuous deformations nor thickness changes within laminae adjoining the veins and in breccia-forming fragments were observed. At places where a complete gypsum profile is visible, fractures filled with breccias do not cut all the laminated gypsum sequence, but only its lower part, reaching as far as half its thickness (4–5 m). At the top they are overlain by an undeformed cover. The breccia veins may represent cataclastic veins of seismic origin, perhaps related to small seismic shocks which caused brecciation of diagenetically altered gypsum laminae along zones of tectonic weakness, before deposition of later sediment. Studies performed by Schreiber *et al.* (1976) confirmed that lithification of gypsum could occur on the basin floor prior to deposition of youger layers.

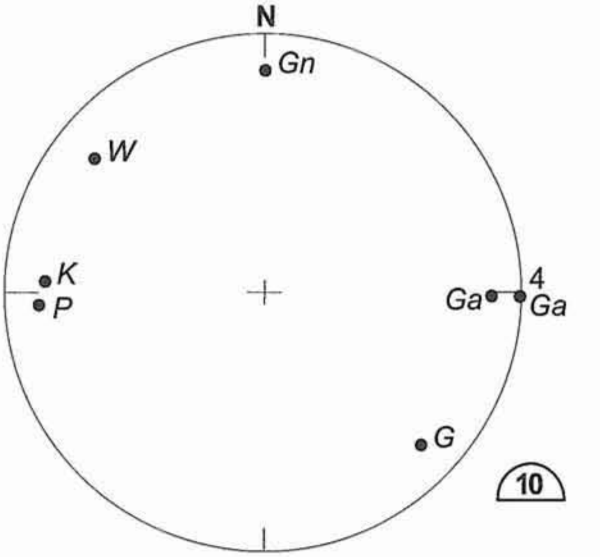


Fig. 9. Diagram of bend's axis orientations in the laminated gypsum

Solec Depression: Ga — Gacki, G — Górki; Działoszyce Depression: Gn — Gniazdowice, K — Koniusza, W — Winiary, P — Posądza; black circle — axis of layer bend

Bąbel (1991, 1996, 1999), however, considered that formation of the breccia veins was related to dissolution of halite crystals.

Breccia complexes up to 5–10 m thick have been found in the Działoszyce Depression (Gniazdowice, Posądza, Koniusza — Fig. 2). Their composition most frequently resembles that of the vein breccias, but they also include fragments showing flow folds, formed by brecciation of neighbouring small fold structures. According to Kwiatkowski (1972), the gypsum fragments with deformed lamination are rollers, and the breccia as a whole represents a submarine slide. Where breccias occur close to medium and small disharmonic folds, internal stress associated with gypsum/anhydrite transformations (Pl. II, Fig. 3) may be a sufficient reason of the gypsum layers's failure. However, such breccias can reach a considerable thickness, e.g. in the Działoszyce Depression up to 20 m (*cf.* Osmólski, 1972). This indicates that these might be tectonic breccias formed as a result of seismic shocks. Breccias in the Działoszyce Depression (*op. cit.*) contain fragments of laminated gypsum together with coarse-crystalline gypsum, most probably derived from the underlying beds. This is not observed in the Solec Depression. This suggests earlier and more intensive tectonic activity in the Działoszyce Depression as compared with the Solec Depression, resulting from the proximity of the former to the Carpathians. The breccias show no roundness indicative of long

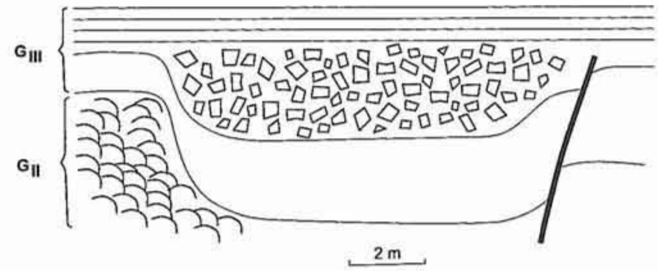


Fig. 10. Synsedimentary trough in laminated gypsum. Gacki quarry — eastern wall

G$_{III}$ — laminated gypsum; G$_{II}$ — sabre gypsum

transport, nor are associated with turbidite sequences. They thus resemble classic submarine slides (*cf.* Peryt and Jasionowski, 1994).

Gypsum and anhydrite deposits form a distinct horizon on seismic profiles. However, detailed exposure mapping has shown that the base only of the gypsum deposits forms a flat surface which may be used for the correlation and measurement of fault throws, particularly as there is a sedimentary continuity at the boundary with the Baranów Beds. The top surface, both at outcrop and below the overlying clays, shows strong relief owing to erosion-karst processes (Nieć, 1970, 1977; Osmólski, 1976, 1977; Pawłowski *et al.*,1985; Kubica, 1992, 1996; Połtowicz, 1998). Between the gypsum and the clay deposits, there is a stratigraphic gap which corresponds to at least the Chodenice Beds (Osmólski, 1972). Połtowicz (1998) considers that the erosion was submarine.

The gypsum deposits themselves (or correlative limestones with native sulphur) have been strongly tilted (Figs. 5, 12) by later tectonic movements of the Late-Styrian (Moldavian) phase — Tab. 1. The primary variable thickness of the gypsum deposits (Bąbel, 1996), and the erosional upper boundary with the Upper Badenian clays hinders the determination of the magnitude of tectonic movement during this phase. Displacements of the Chemical Beds measured in cross-sections (Figs. 5, 7) represents total throw, including effects of earlier phases starting with the Moldavian phase until the present.

## SYNSEDIMENTARY ACTIVITY AT THE BASE OF THE *PECTEN* BEDS

Tectonic movements of the Moldavian phase, which commenced immediately after gypsum deposition, continued during sedimentation of the *Pecten* Beds[1].

This is shown by the following:

1. The thickness variation of the *Pecten* Beds reflects structural position as observed in the Machów open pit (Fig. 12) as well as in the Solec Depression (Fig. 5, sections I–I – III–III), on the Nida Horst (Fig. 5, section IV–IV) and the Słomniki Horst (Fig. 5, section V–V).

2. Breccias and conglomerates which occur in the lower part of the *Pecten* Beds (*Neobulimina longa* Zone) are composed of pebbles of Cretaceous marls and Badenian limestones (lithothamnium and Litava limestones) in the Nida Trough. In the area of the Tarnobrzeg sulphur deposits they consist of fragments of hard grey limestones and cemented masses of *Pecten* shells (Osmólski and Pilichowska, 1978), and fragments of bored limestone (Osmólski, pers. comm.); occasionally rollers, up to 0.5 m in diameter, occur within clays. In the Machów mine, these deposits fill pockets, up to 3 m deep, occurring immediately above the chemical limestones (Fig. 12). The *Pecten* Beds have locally dips of 30–40°, more usually 10–15°.

An angular unconformity with the overlying, almost horizontal Krakowiec Beds, defines the upper limit of the Moldavian phase tectonic movements. These Late Badenian tectonic movements were in fact the first significant tectonic phase in the Miocene of the Carpathian Foredeep. For, longitudinal faults (NW–SE), which in their initial form existed within the Baranów Beds (Fig. 8), were reactivated and their throw slips increased[2].

Numerous new faults, frequently of considerable throws, also came into existence at that time (Figs. 5–7). The structural consequence of those movements were rotations of beds, particularly in footwalls (frequently of antithetic character — Fig. 6; *cf.* Krysiak, 1985, 1986*b*, 1987*a*; Dziadzio and Jachowicz, 1996). Tectonic traps, which became later a location of sulphur deposits, were created this way (Krysiak, 1985, 1987*a*).

Longitudinal faults were active while the *Pecten* Beds were deposited. However, mesostructural observations show that some transverse faults may already have been initiated. Transverse faults were initially dip-slip faults (Pl. I, Figs. 1–3; Fig. 12) and were later reactivated as strike-slip faults. Macro- and mesostructural observations (Fig. 12) show that the longitudinal faults were produced by E–W and ENE–WSW regional extension. This extension probably resulted from regional uplift as there is no evidence for localized horst uplift, and no sign of any horizontal compression, that could have produced the extension. In addition, compaction of Upper Badenian argillaceous deposits which were underpinned by a rigid complex of chemical deposits (limestones, gypsum), might have played an important role in the reactivation of the faults.

## POST-KRAKOWIEC BEDS TECTONICS

The deposition of the Krakowiec Beds has been linked with synsedimentary deepening of the Carpathian Foredeep and the development of minor depressions and troughs (Obuchowicz,

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[1] The term *Pecten* Beds has been used for more than 30-years (Łuczkowska, 1964; Kowalewski, 1966; Osmólski, 1972). However, it has recently been suggested (Studencka, 1999; pers. comm.) that this unit should be termed the Scallop Beds (see Budowa Geologiczna Polski, 1996).

[2] The fault throws measured for the base of chemical deposits are from 2 to 30 times greater than synsedimentary throws of these faults in the Baranów Beds. However, this value involves the recent, but not the Late Badenian, hypsometric position of the beds and reflects the total throw of the fault zones. It is difficult to estimate separately the throws for the Late Badenian and neotectonic movements.

1966; Ney *et al.*, 1974; Rutkowski, 1981). The greatest thicknesses of the Krakowiec Clays are indeed observed in tectonic depressions, where they are better preserved, but this is not a proof of synsedimentary tectonics, because the top of the Krakowiec Clays is erosional. Analysis of borehole sections (Figs. 5, 7) rather shows that the Krakowiec Clays effaced the relief created by the tectonics of the previous phase. The local overlap of the Krakowiec Clays on to Cretaceous deposits (Fig. 2 — borehole 44s; Fig. 5, section IV–IV — borehole 11s), indicates a local increase in basin extent.

The Krakowiec Clays thicken tenfold southwards across the Kurdwanów–Zawichost Zone (the Vistula river line — Figs. 1, 7). This results from enhanced subsidence, whereas the occurrence of abundant sandy intercalations in the upper part of the Krakowiec Clays in the Carpathian Foredeep is interpreted by Karnkowski (1978, 1989) as the delta slope of a river flowing from the Meta-Carpathian Range. I have suggested (Krysiak, 1987*b*) that synsedimentary activity in this part of the basin and in the adjacent Carpathians resulted in clastic supply not from the north but from the south. Karnkowski (1994) has recently invoked both directions of sediment transport.

Following the deposition of the Krakowiec Clays, subsequent tectonic activity (the Attican phase) of postsedimentary character was marked in the area studied. It is difficult to define precisely the upper limit of these movements because the Neogene lithostratigraphic profile is incomplete here (the upper Sarmatian, Pannonian, Pontian and Pliocene are lacking — Tab. 1). Undoubtedly, they took place before the Quaternary, i.e. prior to deposition of the Older Quaternary Witów Gravels (Tab. 1). Rutkowski (1981) and Czapowski and Studencka (1996) suggest tectonic control on the boundary between the Krakowiec Clays and the detrital Sarmatian deposits (data from the Połaniec Trough). The onset of Attican phase movements in the northern marginal part of the Carpathian Foredeep could thus be defined.

A change of the regional stress field took place during the Attican phase in the Carpathian Foredeep. The distribution of complementary shears in this area (Fig. 2 — Gacki) indicates that a system of transverse and oblique faults — of strike-slip and oblique-slip motion with transverse faults dominant (Fig. 2 — E–W, ENE–WSW) — came into existence due to horizontal compression oriented NE–SW (azimuth 30–50°). A shift in mapped geological boundaries and axes of major structures indicates sinistral displacement. Strike-slip displacements were compensated by folding of clays overlying the Chemical Deposits — Fig. 12 (*cf.* Krysiak, 1985, 1986*a*, *b* and detailed mesostructural analysis in Jarosiński, 1992). This sinistral strike-slip wrenching played an important role in the tectogenesis of the Carpathian Foredeep, since it caused a transverse dismembering and mutual displacement of structures striking NW–SE, by up to 0.5–1 km (*cf.* Osmólski, 1972; Osmólski *et al.*, 1978). The Kurdwanów–Zawichost Zone (Figs. 1, 7; *cf.* Jarosiński, 1992) is a fault of this type. Sinistral displacement along this zone seems responsible for the relative displacement of the Piaseczno and Machów sulphur deposits (Figs. 1, 3) as well as the Osiek and Baranów fields (Fig. 4).

Strike-slip and oblique-slip tectonic movements, postsedimentary in relation to the Krakowiec Beds, all have features of this tectonic phase and so should be ascribed to the Attican phase of the Alpine orogeny (Tab. 1).

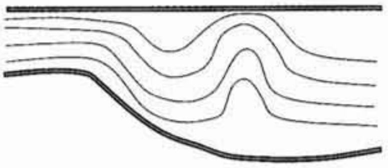

Fig. 11. Compensatory fold in laminated gypsum. Gacki quarry — northern wall

## NEOTECTONICS AND RECENT TECTONIC MOVEMENTS

The Neogene tectonic movements, which ended with the "strike-slip" Attican phase, seem to terminate the Late Alpine cycle, corresponding, in the Carpathians, to the final thrust movements over the foreland.

Starting with the Valahian phase, the next change in the regional stress field took place — from compressional (strike-slip faults) to extensional — caused by uplift of the Carpathians and neotectonics began. Intensive erosion of the uplifting Carpathians resulted in deposition of the Preglacial Carpathian gravels in the foreland. In the area studied these are the Witów Gravels forming thick deposits around the Raba river mouth (Fig. 2; Krysiak, 1987*b*). A series of normal, listric faults of small throws (up to 1 m) occur within the Witów Gravels (Pl. I, Fig. 4). The Witów Gravels are not overlain by younger Quaternary deposits, and their contact with the Miocene deposits is not visible in the outcrops. The age and extent of these faults are therefore difficult to establish. The listric shape of the faults seems to indicate their small extent, perhaps limited only to the gravels. The faulting mechanism may have been mass-movements caused by seismic shocks. Their predominant strike azimuth is 160° (Krysiak, 1987*b*). Such an orientation is typical over the Słomniki Horst (Fig. 1), where measurement was made over its SE slope. Correspondence of fault orientations in the Witów Gravels with major fault zones in the Mesozoic-Miocene basement suggest that tectonic deformation of the gravels formed as a result of the reactivation of basement faults, manifested as seismic shock effects.

It is difficult to decide whether the faults in the Witów Gravels are synsedimentary or formed during inter- and/or postglacial periods. In the latter case, a relaxation mechanism induced by the retreat of an ice-cap cannot be precluded. I consider that the transport direction of these deposits from the west to east (Gradziński and Unrug, 1959; Dżułyński *et al.*, 1974), i.e. outside of the Słomniki Horst, suggests that tectonic movements were initiated as early as the sedimentation of the Witów Gravels. On the basis of height differences in the base of the Witów Gravels, Połtowicz (1967) estimated the neotectonic uplift of the Puszcza Horst (old expression for Słomniki Horst) at about 50 m. According to that author the uplift, associated with tilting of the horst, resulted in a migration of the Vistula river channel from the SE towards the NW between Niepołomice and

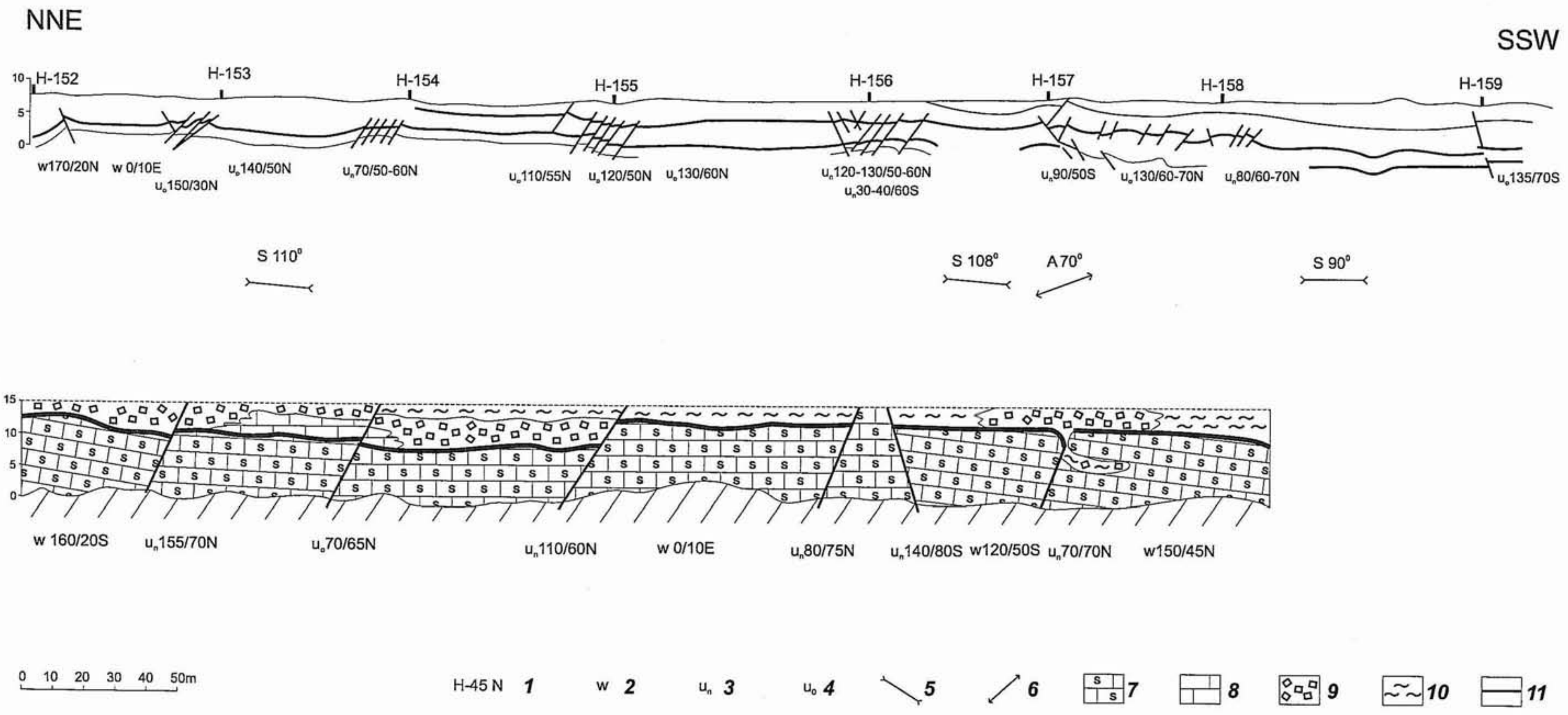

Fig. 12. Tectonic sketch of the Machów-mine eastern wall — sulphur deposit and overburden clay

1 — de-watering wells, 2 — orientation of strata, 3 — normal faults, 4 — reverse faults, 5 — azimuth of synclinal axis, 6 — azimuth of anticlinal axis, 7 — limestones with native sulphur, 8 — barren limestones, 9 – breccias of *Pecten* limestones, 10 — *Pecten* clays, 11 — cover of sulphur deposit; tectonically deformed green clays with *Syndosmya reflexa* and the base of the Krakowiec Clays

Nowy Korczyn by 6–9 km, as well as in displacement of the Raba river bed towards the west by about 6 km. The amplitude of neotectonic movements, estimated from height differences in the base of the Preglacial gravels, are of 40–50 m in the Działoszyce Depression (Połtowicz, 1967; Jonak, 1979; Lindner and Siennicka, 1994), and 40–80 m in the Sandomierz Depression (Laskowska-Wysoczańska, 1981, 1993, 1995). According to Walczowski (1983), a Lower Pliocene base-levelling surface can be distinguished in the central part of the Działoszyce Depression. This surface is lowered southwards by a system of W–E trending fault steps with throws amounting to about 50 m.

## CONCLUSIONS

1. The fault dynamics are contolled by a gravity-driven stress field of varying intensity through time.

2. In general, the fault system affecting Miocene deposits in the area studied comprises two mutually perpendicular fault sets (longitudinal — roughly N–S, and transverse — roughly E–W). Fault azimuths vary from NNW–SSE and ENE–WSW (in the northern part) through NW–SE and NE–SW (in the central part) to WNW–ESE and NNE–SSW (in the southern part). Azimuth changes of the faults depend on their orientation in relation to basement structures and on their distance from the Carpathians. Longitudinal faults are a dip-slip (normal) faults, whereas transverse faults are oblique-slip (oblique-normal-slip) faults and strike-slip faults.

3. The Kurdwanów–Zawichost Zone is of a strike-slip and oblique-slip character. The Miocene activity of this fault zone started in the Early Sarmatian, as a dip-slip fault. The increasing downthrow of gypsum deposits associated with that phase, together with later phases (about 500 m total) might have caused gypsum alteration into anhydrites in Central Carpathian Foredeep. The Kurdwanów–Zawichost Zone was reactivated during the Attican phase (after Early Sarmatian) as a strike-slip fault zone.

4. Normal faults were in large part rejuvenated, and their dips are steep or decreasing with depth. Strike-slip and oblique-normal-slip faults are mostly steep or vertical, although there are also faults of varying dips. At least some of this faults are primary.

5. Most faults transecting the Miocene deposits show post-depositional activity.

6. Fault throws in the area studied do not usually exceed several tens of metres (sporadically reaching about 100 m), as a rule they attain a few to twenty metres.

7. Horizontal displacements along the major strike-slip faults reach a maximum of a few kilometres, commonly attaining several tens of metres or several hundred metres.

8. Deformation observed in gypsum deposits does not show a distinct relationship with regional tectonics. Meso-folds, disharmonic in relation to the upper boundary, are drape or compensation structures. Tectonic breccia veins exhibit a seismotectonic character.

9. Three tectonic phases which affected Neogene deposits in the northern part of the Carpathian Foredeep have been distinguished:

— Styrian phase (Early-Styrian) — coincident with the Paratethys transgression, and controlling facies and thickness variability of the Lower Badenian deposits (the Baranów Beds), by synsedimentary movements, with NW–SE longitudinal faults active;

— Moldavian phase (Late-Styrian) — after deposition of the gypsum and during sedimentation of the lower part of the *Pecten* Beds in the Upper Badenian. Longitudinal faults were reactivated by E–W to ENE–WSW oriented extension, with antithetic rotation of beds in the footwalls of faults, in particular on the eastern side of horsts (Fig. 6).

— Attican phase — after sedimentation of the Krakowiec Beds in the Sarmatian. Strike-slip and oblique-slip faults (NNE–SSW, NE–SW to WSW–ENE) formed in a compressional stress field oriented at an azimuth of 30–50°, with N–S tension. A structural consequence of these movements are transverse displacements of longitudinal structural axes, of up to 1 km.

10. Quaternary neotectonic movements are represented by following phases: the Valahian phase — during deposition of the Preglacial Witów Gravels, and the postglacial Passadenian phase (Tab. 1), during which listric faults transecting the Witów Gravels originated. Rejuvenation of the entire tectonic network of the Carpathian Foredeep took place at that time in an extensional stress field, and the displacements were probably of a seismotectonic character.

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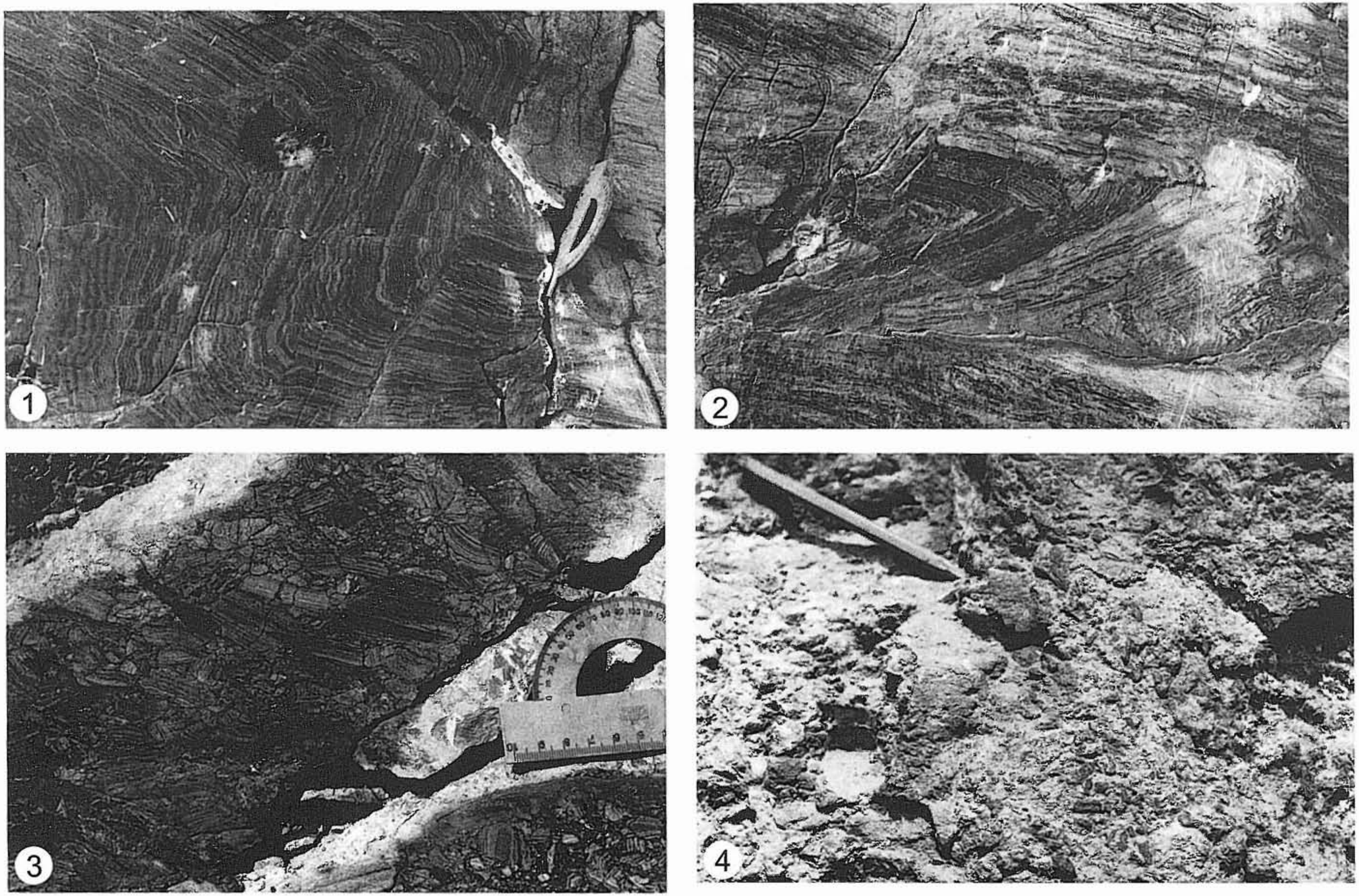
PLATE I

*1.* Normal fault 40/80N, throw 15–20 m; sub-gypsum marls and giant gypsum intergrowths (selenites) visible in hanging wall; Gacki quarry, Nida Trough. *2.* Reverse fault 64/30N; hanging wall — limestone with native sulphur after metasomatically altered giant gypsum intergrowths, foot wall — metasomatized limestone with native sulphur instead of laminated gypsum; Machów-mine (northern wall) near Tarnobrzeg. *3.* Normal fault 30/55N with en echelon pattern of plumage fractures; Gniazdowice. *4.* Normal faults in Witów Gravels — dominant orientation 160/70E; Witów near Brzesko Nowe — Działoszyce Depression

PLATE II



*1.* Kink folds; laminated gypsum; exposure at Gniazdowice near Proszowice. *2.* Small recumbent broken fold (20 cm height); laminated gypsum; exposure at Gniazdowice near Proszowice. *3.* Breccia composed of closely-packed laminated gypsum fragments; the last stage of laminated gypsum diagenesis; exposure at Gniazdowice near Proszowice. *4.* Tectonic breccia of laminated gypsum in a fault fissure; Gacki quarry (pencil as scale)