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The onset of orogenic activity recorded in the Krosno shales from the Grybów unit (Polish Outer Carpathians)

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ABSTRACT:

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The Krosno shales were deposited synorogenically in front of an accretionary prism as interchannel flysch facies on a north-dipping slope that constituted the southern marginal part of the Silesian basin. The turbidite flows originated from channelised currents and probably also as separate sedimentation events from slow, dilute "sheet" flows derived from a linear source (probably a shelf-edge). In the background continuous hemipelagic deposition took place and there were two periods when pelagic sedimentation prevailed. Mean accumulation rate of the entire succession was moderate: from 8 to 11 cm/ky. The source rocks were Jurassic to Eocene sedimentary rocks. Geochemical data (La, Th and Sc contents) point to a continental island arc as the original source of the detrital material. Soft-sediment deformations and methane-related authigenic carbonates that are also present in this succession evidence the migration of methane-charged fluids through the sediment column. This fluid expulsion was probably provoked by orogenic activity, i.e. the formation of an accretionary prism.

Key words: Outer Carpathians, Krosno Formation, Oligocene, Shale sedimentology, Flysch facies, Orogenesis.

INTRODUCTION

Although shales constitute more than 60% of the world's sediments, much more is known about the sedimentology of coarse-grained sediments. The reason for this bias is that "shales are very fine-grained and lack well-known sedimentary structures that are so useful in sandstones" (POTTER & *al.* 1980). Therefore, petrographic examination of shale demands application of more sophisticated methods than just a standard petrographic microscope. The Krosno Formation from the Outer Carpathians is a syntectonic flysch facies deposited in front of an accretionary prism (PICHA & STRANIK 1999, POPRAWA & *al.* 2002, OSZCZYPKO 2004). It consists mainly of shales and sandstones. Large submarine slumps and exotics are numerous in this formation (e.g. KSIĄŻKIEWICZ 1949, DŻUŁYŃSKI & RADOMSKI 1956, KSIĄŻKIEWICZ 1958, ŚLĄCZKA 1961, ŚLĄCZKA & WIESER 1962, MOCHNACKA &

TOKARSKI 1972, BURTAN & *al.* 1984). The sandstones have been described in detail and their sedimentology is quite well recognized, especially from the Central Carpathian Depression, where they accumulate hydrocarbons (KLECKER & *al.* 2001). However, not so much is known about the sedimentology of the Krosno shales. This paper presents the results of detailed petrographic research of the Krosno shales from the Świątkowa Wielka tectonic window obtained by means of routine analyses combined with microstructural and CL analyses. Petrographic examination is supplemented with geochemical data (La, Th and Sc contents). Interpretation of the results in the light of palaeogeographical and geotectonic data allowed reconstruction of the sedimentary and tectonic processes which controlled the development of the shales. A general model of flysch sedimentation, identification of the source rocks and mean rate of accumulation are also shown in this work.

GEOLOGICAL SETTINGS

The rocks examined belong to the Krosno Formation of the Grybów unit, which crops out from beneath the Magura unit in the Świątkowa Wielka tectonic window (KOZIKOWSKI 1956, KARNKOWSKI 1963, KSIĄŻKIEWICZ & LISZKOWA 1972, KOSZARSKI 1985). This small tectonic structure is located in the Beskid Mały mountains, about 5 km west of the village of Krempna (Text-fig. 1). The rocks of the Grybów unit are strongly folded and sliced (MASTELLA & RUBINKIEWICZ 1998). The Krosno Formation occurs within the southernmost slice of the exposed part of the Grybów unit. It crops out along the Krokowy stream and its tributaries. Beds of the Krosno Formation form narrow folds with N to NE vergence, usually overturned.

The Krosno Formation succession examined is developed as dark (grey to black) fine-grained calcareous shales. Their competent rheology allowed

Fig. 1. General map of tectonic units of the eastern part of the Polish Outer Carpathians. The nappes are thrust in a NNE direction. The Grybów nappe (stripes) is covered by the Magura nappe and emerges only within tectonic windows. The Krosno shales examined occur in the Grybów unit, which crops out in the Świątkowa tectonic window (arrow) (after BOJANOWSKI 2007 – with kind permission of Springer Science and Business Media)

strong deformations which, together with the rather homogenous lithology and weak parting, make it hard to trace the bedding in places. What is more, the outcrops are discontinuous, often separated by faults, and macroscopically detectable geopetal structures are rare. Therefore it is difficult to determine the stratigraphic relationships between the outcrops. The only macroscopically observable geopetal structure is cross-bedding, which seldom occurs in the Krosno shales. Normal grading is noticeable only under the microscope. During many years of systematic fieldwork, orientated samples of the shales were collected and examined with the use of a petrographic microscope. This allowed construction of the lithological profile (Text-fig. 2). However, the thickness of the succession examined was hard to determine precisely: it is between 80 and 110 m.

The lower part of the Krosno shales profile hosts dolomites as concretions or nodular beds. The middle part contains numerous "pure" calcite concretions; two beds of laminated limestone were also found: a 3-cm thick layer of marly laminated limestone exhibiting platy parting and a 20-cm thick bed of laminated limestone without parting. The thicker limestone bed grades horizontally into a carbonate build-up composed of seep-related carbonate intraformational breccia (BOJANOWSKI 2007). A kind of a talus is developed on the sides of the carbonate build-up. The Krosno shales from the upper part of the profile contain laminated concretions. The calcite concretions and the carbonate build-up are authigenic rocks which formed as byproducts of hydrocarbon seepage during the sedimentation of the host shales (BOJANOWSKI 2007).

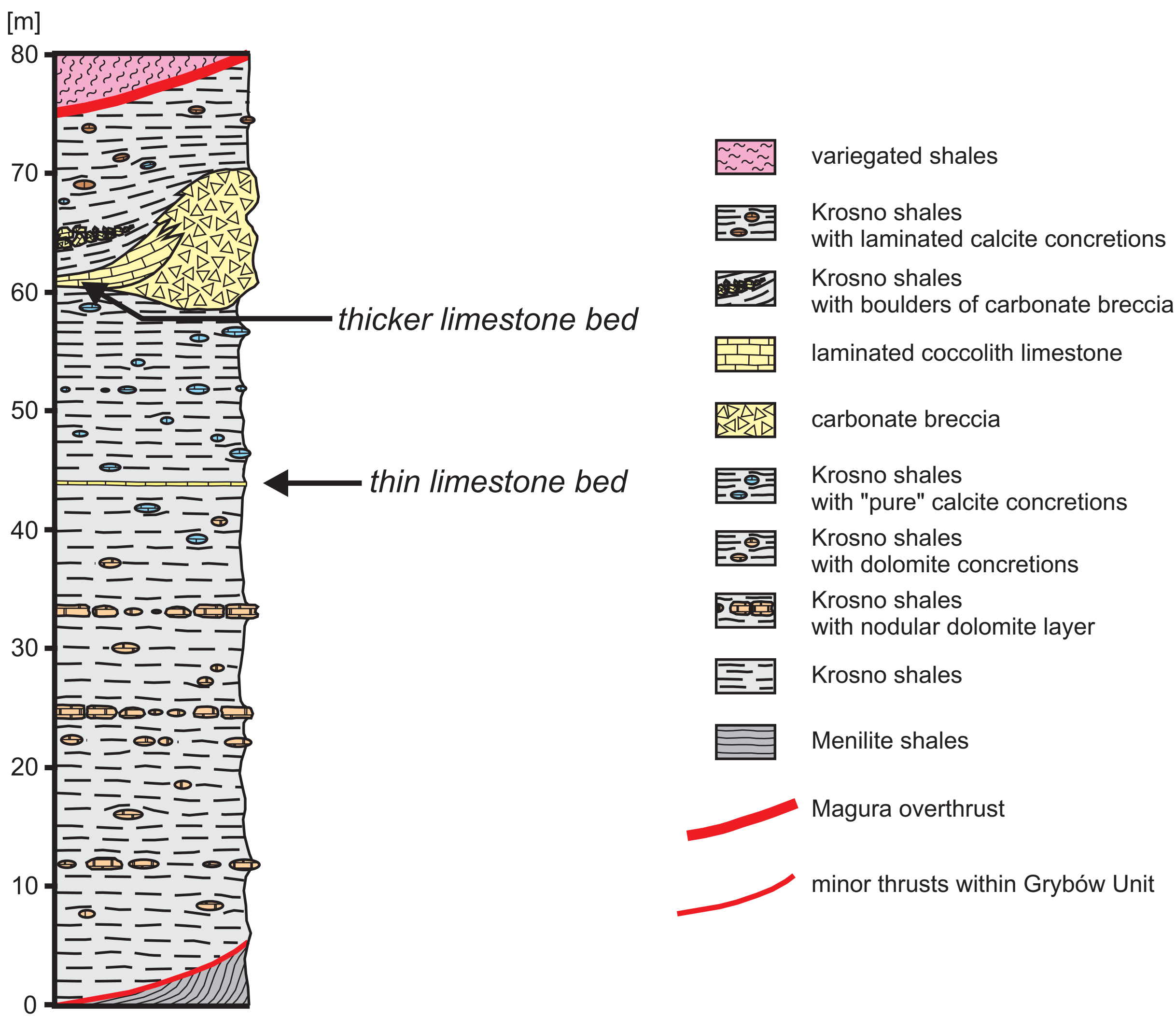

Fig. 2. Lithological profile of the Krosno shale succession examined

REGIONAL AND PALAEOGEOGRAPHICAL BACKGROUND

The Fore-Magura Group comprises a few tectonic units that are thrust on each other in between the Magura and the Silesian nappes (OSZCZYPKO 2004). The Dukla is the largest nappe of the Fore-Magura Group in the eastern Polish Carpathians. In the western Polish Carpathians the Dukla nappe is completely covered by the Magura nappe and is divided into two smaller nappes (the Grybów and the Obidowa-Słopnice units) which are exposed only within tectonic windows (CIESZKOWSKI & *al.* 1985). The Grybów unit is the upper nappe of the western extension of the Dukla unit (ŚLĄCZKA 1971, CIESZKOWSKI & *al.* 1985, ROCA & *al.* 1995, ŻYTKO & MALATA 2001). The entire succession of the Fore-Magura Group exhibits transitional lithofacies that link the Magura and the Silesian basins (OSZCZYPKO-CLOWES & OSZCZYPKO 2004). The Late-Eocene–Oligocene rocks of the succession were deposited in the southern marginal part of the Silesian basin (the Dukla subbasin) (UNRUG 1979, CIESZKOWSKI & *al.* 1985).

The orogenic movements and the accretionary prism formation began in the southern part of the Magura basin in the Late Eocene (PESCATORE & ŚLĄCZKA 1984, POPRAWA & MALATA 1997, PICHA & STRANIK 1999, POPRAWA & *al.* 2002, OSZCZYPKO 2004). Then, syntectonic sedimentation of the Menilite-Krosno series followed after a short period of pelagic deposition of the Globigerina marls (POPRAWA & MALATA 1997, PICHA & STRANIK 1999, POPRAWA & *al.* 2002). The accretionary prism and the associated depocentre prograded to the outer zones of the orogen so that the syntectonic Krosno facies migrated in the same direction (JUCHA & KOTLARCZYK 1961, PESCATORE & ŚLĄCZKA 1984, PICHA & STRANIK 1999, POPRAWA & *al.* 2002, OSZCZYPKO 2004). Diachronism of the Krosno facies is emphasized by the isochronous horizons of Oligocene coccolith limestones (HACZEWSKI 1989, LESZCZYŃSKI & MALIK 1996). The general configuration of the siliciclastic material transport directions in the profile of the Krosno Formation becomes rearranged upwards (DŻUŁYŃSKI & ŚLĄCZKA 1958) due to a significant reduction in the supply of source material from the palaeo-highs ("cordilleras") and the approach of the orogenic front (POPRAWA & *al.* 2002).

The Dukla subbasin constituted the southern marginal part of the Silesian basin during sedimentation of the Krosno Formation (JUCHA & KOTLARCZYK 1961, UNRUG 1968, CIESZKOWSKI & *al.* 1985, POPRAWA & *al.* 2002). High sedimentation rates of the Krosno Formation (20-60 cm/ky) (PESCATORE & ŚLĄCZKA 1984, POPRAWA & *al.* 2002) indicate that erosion of the uplifted source area in the inner zones of the orogen was quite intensive. The Magura nappe was thrust over the Dukla subbasin in the Late Oligocene and reached the outermost parts of the Carpathian basins in the Early Miocene, which caused the cessation of sedimentation of the Krosno Formation (PESCATORE & ŚLĄCZKA 1984, POPRAWA & *al.* 2002, OSZCZYPKO 2004).

The sea level in the Carpathian basins dropped (in places even down to 200-300 m) during the deposition of the Globigerina Marls and the lowermost Menilite beds (latest Eocene–earliest Oligocene), but it rose at the end of the Early Oligocene (KSIĄŻKIEWICZ 1975, OLSZEWSKA 1984, POPRAWA & *al.* 2002). The Carpathian basins became gradually and diachronously shallower during the Late Oligocene and the Early Miocene (POPRAWA & *al.* 2002). In general, the Krosno Formation was deposited in a basin a few hundred metres deep at that time (KSIĄŻKIEWICZ 1975, OLSZEWSKA 1984, POPRAWA & *al.* 2002). The siliciclastic material of the Krosno Formation was mostly derived from older flysch rocks (Cretaceous and Lower Paleogene) of the emerging orogen (DŻUŁYŃSKI & ŚLĄCZKA 1958, PICHA & STRANIK 1999, POPRAWA & *al.* 2002).

METHODS

Standard petrographic microscopy was performed on thin sections. Some thin sections were stained with alizarin red and Feigl's solution according to the procedure described by FRIEDMAN (1959). Because the rocks examined are very fine-grained, detailed observations and analysis of mineral and elemental composition were performed using electron microscopy. Uncovered thin sections, polished and powdered with carbon, were examined using a WDS Cameca SX-100 microprobe (Faculty of Geology, Warsaw University, Poland). Observations were carried out using backscattered electron imaging. Operating conditions:

15 kV accelerating potential, 10-20 nA beam current. Rock chips coated with gold were examined using a JEOL JSM-6380LA scanning electron microscope equipped with EDS analyzer (Scanning Electron Microscope and Microanalysis Laboratory at Faculty of Geology, Warsaw University, Poland). Cathodoluminoscopy was conducted on uncovered, polished thin sections using the Premier American Technologies cold cathode – Luminoscop ELM-3R (Geological Bureau GEONAFTA, Warsaw, Poland). Operating conditions: 5-20 kV beam energy, 1 mA beam current, 20-80 mT vacuum.

Seventeen powdered samples (0.5 gram) were treated with nitrohydrochloric acid (6ml HCl + 2ml HNO_3) in a closed microwave system (Multiwave Perkin Elmer/PAAR Physica) and with hydrogen peroxide solution in order to dissolve organic matter. Deionized water was added to the resultant solution up to 50 ml. The solution was stored in hermetic polyethylene containers at –2 to –4°C. This procedure was conducted at Warsaw University, Faculty of Geology, Laboratory for Water, Soil and Rock Chemistry. The solution was analyzed, after refrigeration, in an inductively coupled plasma mass spectrometer (ICP-MS) Elan 6100 DRC Perkin Elmer at Warsaw University, Faculty of Chemistry. Contents of Th, La and Sc were analyzed with a detection limit of 0.01 ppb. Mean relative error of the measurements was 6% for Th, 2% for La, and 4% for Sc.

PETROGRAPHY

The Krosno shales are dark (grey to black) fine-grained calcareous rocks. They comprise two kinds of beds: thicker grey turbiditic beds (from a few up to 40 mm, usually about 10 mm) and thinner brown to black hemipelagic beds (up to 2 mm, usually about 0.5 mm) (Text-fig. 3A). The turbiditic beds are grain-supported, consist mainly of silt, and exhibit normal grading: they are composed of up to four successive laminae characterized by a decreasing silt/clay ratio from base to top (Text-fig. 3B): (A) well sorted coarse silt with singular very fine sand grains; (B) silt with a wide range of grain sizes; (C) medium and fine silt with singular coarse silt grains; (D) fine and very fine silt with clay. The lowermost lamina A commonly has a sharp erosional base (Text-fig. 3B, C). Lamina B sometimes exhibits cross-lamination indicating an overall northern transport direction (Text-fig. 3B). Lamina C is rather structureless (Text-fig. 3B). Lamina D may grade into a hemipelagic layer (Text-fig. 3C). Detrital material in the middle laminae (B and C) is poorly sorted (Text-fig. 3B, 3D). The normal grading is therefore continuous, but poorly segregated. The grains are usually subangular (Text-figs 3C-D, 4).

The main constituents of the Krosno shales are: platy minerals (mixed-layer clay minerals, micas, chlorites), quartz and calcite (Text-fig. 4). Dolomite, albite, pyrite and organic matter are less abundant (Text-fig. 4). Rutile (Text-fig. 4C) and glauconite (Text-fig. 3D) are found sporadically. Pyrite is authigenic, while quartz, albite, chlorites, micas, rutile and glauconite are detrital constituents. Clay minerals and dolomite may be either authigenic or detrital in origin. Calcite occurs as detrital (yellow in CL), authigenic (light yellow in CL) or biogenic (orange in CL) material (Text-figs 4, 5).

The granulometric variability of the turbiditic beds affects their mineralogical composition: quartz is the main constituent in the lowermost laminae (A and B) (Text-fig. 4B-C), but clay minerals prevail in the upper laminae C and D (Text-figs 3C, 4A). Calcite present in the turbidite beds is mostly detrital (Text-figs 4A, 5B), but biogenic calcite appears in the upper laminae as isolated planktonic foraminifers (Text-fig. 5) or coccoliths (Text-fig. 4A). Intergranular spaces are filled with matrix composed of fine clay minerals – hydromicas according to EDS chemical analyses (Text-fig. 4B-C). Platy minerals present in the matrix are orientated parallel to the outlines of the grains and often oblique to the lamination (Text-fig. 4C).

Some turbidite beds are topped by thin (up to 2 mm) hemipelagic laminae (Text-fig. 3A-C) that contain mainly very fine silt and clay rich in clay minerals, organic matter and biogenic calcite (coccoliths and foraminifers) (Text-fig. 6). Platy minerals are orientated roughly parallel to the lamination in the hemipelagic layers (Text-fig. 6A). The distinction between hemipelagic and turbiditic material can be facilitated by considering the texture of the matrix (O'Brien & *al.* 1980), which is parallel in the hemipelagites, while in the turbidites the matrix is squeezed in between the grains, with the platy minerals orientated parallel

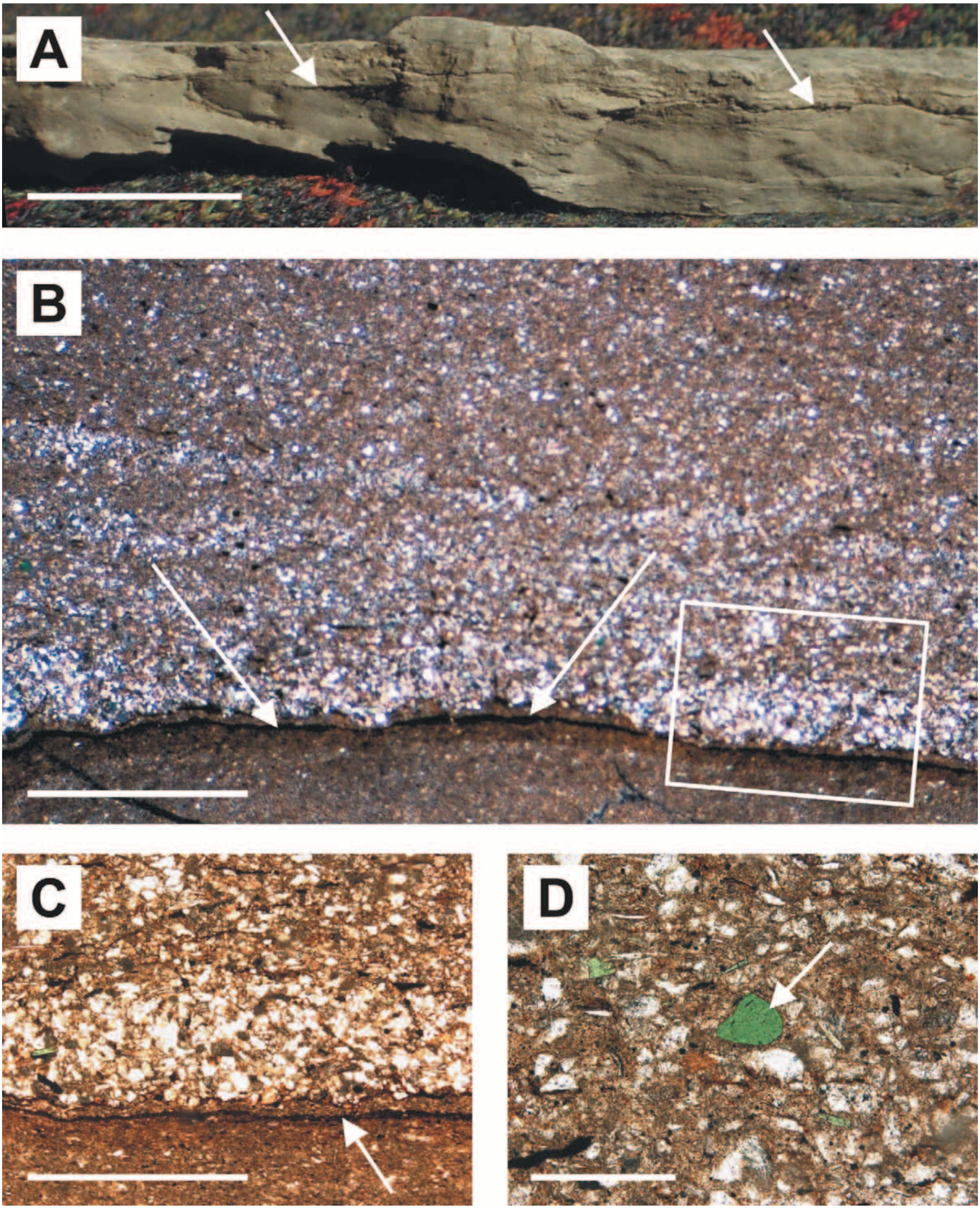

Fig. 3. Images of the Krosno shales. **A** – Three turbiditic beds with cross lamination visible within the middle one. A very thin hemipelagic layer appears between the middle and the upper turbiditic beds as a dark film (arrows). Scale bar – 20 mm. **B** – Thin section (plane polarized light) of two turbiditic beds and a hemipelagic layer in between. The lower part of the image is occupied by the uppermost part of a turbiditic bed (lamina D) topped by a hemipelagic layer (arrows). The overlying material is the lowermost part of another turbiditic bed: lamina A, which covers the hemipelagic layer (note the erosional contact), contains the coarsest material and is rather structurless; lamina B exhibits cross lamination and contains poorly sorted material; lamina C is visible in the uppermost part of the image. Detail of the thin section (in the rectangle) is shown in C. Scale bar – 2 mm. **C** – Thin section (plane polarized light). Detail of B. From base to top of the image: lamina D, hemipelagic layer (arrow), lamina A, lowermost part of lamina B. Scale bar – 1 mm. **D** – Thin section (plane polarized light) of lamina B with large glauconite grain (arrow) in the centre. Scale bar – 0.2 mm

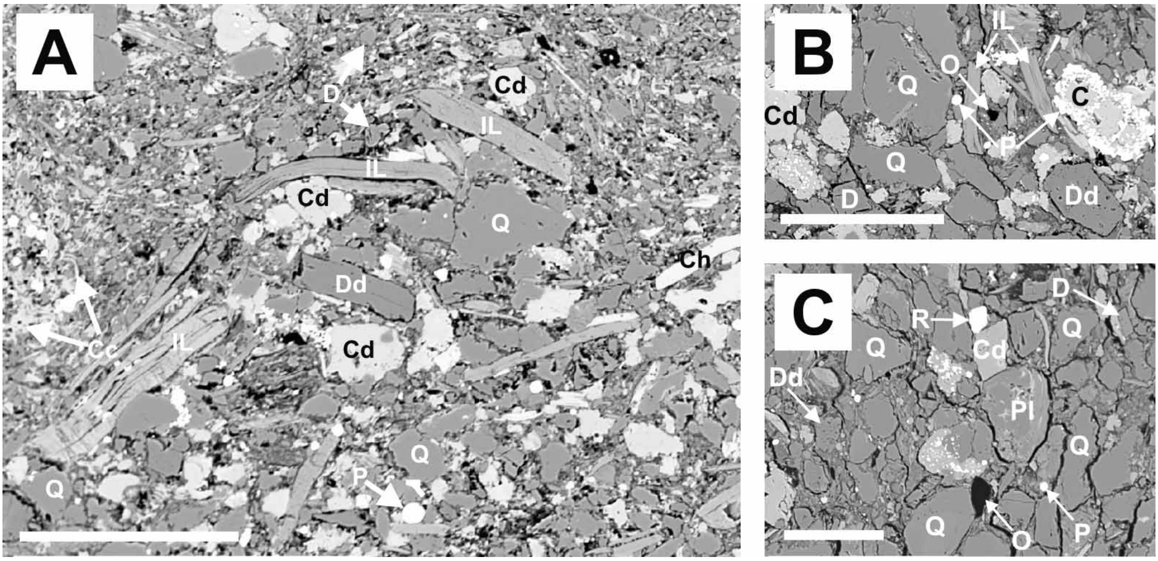

Fig. 4. Back-scattered images of the turbiditic beds. Symbols used: Q – quartz, IL – clay minerals (mostly mixed-layer illite-smectite type), Ch – chlorite, Cd – detrital calcite, C – authigenic calcite, Dd – detrital dolomite, D – authigenic dolomite, Cc – calcite coccoliths, P – framboidal pyrite, Pl – plagioclase, R – rutile, O – organic matter. **A** – Poorly sorted material from lamina C with numerous calcite grains (light grey) and individual coccoliths dispersed in the matrix (arrows on the left). Scale bar – 0.1 mm. **B** – Grain-supported material with very fine matrix composed of hydromicas. Note the pyritized foraminifer filled with calcite on the right. Scale bar – 0.1 mm. **C** – Grain-supported material with very fine matrix composed of hydromicas, which are orientated parallel to the outlines of the grains. Rutile and plagioclase are present in the centre. Scale bar – 0.05 mm

to the outlines of the grains. Moreover, the proportion of clay minerals to quartz is significantly higher in hemipelagic sediments, which is also indicative (SIGHINOLFI & TATEO 1998). According to the classification proposed by POTTER & *al.* (1980), the turbiditic beds should be termed mudstone, and the hemipelagic layers clayshale. According to Mount's classification (1985 *vide* MORSE & MCKENZIE 1990), the mudstone is micritic (detrital material dominates among the $CaCO_3$ constituents), while the clayshale is allochemic (biogenic material dominates among the $CaCO_3$ constituents). The proportion of hemipelagic to turbiditic material generally increases upwards in the profile examined.

In the uppermost part of the profile, the Krosno shales are especially rich in soft-sediment deformations (SSD). Original parallel stratification can be deformed as slump bedding (Text-fig. 7A). Some of the slump folds are "rooted" and indicate a northerly direction of slumping. However, most of the slumped shales have basal décollements and show strong internal deformation (Text-fig. 7B). Nevertheless, the lithology of such slumps does not differ from that of the undeformed strata. Microscopic analysis of these slumps revealed that the rheology of the finer material (the hemipelagic laminae + the upper parts of the turbiditic beds) was plastic, whereas the rheology of the coarser material (the lower parts of the turbiditic beds) was fluidal during deformation (Text-fig. 7C). The Krosno shales sometimes lack clear layering or strong parting and appear to be homogenized (Text-fig. 7D) – they readily disintegrate in the hand. Thin section analysis revealed that they are mud breccias composed of semi-consolidated clasts of Krosno shales from 1 mm up to 2 cm in diameter (Text-fig. 7E). Thanks to early-diagenetic, precompactional cementation (BOJANOWSKI 2001), the original texture of the sediments has been "frozen" in the authigenic carbonates. Most of them precipitated within undisturbed deposits and preserved parallel lamination, but early-diagenetic deformations also appear – a few concretions enclosed already brecciated sediments (Text-fig. 7F).

The Krosno shales contain various carbonate rocks (Text-fig. 2). The laminated limestones were deposited during intensive pelagic sedimentation of coccoliths (see HACZEWSKI 1989). All the other carbonates are clearly authigenic rocks. Except for the dolomites, they formed as by-products of methane oxidation within near-surface sediments (BOJANOWSKI 2007). The thicker laminated lime-

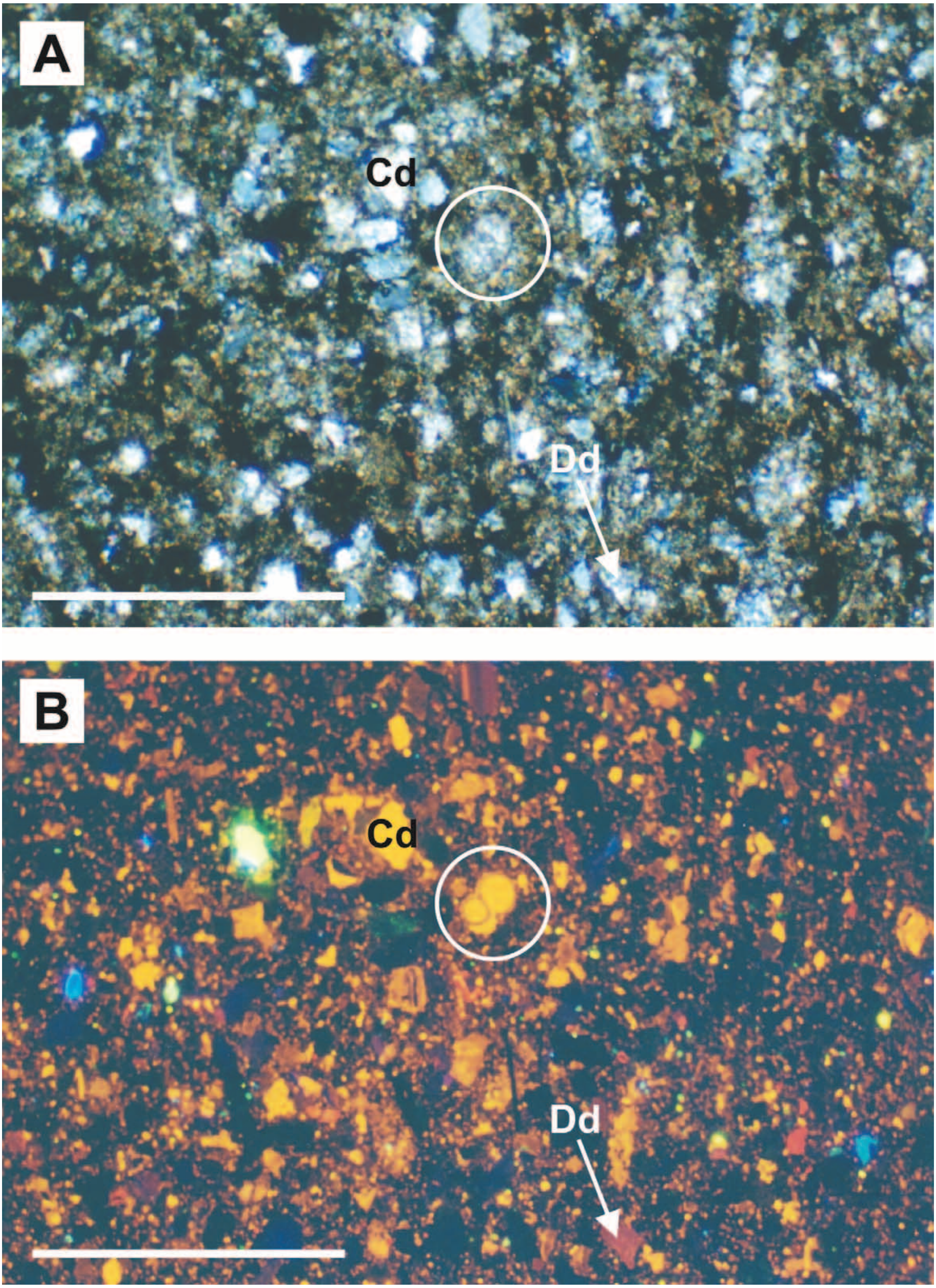

Fig. 5. Images of a turbiditic bed (lamina C). Symbols used: Cd – detrital calcite, Dd – detrital dolomite. Planktonic foraminifer is encircled. Scale bar – 0.5 mm. **A** – Thin section (nicols crossed). The foram is hardly noticeable. **B** – CL image of A. Detrital calcite shows yellow luminescence. The shell of the foraminifer (orange) is filled with authigenic calcite, which exhibits significantly brighter luminescence (light yellow). Scale bar – 0.5 mm

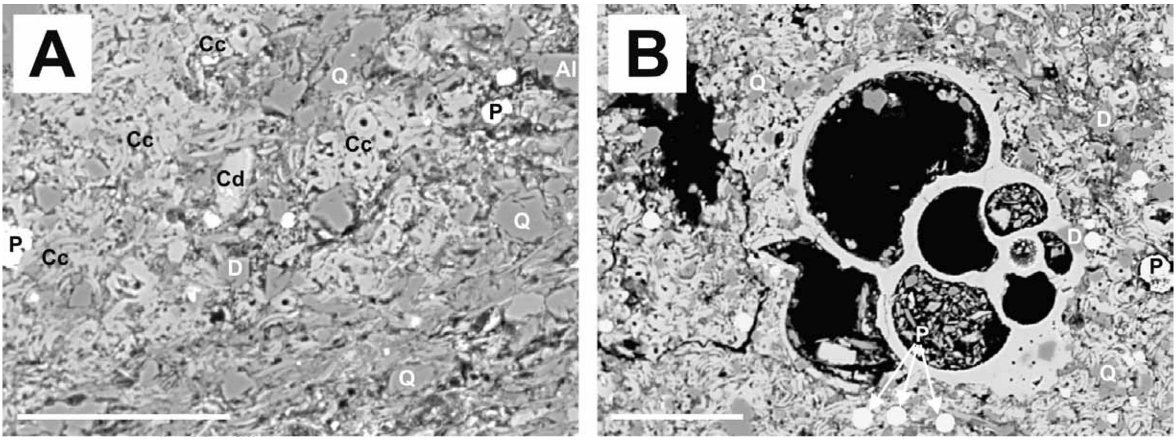

Fig. 6. Back-scattered images of the hemipelagic beds. Symbols used: Q – quartz, Cd – detrital calcite, D – authigenic dolomite, Cc – calcite coccoliths, P – framboidal pyrite, Al – albite. **A** – The main constituent are coccoliths (light grey), but clay minerals (dark grey, very fine platy minerals) dominate in the lower part of the image and are oriented parallel to the lamination. Scale bar – 0.05 mm. **B** – A large planktonic foraminifer is composed of calcite, dolomitized in places, and filled with bitumen. The surrounding material is the same as in A – the main constituents are also coccoliths (light grey). Scale bar – 0.05 mm

stone bed was also indurated by methane-derived calcite cement (*op. cit.*).

The petrographic diversity of the examined succession of the Krosno shales stems from the interplay of three factors: proportion of hemipelagic to turbiditic material – it increases upwards in the profile; intensity and type of sediment deformations; intensity and type of cementation – it led to the formation of various authigenic carbonates.

DINOFLAGELLATES

Biostratigraphical analysis of the succession was based on dinoflagellates (BARSKI & BOJANOWSKI in prep.) and showed that the profile was deposited in the middle of the Rupelian. This dating allowed a rough estimation of the time span for the deposition of the entire profile (about 1 my). The ages of the reworked species (Jurassic, Late Cretaceous, Paleocene, Eocene) provide some indication of the stratigraphy of the source rocks.

Among the species identified, there are two which are used as palaeoecological indicators (KÖTHE 1990): *Deflandrea phosphoritica*, indicating a proximal environment and *Spiniferites ramosus*, indicating a distal environment (BARSKI & BOJANOWSKI in prep.). These two species are found together in the samples, although in different proportions to each other.

PROVENANCE OF DETRITAL MATERIAL

The provenance of the Krosno shales has been investigated on the basis of relative La, Th and Sc contents depicted on a standard triangle diagram (Text-fig. 8) (according to BHATIA & CROOK 1986). Projection points of all the samples in the diagram indicate that the detrital material from the Krosno shales was originally derived from rocks forming a continental island arc.

SEDIMENTATION OF THE KROSNO SHALES

The succession of the Krosno shales examined is rather monotonous, because it was deposited under relatively stable conditions. Fine-grained flysch sediments were laid down episodically while in the background, continuous, calm hemipelagic sedimentation took place. Turbidity currents caused deposition of few-cm-thick turbiditic beds (Text-fig. 3A). Hemipelagic sedimentation resulted in deposition of finer, few-mm-thick beds in between successive turbidity beds (Text-fig. 3A-C) when the pause in flysch sedimentation was long enough. Apart from this, there were periods when pelagic sedimentation prevailed and deposition of calcareous ooze (coccoliths) was very intensive, which gave rise to the laminated limestones.

The biostratigraphical analysis (BARSKI &

BOJANOWSKI in prep.) allowed estimation of mean accumulation rate of the Krosno shales succession to be between 8 and 11 cm/ky by taking the thickness (between 80 and 110 m) and the time span for the deposition (about 1 my) into account.

The co-occurrence of dinoflagellate species indicative of separate, mutually exclusive ecological settings in the Krosno shales resulted from redeposition of proximal material by turbidity currents into a more distal environment, where

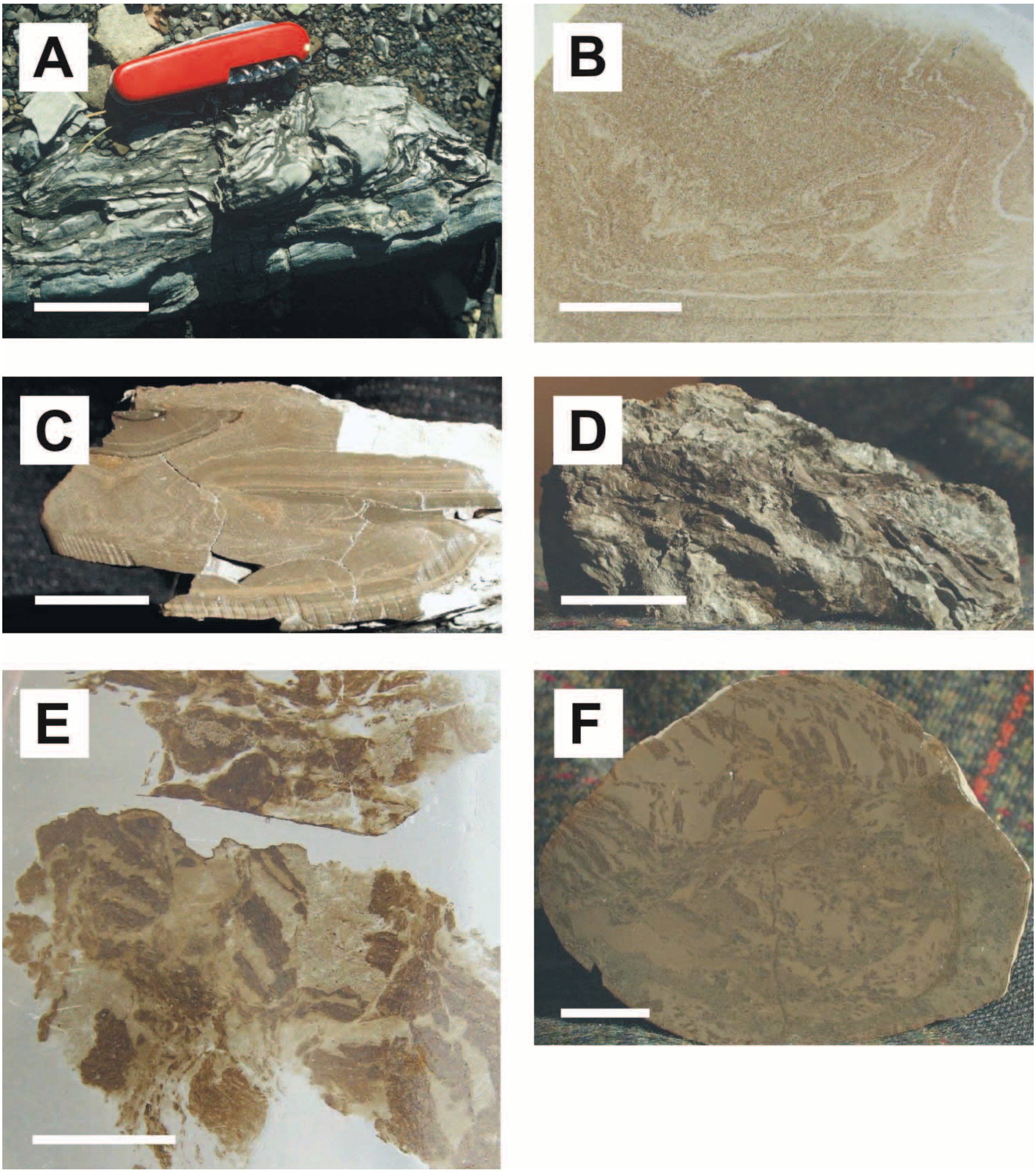

Fig. 7. Images of the soft-sediment deformations from the Krosno shales. **A** – Shale with slump bedding. Scale bar – 50 mm. **B** – Thin section (plane polarized light) of the shale with slump bedding showing fluidization deformations of the turbiditic silt. Scale bar – 5 mm. **C** – Cross-section of the shale with slump bedding. The brown laminae (abundant in clay) are deformed plastically, while the grey laminae (abundant in silt) lost their coherence and were fluidized. Scale bar – 20 mm. **D** – Apparently homogenized mudstone, which readily disintegrates in the hand. Scale bar – 20 mm. **E** – Thin section (plane polarized light) of the specimen from D showing that it is a mud breccia. The brown laminae (abundant in clay) are deformed plastically, while the grey laminae (abundant in silt) lost their coherence and were fluidized. Scale bar – 5 mm. **F** – Cross-section of the concretion, which cemented brecciated mudstone. Scale bar – 20 mm (F – after BOJANOWSKI 2007 – with kind permission of Springer Science and Business Media)

hemipelagic sedimentation persisted (BARSKI & BOJANOWSKI in prep.).

The northern vergence of the "rooted" slump folds together with the northerly transport direction of flysch material in the Krosno shales implies that deposition took place on a north-dipping, fairly steep slope (see MARTINSEN & *al.* 2003). This slope constituted the southern marginal part of the Silesian basin. Coarse-grained flysch facies, which are absent in the succession examined, were deposited in more distal environments of the Silesian basin at that time (JUCHA & KOTLARCZYK 1961, ŚLĄCZKA 1969, PESCATORE & ŚLĄCZKA 1984, PICHA & STRANIK 1999, POPRAWA & *al.* 2002, OSZCZYPKO 2004). It is proposed that the succession was deposited on a slope between submarine canyons via which turbidity currents carried coarse-grained material to more distal and deeper parts of the Silesian basin. The channels constituted a kind of bypassing system and acted as sediment conduits for turbidity currents flowing downslope. The fine-grained turbidite beds deposited in such settings represent interchannel flysch facies (see STOW & BOWEN 1980, STOW & *al.* 2001).

Weak segregation of the normal grading in the turbiditic beds (Text-fig. 3B, 3D) indicates that clay material flocculated in seawater and that clay flocs were deposited together with silt grains at relatively low settling velocity (STOW & BOWEN 1980). The relatively high content of clay suggests that the current density and velocity at the moment of deposition were also relatively low (*op. cit.*). These slow-moving dilute interchannel flows could have originated from higher-velocity turbidity currents on their passage downslope through canyons – much of the fine-grained material that arose from the "classical" turbidity currents could have been flushed out of the canyons and spread on the nearby slopes, resulting in deposition of fine-grained interchannel flysch facies (see BOUMA 2000, WYNN & *al.* 2000, STOW & *al.* 2001). However, some flows could not have been associated with the channelized currents and they represented separate sedimentation events. They were derived from a linear source, probably a shelf-break, that was composed of fine-grained material.

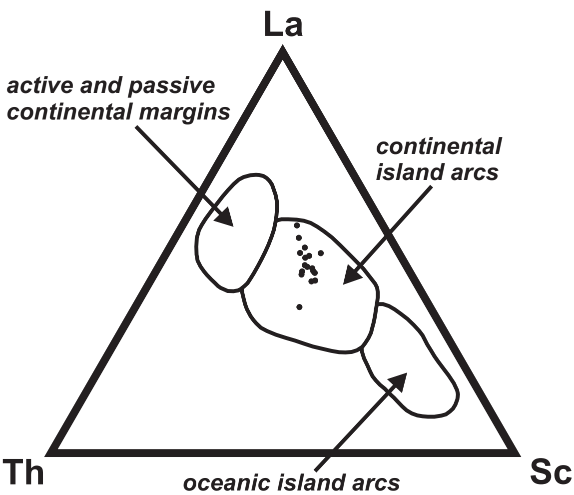

Fig. 8. La – Th – Sc discrimination diagram (according to BHATIA & CROOK 1986) of the Krosno shales. All the projection points (17 samples) fall within the field for continental island arcs

This fine-grained sedimentation of the Krosno shales took place along the orogenic front (PICHA & STRANIK 1999, POPRAWA & *al.* 2002, OSZCZYPKO 2004) which was approaching from the south. The observed general attenuation in flysch sedimentation upwards in the profile was probably caused by gradual reduction of regional accommodation related to the drop in relative sea level (KSIĄŻKIEWICZ 1975, OLSZEWSKA 1984). This shallowing of the Dukla subbasin resulted from the gradual progradation of the accretionary prism (POPRAWA & *al.* 2002). This interpretation links very fine-grained sedimentation with the proximity of the source area and the attenuation of flysch sedimentation with the increasing tectonic instability, phenomena that are usually considered to be in opposition to each other (see e.g. MARTINSEN & *al.* 2003).

The relative contents of La, Th and Sc in the Krosno shales (Text-fig. 8) indicate that the detrital material was originally derived from rocks forming a continental island arc. The fine-grained development, presence of carbonate clasts and redeposited dinoflagellates (Jurassic–Eocene in age) (BARSKI & BOJANOWSKI in prep.) in the shales examined suggest that the source rocks were at least partly sedimentary rocks: carbonates and clastics (probably flysch). It can be envisaged that products of erosion of crystalline rocks forming a continental island arc were deposited as flysch. Later the flysch rocks were incorporated into the orogenic front, elevated, eroded and redeposited as the synorogenic Krosno shales.

Intensive synorogenic redeposition and slope destabilization on a regional scale in the Silesian basin is proved by the high content of exotics (often as olistostromes or olistoliths) (e.g. ŚLĄCZKA 1961, ŚLĄCZKA & WIESER 1962, MOCHNACKA &

TOKARSKI 1972, BURTAN & *al.* 1984), soft-sediment deformations and fluidization structures in the Krosno Formation (KSIĄŻKIEWICZ 1949, DŻUŁYŃSKI & RADOMSKI 1956, KSIĄŻKIEWICZ 1958). The Krosno shales examined were also prone to early-diagenetic deformations (Text-fig. 7), which implies that the sediments were semi-consolidated and strongly fluidized during deformation. Slumps travelled some distance, but were derived from an area with similar lithology, which is quite typical of interchannel flysch facies (WYNN & *al.* 2000, MARTINSEN & *al.* 2003). This slope instability was probably partly caused by an upward migration of methane-bearing fluids which resulted in overpressurization and fluidization of shallow, poorly consolidated sediments (BOJANOWSKI 2007). It is likely that this process also contributed to the widespread slope destabilization in the Silesian basin during sedimentation of the Krosno Formation. The fluids were probably liberated from gas-hydrates which dissociated due to the Late Oligocene sea level drop (BOJANOWSKI 2007). All these processes were triggered by orogenic activity.

CONCLUSIONS

The Krosno shales examined were deposited on a north-dipping, fairly steep slope that constituted the southern marginal part of the Silesian basin. Three kinds of sedimentation contributed to the formation of the shales: flysch, hemipelagic and pelagic. Hemipelagic sedimentation was rather slow and continuous. Turbiditic material was laid down episodically by turbidity currents, which significantly raised the overall sedimentation rate. Additionally, two periods of intensive pelagic sedimentation occurred during which coccolith ooze was deposited. The mean accumulation rate of the entire succession was moderate: from 8 to 11 cm/ky. Some parts of the slope were destabilized shortly after deposition and slumping occurred due to significant overpressurization and fluidization of semi-consolidated sediments. The source rocks were older (Jurassic – Eocene) sedimentary rocks. Geochemical data point to a continental island arc as the original source of the detrital material.

Large turbidity currents carried the coarse-grained material via the submarine canyons to more distal and deeper parts of the Silesian basin where the sandstone facies of the Krosno Formation were deposited. Some of the fine-grained material was suspended above the turbidity currents and deposited on the slopes between the submarine canyons. However, some the turbidite beds were probably deposited as separate sedimentation events from slow, dilute "sheet" flows derived from a linear source (probably a shelf-edge). Such fine-grained interchannel flysch facies were laid down synorogenically, and the general attenuation of turbiditic deposition in the succession examined resulted from lowering of sea level probably related to the accretionary prism formation in this part of the Silesian basin. Authigenic seep-carbonates present in this succession (BOJANOWSKI 2007) are also evidence of the orogenic activity, which caused methane-charged fluid expulsion.

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