Geological Quarterly, 2019, **63** (1): 106–125
DOI: http://dx.doi.org/10.7306/gq.1460

Occurrence of clastic injectites in the Oligocene strata of the Carpathians and their significance in unravelling the Paleogene and Neogene evolution of the Carpathian orogeny (Poland, Ukraine and Romania)

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Jankowski, L., Wysocka, A., 2019. Occurrence of clastic injectites in the Oligocene strata of the Carpathians and their significance in unravelling the Paleogene and Neogene evolution of the Carpathian orogeny (Poland, Ukraine and Romania). Geological Quarterly, **63** (1): 106–125, doi: 10.7306/gq.1460

The irregular distribution of sand injections, traditionally termed “dykes” in the Polish geological literature, within individual Carpathian units and within individual lithofacies were observed during long-lasting field works. Injectites have been observed in the Magura Beds and in the Inoceramian Beds of the Polish and Romanian Carpathians, and in the Central Carpathian Paleogene deposits. However, they are most common in the Oligocene-Miocene Menilite Beds, where they are typical and abundant, particularly in the Skole Unit. Two clastic injectite types were distinguished: sedimentary (S-type) and tectonized (T-type). Based on the occurrence and interpretation of these injectites a new two-stage conceptual model is proposed for the Polish segment of the progressive Oligocene-Miocene Carpathian orogenic belt evolution. Type S clastic injectites are interpreted as having formed in the compressional stage, during foredeep basin migration while depositional slope changes were taking place in the Late Oligocene to Early Miocene. Type T injectites are interpreted as having formed by reactivation of S-type injectites in the last, mainly strike-slip, phases of Carpathian orogenic belt formation.

Key words: injectites, dykes, Carpathians, Oligocene, Poland, Ukraine, Romania.

INTRODUCTION

Veins unconformably cutting successions of sedimentary strata and composed of material lithologically different than that of the host rocks have been observed in many regions of Poland (e.g., the Carpathians, the vicinity of Kraków, the Sudety Mountains or Kleszczów Graben) and described by Polish (e.g., Dżułyński and Radomski, 1957; Dżułyński, 1965a, b; Dżułyński and Walton, 1965; Teisseyre, 1967; Łuczyński, 2001; Wieczorek and Olszewska, 2001; Barski and Ostrowski, 2006; Goździk and Van Loon, 2007; Haluszczak, 2007; Krobicki et al., 2008; Wojewoda, 2008; Wojewoda and Burliga, 2008; Kołodziej et al., 2010; Barski, 2012; Barmuta et al., 2014, Matyszkiewicz et al., 2016) and foreign authors (e.g., Aubrecht and Túnyi, 2001; Udič and Jacko, 2008; Tămaş et al., 2016).

These forms are traditionally referred to s dykes. This general term is widely used in the literature and refers to the unconformable geometric relationship of the dyke structure to the bedding. Dykes may occur on various scales, ranging in width from several millimetres to several metres, and have variable lengths. Early reports on dykes date back to the beginning of the 19th century (Strangeways, 1821; Couvier and Brongniart, 1822; Murchison, 1827; Darwin, 1833–1834; Strackland, 1838, 1840; Leyell, 1839; Dana, 1849; Prestwich, 1855; Kirkby, 1860 – for references see Cooley, 2011). With advancing research, various terms describing the composition and geometry of dykes, their relationship with the host strata and the surrounding rocks, and formation mechanisms have been proposed. As a result, a great variety of terms describing the geometry and formation mechanism of dykes can be found in the geological literature (see Cooley, 2011), including: *clastic intrusion*, *sandstone dyke*, *fissure fill*, *injectite*, *neptunian dyke*, *pseudo ice-wedge cast*, *sedimentary insertion*, *sheeted clastic dyke*, *syn-sedimentary filling*, *tension fracture*, or *hydraulic injection dyke.* Moreover, there are many publications reporting occurrences and discussing these forms, especially in relation to their formation mechanisms, role of tectonic regime or potential as pathways for hydrocarbon migration (e.g., Collinson, 1994; Niell et al., 1997; Jolly and Lonergan, 2002; Surlyk and Noe-Nygard, 2003; Jonk et al., 2007; Hurst et al., 2011; Alsop et al., 2017). Therefore, to avoid any misunderstandings, we decided to use the term injectites for the actively filled structures. While, for the passively filled structures, we propose to use the term dykes.

Our in-house data, as well as data available in publications, indicate that in the Carpathians the occurrence of injectites is limited only to certain sedimentary units. They have been reported from the mudstones of the Magura Beds (Žlin Beds of Eocene age in Slovakia; Udič and Jacko, 2008) where their for-

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Received: August 20, 2019; accepted: November 27, 2018; first published online: March 22, 2019

mation is linked to active slope areas within sedimentary basins and to the opening of fault fractures. This conclusion is based on the presence of synsedimentary deformation structures, such as slumps, in the host strata (Udič and Jacko, 2008).

Our detailed geological fieldwork (mapping at the scale of 1:50 000) revealed an irregular distribution of injectites within Carpathian units. Apart from the Magura Beds, injectites have also been observed in several other units, including the Inoceramian Beds of the Polish and Romanian Carpathians, but are the most common in the Oligocene-Miocene Menilite Unit. In fact, their occurrence within the Menilite Beds is a peculiar feature, particularly in the Skole Unit. It is therefore intriguing that studies of clastic injectites in the Carpathian units are only rarely reported in the Polish geological literature. There are only a few reports discussing their occurrence and origin (Dżułyński and Radomski, 1957; Dżułyński, 1965a, b; Dżułyński and Walton, 1965; Haczewski and Tokarski, 1986; Krobicki et al., 2008; Barmuta et al., 2014). The most comprehensive description was provided by Dżułyński and Radomski (1957), with an overview of injectites (termed dykes in the paper) in Oligocene rocks, mainly in the Menilite Beds and their stratigraphic equivalents such as the Zakopane Beds. However, when discussing the spatial relationships and origin of these forms, these authors did not comment on the significance of the abundance of them in the Oligocene strata and the importance of palaeogeography; furthermore they did not recognize their genetic relationship with Early Oligocene tectonic processes and did not discuss the importance of geodynamic factors during the development of the injectites. Mechanical factors were suggested as triggering mechanisms but Dżułuński and Radomski (1957) did not fully explain the processes leading to injectite formation.

The occurrence of injectites in the Menilite Beds was also noted by Haczewski and Tokarski (1986) (in the paper called dykes). They described the Kliwa Sandstones as intruding the cherts of the Menilite Beds. The authors indicated that the injectites were not folded and that there were no slip surfaces along the contact with the cherts. According to them, the emplacement of injectites positively correlates with the direction of the regional stress field. Injectites were formed when the cherts were lithified and folded. Furthermore, the evidence of brittle deformation structures within the injectites imply that folding processes were still active when the injectites were lithified, which may suggest that the intrusions formed when their source was under a 1–2 km overburden and that this process took place about 20 my after sedimentary deposition (Haczewski and Tokarski, 1986).

Grasu (1996) reported the presence of sandstone injectites in the Late Eocene (Bisericani Formation) to Early Oligocene (Bituminous Marl Formation) deposits of the Vrancea Nappe, exposed in the Bistri a Halfwindow. He recognized two types of injectite: an undeformed one with sharp walls and a deformed type showing boudinage from being folded.

Considering the low number of reports of injectites occurring within the Menilite Beds, a lack of comprehensive injectite classification and of interpretation of mechanisms of their formation, particularly with respect to the evolution of the Outer Carpathians during the Oligocene, detailed fieldwork was undertaken to document injectite occurrences, to describe their types and spatial relationship with the host strata and the surrounding rocks, as well as to discuss the possible mechanisms of their formation and infilling. Most of this fieldwork was performed in the Outer Carpathians (Outer Flysch Carpathians) and within the Central Carpathian Paleogene in the Inner Carpathians (Central or Western Carpathians in the Slovak literature; Fig. 1) in Poland (within the outcrops of the Central Carpathian Paleogene, Silesian and Skole units; Figs. 2 and 3), Ukraine (Boryslav-Pokuttya Unit; Figs. 1 and 3) and Romania (Vrancea Unit, so-called Marginal Folds; Figs. 1 and 3). As injectites were documented in a total of 21 outcrops (Figs. 2 and 3) over a relatively broad area, the amount of data collected is large enough for a regional-scale analysis and to propose a model that correlates the origin of injectites with the stages of the evolution of the Carpathian orogenic belt.

GEOLOGICAL SETTING

The study area is located mainly within the Outer Flysch Carpathians (Fig. 1). This part of the Carpathian belt is separated (in the western region of the folded belt) from the Inner Carpathians (occasionally referred to as the Central Carpathians in the Slovak segment) by a narrow boundary zone, the Pieniny Klippen Belt. The Pieniny Klippen Belt was closely associated with one of the subduction zones that resulted in the development of the suture zone between the Outer and Inner Carpathians (e.g., Birkenmajer, 1976, 1977). The Outer Carpathian orogeny is typically interpreted as a thin-skinned accretionary wedge, formed in a piggy-back mode, with subsequent tectonic elements accreted to the orogenic wedge (e.g., Mahel and Buday, 1968; Książkiewicz, 1972; Koszarski and Ślączka, 1976). The present study, however, proposes an alternative interpretation of the tectonic evolution of the Carpathians Mountains.

Both the Inner and Outer Carpathians form parts of an orogenic belt made up of sediments deposited in the accommodation spaces created at the margins of the European platforms after the breakup of the Pangea supercontinent (Wieczorek, 1990, 1993); this partition process is well-recorded in the Triassic to Cretaceous sedimentary successions. Analysis of distributions of sedimentary facies, particularly of Mesozoic age, indicates the existence of a half-graben system geometry that formed in an extensional regime. The marine transgression over the half-graben areas was gradual, initially within the Inner Carpathians area (Jankowski et al., 2012c; Jankowski, 2015a). The existence of the half-graben geometry has also been documented in some regions of the Carpathian foreland area, for example in the northernmost (in the Polish segment) Lublin half-graben that remained unaltered even in a compressional regime (Malinowski and Mojski, 1981). The presence of such a geometry during the development of the Carpathian basins is further supported by the most recent geophysical data (Malinowski et al., 2013, 2015; Probulski and Maksym, 2015). The Lublin half-graben was not incorporated into the structure of the Carpathian orogenic belt and as such represents an element of the undeformed foreland of the Carpathians, infilled with Mesozoic deposits unconformably overlain by Miocene deposits.

The pattern and geometry of the half-grabens were controlled by tectonic structures inherited from the “pre-Alpine” orogenic stages, such as fault zones developed during the Variscan or older orogenies (Jankowski and Probulski, 2011). Some of the pre-Alpine fault zones were also reactivated during the orogenic collapse of the Carpathians and, jointly with the overthrusted Carpathian orogenic belt, subjected to extensional deformation (Jankowski, 2015a). The process of half-graben infilling by sediments likely continued from the Triassic until the formation of the foreland basin and the beginning of the closure of the Carpathian basins (Jankowski, 2015a) coupled with the process of inversion of extensional structures. In the Polish segment of the Carpathians, the oldest deposits infilling the half-grabens occur in the Inner Carpathian region (Jankowski, 2015a).

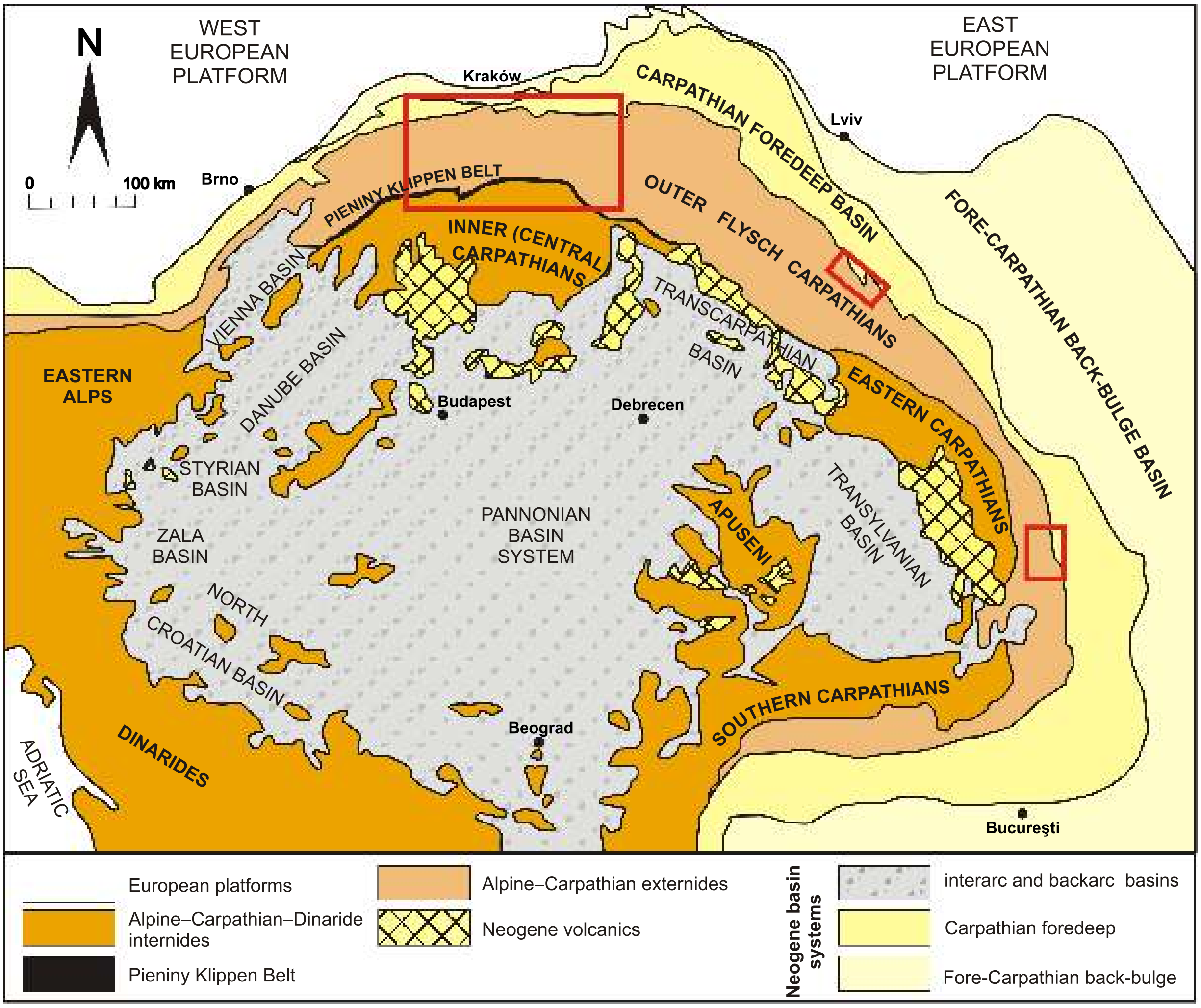


Fig. 1. Sketch map of the Alpine-Carpathian-Pannonian-Dinaride domain (after Kováč et al., 2007, adapted)

Study areas marked by red rectangles

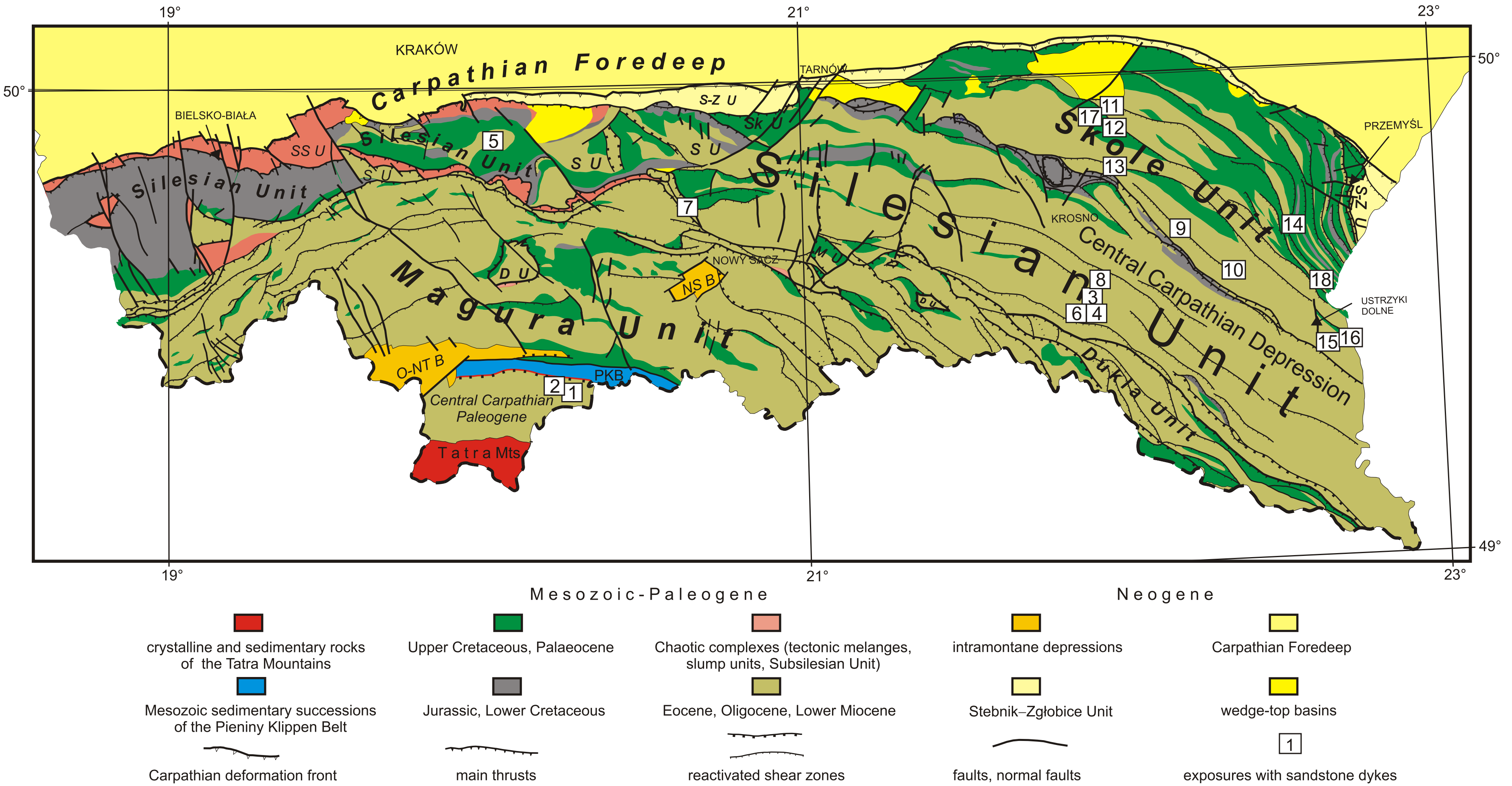
The final stage of the orogenic belt formation was related to the closure of the basinal zone of Paratethys (a part of Tethys) and its subdivision during the last stage of its development into a series of isolated basins (Popow et al., 2002).

Shortening and closure of the Carpathian basins are processes traditionally linked with the concept of the collision of European platform with a series of approaching more or less hypothetical microplates, moving in variable directions. According to this concept, the Outer Carpathians are an orogenic belt related to the collision of the East European Platform with the ALCAPA (Alps-Carpathians-Pannonian) and Tisza-Dacia blocks (terranes, microplates) (e.g., Ratschbacher et al., 1991; Kováč et al., 1997, 2007; Grad et al., 2006).

The last stage of the orogenic belt formation and basin shortening has been related to subduction (e.g., Birkenmajer, 1976, 1977; Kováč et al., 2017). The concept of subduction in the Carpathians has long been criticized (Książkiewicz, 1977), due to the lack of volcanism and of the geometry typical of a subduction zone. Data collected during recent seismic surveys in the Polish segment of the Carpathians also contradict the subduction hypothesis (Malinowski et al., 2013; Probulski and Maksym, 2015).

It has been postulated based on palaeomagnetic surveys that the arcuate, oroclinal shape of the Carpathians resulted from differences in the directions of rotations of individual structural blocks in the western and the Romanian sectors (Márton and Márton, 1996); field mapping, however, does not provide convincing evidence for such processes. Instead, some concepts (Burtman, 1986; Jankowski, 2015a) associate the formation of the Carpathian orocline with the process of bending of the orogenic belt and basin space due to the pressure of the Moesian Plate, which caused eastward (so-called) extrusion and tectonic escape of the westernmost segment of the Carpathians. This process of tectonic escape coincided with the formation of pull-apart basins and development of fault zones parallel and perpendicular to the orientation of the main tectonic structures (Jankowski, 2015a).

In the Polish segment, the pre-Cretaceous phase of accretion of the southernmost fragments of the orogenic belt led to the formation of the Inner Carpathians, with its northern boundary running along the margin of the Laramide front and overlapping with the tectonic boundaries of the Manín and Haligovice units (Jankowski, 2015a). The Pieniny Klippen Belt developed as a chaotic complex in the foredeep of these units. More recent



Exposures: 1 - Kacwińska river, 2 - Łapsze,3 - Besko, 4 - Hamry, 5 - Kobielnik, 6 - Rudawka Rymanowska, 7 - Znamirowice, 8 - Zmiennica;9 - Bachórz, 10 - Bircza, 11 - Futoma, 12 - Hermanowa, 13 - Hłudno, 14 - Krępak, *15* - Łodyna, *16* - Stebnik, *17* - Straszydle, *18* - Wojtkowa

Abb: D U - Dukla Unit; M U - Magura Unit; NS B - Nowy Sącz Basin; O-NT B - Orawa–Nowy Targ Basin; PKB - Pieniny Klippen Belt; S U - Silesian Unit; SS U - Subsilesian Unit; S-Z U - Stebnik–Zgłobice Unit; W U - Węglówka Unit

Fig. 2. Geological map showing locations of the studied outcrops in the Polish part of the Carpathians (after Jankowski, 2017, adapted)

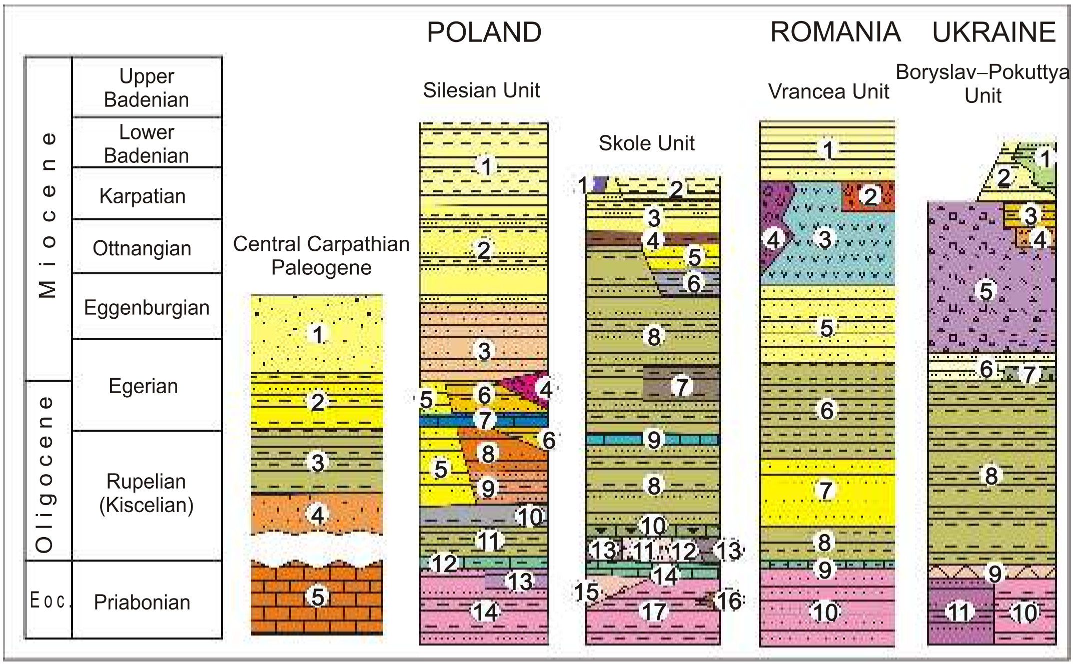


Fig. 3. Correlation of regional litostratigraphic units (based on Jankowski et al., 2012b) in which outcrops with sand injectites were documented

Central Carpathian Paleogene: 1 – Ostrysz Beds, 2 – Chochołów Beds, 3 – Zakopane Beds, 4 – Szaflary Beds, 5 – Nummulite Beds; **Silesian Unit**: 1 – Upper Krosno Beds, shale member, 2 – Upper Krosno Beds, sandstone-shale member, Niebylec Shales, 3 – Ostre Sandstones, 4 – Gorlice Beds, 5 – Lower Krosno Beds, sandstone member (Lesko Sandstones), 6 – Lower Krosno Beds, sandstone-shale member, 7 – Jasło Limestone, 8 – Lower Krosno Beds, Otryt Sandstones, 9 – Zatwarnica Beds, 10 – Transition Beds, 11 – Menilite Beds, 12 – Globigerina Marls, 13 – Green Shales, 14 – Hieroglyphic Beds; **Skole Unit**: 1 – Leszczawka Diatomites, 2 – Upper Krosno Beds, sandstone-shale member, 3 – Upper Krosno Beds, sandstone-shale member, 4 – Niebylec Shales, 5 – Lower Krosno Beds, 6 – Transition Beds, 7 – Łopianiec Beds, 8 – Menilite Beds, Kliwa Sandstones, 9 – Jasło Limestone, 10 – Dynów Marls and Cherts, 11 – Siedliska Conglomerates, 12 – Boryslav Sandstones, 13 – Subcherts Beds, 14 – Globigerina Marls, 15 – Popeli Beds, 16 – Pasichna Beds, 17 – Hieroglyphic Beds; **Vrancea Unit**: 1 – Doftana Beds, 2 – Brebu Conglomerates, 3 – Salt Beds, 4 – Slon Beds, 5 – Gypsum Beds (Gura Soimului Beds, Gura Misina Beds), 6 – Upper Dysodilic Shales, 7 – Kliwa Sandstones, 8 – Lower Dysodilic Shales, Bituminous Marls, Menilite Beds, 9 – Globigerina Marls, Lucăceşti Sandstones, 10 – Bisericani Beds; **Boryslav-Pokuttya Unit**: 1 – Balychi Beds, 2 – Stebnik Beds, 3 – Dobrotiv Beds, 4 – Sloboda Conglomerates, 5 – Vorotyscha Beds, 6 – Polanitsia Beds, 7 – Rusiv Conglomerates, 8 – Menilite Beds, Boryslav Sandstone, 9 – Popeli Beds, 10 – Bystrytsia Beds, 11 – Vytkivtsi Beds

concepts (Plašenka and Mikus, 2010; Casteluccio et al., 2015) define the Pieniny Klippen Belt as a "wildflysch" complex. However, the Polish segment of the Pieniny Klippen Belt was interpreted by Jankowski (2015a) as a slump complex that was incorporated into the Late Cretaceous-Paleocene basin from the Laramide front of the margin of the Inner Carpathians. Accordingly, the formation of the Inoceramian facies marks the beginning of the formation of a sedimentary basin in the peripheral foreland basin system (De Celles and Giles, 1996) of the Inner Carpathians, traditionally known as the Outer Carpathians flysch basin.

During the Late Cretaceous the depocentre of the foreland basin began migrating towards the European platforms. This event is well-reflected in the distribution of the Paleocene, Eocene, and younger Upper Oligocene and Miocene deposits (Książkiewicz, 1974). The migration of the depocentre was terminated during the Early Oligocene extension stage. As a result, new sedimentary basins such as the Central Carpathian Depression and the Central Carpathian Paleogene Basin were formed (Jankowski, 2015a). In the Polish and the Romanian segments, shifting of the foredeep basin depocentre, forebulge and backbulge zones persisted until the Late Miocene and the Pliocene, respectively (Săndulescu, 1988, 1994). As a result of foreland basin shifting, the half-grabens of the rifting stage became tectonized and gradually incorporated into the orogenic belt structure. This process was associated with the formation of chaotic complexes such as olistostromes, slides and slumps in the foreland of the active thrusts (Jankowski, 1995, 2004).

Facies distribution in the Carpathian Basin was also highly influenced by intervals of relatively high sea level, resulting in periodic lateral extensions of wedge-top sedimentation, often over the entire area of the already-formed Carpathians. In the hinterland, the wedge-top sediments unconformably overlie the already formed orogenic belt; sedimentary gaps and unconformities are quite common here (Jankowski, 2015a).

Processes of shortening and tectonization led to the formation of several tectonic units that can be correlated based on litho- and biostratigraphy (Fig. 3). Sedimentary sequences within each of these units consistently become younger towards the foreland area (Fig. 3). Traditionally, these units have been considered as tectono-facies units and their sedimentary sequences were thought to have originated within separate basins/sub-basins (Książkiewicz, 1972; Oszczypko et al., 2006). However, some studies indicate these traditionally distinguished units are only tectonic in nature (Săndulescu, 1988; Jankowski et al., 2012b) and should not be cross-correlated with sedimentary basins, and consequently that the development of the Carpathian sedimentary basin and the tectonic deformation occurred with no relationship to each other.

The most recent investigations have also recognized the great importance of gravitational processes in incorporating many fragments of Carpathian tectonic units into the orogenic belt structure, especially within the most internal Magura Unit (Jankowski, 2007). Based on data collected during field studies and thermochronological investigations, several additional stages of tectonic deformation have been recognized, hence refining the evolution of the Carpathian orogenic belt (Mazzoli et al., 2010; Jankowski and Probulski, 2011; Andreucci et al., 2013; Castelluccio et al., 2015). It has been shown that reactivations of thrust zones as strike-slip zones were amongst the most important factors shaping the Carpathians (Jankowski and Probulski, 2011) and are responsible for the formation of flower or horse-splay structures. Tectonic development terminated with the stage of the orogenic collapse, a process well-documented by field observations and thermochronology, and recorded on geological maps (Mazzoli et al., 2010; Jankowski and Probulski, 2011; Andreuccci et al., 2013; Jankowski and Margielewski, 2014). The evidence of this orogenic collapse can also be observed in the northern foreland of the Carpathians (Jankowski and Margielewski, 2014). The tectonic history of the Carpathians, in the Polish segment, ends with a phase of large block movement and tectonic denudation, a stage still ongoing at the present time (Jankowski et al., 2012c).

VARIATIONS IN SAND INJECTITIES

Studies of sand injectites were performed in several tectonic units of the Outer Flysch Carpathians (Figs. 1–3). Most of the injectites investigated occur within the Oligocene to Miocene sedimentary successions. These features were generally noted in the most external tectonic elements of the orogenic belt, and are particularly abundant within the Menilite Beds of the Skole Unit and its Romanian equivalent, the Vrancea Unit, and in the Silesian Unit (Figs. 2 and 3). They were also found within the Borislav-Pokuttya Unit (its equivalent in Romania is referred to as the Vrancea Unit, the so-called Marginal Folds). Despite being previously noted by Udič and Jacko (2008), during this study injectites were rarely observed in the sedimentary succession of the Magura Unit.

Burial under younger sediments caused most of the strata to undergo some deformation directly after deposition and before the final stages of diagenesis. The main process, compaction, causes the expulsion of pore fluids and tighter grain packing. Apart from overpressure, various soft-sediment deformation may develop as a result of liquefaction and fluidization of the clastic beds. These processes introduce intrusions of remobilized clastic sediment into the surrounding strata, forming sheets concordant (sills) or discordant (dykes) to bedding (Fig. 4) (e.g., Jolly and Lonergan, 2002; Collinson, 2003; Jonk et al., 2007). Therefore it is quite likely that most dykes formed *via* injection of fluidized sand from source beds (parent bodies) into the host rock (Figs. 4 and 5). In that case they should be called injectites.

In general, sand injectites may be formed during the course of active hydraulic fracturing as a result of injection of pressurized fluids into the host rock, both into the overlying beds (Fig. 4A, C; Jolly and Lonergan, 2002; Jonk et al., 2007; Hurst et al., 2011), laterally, and even into the underlying beds (Fig. 4B; Dreimanis, 1992; Niell et al., 1997; Surlyk and Noe-Nygard, 2003; Hurst et al., 2011). Underwater neptunian dykes and clastic dykes formed in terrestrial settings are other types of structures passively infilled with sediment that is lithologically different from the host rocks, that formed due to passive, gravitational infilling of fractures that opened towards the top (e.g., Amårk, 1986; Collinson et al., 1989; Obermeier, 1996; Aubrecht and Túnyi, 2001; Łuczyński, 2001; Jasionowski et al., 2012; Matyszkiewicz et al., 2016). In certain types of sediments, dykes can also form as a result of early diagenetic processes when diagenetic transition of opal A into opal T causes significant decrease in porosity and sediment volume (Richard et al., 2006).

The most typical triggering mechanisms causing sediment liquefaction and mobilization are (1) direct seismic shocks (Allen, 1975; Audemard and de Santis, 1991; Peterson, 1997; Wojewoda and Burliga, 2008), indirect seismic shocks (tsunami effects; Olson, 2007), and shocks caused by physical impacts (Hunton and Shoemaker, 1995; Sturkell and Ormo, 1997; Kenkmann, 2003; Wittmann et al., 2004; Hudgins and Spray, 2006; Levi et al., 2006); (2) increase in pore pressure related to overburden thickness and sedimentary slides (Dżułyński and Radomski, 1957; Holzer and Clark, 1993; Jolly and Lonergan, 2002) or liquefaction caused by wave action (Dalrymple, 1979; Martel and Gibling, 1993; Alfaro and Soria, 1998); and (3) increase in fluid volume resulting from sedimentary mechanisms, e.g., influx of fluids from deeper horizons to shallower sand bodies causing overpressure in the bedding (Truswell, 1972).

In the first case, the occurrence in sedimentary successions of injectites triggered by seismic shocks may be considered as evidence of tectonic activity in the area, an indicator for determining palaeostress (Obermeier, 1996; Beacom et al., 1999; Boehm and Moore, 2002; Obermeier et al., 2005; Wojewoda, 2008; Wojewoda and Burliga, 2008), and a tool for determining the geometry of faults during stages of orogenic belt development (Harms, 1965; Hurst et al., 2011; Alsop et al., 2017).

In general, the formation of sand injectites is associated with sediments deposited in marine settings characterized by high sedimentation rate, dominance of mud and mud-sand deposits, and synsedimentary tectonic activity (Collinson, 1994; Jolly and Lonergan, 2002; Hurst et al., 2011). Tectonic activity that is clearly evident in a directional stress field may cause increase in pressure within sediments and, in addition, rapid increase in pore fluid pressure, resulting in liquefaction and fluidization of unconsolidated sediments. Such events usually take place within active slope depositional settings where fold-and-thrust belts are being formed (Winslow, 1983; Jolly and Lonergan, 2002). Hence, these processes occur most commonly in tectonically active parts of sedimentary basins, at their margins (Boehm and Moore, 2002; Jonk et al., 2003), and particularly in foredeep basins characterized by a migrating depocentre.

Deep marine settings are not the only environments where sand injectites have been observed. These structures have also been noted in salt domes (Amårk, 1986; Marco et al., 2002), glacial sediments (Kruger, 1938; Larsen and Mangerud, 1992; Rijsdijk et al., 1999; Le Heron and Etienne, 2005; Van Der Merr et al., 2009), fault fractures within crystalline rocks (Cross, 1894; Birman, 1952), shallow marine, deltaic, fluvial and lake settings (Jolly and Lonergan, 2002), and in the vicinity of volcanic and impact craters (Mashchak and Ezersky, 1980, 1982).

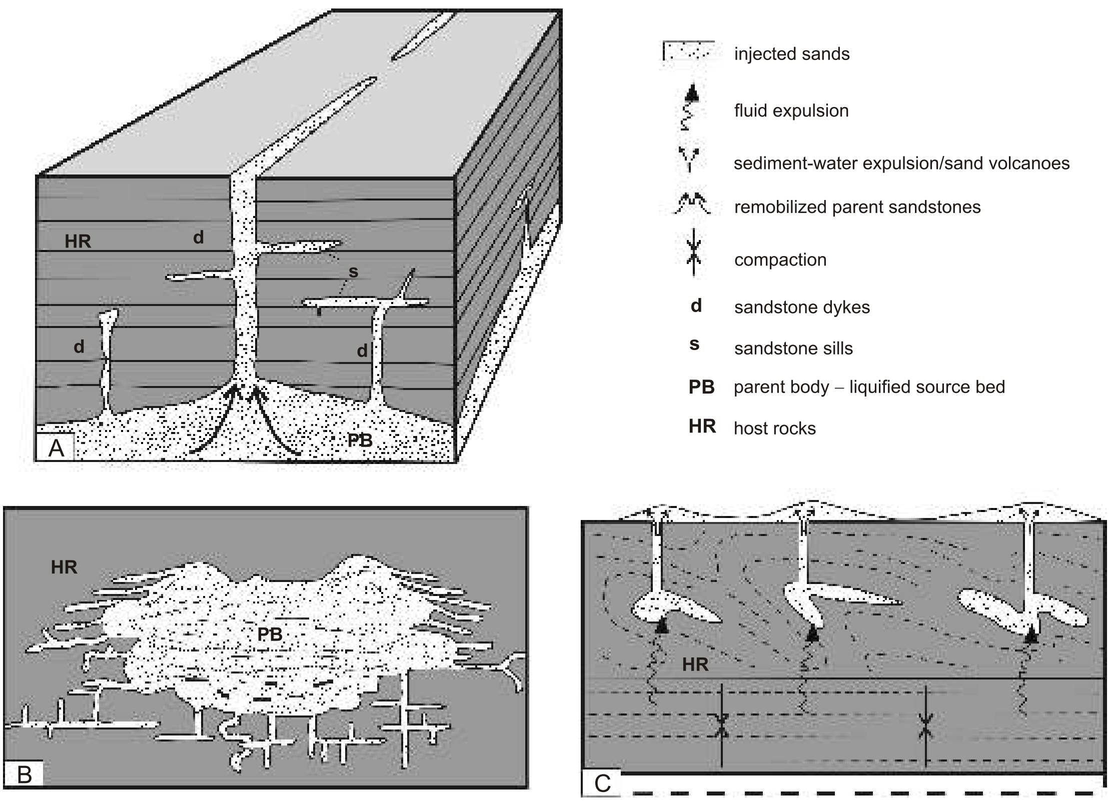


Fig. 4. Variations in sediment injections (not to scale)

A – schematic diagram illustrating the principal features of clastic sills and dykes (based on Jolly and Lonergan, 2002); **B** – final stage of remobilization, fluidization, liquefaction, intrusion and diapirism of sandstone under deep burial of the base-of-slope sand bodies loaded into the slope muds (based on Surlyk and Noe-Nygaard, 2003); **C** – compaction of sediment underlying a slump, causing, because of loading, fluid expulsion and remobilization, injection, and extrusion of the sandy and/or silty fraction (after Jonk et al., 2007, adapted)

DESCRIPTION OF THE INJECTITES STUDIED

Our recent fieldwork data have indicated that, in the Outer Carpathians and in the Central Carpathian Paleogene rocks, sand injectites occur most frequently within the sandstone-mudstone and siliceous-mudstone sedimentary successions. In the Outer Carpathians their occurrences have been most commonly observed within the outcrops of the Menilite Beds in the Silesian and Skole units in Poland (Figs. 2 and 3), the Boryslav-Pokuttya Unit in the Ukraine, and the Vrancea Unit in Romania (Figs. 1 and 3).

Field data, including geometric relationships with the host strata, as well as regarding the texture and structure of the sand injection infills, indicate that these forms unconformably cut the host rocks. However, the tectonic component can vary considerably from one injectite to the other. Based on the degree of tectonic influence two main categories of sand injections are distinguished: post-sedimentary (type S sand injectites) and strongly tectonized (type T sand injectites).

The type S post-sedimentary sand injectites are characterized by varying orientation to bedding (Figs. 5, 6A, C, D and 7C–G), uneven pattern of the injection walls (Fig. 6C–D), variable width (Fig. 5), variable degree of compaction-related shortening (Fig. 6D) and massive sand infilling (Fig. 7B). In some instances, a direct connection of the sand injection with the parent body was evident (Fig. 5).

Outcrops with such types of sand injectites are characterized by a moderate extent of fault-related deformation. In most cases, the observed synsedimentary, fold-related deformation structures (Fig. 5) and flow deformation structures (Fig. 6C) appeared to be genetically associated with submarine slides active during sedimentation. We believe that these injectites were formed as a result of fluidization and displacement of fluidized sand within loosened cohesive mudstone-sandstone and siliceous deposits that were likely slightly lithified. According to Jonk et al. (2007), the stress field that caused increase in the pore pressure and sediment fluidization occurred under the overburden of younger sediments due to sediment deformation related to the displacement of a sediment mass caused by submarine slides occurring on the slopes of depositional forms and during the formation of the slopes of fold-and-thrust forms. The considerable size of the infillings of the type S sand injections and their relatively small transverse cross-section, may indicate that they were formed as a result of a high-density laminar flow, without hydraulic contact with the sedimentary basin bottom (Cobain et al., 2015). Assessment of the sediment thickness under which fluidization and injection processes occurred is extremely difficult because it depends on a number of factors, such as the magnitude and orientation of the stress field, proportion of the sand to mud interbeds, initial porosity, presence of silica, and so on (Jolly and Lonergan, 2002; Richard et al., 2006; Hurst et al., 2011).

The type T strongly tectonized sand injectites are characterized either by perpendicular or sharply angular orientation to

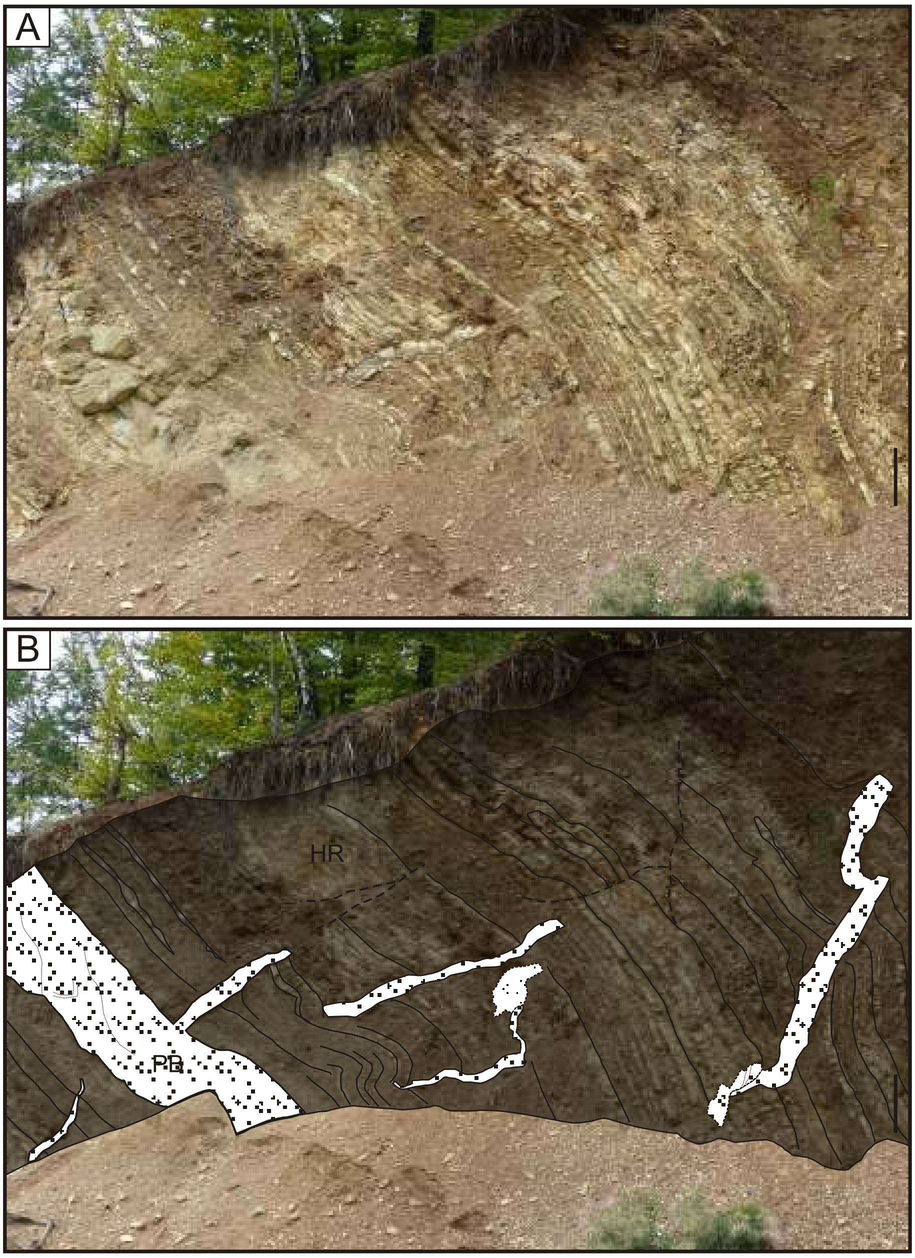


Fig. 5. Assemblage of clastic injectites in the Menilite Beds at the Hłudno 2 outcrop, Skole Unit (for location see Fig. 2)

A – general view of the outcrop; **B** – interpretative drawing of the clastic injectite pattern; note the sandy parent body (PB), joints (dashed lines), bedding (solid lines) and fold-like deformation within the host rock (HR)

bedding (Figs. 6A, B and 7A), flat planes bounding the injection (Fig. 6A, B), generally constant injection widths (Figs. 6A, B and 7A), massive, strongly fractured sand infills (Figs. 6A, B and 7A), and the presence of slickensides on the planes bounding the injections and on internal fracture planes within the infills (Fig. 7A). The slickensides were covered by striae and mineralization pointing to the horizontal, strike-slip orientation of the displacements.

Although the outcrops in which such sand injectites were observed, e.g., Futoma (Poland), Hłudno (Poland) (for location see Figs. 2 and 3) and Trochaniv (Ukraine) were characterized by significantly extensive fault-related deformation, synsedimentary, fold-type deformation structures (Fig. 5) and slump-slide deformation structures were still present (Fig. 6C). Furthermore, these outcrops also revealed that some of the type T injectites had features typical of the type S injectites, such as uneven bounding planes of the injection and variable width (Fig. 7C).

The co-occurrence of sedimentary and tectonic features indicates the occurrence of at least two phases during sand injectite development. Structures that initially developed as type S sand injections, and as such became ideal zones compensating tectonic stress during the evolution of the orogenic belt, were subsequently transformed into type T sand injectites. This reactivation process of the type S injectites into the type T injectites is interpreted as associated with zones of strike-slip faults.

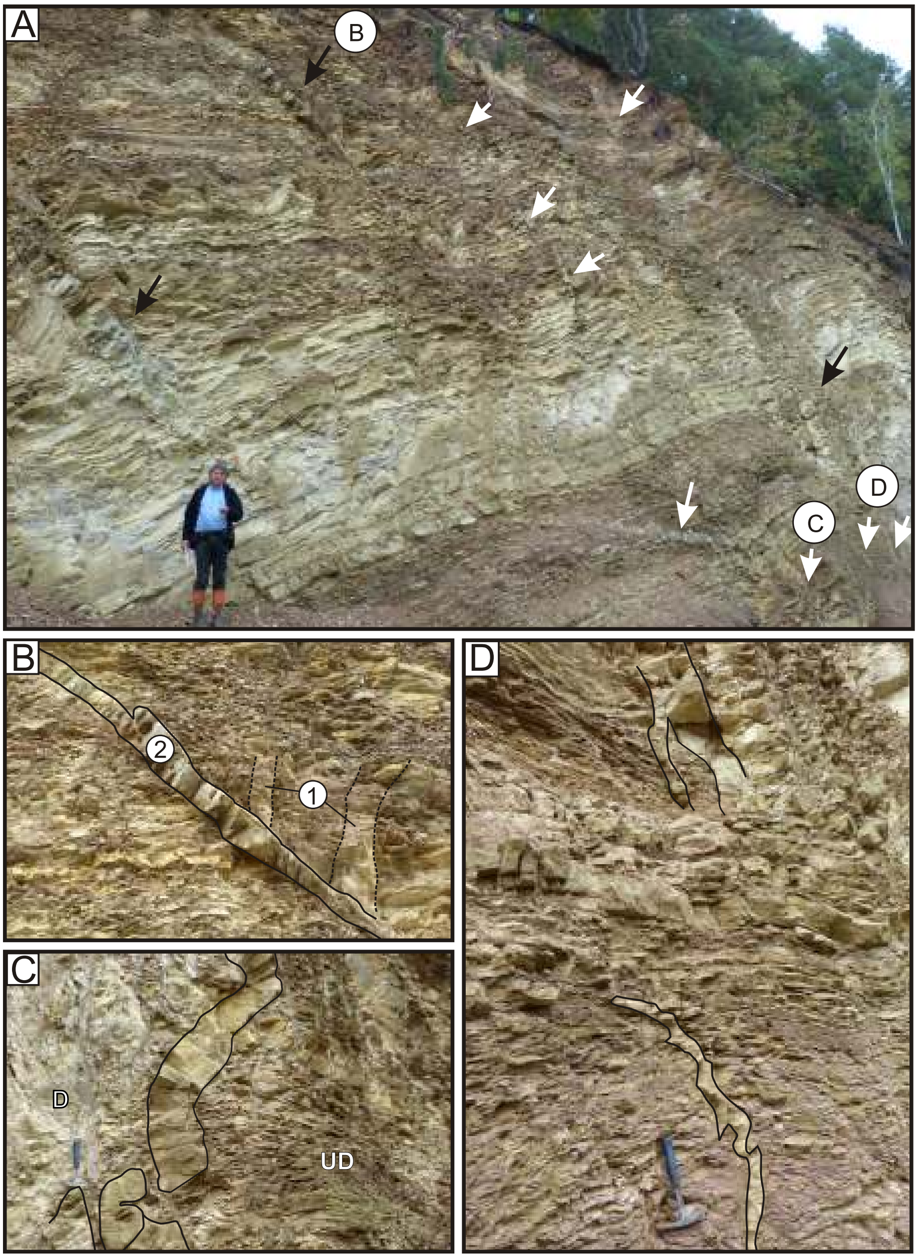


Fig. 6. Clastic injectite pattern in the Menilite Beds, Hłudno 1 outcrop, Skole Unit (for location see Fig. 2)

A – general view of the outcrop; white arrows point to clastic injectites, black arrows point to sand injections bordered by fault zones, circled letters – location of insets B, C and D; **B** – two oblique injectite systems, the older system is sub-perpendicular to bedding (1), and is dissected and displaced by the younger injectite (2); **C** – deformed injectite between strongly deformed unit (D) and undeformed menilites (UD); **D** – two slightly deformed injectites perpendicular to bedding

SIGNIFICANCE OF CLASTIC INJECTITES IN UNRAVELLING THE PALEOGENE AND NEOGENE EVOLUTION OF THE POLISH SEGMENT OF THE CARPATHIAN OROGENY

As already stated, most clastic dykes observed during this study occur within the Oligocene-Miocene Menilite Beds of the sedimentary successions of the Skole Unit and its Romanian equivalent, the Vrancea Unit (Figs. 2 and 3). The Menilite Beds, and its eastern and western stratigraphic equivalents respectively, contain the most lithologically diverse successions, including for example black bituminous shales, the Dysodilic Shale, Kliwa sandstones, etc. (Fig. 8A–D). The Menilite Beds and their equivalents have the widest lateral extent and are present not only in the area traditionally referred to as the Outer Carpathians but also within the Eastern Carpathians and in the Molasse Basin. According to Jankowski (2007), the Menilite facies should also encompass deposits which begin the sedimentary infill of the Central Carpathian Paleogene Basin, including the Podhale Basin of the Polish sector (Jankowski, 2015a) and its equivalents in the Romanian segment (Jankowski, 2007). The common occurrence of the Menilite Beds in both the Outer and Inner Carpathians indi-

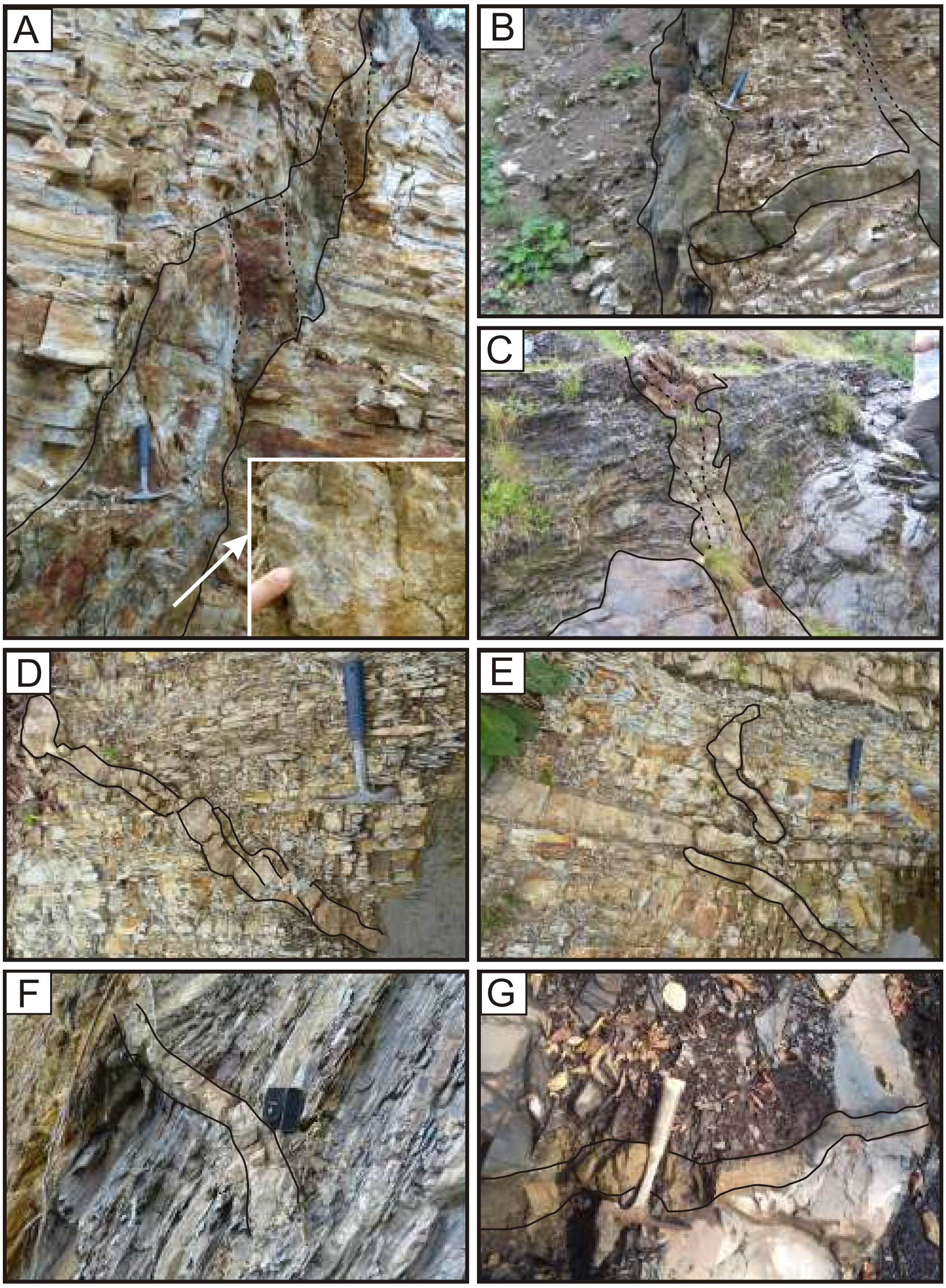


Fig. 7. Clastic injectites from different localities (for location see Figs. 1 and 2)

A – part of a giant-scale, strongly tectonized clastic injectite at Futoma (PL); white arrow points to surface with slickensides, enlarged in white box; **B** – sand-glauconite clastic injectite in Futoma (PL); **C** – strongly tectonized injectite in Trochaniv, Verkhnje Synovydne (UA); **D** – top view of the menilites succession cut by clastic injectite in the Nechit River section (RO); **E** – folded clastic injectite in the Nechit River section (RO); **F** – undeformed clastic injectite in Znamirowice (PL); **G** – coarse-grained clastic injectite at Kobielnik (phot. Anna Filipek)

cates that there must have been a very close relationship between these two regions during Oligocene-Miocene time. Furthermore, the Menilite-like deposits also occur outside the Carpathian area (Jarmołowicz-Szulc and Jankowski, 2011) and reach areas of the foreland that have not been subjected to the Carpathian deformation, as far as the area of the European Platform, e.g., Roztocze (Myśliwiec and Śmist, 2006; Jankowski, 2015a; Jankowski and Margielewski, 2015).

The deposition of the Menilite Beds began in the Early Oligocene (Olszewska, 1985; Garecka, 2008, 2012). The extent of their distribution was closely controlled not only by the morphology of the basin bottom but also by relative sea level

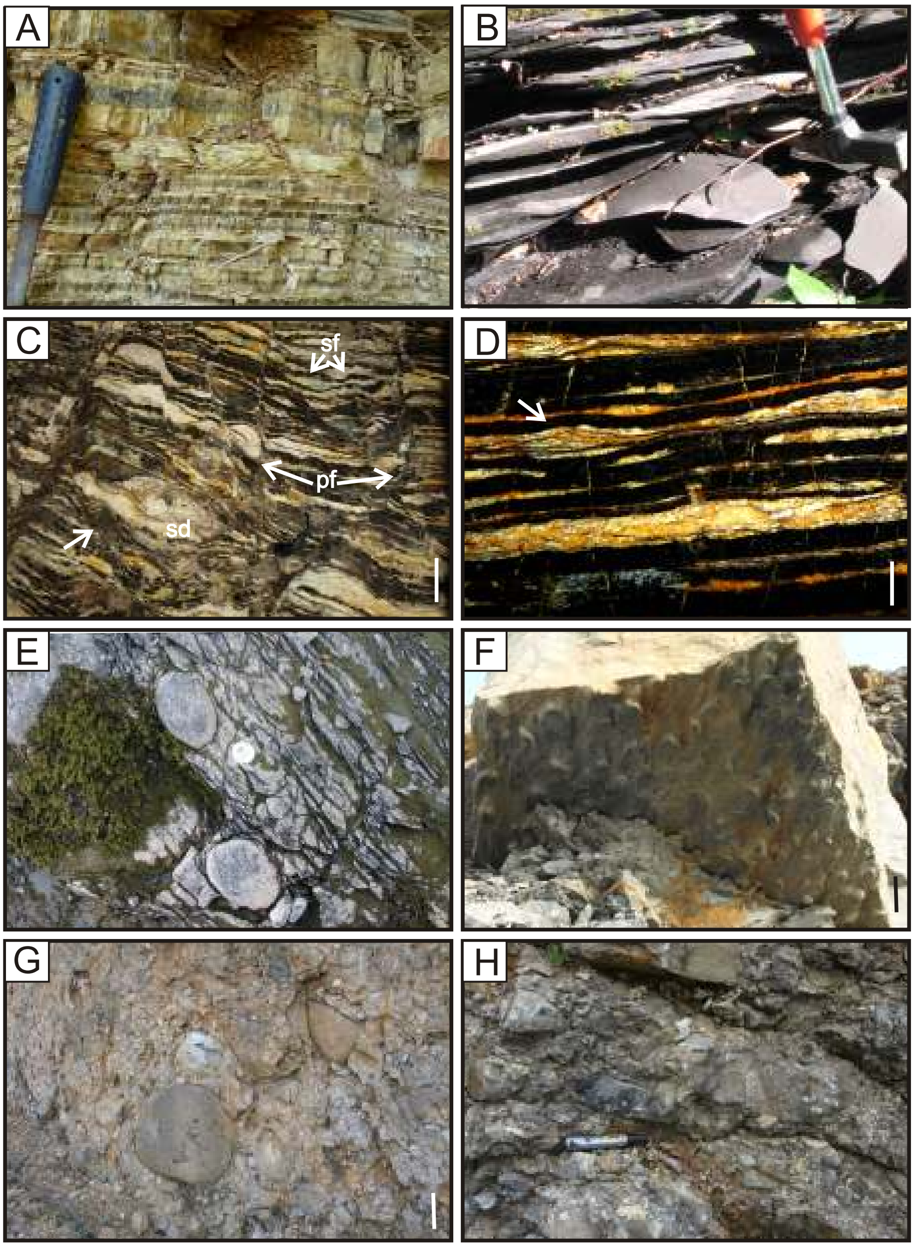


Fig. 8. Lithology of the Menilite Beds

A – thin- and medium-bedded cherts, Leszczawa Górna outcrop; **B** – black clayey shales, Smilno outcrop (Slovakia), hammer for scale; **C** – thin-bedded black muddy shales with sandy, ripple flaser- and wavy-cross-bedded interlayers; note synsedimentary deformation (sd), synsedimentary faults (sf) and post-sedimentary faults (pf), Aksmanice outcrop, white bar is ~5 cm; **D** – thin-bedded black muddy shales with sandy, rippled wavy and lenticular cross-bedded interlayers; note chevron-like pattern of wave ripples (white arrow), white bar is ~3 cm; **E** – medium-grained clasts at the base of muddy shales, Kalnica outcrop; **F** – flute casts on the lower surface of a sandstone bed, Cergowa Beds, Barwałd outcrop, black bar is ~10 cm high; **G** – debrite-type conglomerate, Skrzydlna outcrop, white bar is ~10 cm; **H** – poorly sorted, debrite--type breccia built of Mesozoic clasts, Zdiar exposure (Tatra Mts., Slovakia), pencil is ~10 cm

changes occurring periodically in the Carpathian Basin from the Early Oligocene to the Early Miocene (e.g., Amadorii et al., 2012; Kováč et al., 2017). In the sedimentary succession, the Menilite Beds occur above the Lower Oligocene Globigerina Marls horizon which marks the maximum flooding surface of the Early Oligocene Carpathian Basin (Watkinson et al., 2001). In some areas and sections, the Menilite Beds are found above the Hieroglyphic Beds or the Mszanka Sandstones. Interestingly, the Menilite Beds appear simultaneously across the entire area of their occurrence. The onset of black shale appearance within the Menilite Beds correlates to a period of water anoxia and a lack of sea-water circulation that was followed by a significant shallowing caused by an isolation of the most marginal parts of the Tethys Ocean, i.e., the Paratethys (e.g., Popow et al., 2002). A significant feature of the Menilite Beds system is the strongly diachronous termination of their sedimentation and of the lateral transition into the Krosno Beds during the final Miocene stage of their development (Olszewska, 1985; Garecka, 2008, 2012; Jankowski et al., 2015).

The Menilite Beds have been traditionally interpreted as "deep flysch" sediments (e.g., Uchman et al., 2006), sometimes even as deep-marine "oceanic" sediments (e.g., Oszczypko et al., 2006; Golonka et al., 2006). More recent studies, however, suggest that these views should be revised. Evidence of wave ripple-marks and hummocky cross-stratification in the Menilite Beds of the marginal tectonic elements of the Carpathians, e.g. the Boryslav-Pokuttya Unit (Jarmołowicz-Szulc and Jankowski, 2011), imply shallow-marine settings. The presence of structures indicating shelfal environments was also documented within the Menilite Beds near Gorlice by Watkinson et al. (2001) and confirmed recently by Dziadzio (2015). Evidence of a generally shallow sedimentary setting and relatively proximity to land occur in the Menilite facies of the Central Carpathian Paleogene Basin, i.e. the Zakopane Beds (Filipek et al., 2017). Additionally, geochemical data collected at a number of the Menilite Beds localities also indicate that the deposition was occurring in close proximity to land masses within shallow sedimentary settings (Matyasik et al., 2012).

The reconstruction of sedimentary-tectonic evolutionary stages, including the area of initial deposition that was subsequently subjected to partly synsedimentary stages of tectonic deformation, is required to explain both formation mechanisms of the sand injections within the Menilite Beds and the extent of these injectites. Therefore, we propose a novel model of development of the Carpathians and their foreland. The model is based on our data collected during extensive and detailed fieldwork, including systematic mapping (e.g., Gucik et al., 1991; Jankowski, 2004, 2007, 2013, 2015b; Jankowski and Ślączka, 2015) combined with results of seismic analyses of many sedimentary sections (Jankowski and Probulski, 2011) and data available in earlier scientific reports (e.g., Jankowski et al., 2012b, 2015; Jankowski, 2015a). The proposed model of the evolution of the Polish segment of the Carpathians applies to the area of the extent of the Menilite facies and its synchronous stratigraphic unit – the Krosno Beds – during the Oligocene and Miocene.

Early Oligocene extension stage (Fig. 9B). The model assumes the existence of zones that periodically separated Carpathian sedimentary sub-basins, commonly referred to as the Carpathian cordilleras. Their development is closely associated with the Early Oligocene extension phase (Jankowski and Probulski, 2011; Jankowski, 2015a), as a result of which the sedimentation of the Menilite Beds expanded over a broad area that included intra-montane basins, the Central Carpathian Depression and the Central Carpathian Paleogene Basin (Fig. 9B). Within this latter basin, the Menilite Beds are referred to as the Zakopane Beds or the Huty Formation.

The occurrence of the sub-basins was related to the area of half-graben boundaries formed after the breakup of the platform margin (Fig. 9A), as well as to their uplift during the forebulge migration that was already taking place during the Oligocene and later in the Miocene. The existence of the tectonic half-grabens during the Early Oligocene (Fig. 9B) is documented by borehole data (Zawoja 1, Sucha Beskidzka 1, Potrójna 1, Ropa 1). The process of the Early Oligocene breakup of a segment of the already formed orogenic belt was followed by the formation of the Menilite Beds cover on the orogenic wedge (e.g., Stronie structure).

The presence of the Menilite Beds on various components of the pre-Oligocene basement in the entire area of the Carpathians studied has been documented by many examples of direct contacts of the Menilite Beds with older strata, including the Oligocene/Cretaceous contacts (Latoś, 1960–1980) where Paleocene and Eocene sedimentary successions are missing. The Menilite Beds cover represents thus a sedimentary succession that was not associated with the older basement, that not only partly developed within tectonic grabens in sedimentary and stratigraphic continuity with the Paleogene strata, but also covered horsts that were subjected to Oligocene erosion, such as the Fredropol and Gorlice horsts (Fig. 9B, C). The processes of horst and graben extension and formation could also influence relative sea level fall. It has been concluded, based on sedimentological observations (e.g., Watkinson et al., 2001; Dziadzio, 2015; Filipek et al., 2017), that the sedimentary basin in which the deposition of the Menilite Beds occurred during the Early Oligocene, was an extensive and relatively shallow sedimentary basin characterized by restricted water circulation and restricted oxygenation where fine clastic sedimentation prevailed.

Late Oligocene – Early Miocene compression and inversion stage (Fig. 9C). The change of tectonic regime from extension to compression during the sedimentation of the Menilite Beds and the shifting of the orogenic belt-foreland basin system caused successive synsedimentary transformation of the sedimentary basin. At this time, the sedimentary facies were becoming more diverse, with the Menilite Beds transitioning laterally into the Krosno Beds. The depositional system of these units is referred to herein as the Menilite-Krosno system. There is an apparent diachroneity in the relationships of the Menilite-Krosno system. In southern regions, the Menilite facies were replaced by the Krosno facies during the Late Oligocene. The occurrence of the isochronous Menilite and Krosno facies is documented by the Jasło Limestone Chronozone (NP 24) (Jucha and Kotlarczyk, 1961; Garecka, 2012).

The process of basin shortening and closing caused partial destruction and zonal separation, and a transformation of the half-graben geometry of the basement. As a result, these zones inherited a particularly complex geometry, further complicated by their reactivation during the secondary tectonic deformation (Jankowski and Probulski, 2011; Jankowski, 2015a). Moreover, the original half-graben geometry likely influenced the inversion process during basin shortening and the placement of overthrusts of the main tectonic elements, whereas zones of tectonic boundaries of the half-grabens could have influenced the formation of out-of-sequence and in-sequence overthrusts (Fig. 9C).

According to the model presented (Fig. 9C), several depositional zones may be distinguished within the Menilite-Krosno system. These zones were developing during the final stage of the Carpathian Basin evolution and the orogenic belt formation depending on the progress of the shifting of the orogenic belt-foreland basin system, of synsedimentary folding processes and of ongoing sedimentary infill of the inherited half-grabens. Deposits of the Menilite-Krosno system also occur as wedge-top facies, in De Celles and Giles' terminology (DeCelles and Giles, 1996), on the already-formed Carpathian orogenic belt, overlying sequences of the Magura Unit and the Pienniny Klippen Belt.

During the Late Oligocene to Early Miocene evolutionary stage of the Carpathian Basin development, the Menilite-Krosno Beds were preferentially deposited within the accommodation space of the migrating Oligocene-Miocene entire foreland basin. They spread over the region of the synchronous backbulge, where they extend as far as the Roztocze region (Jankowski and Margielewski, 2014). As a result, the Lower Miocene sedimentary cover (e.g., Przemyśl Beds, Krakowiec Clays) is lithologically similar to the Krosno Beds and retains a

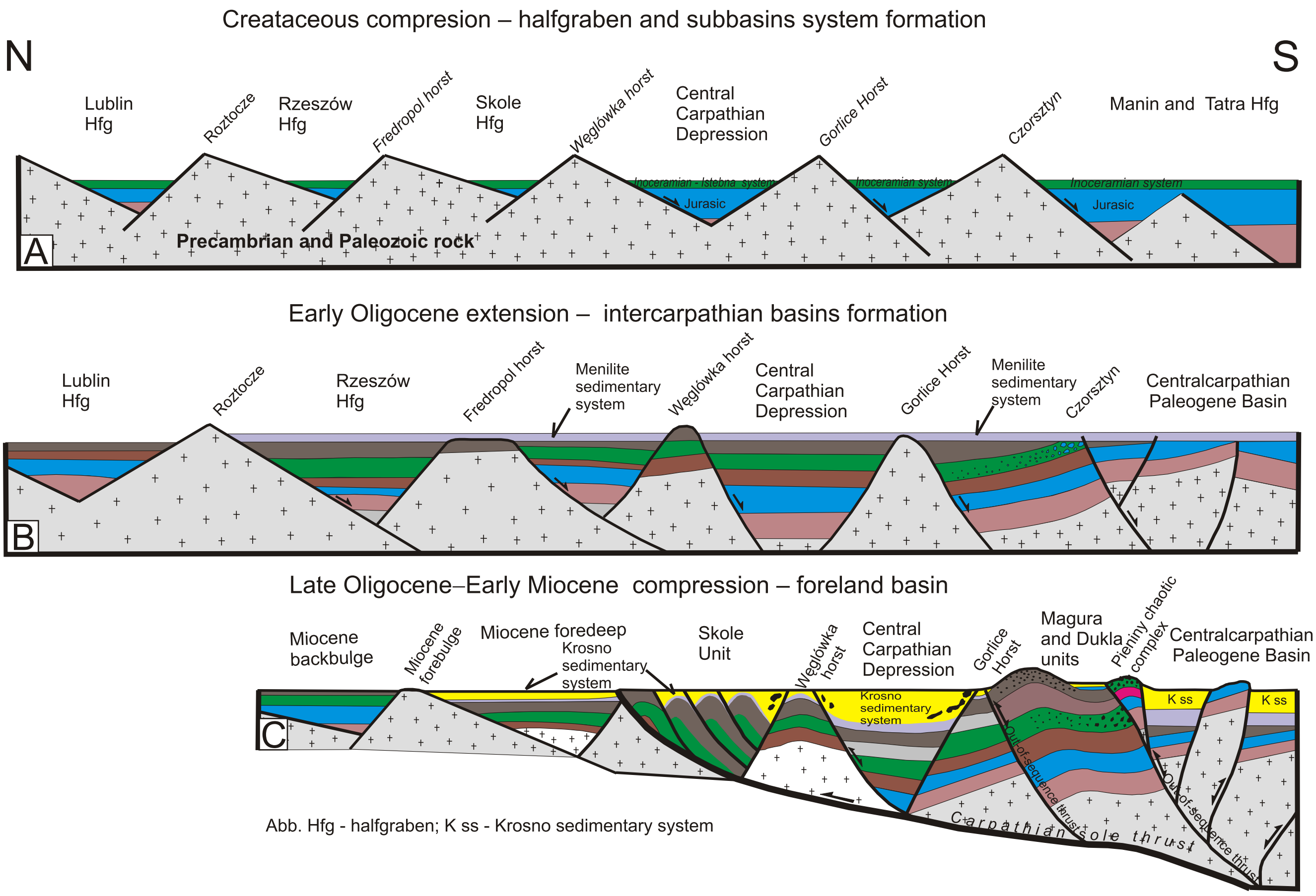


Fig. 9. Conceptual model of the last stages of the Polish part of Carpathian orogenic belt building

A – Cretaceous extension; **B** – Early Oligocene extension; **C** – Late Oligocene–Early Miocene compression and inversion

westward direction of transportation, parallel to the orogenic belt (Jankowski et al., 2015). Therefore, the most northern external depositional zone of the Menilite-Krosno facies, that subsequently became the Miocene Fore-Carpathian Basin, is the area of present-day Roztocze.

RELATIONSHIP OF THE CLASTIC INJECTITES TO THE EVOLUTIONARY STAGES OF THE POLISH SEGMENT OF THE CARPATHIAN OROGENIC BELT

The clastic injectites investigated during the present study occur typically in the Skole Unit (Figs. 2 and 3) within deposits assigned to the Menilite and Menilite-Krosno systems. These strata contain much evidence of shallow marine sedimentation (Jarmołowicz-Szulc and Jankowski, 2011; Matyasik et al., 2012). Well-sorted sandstones, with shallow marine sedimentary structures such as wave cross-bedded interlayers and hummocky cross-stratification (Fig. 8 C, D; Jarmołowicz-Szulc and Jankowski, 2011), as well as amber blocks (Teisseyre, 1922) have been observed within the shelfal Kliwa Sandstones that occur in the Skole Unit.

The depositional zone of the Menilite Beds (in the area of the present-day Skole Unit) was tectonically unstable during the deposition of the Menilite-Krosno system (Fig. 9C). At the end of the Oligocene, it functioned as a forebulge during the northward migration of the basin depocentre. At that time, extension was quite likely to occur on the forebulge slope, as suggested by the common occurrences of injectites and slope facies (Fig. 10A–F).

The compressional stage of the basin development described above (Fig. 9C) implies a very high facies variability and, moreover, the existence of zones of variable tectonic regimes. This, however, does not indicate great sedimentation depths. In fact, deposits of the Menilite-Krosno facies consist of pelagic and hemipelagic sediments, slope facies indicating well-developed processes of gravitational mass movements (Fig. 8G), and channels distributing sedimentary material from depositional slopes of the accretionary prism into the basin. As a result, the compressional stage provides good conditions for processes of sand fluidization and displacement within fractures formed in cohesive mudstone-sandstone and siliceous deposits. The stress field causing increase in pore pressure and the sediment fluidization developed under an increasing overburden of the accumulating deposits, leading to sedimentary deformations related to displacement of sediment masses within submarine slides on the slopes of depositional forms and during the formation of slopes of fold-and-thrust forms.

Numerous sand injectites that occur within the Menilite Beds of the Silesian Unit (Figs. 2 and 3) have also been investigated during this study. These injectites are of the type S, indicating that their origin is closely related to sedimentary processes (Fig. 7F, G). The geotectonic position of the Silesian Unit suggests that, during both the Early Oligocene extensional stage and the subsequent compressional stage, sedimentary

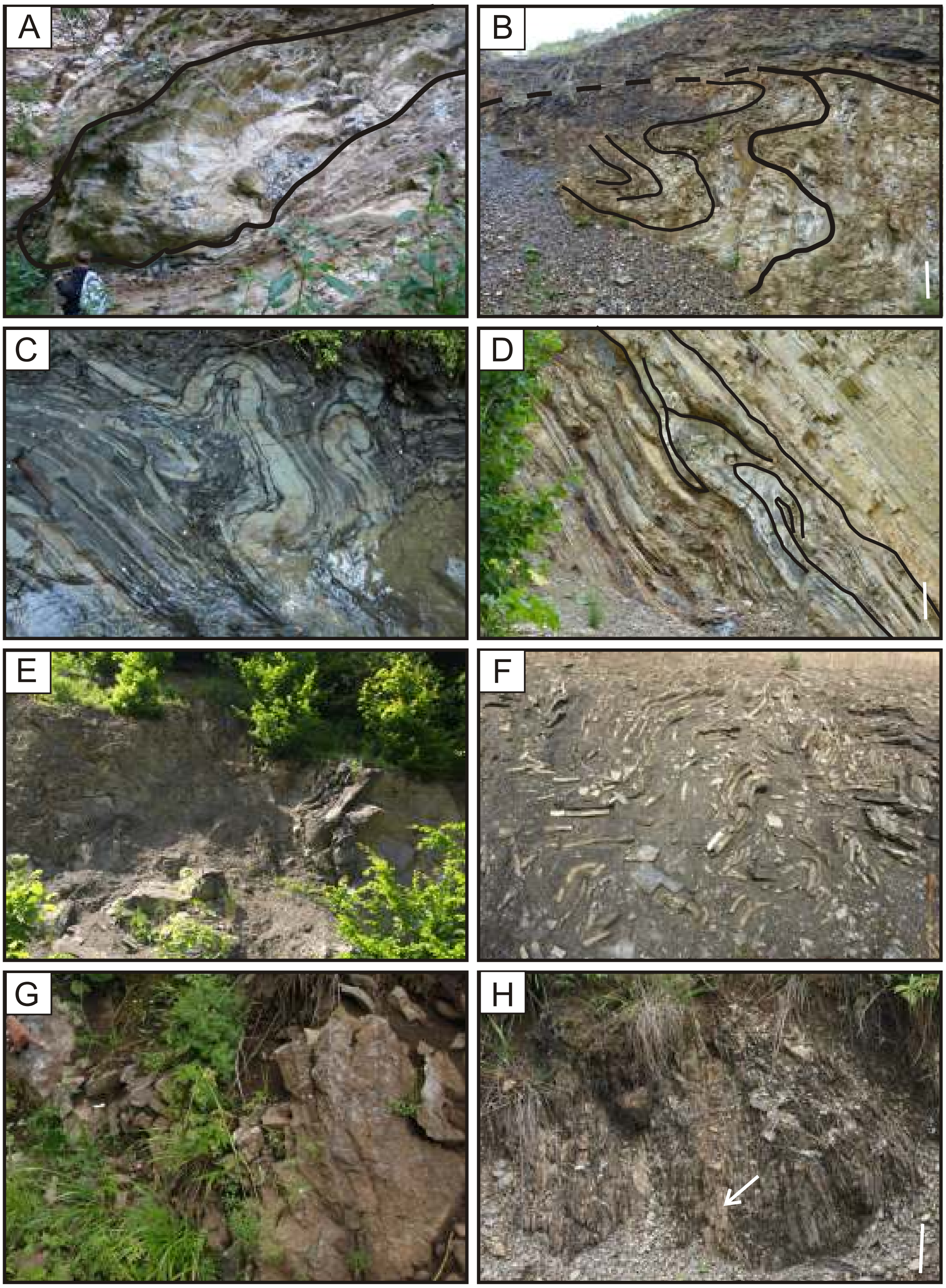


Fig. 10. Synsedimentary fold deformations of the Menilite Beds (A–D photos) and characteristic of the Menilite Beds in the Inner Carpathians (E–H photos)

A – large-scale slump deformation within medium-bedded menilites with sandy interlayers and diatomites, Hermanowa outcrop; **B** – large-scale slump deformation overlain by undeformed silty and sandy menilites, Futoma outcrop, white bar is ~1 m; **C** – medium-scale slump deformation within thin- to medium-bedded menilites, Serednica outcrop; **D** – medium-scale slump deformation inside undeformed chert succession of the Menilite Beds; slumping direction points to sequence overturning, Krościenko outcrop, white bar is ~1 m; **E** – general view of strongly deformed menilites of the Central Carpathian Depression area, Wrota Podpolozia (UA); **F** – slump deformation structures within thin-bedded menilites of the Carpathian Depression area, Nizne Vorota (UA); **G** – small exposure with conglomerates and breccias with a prevalence of Mesozoic clasts, Zdiar outcrop (Tatra Mts., Slovakia), hand for scale; **H** – the Menilite-type (Zakopane Beds) shales with chert horizons, Zdiar outcrop (Tatra Mts., Slovakia), white arrow points to chert layer, white bar is ~20 cm material was accumulating over the areas of unstable and tectonically active slopes (Fig. 9B, C). During the sedimentary infilling of the Central Carpathian Depression, which in the eastern segments of the Polish Carpathians may be correlated with the Silesian Unit, slump processes dominated in the basin (Fig. 10E, F). The formation of the type S sand injectites is interpreted herein as one of the indicators of the presence of unstable slopes; it was also accompanied by extensional processes occurring on the slope. Shifting of the sedimentary zone depocentre of the Skole Unit also caused tectonic instability of this unit. As a result, injectites are moderately abundant within the Skole Unit. Field observations also indicate that synsedimentary extension and slumping processes occurred frequently in this zone (Figs. 10A–D). These processes were accompanied by the formation of injectites. During the Late Oligocene, the sedimentation area of the Skole Unit was located in the vicinity of a rapidly uplifting forebulge, hence creating favourable conditions for the occurrence of tectonically unstable slopes and processes of synsedimentary deformation, slumping and slides. The analysis of injectite geometry within the Skole Unit shows constant trends related to zones of normal faults pointing to N–S extension. Most observed injectites are related to areas of south-dipping normal faults, a feature that is particularly clearly visible in the outcrops such as Hłudno. This, however, may also be linked with the instability of E–W-oriented slopes pointing to a typically northwards shifting of the depocentre and foreland basin system.

Another process that requires explanation is the tectonic reactivation of the type S clastic injectites occurring within the Skole Unit and their transformation into the type T injectites. Evidence of this reactivation process has been seen in many exposures. The structures observed indicate that the reactivation of initially normal fault zones infilled by injectites occurred during the changing conditions of the strike-slip regime. The analysis of reactivation surfaces reveals sinistral strike-slip faults as the most common features (Fig. 7A). Thus, zones where injectites initially formed resulted from extensional conditions on unstable slopes, although the reactivation process occurred as a consequence of changes in tectonic regime and reactivation of discontinuity zones such as strike-slip zones.

The degree of consolidation of the injectite infills and their spatial orientation relative to bedding indicate that the timing of the formation of the type S injectites coincides with the migration of basin depozones. In contrast, the reactivation of these injectites took place during the secondary, strike-slip stage of the Carpathian deformation (Jankowski and Probulski, 2011). It post-dates the compressional stage, which caused closing of the basin and movement of the depozone, as well as the shift of the foredeep-forebulge-backbulge system of the Fore-Carpathian Basin.

DISCUSSION

The proposed conceptual model of the basin evolution (Fig. 9) applies only to the Polish segment of the Carpathians near the Przemyśl meridian. The model explains the zonal pattern, the basin geometry changes and the shift of sedimentary facies occurring on the shelf and slope during the final Oligocene-Miocene stage of the Carpathian Basin development. According to this model, stages of development of the type S clastic injectites can be correlated to stages of basin infilling by a thick sedimentary succession and to synsedimentary formation of depositional slopes during the Late Oligocene to Early Miocene (Fig. 9C), while tectonic reactivation of the type S injectites into the type T clastic injectites are related to tectonic rearrangement of the orogenic belt during the terminal stages of its evolution, relaxation and development of strike-slip faults.

The migration of the Menilite and Menilite-Krosno systems, with subsequent northeastwards shifting of the forebulge and backbulge zones, led to the rearrangement of the half-graben system inherited after the breakup of the platform margin (Fig. 9A) and the formation of accommodation space of the Carpathian Basin Oligocene-Miocene depositional system (Fig. 9B, C). As mentioned above, the half-graben system geometry was derived from older pre-Alpine tectonic structures present along the margin of the European platforms.

Our model disagrees with some older theories of the evolution of the Carpathians and questions some the previous interpretations, for example those directly linking tectonic units to basinal units (e.g., Oszczypko, 2004). This commonly accepted concept, frequently discussed in the Carpathian literature, of restricting the association of sequences of single tectonic units to areas of separate basins is nowhere reflected in the distribution of sedimentary facies (Jankowski et al., 2012a; Jankowski, 2015a). Thus, rock sequences of so-called tectono-facial units completely do not reflect the pattern of basinal zones within the Oligocene and Early Miocene strata. Instead, thrusting, shearing and closing directions were more crucially important for the composition of the sequences in these tectonic units. Therefore, the pre-Miocene basin pattern and the present-day boundaries of tectonic units show no relationship, as clearly visible in the distribution style of some sedimentary facies, such as the Cergowa Sandstone and the Mszanka Sandstone (Cieszkowski et al., 1990). The facies distribution of the Oligocene Cergowa Sandstone, occurrence of which is traditionally restricted to the Dukla Unit (Wagner, 2008), is not apparent; the distribution of this sandstone was likely originally related to an intrabasinal high that was obliquely sheared during the Miocene evolution of the orogenic belt, and the orientation of this high corresponds to the basin geometry and not to the orientation of tectonic units (Jankowski, 2015a).

The gravimetric data and analysis of facies distributions in both Cretaceous and younger deposits further indicate that transformational fault zones separating the half-grabens were equally significant for the geometrical pattern and number of half-grabens in individual segments of the developing Carpathian Basin. For example, the Gorlice meridian fault zone, that is reflected in gravimetric images, represents a transformational fault responsible for the westward truncation of the Central Carpathian Depression (CCD) (Fig. 2). The analysis of geological evolution of the CCD shows how the changes in tectonic regime, from extensional to compressional, influenced the development of the sub-basin and the types of infilling sedimentary facies. This region represents a tectonic graben formed during the Early Oligocene (Fig. 9B). Throughout the Early Oligocene, the graben was filled by deposits of the Menilite Beds and afterwards, during the Late Oligocene, by syntectonic deposits of the Krosno Beds, mostly through gravitational slumps (Fig. 10E, F) and apron-like clastic deposits, e.g., the Otryt Sandstones. During the Early Oligocene, the CCD functioned as the depocentre of the Carpathian Basin. Interestingly, due to the rejuvenation of the Lower Oligocene structures during the Late Miocene collapse stage (Fig. 2), the graben structure of the CCD has been preserved up to the present day, and manifests itself perfectly in the morphology of the area. Depending on the distance to the CCD, there are some differences in the development of the Menilite Beds, as well of as the Eocene and Cretaceous strata occurring in the area south of the CCD (Fig. 9C), suggesting the presence of two half-grabens. These sedimentary sequences were subsequently incorporated within the Dukla and Magura units (Fig. 9). During the Oligocene, sedimentation became homogeneous over the entire area (Fig. 9B, C). The Menilite Beds of the Dukla Unit display characteristics of a slope cover, e.g., the Grybów Beds, elsewhere referred to as the Dusin Beds (Jankowski et al., 2012a), with debrite-type sedimentation (Fig. 8E) incised by channels infilled with the Cergowa Sandstone (Fig. 8F). The sedimentary character of the Cergowa Sandstone indicates that during its deposition a locally elevated zone must have existed. The occurrence of the Cergowa Sandstone west of the Gorlice zone suggest that this elevated zone extended to the margins of the Carpathians. To the south, deposits of the Menilite Beds, which at some localities overlie deposits of older age, e.g., similar to the Magura Beds (known as the Supra-Magura Beds here), occur in a position indicating that they developed as part of wedge-top sedimentation.

Additionally, the occurrence of the Węglówka zone separating the Skole Unit from the southern part of the CCD and its role during the development of the Przemyśl meridian area is noteworthy. Throughout the sedimentation of the Menilite system during the Early Oligocene, this zone was a tectonic horst, most likely elevated above sea level (Fig. 9B). However, during the Late Oligocene to Early Miocene migration of the depocentre of the Menilite-Krosno system (Fig. 9C) this area subsided temporarily and was covered with sediment.

The absence of the Menilite Beds near Przemyśl, within the Fredropol Horst (Fig. 9B), may be related to the uplift of the Oligocene forebulge zone and the boundary of the older half--graben structures. This zone is also associated with the boundaries between the Skole Unit and the Stebnik and Boryslav-Pokuttya units, and distinguished as a narrow boundary belt (Kotlarczyk, 1988; Jankowski et al., 2004). Farther to the south, this area, that was subsequently incorporated within the Skole Unit, denotes a shallow sea zone related to a remnant sedimentary basin, which can be referred to as the Late Oligocene foreland basin associated with a gradual shifting of the depozone towards the Miocene foreland area.

This theory of the northward extension of the Menilite--Krosno sedimentary system has been introduced into the scientific debate only relatively recently (Jankowski and Margielewski, 2014, 2015; Wysocka et al., 2016) and represents a novel concept that is not yet commonly accepted in the Polish geological literature. As already introduced in the discussion of in the Late Oligocene-Early Miocene stage of the Carpathian orogenic belt evolution (Fig. 9C), the deposition of the Menilite--Krosno facies extented to the Roztocze area, a region located far north into the platform (Myśliwiec and Śmist, 2006; Jankowski and Margielewski, 2015; Wysocka et al., 2016). The Menilite Beds deposited during the Early Oligocene depositional stage (Fig. 9B) were eroded from the area of the present-day Carpathian foreland due to subsequent migration of the uplifted forebulge zone, as were the older sedimentary sequences overlying the Precambrian and Mesozoic rocks in the basement of the Carpathian Foredeep. Remnants of the Menilite Beds and the older Eocene basement have been observed in the inner zone of the Carpathian Foredeep. Here, these deposits formed the basement for the Miocene strata within the foreland basin succession during the Middle and Late Miocene stage of its development (Fig. 9C). They were deformed and, together with the basement of the Menilite Beds and the Eocene strata, incorporated as the Stebnik Unit, as documented in the Jaksmanice borehole.

Furthermore, the proposed model designates the Zakopane Beds of the Central Carpathian Paleogene Basin as part of the Menilite system. It is hypothesized that the evolution of the sedimentary infill of the Central Carpathian Paleogene Basin, which formed on the older accreted fragments of the Mesozoic basement, represents the best fit and started at the beginning of the extensional stage during the Early Oligocene. This process is documented by coarse-grained sedimentary systems, i.e., the Tokaren conglomerates (Janočko and Soták, 2001), characterized by a predominance of clastic materials of Mesozoic age (Figs. 8H and 10G), that are overlain by Menilite--type deposits (Zakopane Beds) in which chert horizons have been recently discovered (Fig. 10H), and finally by Krosno-type deposits (Chochołów Beds). The Menilite-Krosno sedimentary cover, common for the entire Carpathian Basin, and also overlying the Tatra and the Magura elements (Fig. 9B), was shattered during the Oligocene extension, the collapse of the orogenic belt (Jankowski, 2015a) and during the formation of the strike-slip faults (Pieniny flower structure) (Żytko, 1998; Fig. 9C). The remnants of the Eocene and Oligocene sedimentary succession once covering the Cretaceous chaotic complex of the Pieniny Klippen Belt occur in Zaskale in close proximity to the klippen belt (Szydło et al., 2015) as well as in the Udol outcrops (Książkiewicz, 1972). The strata of the Podhale region were partly folded following the Early Oligocene compressional phase (Fig. 9B, C). The present-day Podhale Basin, with its clearly noticeable separation of the sedimentary cover of the Central Carpathian Paleogene Basin due to the elevated morphology of the Tatra Mountains, has resulted from the rapid exhumation of the Tatra Massif over the last several million years (Jankowski, 2015a; Śmigielski et al., 2016).

CONCLUSIONS

The results of detailed geological field investigations, data interpretations and modelling of the Paleogene and Neogene evolution of the Polish segment of Carpathians suggest the following conclusions:

– in the Outer Carpathians, clastic injectites occur commonly within the Menilite Beds of the Silesian and Skole units in Poland, the Vrancea Unit in Romania and the Boryslav--Pokuttya Unit in Ukraine;
– the clastic injectites occur also in the sedimentary succession of the Central Carpathian Paleogene Basin within the Inner Carpathians, although less frequently;
– these injectites typically formed during post-sedimentary processes (type S injectites) with some being subsequently reactivated (type T injectites);
– a two-stage model of the evolution of the Carpathian Basin is proposed; it includes the Early Oligocene extensional stage followed by the Late Oligocene-Miocene compressional and inversion stage;
– the formation of the clastic injectites is interpreted herein as related to the compressional stage, the movement of the foredeep basin and associated migration of slopes of depositional forms (the type S injectites) occurring during the Late Oligocene to Early Miocene interval;
– the formation of the type T injecties is interpreted as a result of tectonic reactivation of the type S injectites that occurred during the final predominantly strike-slip phases of Carpathian orogenic belt formation.

Acknowledgements. The inspiration for this paper was an inspirational scientific discussion with Professor Andrzej Radwański just before his unexpected death. Therefore, authors would like to cordially give thanks for his open-minded and non-conformist ideas, and so dedicate the paper to the memory of Professor Andrzej Radwański. We would like to cordially thank C. Miclaus (Alexandru Ioan Cuza University, Romania) and anonymous reviewer for their constructive comments on the manuscript and scientific discussion. We also wish to thank L. Fodor (Hungarian Academy of Sciences) for his comments on an earlier version of the manuscript. We are also grateful to M. Obermajer for providing language corrections.

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