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SOURCE ROCK PREDICTION VALUE: WORLD PROVINCES DURING LATE JURASSIC–EARLIEST CRETACEOUS TIMES AND POSITION OF WEST CARPATHIANS IN SRPV PREDICTION

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Abstract: Thirty-six Late Jurassic–Early Cretaceous regions were evaluated to obtain the Source Rocks Prediction Value (SRPV). We focused on three major processes, which control the organic richness in a specific paleogeographic, climatic and tectonic setting. These three processes are biologic productivity, background sedimentation rates with non-dilution of organic richness by clastic sedimentation, and preservation of organic matter. A high or increased level of primary biologic productivity supports an increased flux of organic carbon to the sediments of the sea floor. When sedimentation rate increases, especially of fine-grained sediment, the organic matter content of the sediment also increases. Preservation of organic matter depends on domination of anoxic conditions during periods of stagnation of Carpathian basins. The debate over which of the three primary processes is the most important control on the accumulation of organic-rich facies is inconclusive. We assume that the three processes are equally important, and that the balance between them has the overriding control. The amount and richness of organic matter buried in marine sediments then depends on the balance between production and destruction, where the latter includes consumption, decomposition, and dilution. The modeling of the Source Rocks Prediction Value has placed the marginal Tethyan Ocean (Carpathian basin) among the basins, which contain the richest Late Jurassic–Early Cretaceous source rocks in the world. Using the semi-quantitative Delphi method for 36 Late Jurassic regions, which represents a single tectono-depositional province in this time, we evaluated the assessment of SRPV for each of these. The south-Caspian and Central Asia basin was ranked eighth, while the Carpathian basin ninth. The paleogeographic and paleoclimatic settings are indicated as main factors in distribution by basins of known organic-rich rocks. The high organic productivity of the Carpathian basins was caused by upwelling, as well as restricted conditions in the narrow rift basins. The Upper Jurassic organic-rich Mikulov marls representing world-class source rocks (in the southeastern Czech Republic and north-eastern Austria) and Upper Jurassic–lowermost Cretaceous Vendryně Formation rocks were used as local example in analysis of oil source deposits within West Carpathian arc. The average measured Source Potential Index (SPI) for both investigated Upper Jurassic organic rich formations is around 10 and this value fits very well the SPI predicted for Carpathian Upper Jurassic using Source Rocks Prediction Value method.

Key words: source rocks, oil, gas, Jurassic, Cretaceous, Carpathians,

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INTRODUCTION

Understanding the critical processes controlling the accumulation and preservation of organic matter to eventually form source rocks and how these processes vary depending on the specific paleogeographic, tectonic, and climatic setting is the starting point for source rock prediction. Numerous studies have shown that nearly all of the world's hydrocarbons originate from marine source-rock, usually deposited during major periods of eustatic sea level rise (Brooks & Fleet, eds., 1987; Klemme & Ulmishek, 1991; Huc, ed., 1995). The Late Jurassic–Early Cretaceous supersequences (Golonka & Kiessling, 2002) were selected as a test case for this study because they encompass long periods of transgression and regression. Marine source rocks of this age are globally distributed in a variety of geographic settings and have contributed more oil and gas reserves (25% of the world's discovered hydrocarbons) than any other source rock interval (Klemme & Ulmishek, 1991; Klemme, 1994). Although non-marine source rocks of this age locally contribute significant hydrocarbons, their global significance is minor and their prediction is not addressed in this project.

The authors already presented some of the most interesting results of their studies during international conferences (*e.g.*, Golonka *et al.*, 2004). Also, the partial results concerning the position of the Carpathian Oil and gas province were published in Polish (Golonka *et al.*, 2001). These published abstracts and longer papers generated a significant interest among the petroleum geologists. We decided to present the paper containing results of the global Late Jurassic–Early Cretaceous studies of prediction of the source rocks and hydrocarbon potential based on paleogeography and regional geology as well as the practical application toward the prediction of hydrocarbons within the Carpathian realm in Poland and adjacent countries. This studied are also relate to the recently published book describing various aspects of Carpathian Geology and hydrocarbon resources (Golonka & Picha, 2006).

PROCESSES FOR ENHANCED ORGANIC RICHNESS

A variety of global scale maps (e.g., Golonka, 2007) were used to help determine the efficiency of the processes controlling the organic richness in each specific paleogeographic, climatic, and tectonic setting where Upper Jurassic marine source rocks are known or thought to be present. The major processes that operated in these settings were high biologic productivity, non-dilution of organic richness by clastic sedimentation, and preservation of organic matter within its depositional environment (Table 1).

Table 1

Processes for enhanced organic richness

BIOLOGIC PRODUCTIVITY (Nutrient concentrating processes and settings)
A. Enhanced nutrient concentrations 1) Terrestrial input of nutrients 2) Coastal upwelling 3) Open water upwelling
B. Evaporitic settings 1) Silled basins 2) Shelf/platform depressions 3) Rifts on flooded continental platforms 4) Mid and high latitude deserts
C. Restricted geographic configuration
D. Terrestrial kerogen influx
E. High latitude effect (oceanic convergence)
DEPOSITIONAL PRESERVATION (of organic material in depositional environment)
A. Actively subsiding depocenter at time of deposition
B. Maintenance of anoxia 1) Positive water balance (fresh-water influx) 2) Salinity stratification 3) Thermal stratification 4) High productivity 5) Restricted circulation (deep, narrow trough or silled basin)
C. Isolation factor 1) Distance of basin from paleo-shoreline (coastlines, shelves, epeiric seaways) 2) Local uplift deflecting drainage away from basin (rifts)
NON-DILUTION OF SEDIMENTED ORGANIC MATTER (low sedimentation rate)
A. Proximity to orogenic belts during interval of source rock deposition
B. Drainage conduits into depocenter from uplifted areas
C. Rate influence by climatic belts, e.g. wet zones

Much of the organic matter produced in the oceans eventually settles into deeper water. A great part of this material is oxidized during setting, consumed by benthic or planktonic organisms, or undergoes strong degradation in the sediments. A variable, part of the primary production, however is buried and preserved. The relative importance of these processes depends on the level of organic matter production, the depth of the water column, the rate of sedimentation, and the availability of oxidants. These processes ultimately controlling the occurrence of organic-rich facies are discussed in the following chapter.

Biologic productivity

High or increased levels of primary production of organic matter by photosynthesis within single-celled marine algae in the surface waters of the ocean supports an increased flux of organic carbon to the sea floor. This process is invoked as one of the primary controls on the origin of organic-rich sediments.

Primary production of organic matter by planktonic organisms is governed by solar radiation and nutrient supply. Light attenuation restricts photosynthesis to the euphotic zone, which ranges in depth from 100–120 meters in clear, open oceans to only a few meters in turbid and nearshore areas (Szeligiewicz, 1999). The euphotic zone is also limited to only a depth of about 20–35 meters in plankton-rich, stagnant areas like the Azov and Black seas (Özsoy & Ünlüata, 1997; Bakan & Büyükgüngör, 2000; Mc Carthy *et al.*, 2007). The euphotic zone usually has low concentrations of dissolved nutrients, because these are consumed by the phytoplankton. Deeper water, below the euphotic zone, is enriched in nutrients by bacterial degradation of organic debris (fecal pellets and dead organisms) as it sinks to the ocean floor. Sustained primary production can only occur if the nutrient supply into the euphotic zone is maintained. Nutrients can be supplied to the euphotic zone by wind-driven mixing of deeper water, by upwelling of intermediate water beneath areas of surface water divergence, and in coastal areas by lateral inflow of nutrient-rich river waters.

Regional and seasonal variations of the photosynthetic production of organic carbon in the oceans reflect these controls. At high latitudes, low levels of sunlight severely limit production in the winter months. Winter winds and gravity-driven cold and dense current circulation, however, mix the nutrient-rich deeper water with the surface water. Consequently, high production rates usually occur immediately following this mixing period in spring and early summer, when the euphotic zone has expanded and the surface water has been stabilized by solar warming. Nutrient availability can become a limiting factor by late summer. This process is known as the high latitude effect and occur both recently and presumably in the past (*e.g.*, Gnanadesikan *et al.*, 2002; Schumacher & Lazarus, 2004 and references therein). At low latitudes, solar radiation to the water surface is not a limiting factor. At the equator, global wind patterns cause a divergence in ocean surface water currents. This, in turn, causes upwelling, which allows sustained production at a moderate rate throughout the year. Mid-latitudes characteristically have low production rates because the warm surface water is separated from deeper, nutrient-rich water throughout the year by a stable thermocline. The only way for mixing to occur in the mid-latitude open oceans is by brief, non-sustainable storms. High production rates are also common along continental coastlines where trade winds cause coastal upwelling by driving surface water away from the coast.

The origin of many organic-rich rocks has been attributed to upwelling (*e.g.*, Brooks & Fleet, eds., 1987; Huc, ed., 1995; Summerhayes *et al.*, eds., 1995). This is because upwelling zones are rich in dissolved nutrients necessary to sustain high organic productivity. If the organic matter is sedimented in a basin and prevented from oxidation such that not all of it is recycled into the water column, it can then become a source rock. Recent and ancient lithofacies impacted by upwelling contain not only increased amounts of organic carbon but also a biogenic siliceous component (*e.g.*, De Wever & Baudin, 1996; Rais *et al.*, 2007) and phosphorite and glauconite, especially famous phosphorites in Jordan (and "Phosphoria sea") (*e.g.*, Abed *et al.*, 2007, Piper *et al.*, 2007 with references cited therein). The so-called General Circulation Model suggests connection between paleo-upwelling regimes and marine petroleum source rock occurrence (see Huc, ed. 1995; Selwood & Valdes, 2007 with references therein). Nevertheless, the "productivity versus preservation" problem is still matter of hot discussion. First option indicates biologic productivity in the water column as main factor of origin of organic-rich petroleum source rocks (*e.g.*, Huc, ed., 1995; Handoh *et al.*, 2003; Vandenbroucke & Largeau, 2007 with references cited therein) including Late Jurassic case, the second one suggests that accumulation and preservation processes in protection of organic matter in sediments are strictly connected with anoxic bottom-water (*e.g.*, Demaison & Moore, 1980; Brooks & Fleet, eds., 1987; Tyson & Pearson, 1991). Recently, many workers supposed that productivity, preservation and dilution processes are not independent and the accumulation of organic matter in sediments is mixing in origin (Tyson, 2005; Huc *et al.*, 2005; Sellwood & Valdes, 2006; Vandenbroucke & Largeau, 2007). The mechanisms to produce upwelling for the generation of organic-rich rocks are briefly discussed below in decreasing order of importance. The Late Jurassic Mediterranean Tethys has certainly been one of the best places to origin of petroleum source rocks system according to paleogeographic position of this region and tectono-eustatic cycles which produced major organic-rich intervals, including Late Jurassic one (Brooks & Fleet, eds., 1987; Klemme, 1994; Handoh *et al.*, 2003; Golonka, 2007; Vandenbroucke & Largeau, 2007).

Coastal upwelling is caused by steady, prevailing winds parallel to the coast and is known both in the modern and in geological record (Summerhayes *et al.*, eds. 1995; Handoh *et al.*, 2003; Sellwood & Valdes, 2006; Piper *et al.*, 2007 with references cited therein). In the Northern Hemisphere, surface water is driven 90° to the right of the wind direction, and in the southern hemisphere, 90° to the left. When this causes surface water to be driven away from the coast, it is replaced by cooler, more nutrient-rich water brought up vertically from below, usually from the intermediate zone. Coastal upwelling zones are most common today along coastlines with a north-south orientation and are of the highest rates of primary production in the oceans. Latitudinally oriented coastlines may have been more productive in the past when continents and oceans had a more predominant east-west trending component, *i.e.* Tethys in the Late Paleozoic and Mesozoic.

Open water upwelling due to open ocean current divergence is divided into two types, radial (circular) and symmetrical (zonal and equatorial). These are produced in areas of low pressure, where the wind spirals into the area because of the Coriolis Force of the rotating earth and the oceanic surface currents are deflected approximately 90° by Ekman transport. This deflection is to the right in the Northern Hemisphere and to the left in the Southern Hemisphere. The net result is that surface water is moved out of the area and replaced by cool nutrient-rich water from below. Channelled-flow upwelling occurs when the wind is channeled down a narrow seaway and blows the surface water ahead of it to be replaced by deep nutrient-rich water. One side of the channel may experience upwelling and the other side downwelling.

Certain geographic and environmental settings are also known to effectively concentrate nutrients in seawater. Evaporitic settings, areas where evaporation of seawater is greater than fresh-water replacement, typically concentrate ions including nutrients such as potassium, magnesium, and nitrates and often result in salt deposition (Piper *et al.*, 2007). Restricted seas, cut-off from circulation with the open ocean, can concentrate nutrients in two ways: by riverine discharge, where nutrients are continuously added to the system with no mixing to the open ocean, and by internal circulation of nutrient-enriched deeper water to supply the shallow productive zone. The latter method can eventually burn itself out.

Depositional preservation

Destruction of organic matter by oxidation depends on the particle size of the organic debris and the residence time in the water column and upper few meters of the sediment column. The finer the debris and the deeper the water, the longer it is exposed to oxidation as it falls through the water column. However, dissolved oxygen does not easily move into the interior of larger particles which are heavier and sink more rapidly to the sea floor. In both the water column, and in pore waters in the uppermost part of the sediment column, dissolved oxygen is the preferred oxidant. When this is exhausted, bacteria use the oxygen in nitrate and sulphate, and then carbon dioxide. When these are exhausted, degradation of some organic matter can slowly occur by fermentation (methanogenesis) (*e.g.*, Brüchert *et al.*, 2000; Krüger *et al.*, 2005).

Dissolved oxygen is the active oxidant in oxic water and the upper, bioturbated zone of the sediment column. Combined oxygen in nitrate and sulphate is the active oxidant in anoxic water and anoxic sea floor sediment (*e.g.*, Jørgensen *et al.*, 2001). It is commonly assumed that organic matter oxidation occurs much more rapidly in oxic than in anoxic conditions. However, Pedersen and Calvert (1990) have assembled a substantial body of evidence indicating that differences in oxidation rates between oxic and anoxic conditions are small, and both processes are very efficient in the degradation of organic matter. Yet organic matter often escapes destruction in anoxic settings. Several factors help to overcome the natural process of oxidation and promote the preservation of organic matter. These are the formation of non-metabolizable organic matter, occurrence of low temperature/high pressure, presence of a large content of organic matter, and development of anoxic, suboxic, or hypersaline bottom water. These will be briefly discussed in order of least to most important in aiding the predictability of areas where organic-rich facies may have been preserved.

A fraction of the sedimentary organic matter is non-metabolizable by micro-organisms responsible for oxidation. This fraction is formed by selective decomposition and/or abiotic reaction, and seems fairly ubiquitous wherever organic matter is deposited.

The combined effect of low temperature and high pressure, as found in deep-marine basins, significantly slows the rate of metabolism of oxidizing (aerobic) bacteria. For example, at a water depth of one kilometer, where most ocean water is about 4°C or less and pressure is about 100 atm., the metabolic rate of deep-sea bacteria is commonly 100 times lower than the rate at normal atmospheric pressure and temperature of ocean surface waters (Jannasch & Wirsen, 1977; Turley, 2000). However, given enough time, even at slow rates, sediment will eventually be oxidized, unless other processes or burial preserves it.

As the content of organic material in sea floor sediments increases, the rate of degradation decreases. This is true for both oxic and anoxic environments. There may be several possible reasons for this, including that the supply of organic matter is large enough to overwhelm the system (high biologic productivity) both in open ocean condition and restricted sea, respectively. Source rocks are known that have originated in conditions sufficiently oxidic to support burrowing invertebrates (Pedersen & Calvert, 1990). In these cases, the organic matter content is large compared to the inorganic/mineral content of the sediment. These last two factors, temperature/pressure and large amounts of organic material, may especially operate in areas of organic-rich facies formed by upwelling (Brooks & Fleet, eds., 1987; Summerhayes *et al.*, eds. 1992, 1995; Huc, ed., 1995).

The most important factor that operates to preserve organic matter in sediment is the reduction or removal of oxygen from the bottom layers of water on the sea floor. Most major source rocks, with the exception of prodelta shales and some turbidites and upwelling-related source rocks, show evidence of having been deposited in anoxic or suboxic conditions (Brooks & Fleet, eds., 1987; Tyson & Pearson, 1991; Huc, ed., 1995; Meyers, 2006), known also from Carpathian basins (*e.g.*, Golonka & Picha, 2006). Even though oxidation occurs in both oxic and anoxic conditions at similar rates, anoxia at the bottom of the water column can help to preserve organic matter. This is done through restriction (due to lack of oxygen) of deposit feeders, which represent a major catalyst to oxidation efficiency. Another way anoxic pore water aids preservation is that many organic molecules (liquid hydrocarbons, lipids, lignins) are more stable in anaerobic conditions and are resistant to anaerobic degradation (*e.g.*, Meyers, 1997; Vandenbroucke & Largeau, 2007).

Anoxia develops where the demand for oxygen exceeds supply (Demaison & Moore, 1980). When an oxygen demand develops in the bottom water, the oxic-anoxic boundary moves out of the sediment column and into the water column. Such situations occur in the marine environment in tectonically and climatically controlled settings, which prevent the transportation of oxygen-rich surface waters to the sea floor or where the quantity of organic matter exceeds the supply of dissolved oxygen. Typical settings are described below.

1) Topographically restricted basin configurations: silled or enclosed basins. In silled basins, circulation of the water mass generally takes place *horizontally* and *above* the depth of the sill (*e.g.*, Algeo *et al.*, 2007; Piper *et al.*, 2007). As long as *vertical* currents do not develop, the basin below the sill can become anoxic. In land-locked basins, only those with positive water balance tend to become anoxic and develop organic-rich sediments; enclosed basins with negative water balance tend to develop vertical circulation that brings oxygenated water to the sea floor (*e.g.*, Özsoy & Ünlüata, 1997; Bakan & Büyükgüngör, 2000). For example, Oligocene Carpathian basins have been organic-rich land-locked flysch basins, produced oil-bearing deposits (Golonka & Picha, 2006). Also, in restricted seas within arid regions (Red and Mediterranean Seas), evaporation far exceeds river inflow causing such a negative water balance that the resulting circulation often oxygenates the bottom water (Krijgsman, 2002). Anoxic conditions in silled basins

on oceanic shelves also depend upon overall climatic and circulation patterns within the region.

2) Stratification of the water mass: density stratification of the water mass tends to prevent vertical circulation (*e.g.*, Huc, ed., 1995). Density stratification develops when a layer of low-salinity water floats on a layer of high-salinity water, or when a layer of warm water floats on a layer of cold water. This process can operate in any body of water, from enclosed basins to the open ocean. When this stratification is persistent, oxygen-rich surface water does not mix with deeper water and bottom water can become anoxic. Storms can disturb this balance, and horizontal currents within the dense bottom-water layer can laterally transport oxygen-rich water into an area. So when persistent stratification is established in restricted (silled) basins, the combination is extremely favorable for the preservation of organic matter.

Open-ocean oxygen-minimum zones are related to thermal stratification. They occur below the permanent thermocline, at the base of the transition zone between low-density and high-density water (pycnocline). These zones are deficient in oxygen, but not always anoxic. Where they impinge on the sea floor, organic-rich sediments can be preserved (*e.g.*, see reviews by Cornford, 1979; Demaison & Moore, 1980; Tyson, 1987; Huc, ed., 1995; Rogers, 2000; Handoh *et al.*, 2003). Oxygen in this zone is depleted by bacterial degradation of phytoplankton settling from the euphotic zone and by zooplankton consumption. Because oxygen is consumed and not easily renewed (due to density stratification), the water in this zone may become anoxic. The thickness, lateral extent, and oxygen content of this zone depend primarily on the biologic productivity of the surface water, global climatic conditions and the large-scale advection patterns in the deep ocean. Open oceanic anoxic layers are currently present in the oxygen-minimum layers of the eastern tropical Pacific and northwestern Indian Oceans, far from deep, oxygenated polar water sources. A global expansion of the oxygen-minimum zone in ancient oceans has been suggested for extended time periods within the Ordovician Silurian, Devonian, Jurassic and Cretaceous (*e.g.*, Klemme, 1994; Huc, ed., 1995; Rogers, 2000; Golonka & Krobicki, 2001; Handoh *et al.*, 2003; Algeo *et al.*, 2007; Cramer & Saltzman, 2007; Jones *et al.*, 2007; Stein, 2007 with references cited therein). Usually, these correspond to periods of widespread black-shale deposition.

Hypersalinity causes an extreme form of salinity stratification, and develops from dissolution of salt or intense evaporation of sea water. Hypersalinity results in a significant density boundary, which prevents circulation of oxygenated surface-water to the sea bottom. However, there are other preservational advantages in hypersaline water. The solubility of oxygen in brine decreases as salinity increases, and most organisms responsible for organic matter degradation cannot live in highly saturated brines (*cf.* Evans & Kirklund, 1988; Sarg, 2001 with references therein).

3) Anoxia caused by high productivity: This develops below zones of high surface biologic productivity when the oxygen in bottom-water layers is depleted because more organic matter is produced than can be oxidized with the available oxygen supply. Such situations typically develop on the western sides of continents at middle latitudes, such as offshore Peru and Angola, where upwelling is currently strong. Preservation of organic matter in high-productivity areas is often further enhanced where the oxygen-minimum zone impinges on the sediment-water interface.

Non-dilution of organic matter

A critical control on the organic matter content of marine sediments is the *rate* of organic matter accumulation on the ocean floor versus the *rate* of sedimentation of terrigenous and skeletal mineral matter. As described above, the rate of organic matter deposition is mainly controlled by primary production rates and the amount of oxidation that takes place in both the oxic and anoxic water column through which the organic matter must settle. The organic carbon concentration in sediments is then ultimately determined by the amount that the organic matter is diluted by inorganic sediment. For rich source rocks the sedimentation rate of organic matter exceeds that of the mineral matter, which is usually very low.

The type of sediment within which organic material is buried also affects organic carbon richness. Studies (Brooks & Fleet, eds., 1987; Klemme & Ulmishek, 1991; Huc, ed., 1995) have shown that, on average, clay contains about twice as much organic matter as silt, and silt contains about twice as much as sand. Organic matter is more easily adsorbed to fine-grained sediments (clays) than to coarser-grained sediments, because clay minerals, especially montmorillonite, have nearly 1000 times more surface area by weight than quartz and other sand components. Grain size also helps determine how easily sediment is oxidized. It is easier to disperse oxygen into sand than silt, which in turn is more susceptible to oxygen transport than clay. This is because clays, even though they have yield initial porosity, have porosity with greater tortuosity than silts or sands. Oxidizing agents therefore travel slower and have a longer path through finer-grained sediments.

In general, as the rate of deposition of fine-grained sediment increases, the organic matter content of the sediment also increases. This general rule holds when the sedimentation rate is not excessive and follows from the above relationships and because rapid sedimentation decreases the time organic matter is exposed on the sea floor or in the top few meters of the sediment column. This is why pro-delta shales can become source rocks. It is important to remember, though, that while rapid deposition can lead to better preservation, *richness* is still mainly determined by the amount of dilution of the organic material by other sediment.

Several of factors have reinforcing effects, while others may operate antagonistically. For example, silled basins may help restrict circulation to maintain anoxia and have the added benefit of restricting clastic sedimentation, which may dilute organic richness, while drainage into a basin may provide nutrients for increased biologic productivity but also be a source of oxygen and clastics that destroy organic richness. The balance of these factors must be assessed within the overall context of the basin's tectonic and depositional environment.

DELPHI METHOD

The debate over which of the three primary processes described above is the most important control on the accumulation of organic-rich facies is inconclusive. In this project we assume that the three processes are equally important, and that the balance between them has the overriding control. The amount and richness of organic matter buried in marine sediments then depends on the balance between production and destruction, where the latter includes consumption, decomposition, and dilution.

A Delphi approach (Linstone & Turoff, 1975) is a good method for semi-quantitatively assessing three equal but intangible or uncertain variables. The Delphi method is based on achieving a reasonable consensus of expert opinion in a systematic and objective manner by surveying a group of people who have knowledge, experience, and access to detailed information about the phenomena being investigated. In this study, a survey was conducted as to the efficiency of the processes controlling organic richness in specific paleogeographic, climatic, and tectonic settings where Upper Jurassic marine source rocks are known or thought to be present. The three major processes operating in these settings are biologic productivity, depositional preservation, and non-dilution. These were discussed above and shown in Table 1. The overall efficiency of each of these processes, including their sub-processes, was assigned a value between 1 and 5 (5 being most effective) as judged by at least five experts and the results combined into an average value. The experts were from Mobil Exploration and Producing Center; the team included the present authors (Golonka, Bocharova, Edrich, Pauken, Wildharber). The three values, one for each of the major processes, were multiplied together to produce a Source Rock Prediction Value (SRPV, range 1–125).

Thirty-six Late Jurassic regions were evaluated in this. The assessment of SRPV for each of these specific Late Jurassic settings is included in Table 2 and the areas are depicted on Figure 1. Each of these large areas represents a single tectono-depositional province in the Late Jurassic. They each encompass one or more present-day sedimentary basins, which may or may not have similar or related tectonic and stratigraphic development since the Jurassic.

Table 2

Late Jurassic SRPV assessment and predicted SPI

Regions	SRPV	SPI
1. North Sea*	75	15*
2. Gulf of Mexico	72	14.5
3. E. Greenland/N. Atlantic/Barents	72	14.5
4. Saudi Arabia*	70	14*
5. Yemen/Somalia	70	13.9
6. East. Mediterranean	64	11.9
7. Falkland/Weddell Sea	64	11.9
8. S. Caspian/Central Asia	63	11.5
9. Carpathian/N. Black Sea	56	9.8
10. Chukchi/E. Siberia /Beaufort	56	9.8
11. West Siberia/(Kara Sea?)*	55	8*
12. Greater Newfoundland, Galicia	49	7.7
13. Neuquen/Chilean Back-arc	48	7.6
14. Ross Sea	35	4.8
15. Tofino Basin, B.C. Canada	34	4.4
16. West Africa/Guinea, Morocco	31	4
17. North India	26	3.5
18. Bogota Basin	26	3.5
19. Proto-Caribbean	26	3.5
20. SE Africa	26	3.5
21. Papua/NE Australia	26	3.5
22. Apulia	25	3.2
23. S. Iberia/Corsica/Sardinia	25	3.2
24. Tibet/Qiangtang and Lhasa	24.5	3.1
25. East African Seaway	22.5	3
26. NW Shelf Australia*	22	6*
27. NW India	19	2.7
28. Khatanga/Yenisey	18	2.5
29. Greater Sverdrup	18	2.5
30. Ogaden*	16	2*
31. Algeria/Tunisia/Morocco	16	2.1
32. Chilean Fore-arc	14	2
33. Thai/Malay	12	1.9
34. Central Atlantic/E. Coast U.S.	8	1.8
35. Sumatra Borneo	6	1.6
36. Baja Borderland	4.5	1.5

Measured SPI from Demaison and Huizinga (1991)

RANKING REGIONAL SOURCE ROCK PETROLEUM POTENTIAL

The SRPV gives a semi-quantitative assessment of the efficiency of accumulation and preservation of organic matter within a depocenter. It does not actually indicate whether the organic facies will be rich enough to be a good source rock. However, the processes described above and evaluated to determine SRPV also control source rock richness, and source rock richness can be measured. By calibrating SRPV with known, measured values of source rock richness (SPI), the *presence and richness* of potential source rocks can be predicted.

The source potential index (SPI) is a measure of cumulative petroleum potential (Demaison & Huizinga, 1991). It

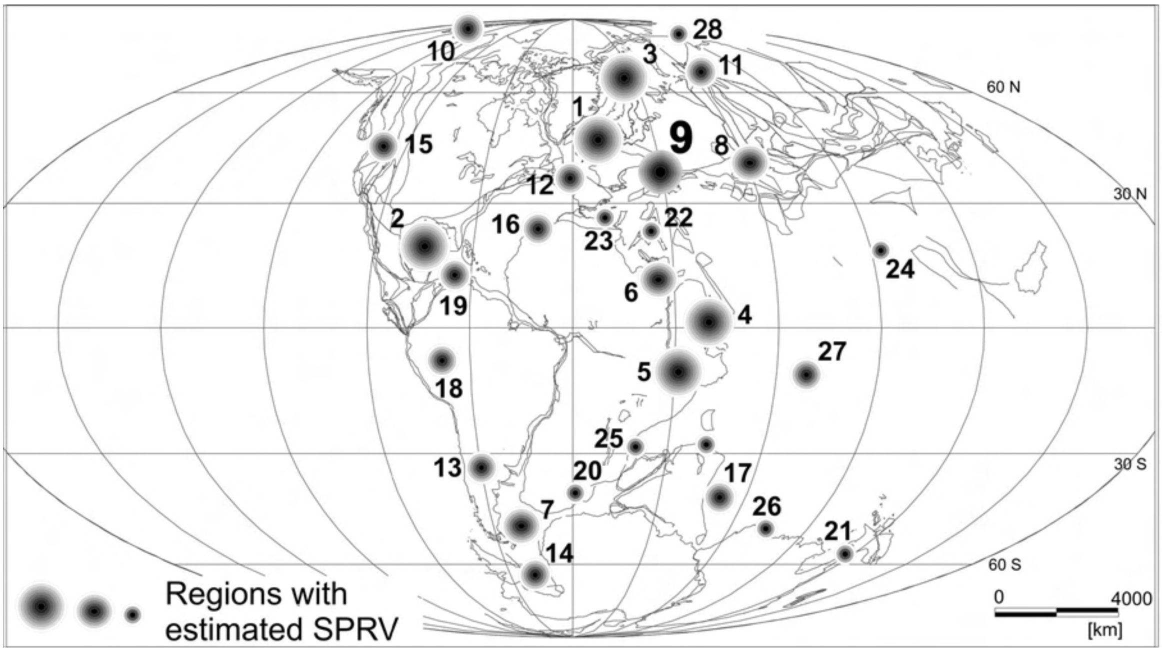

Fig. 1. Regions with evaluated SRPV. Base map – Late Jurassic–Early Cretaceous paleogeographic map (modified from Golonka *et al.*, 2001). 28 highest ranked region have been depicted: 1 – North Sea, 2 – Gulf of Mexico, 3 – East Greenland, N. Atlantic/Barents, 4 – Saudi Arabia, 5 – Yemen/Somalia, 6 – Eastern Mediterranean, 7 – Falklands/Weddell Sea, 8 – South Caspian/Central Asia, 9 – Carpathians, 10 – Chukchi Sea/East Siberia/Beaufort Sea, 11 – West Siberia/Kara Sea, 12 – New Foundland/Galicia, 13 – Neuquen, 14 – Ross Sea, 15 – Tofino (B.C., Canada), 16 – West Africa/Guinea, Morocco, 17 – North India, 18 – Bogota Basin, 19 – Proto-Caribbean, 20 – South-East Africa, 21 – Papua/NE Australia, 22 – Apulia, 23 – South Iberia(Spain)/Corsica/Sardinia, 24 – Tibet/Qiangtang & Lhasa, 25 – East African Seaway, 26 – NW Australian shelf, 27 – North-West India, 28 – Khatanga/Yenisey

is defined as the maximum quantity of hydrocarbons (in metric tons) that can be generated within a column of source rock under 1 m^2 of surface area (Fig. 2). This index combines thickness and richness into a single parameter, which does not distinguish between gas or oil or depend on maturity or source rock type. In Fig. 2, S_1+S_2 is source rock richness or genetic potential, measured from Rock-Eval pyrolysis. The average genetic potential is calculated from systematic sampling of the source rock section.

Average SPI measurements have been published for 6 Late Jurassic marine source rock sequences (Demaison & Huizinga, 1991). These sequences are found in 5 independ-

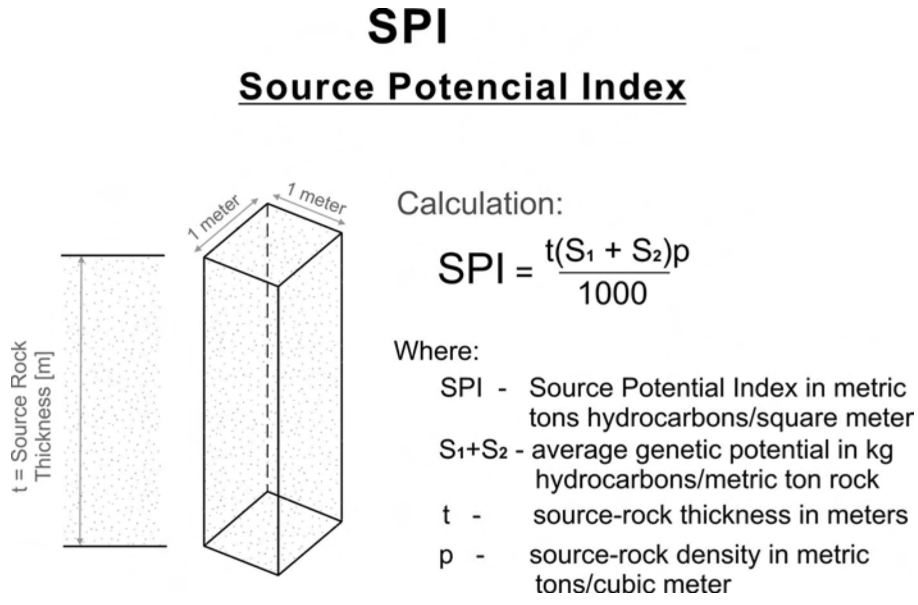

Fig. 2. The source potential index (SPI) – a measure of cumulative petroleum potential (Demaison & Huizinga, 1991)

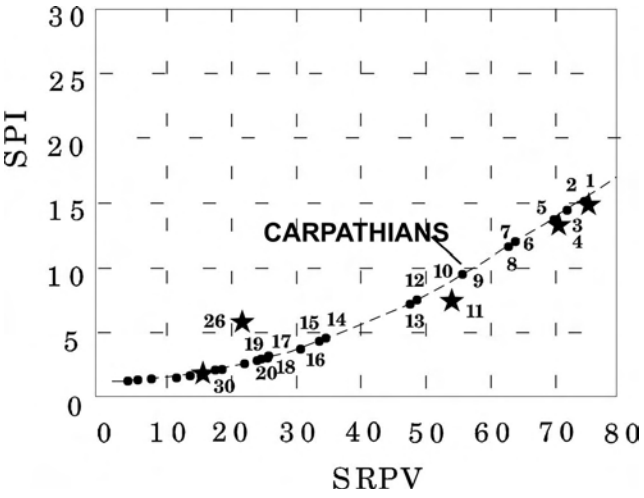

Fig. 3. Graph showing relation between source potential index (SPI) and Source Rock Prediction Value (SRPV) for Upper Jurassic–Lower Cretaceous provinces (modified from Golonka *et al.*, 2001): 1 – North Sea, 2 – Gulf of Mexico, 3 – East Greenland, N. Atlantic/Barents, 4 – Saudi Arabia, 5 – Yemen/Somalia, 6 – Eastern Mediterranean, 7 – Falklands/Weddell Sea, 8 – South Caspian/Central Asia, 9 – Carpathians, 10 – Chukchi Sea/East Siberia/ Beaufort Sea, 11 – West Siberia/Kara Sea, 12 – New Foundland/ Galicia, 13 – Neuquen, 14 – Ross Sea, 15 – Tofino (B.C., Canada), 16 – West Africa/Guinea, Morocco, 17 – North India, 18 – Bogota Basin, 19 – Proto-Caribbean, 20 – South-East Africa, 26 – NW Australian shelf, 30 – Ogaden

ent Late Jurassic tectono-depositional provinces. Two of the SPI measurements came from separate basins on the northwest Australian margin. These basins were part of the same tectono-depositional province in the Late Jurassic; and SPI values were equal. SRPV's were assessed for each of the five independent areas and plotted against the measured SPI values to calibrate the method and generate the curve in Fig. 3. These five values were fit to a second-order polynomial curve with an r^2 of 0.95 (shown as stars in Fig. 3). This graph allowed the calculation of SPI values for the other Late Jurassic depositional settings for which SRPV's were assessed. These regions, which are outlined in Fig. 3, are shown as open circles in Figure 1 and ranked in order of SRPV in Table 2.

When the geochemical data used to calculate SPI are not available, the SRPV method may be used to predict the presence and the quality of source rocks in a region. However the presence of good source rocks is only one of the factors required for petroleum accumulation. The other components of the petroleum system must also be appropriate.

Demaison and Huizinga (1991) published their paper 18 years ago. However, their SPI values are still most actual, since only few papers with measured SPI were later published (*e.g.*, Moretti *et al.*, 1995, Ritts *et al.*, 1999, Schoellkopf & Patterson, 2000, Banerjee *et al.*, 2002) none of them concerning Upper Jurassic source rocks. Therefore, the present authors decided to check and validate method using their own research in the West Carpathians.

POSITION OF WEST CARPATHIANS IN SRPV PREDICTION

The Polish and Slovak West Carpathians form the northern part of the great arc of mountains, which stretch more than 1 300 km from the Vienna Forest to the Iron Gate on the Danube. Traditionally the Carpathians are subdivided into an older range known as the Inner Carpathians and the younger ones, known as the Outer Carpathians. From the point of view of the plate tectonic evolution of the basins the following major elements (Figs 4–6) were distinguished in the Outer Carpathians and the adjacent part of the Inner Carpathians (Golonka *et al.*, 2005, 2006, 2008b, Ślączka *et al.*, 2006):

Inner Carpathian Terrane – continental plate built of the continental crust of Hercynian (Variscan) age and Mesozoic–Cenozoic sedimentary cover. The Inner Carpathians form a prolongation of the Northern Calcareous Alps, and are related to the Apulia plate. The uppermost Paleozoic – Mesozoic continental and shallow marine sedimentary sequences of this plate are folded and thrust into a series of nappes. They are divided into the Tatric, Veporic and Gemeric nappes that are the prolongation of the Lower, Middle and Upper Austroalpine nappes respectively. The nappes and the Hercynian basement are unconformably covered by mid-Eocene/Oligocene flysch and Early/Middle Miocene marine and terrestrial (continental) molasses. The Jurassic rocks within the Inner Carpathian terrane are represented by platform and basinal deposits.

Alpine Tethys (*e.g.*, Golonka *et al.*, 2005, 2006, 2008b) developed as a basin during Jurassic times between Inner Carpathian–Eastern Alpine terrane and North European Platform. In the western part it contains the ophiolitic sequences indicating the truly oceanic crust. In the eastern part the ophiolitic sequences are known only as pebbles in flysch, the basement of the realm was partly formed by the attenuated crust. In Poland, Slovakia and Ukraine, the Alpine Tethys realm is represented by the sedimentary sequences belonging to the Pieniny Klippen Belt and the Magura Unit (Golonka *et al.*, 2006). The Czorsztyn submerged ridge divided the oceanic realm into two basins. The Magura Basin (Fig. 4) constitutes the northwestern part of the Alpine Tethys (Golonka *et al.*, 2008b). The Jurassic, Cretaceous, Paleogene and Miocene sequences of this basin are involved in the allochtonous units covering the North European platform (Magura Nappe). The Jurassic rocks of the Magura Basin are represented by basinal, slope and ridge facies.

North European Platform – large continental plate amalgamated during Precambrian–Paleozoic time. Proterozoic, Vendian (Cadomian), Early Paleozoic (Caledonian), Late Paleozoic (Hercynian) fragments could be distinguished within the folded and metamorphosed basement of this plate. Beneath of the Outer Carpathians the sedimentary cover consist of the autochthonous Upper Paleozoic, Mesozoic and Cenozoic sequences covered by the allochtonous Jurassic–Neogene rocks. The autochthonous Jurassic rocks within North European Platform are represented by mainly platform facies. They are known from bore-holes as well as from the exotic pebbles and olistoliths in the Outer Carpa-

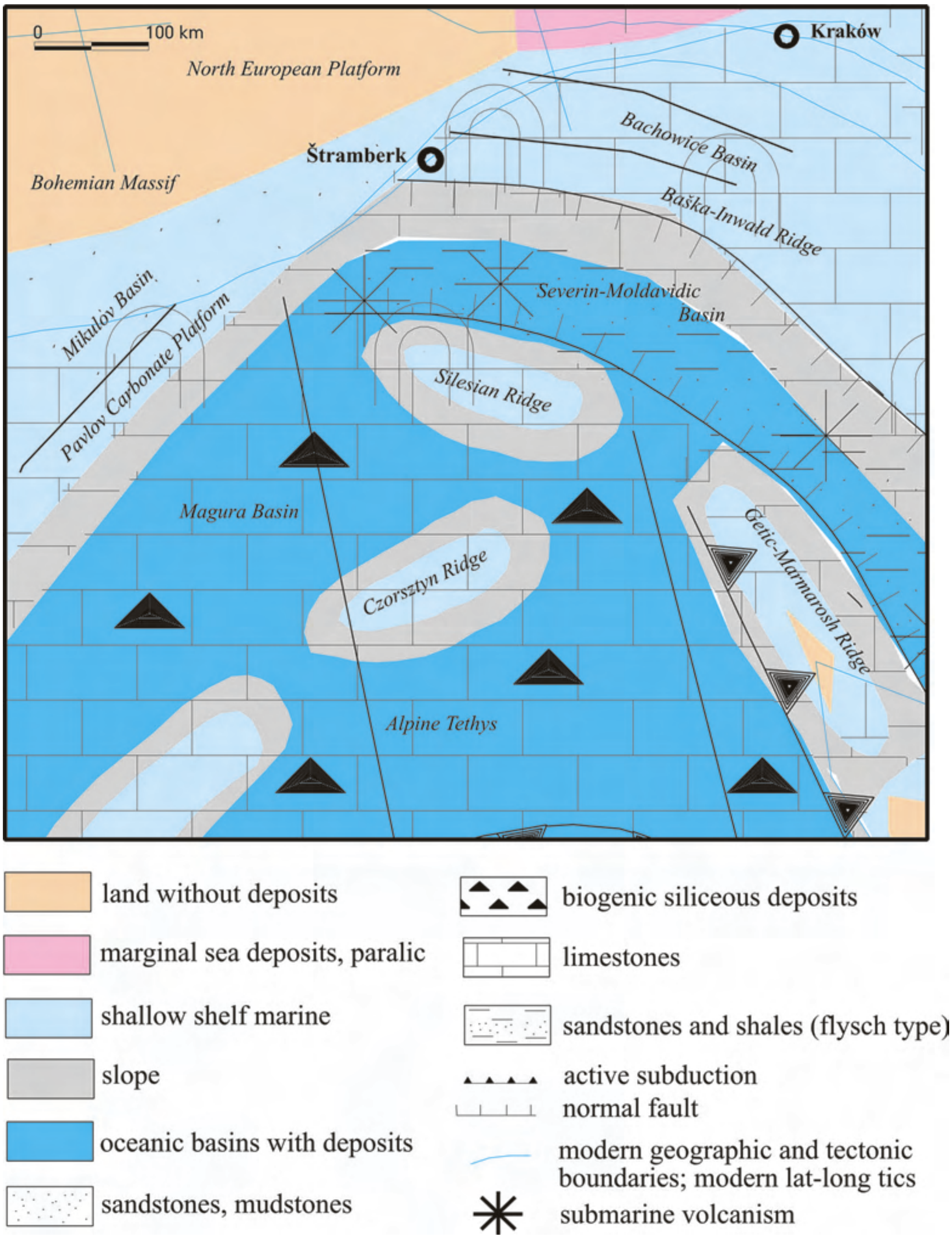

Fig. 4. Paleoenvironment and paleolithofacies maps with main paleogeographical element of the West Carpathians and adjacent areas during the Late Jurassic (from Golonka *et al.* 2008b; modified). Plate position 140 Ma

thian flysch sequences. The southern part of the North European Platform, adjacent to the Alpine Tethys is known as Peritethys. The uplifted and basinal zones were distinguished within the platform. The Baška–Inwałd Ridge and Pavlov Carbonate Platform belong to the uplifted elements, while Bachowice and Mikulov basins represent the basinal zones (Fig. 4). The uplifted zones originated during the platform break-up.

Severin-Moldavidic (proto-Silesian) realm (SM), developed within the North European Platform as rift and/or back-arc basin. The SM basement is represented by the attenuated crust of the North European plate with perhaps incipient oceanic fragments. The sedimentary cover is represented by several sequences of Late Jurassic–Early Miocene age belonging recently to Dukla, Silesian, Subsilesian, Ždanice, Skole tectonic units in Poland and the Czech Republic (Ślączka *et al.*, 2006). The Severin-Moldavidic basinal realm ends in Moravia while slope sequences extend further westwards. The *Silesian Ridge* (Golonka *et al.*, 2006, 2008b), separated the Alpine Tethys and the Severin-Moldavidic realm. The western part of the ridge was reorganized during Late Cretaceous forming the basement of the uplifted Silesian Cordillera. In Poland, Czech Republic and Slovakia, the ridges sedimentary sequences are represented only by pebbles and olistoliths in the flysch of the Pieniny Klippen Belt and Magura Unit. It is known only from

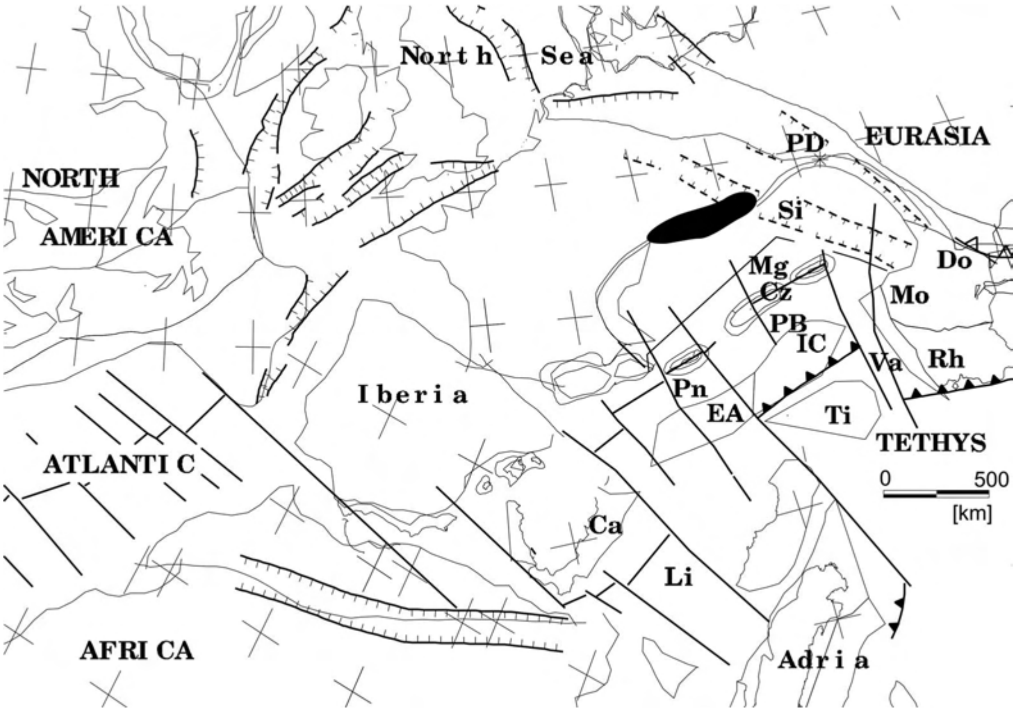

Fig. 5. Paleogeography of Central and Western Europe during Late Jurassic–earliest Cretaceous with the Carpathian source rock area marked in black (modified according to Golonka, 2007). Abbreviations of oceans and plates names:, Ca – Calabria-Campania, Cr – Czorsztyn Ridge, Do – Dobrogea, EA – Eastern Alps, IC – Inner Carpathians, Mg – Magura Basin, Mo – Moesia, PD – Polish-Danish Trough, Pn – Penninic Ocean, PKB – Pieniny Klippen Belt Basin, Rh – Rhodopes, Si – Silesian Basin, Ti – Tisa, Va – Vardar Ocean

exotics and olistoliths occurring within the various allochthonous units of the Outer Carpathians. The shallow-water marine sedimentation prevailed on the Silesian Ridge during Late Jurassic and earliest Cretaceous times. The carbonate material was transported from the ridge toward the Severin-Moldavidic Basin.

Detailed discussion of the plate tectonics of the area and full references are given by Golonka *et al.* (2006). The Jurassic – Early Cretaceous geodynamic evolution of the Circum-Carpathian Tethyan region reflects the plate tectonic history of the Earth during the break-up of Pangea. The Triassic and Jurassic rifting events resulted in the formation of the oceanic type basins along the northern margin of the Tethys Ocean. The Triassic Meliata and Triassic–Jurassic Pieniny Klippen Belt Oceans were formed in the western part of the region (Golonka *et al.*, 2006 and references therein). Tauric and Greater Caucasus-Caspian Oceans were located east of the Moesian Platform. The Central Atlantic was in an advanced drifting stage during the Middle – Late Jurassic. Rifting continued in the North Sea and in the northern Proto-Atlantic. The progressive breakup of Pangea resulted in a system of spreading axes, transform faults, and rift systems. This system connected the ocean spreading in the Central Atlantic and Ligurian-Piedmont Ocean to the opening of the Pieniny Klippen Belt Basin and to the rifting that continued through the Polish-Danish graben to Mid-Norway and the Barents Sea. The Pieniny Klippen Belt Basin was fully opened by the Middle-Late Jurassic (Golonka *et al.*, 2006, 2008b). The basinal parts of northern Tethys are surrounded by several carbonate platforms. These platforms cover, among others, the shallow parts of Adria, Umbria-Marche, Eastern Alpine, Inner Carpathian, and Moesian plates. The spreading in the Ligurian-Pieniny Klippen Belt Ocean continued until the Tithonian. Major plate reorganization took place during the Tithonian (Fig. 5). The Central Atlantic began to propagate into the area between Iberia and the New Fundland shelf (Golonka, 2007). The Ligurian-Pieniny Ocean reached its maximum width and the oceanic spreading stopped. The subduction of the Meliata-Halstatt Ocean and the collision of the Tisa block with the Inner Carpathian terranes were concluded in the latest Jurassic–earliest Cretaceous. During the Tithonian, subduction jumped to the northern margin of the Inner Carpathian terranes and began to consume the Pieniny Klippen Belt Ocean. The Tethyan plate reorganization resulted in extensive gravitational fault movement. Several tectonic horsts and grabens were formed, rejuvenating some older, Eo- and Meso-Cimmerian faults. The Outer Carpathian (Silesian and Magura) basin had developed with extensional volcanism (Golonka *et al.*, 2006). To the west, these troughs extended into the Valais ocean, which entered a seafloor-spreading phase, and further on into the area between Spain and France and to the Biscay Bay. Eastward it was continued into the Eastern Carpathians in Ukraine and Romania (Golonka *et al.*, 2006 and references therein).

Breakup of Pangea during the Jurassic and Early Cretaceous created a system of rifts along the northern Tethyan margin extending from France and Switzerland to Afghanistan (Golonka, 2007). Some of these rifts developed into oceanic basins with underlying oceanic crust. Others developed on attenuated/transitional continental crust and turned into aulacogens. The basins were separated from the main

Tethys Ocean by several plates and ridges like Briançonnais, Czorsztyn Ridge, Moesian and Rhodope plates, Pontides, Armenian and Lut plates. Partial uplift of mainland Europe and the late Cimmerian orogeny resulted in establishment of restricted conditions in the marginal Tethyan basins (Golonka *et al.*, 2004).

The paleogeographic and paleoclimatic setting favored upwelling along the ridges and continental margins (Fig. 6). Firm evidence for upwelling was provided by detailed studies of Berriasian sediments on the Czorsztyn Ridge in Poland (Golonka & Krobicki, 2001).

During Cretaceus–Cenozoic times the sedimentary sequences deposited within the Magura Basin and Severin-Moldavidic basins as well as within the marginal part of the North European platform were uprooted and thrust over the platform at a distance of at least 60–100 km (Golonka *et al.*, 2005, Ślączka *et al.*, 2006). In Poland and Czech Republic these allochthonous, mainly flysch units are being regarded as Outer Flysch Carpathians (Fig. 7). They form a stack of nappes and thrust-sheets arranged in several tectonic units (Figs 8, 9). Along the frontal Carpathian thrust, a narrow zone of folded Miocene deposits was developed.

Source Rock Prediction Value (SRPV) modeling placed Tethyan marginal basins among the best source rocks containing Jurassic basins of the world. South-Caspian and Central Asian Basin took eighth position, while the Carpathian basin ninth position in Table 2. The authors are evaluating now the actual possibility to find new Jurassic/Early Cretaceous source. The high organic productivity in these basins was caused by upwelling (Golonka & Krobicki, 2001), as well as restricted conditions in the narrow rift basins.

According to Golonka and Picha (2006), the Upper Jurassic organic-rich Mikulov marls representing world-class source rock were found in the wells in the southeastern Czech Republic and northeastern Austria. In this area the Jurassic rock lay unconformably on the pre-Mesozoic basement deformed during the Cadomian and Variscan orogenies (Adamek, 2005). The Jurassic sedimentation started in the Middle Jurassic. The fault-block movement differentiated the Late Jurassic paleoenvironment into the shallow and deeper, basinal parts (Figs 4, 10).

Predominantly marly, organic-rich sediments of the Mikulov Formation were deposited in a deeper, more restricted marine environment (Ladwein, 1988; Adamek, 2005; Golonka & Picha, 2006). The basin grew shallower during the Late Jurassic time and deeper water marls passed gradually into the clastic Kurdejov Formation and with the regression into the carbonates of the Ernstbrun Formation.

Today the thickness of the Mikulov Formation exceeds locally 1500 meters. These rocks were indicated as the good source rocks based on a geochemical studies (Ladwein, 1988; Golonka & Picha, 2006). Their TOC value vary between 0.2–10% with an average value of 1.9%. They sourced oils in the Vienna Basin and Carpathian subthrust in Austria and the Czech Republic. They were deposited on the North European Platform but presently are below the tectonic units (nappes) of the Outer Flysch Carpathians (See Mikulov area on Fig. 11, Bulhary and Kobyli boreholes for examples). The genetic potential (S1 + S2 of Rock Eval pyrolysis) of Mikulov marls ranges from 2 to 10 kg of hydrocarbons per Gg of rocks (Krejči *et al.*, 1996). Assuming average density of 2.5 kg per cubic meter (the value used by Demaison & Huizinga, 1991 in their SPI calculations) and

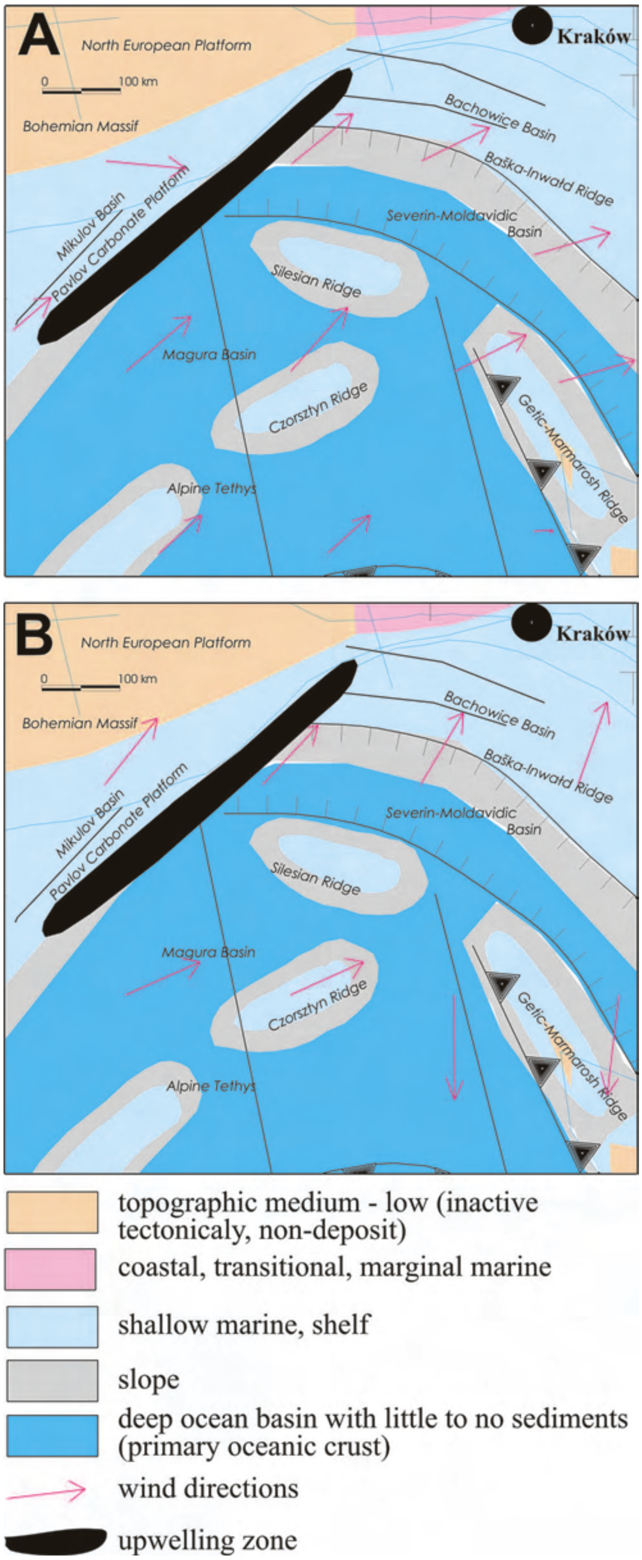

Fig. 6. Paleoenvironment, wind direction and upwelling zones of the Carpathian area during Tithonian–Berriasian. **A.** Winter Northern Hemisphere. **B.** Summer Northern Hemisphere

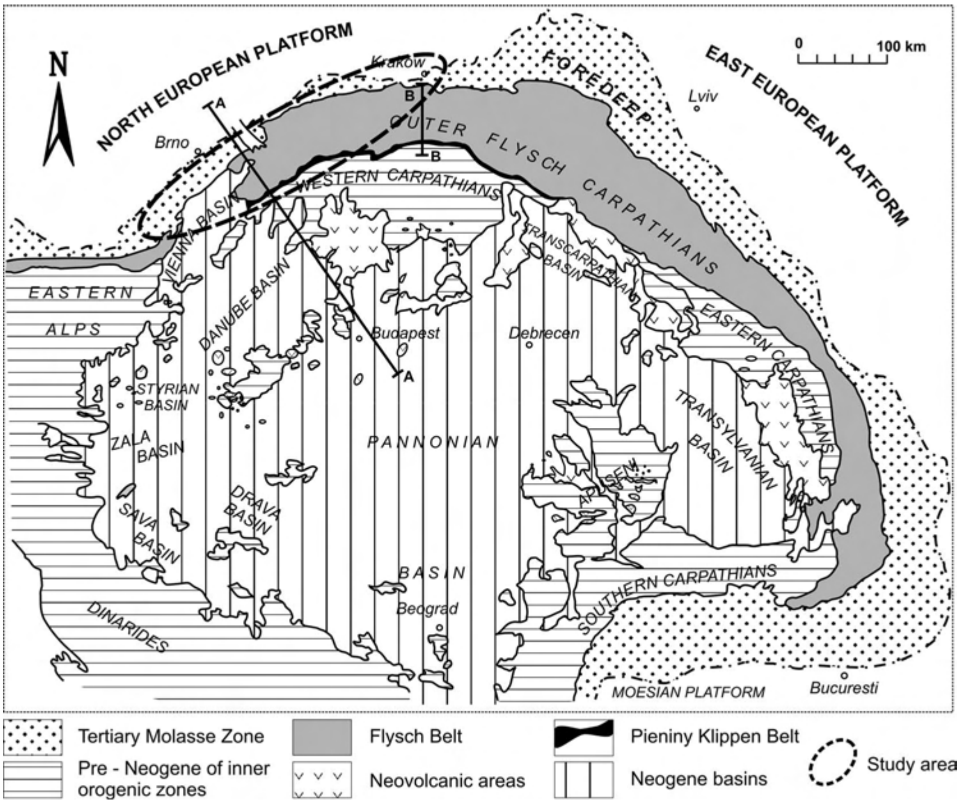

Fig. 7. Tectonic sketch map of the Carpathian area (modified from Golonka and Picha, 2006) with location of cross-sections on Figs 8 and 9.AA and BB – lo

1000 m thickness the calculated SPI is 5 for lowest S1 + S2 value, and 25 for the highest one. The SPI predicted from the SRPV method is around 10. It looks like the predicted SPI fit quite well the actual measured and calculated data. According to Krejči *et al.* (1996), the reactive part of kerogen in Mikulov Marls is type II, while the abundant intertinite makes the bulk hydrogen Index lower, as if the kerogen was type II/III given by Ladwein (1988). According to Picha and Peters (1998), biomarker oil-to-source correlation indicate link of subthrust oils in the Czech Republic, West Carpathians subthrust plays with Upper Jurassic or Paleogene (mainly Menilitic shales) spurce rocks. The Menilitic shales source oils for most deposits in Poland (see geochemical characteristics *e.g.*, Lafargue *et al.*, 1994; Kotarba & Koltun, 2006). Some oils in Poland, especially in Carpathian foreland east of Krakow are perhaps also linked to Jurassic sources.

Organic-rich sediments were also deposited within the Severin-Moldavidic (proto-Silesian). The Vendryně Formation (formerly known as Upper Cieszyn Shales) was deposited during Late Jurassic times (Golonka *et al.*, 2008a, c). This formation occurs in Czech Republic and Poland between Frydek and Bielsko-Biała area (Fig. 11). It is represented by dark-grey, dark-brown or black marly shales often with submarine slumps and with rare intercalations of pelitic or detritic limestones. The formation thickness is 300 meters, age Oxfordian–Tithonian, perhaps up to earliest Berriasian. The genetic potential S1 + S2 (Rock Eval pyrolysis performed by Dr. Irena Matyasik from Oil and Gas Institute in Krakow) of Vendryně Formation was investigated for samples from formation type locality Wędrynia (Czech name Vendryně) in the Czech Republic and from Goleszów and nearby Gumna in Poland (Fig. 11). It range from of 2 to 6 kg of hydrocarbons per Gg of rocks. Assuming average density of 2.5 kg per cubic meter (the value used by Demaison & Huizinga, 1991 in their SPI calculations) SPI is 1.5 for lowest S1 + S2 value, and 7.5 for the highest one. It is below the actual potential of Mikulov Marls and predicted from SRPV calculations. The average actual SPI for both investigated Upper Jurassic organic rich rocks is around 10 like the SPI predicted for Carpathian Upper Jurassic. It validates the SRPV method, at least for this area, where our knowledge about geology and paleogeography is quite good.

CONCLUSIONS

Late Jurassic/Early Cretaceous rocks from platform below thrust belt and from the allochtonous units possibly sourced some Carpathian oils. The SRPV method is a tool to predict regional source rock petroleum potential. A definable relationship between SRPV assessments and measured SPI values was successfully shown for the Late Jurassic–earliest Cretaceous interval. However, other important source rock intervals should be evaluated to further validate

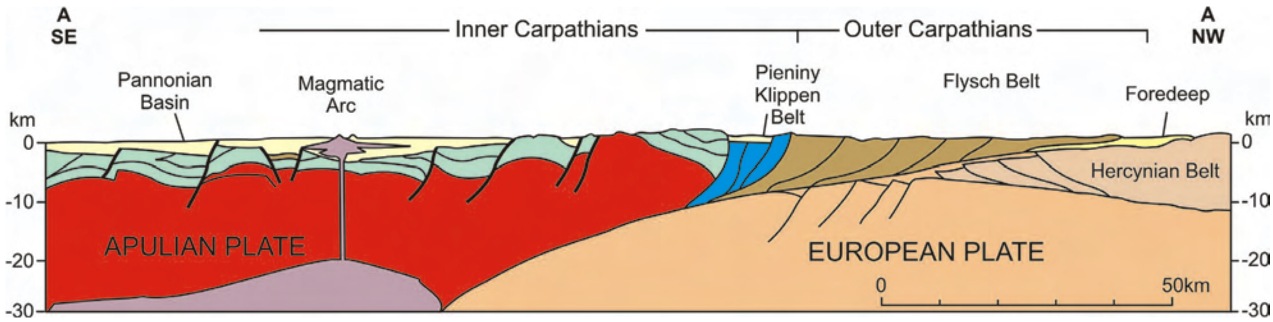

Fig. 8. Generalized cross-section A-A (for location see Fig. 7) across Carpathian-Pannonian region (modified from Golonka & Picha, 2006)

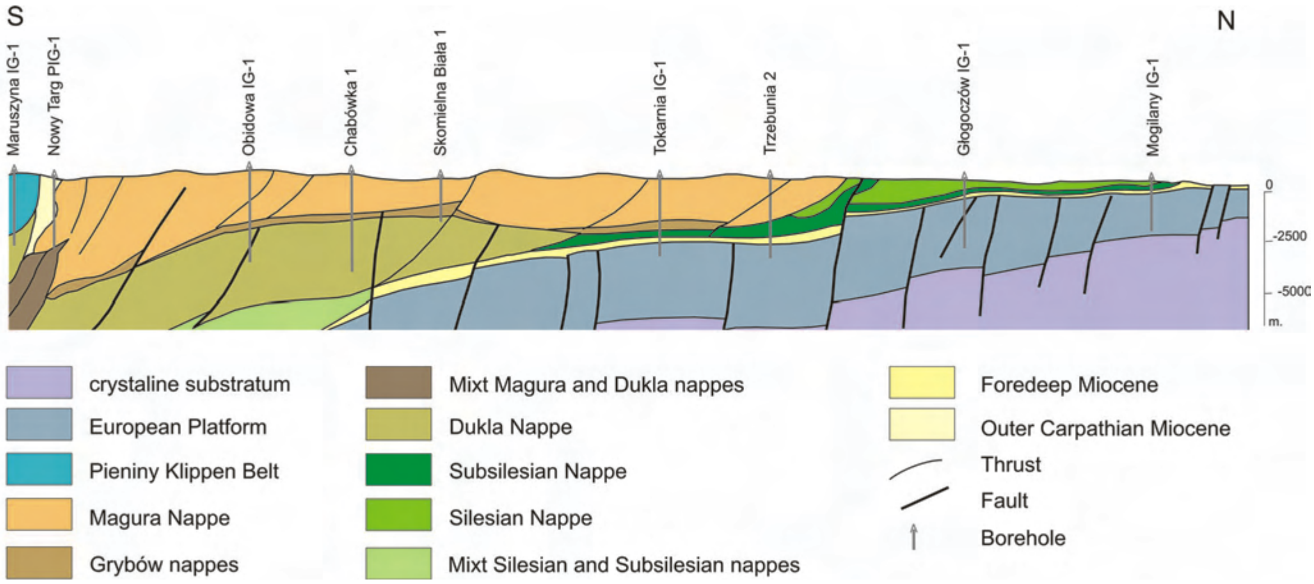

Fig. 9. Generalized cross-section B-B (for location see Figs 7, 11) across Outer Flysch Carpathians in Poland (modified from Golonka *et al.*, 2005)

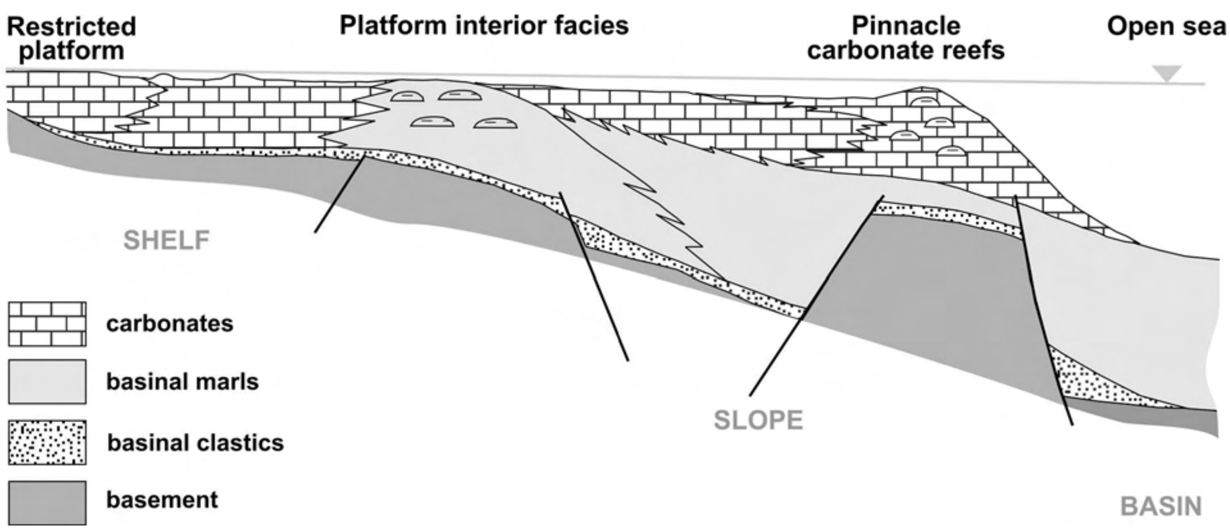

Fig. 10. Distribution of Jurassic facies on the SE margin of the European Platform (after Adámek, 2005; modified)

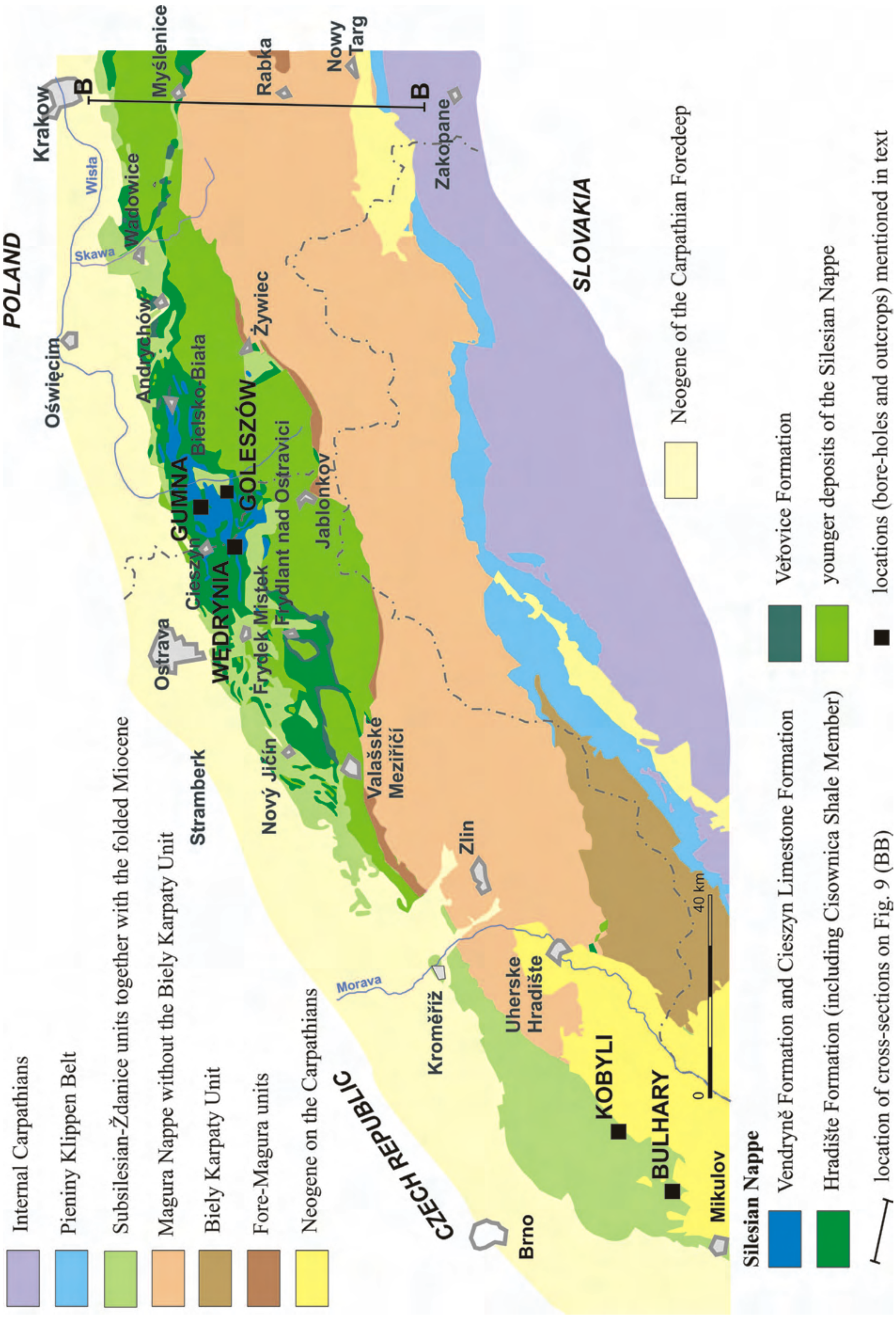

Fig. 11. Schematic geological maps of the Outer Carpathian West of Kraków (modified from Golonka *et al.* 2008c)

and refine this method and determine the extent of its application. Also, much work is still required in the Upper Jurassic areas predicted to have significant under-explored potential to test our hypotheses.

Although SPI is typically determined for a specific exploration well or outcrop section, a single SPI value may represent the basin-wide source rock interval because many source rocks are considered to have a fairly uniform distribution of richness and thickness. The single SRPV, like the single SPI value, attempts to evaluate a wide area as if the lithofacies and depositional environment were uniform throughout. The next step would be to determine the variability of both SPI and SRPV within a region or basin by careful analysis of more local data. Examples include studying thickness changes interpreted from regional seismic, refining or utilizing more appropriate tectono-depositional models, using more detailed paleogeography, and sampling and incorporating more well data. This type of analysis could lead to the prediction of intra-basin source rock distribution and potential and the better understanding of play fairways.

Acknowledgements

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