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**Nd isotope signatures and model ages of Lower Paleozoic
clastic rocks of the Holy Cross Mountains:
a key to understanding the geotectonic history**

Skład izotopowy neodymu i wiek modelowy materiału detrytycznego
skał klastycznych dolnego paleozoiku Gór Świętokrzyskich
kluczem do rekonstrukcji historii geotektonicznej

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STRESZCZENIE

Wstęp

Góry Świętokrzyskie stanowią unikatowy obszar na tle Europy, ponieważ ich geotektoniczny rozwój ma ścisłe powiązania z historią dwóch wielkich paleozoicznych kontynentów, Baltyki i Gondwany. W skład współczesnych Gór Świętokrzyskich wchodzi dwie jednostki litosferyczne: tzw. blok Małopolski oraz blok Łysogór. Jednostki te cechuje odmienna budowa geologiczna skał paleozoiku, polegająca na różnej sekwencji osadów, różnej faunie, różnym czasie pojawiania się luk stratygraficznych oraz na nieco odmiennym stylu tektonicznym. Jednostki Małopolski i Łysogór, które swoim zasięgiem wykraczają poza obszar Gór Świętokrzyskich, interpretowane są od szeregu lat jako terrany, lecz kwestia ich pochodzenia oraz czasu ich amalgamacji nie zostały do końca w pełni wyjaśnione (np. Belka *i in.* 2000; Belka *i in.* 2002; Pharaoh *i in.* 2006; Nawrocki *i in.* 2007; Narkiewicz *i in.* 2011; Żylińska, 2013). Obydwie jednostki wchodzi w skład strefy Szwu Transeuropejskiego (Trans-European Suture Zone, TESZ), najbardziej enigmatycznej części litosfery Europy, który oddziela prekambryjską Platformę Wschodnioeuropejską od waryscyjskich i alpejskich orogénów Europy Zachodniej (np. Pharaoh 1999; Winchester *i in.* 2002; Nawrocki & Poprawa 2006; Janutyte *i in.* 2014). Położenie jednostek Małopolski i Łysogór w obrębie Szwu Transeuropejskiego powoduje, że poznanie ich wczesno-paleozoicznej historii odgrywa kluczową rolę dla zrozumienia ewolucji geotektonicznej europejskiej litosfery, którą od przełomu prekambru i kambru charakteryzuje proces postępującej akrecji. Ważna rola obszaru Gór Świętokrzyskich dla zrozumienia ewolucji Europy polega również na tym, że tutaj występują największe wychodnie skał paleozoicznych w obrębie Szwu Transeuropejskiego. W przeszłości dla rozwoju geotektonicznego Małopolski i Łysogór proponowano różne scenariusze. W jednej ze skrajnych hipotez oba bloki traktowano jako integralne części kontynentu Baltyki, usytuowane w paleozoiku w bezpośrednim sąsiedztwie siebie i w pozycji analogicznej do współczesnego położenia (np. Dadlez, 2001; Jaworowski & Sikorska, 2006). Od ponad dwudziestu lat przeważają jednak poglądy, że bloki Małopolski i Łysogór były egzotycznymi terranami, wykazującymi powinowactwo zarówno względem kontynentu Baltyki, jak i Gondwany (Pożaryski, 1991; Lewandowski, 1993; Belka *i in.* 2000; Belka *i in.* 2002; Winchester *i in.* 2002; Nawrocki and Poprawa 2006; Winchester *i in.* 2006; Nawrocki *i in.* 2007; Narkiewicz *i in.* 2011). W celu odtworzenia pozycji paleogeograficznej Małopolski i Łysogór we wczesnym paleozoiku stosowano różne metody badawcze,

począwszy od badań paleomagnetycznych (np. Lewandowski 1993, Nawrocki 1995, Nawrocki *i in.* 2007), poprzez analizę biogeograficzną fauny (np. Żylińska 2001; 2002; 2013), aż do badań proveniencji ziaren detrytycznych muskowitów i cyrkonów wchodzących w skład skał okrucowych (Belka *i in.* 2000; 2002; Nawrocki *i in.* 2007). W studiach proveniencji nie zajmowano się dotychczas składem osadów drobnoziarnistych i łupków ilastych, które stanowią znaczną część dolnopaleozoicznych sekwencji Gór Świętokrzyskich.

Cele i metody

Głównym celem prowadzonych badań była identyfikacja źródeł alimentacyjnych dla materiału klastycznego dolnopaleozoicznych osadów Gór Świętokrzyskich przy wykorzystaniu składu izotopowego neodymu, samaru i strontu. Badania tego typu prowadzone były po raz pierwszy w Górach Świętokrzyskich. Samar i neodym należą do grupy pierwiastków ziem rzadkich (REE) a ich stosunki izotopowe mogą być używane do badań proveniencji detrytusy. Chociaż metoda znana jest od wielu lat, jej stosowanie nie jest proste i częste, gdyż wymaga bardzo zaawansowanego zaplecza laboratoryjnego. Największą zaletą neodymu i samaru jest ich nieaktywność podczas procesów hipergenicnych, co powoduje, że wietrzenie, transport, sedymentacja i diagenaza nie powodują frakcjonowania tych pierwiastków (np. Goldstein *i in.* 1984). Dlatego też materiał detrytyczny skał klastycznych zachowuje skład izotopowy pierwotnych skał wyjściowych. Od strony praktycznej każda skała klastyczna czy ilasta może zostać poddana badaniom składu izotopowego neodymu i samaru, niezależnie od jej składu mineralnego. W tym aspekcie badania neodymu i samaru mają generalną przewagę nad innymi wskaźnikami proveniencji (detrytyczne łyszczyki i cyrkonony), które nie występują powszechnie we wszystkich skałach klastycznych. Istotną zaletą metody w porównaniu z innymi metodami proveniencji jest również fakt, iż do badań wykorzystuje się bardzo małe, najwyżej kilkuset miligramowe próby. Na podstawie pomiarów izotopowych przy użyciu spektrometru masowego (TIMS) oblicza się dwa główne parametry: epsilon neodymu (ϵNd) oraz tzw. wiek modelowy ($T_{2\text{DM}}$). Pierwszy z tych parametrów zmienia się w skałach skorupy ziemskiej w stosunkowo dużym zakresie od +10 do -30, a jego wartość zależy w głównej mierze od reżimu geotektonicznego obszaru, charakteru skał oraz ich wieku. Wiek modelowy, z kolei, określa długość okresu jaki minął od momentu, gdy materia budująca daną skałę wytopiona została z płaszcza i po raz pierwszy stała się częścią skorupy ziemskiej. Jako informację uzupełniającą w prowadzonych badaniach wykorzystano również izotopy strontu. Do określenia proveniencji materiału

można bowiem również użyć stosunek izotopów $^{87}\text{Sr}/^{86}\text{Sr}$, który w skałach skorupy ziemskiej jest bardzo zmienny, ale wyraźnie zależy od charakteru litologicznego skał.

Podczas przygotowywania niniejszej rozprawy badaniom izotopowym poddano 87 prób skał klastycznych, które pochodziły z 34 odsłoneń osadów kambru, ordowiku i syluru obydwu jednostek Gór Świętokrzyskich, bloku Małopolski i bloku Łysogór. Analityczna część pracy przeprowadzona została samodzielnie przez doktorantkę w Laboratorium Izotopowym UAM w Poznaniu. Obejmowała ona chemiczną separację pierwiastków przy pomocy wymienników jonowych oraz pomiary stosunków izotopowych przy pomocy termojonizującego spektrometru masowego FINNIGAN MAT 261.

Wyniki

System izotopowy Sm-Nd

Przeprowadzone badania pokazały dobitnie, że skład izotopowy neodymu kambryjskich skał klastycznych Łysogór różni się zasadniczo od składu skał kambryjskich w jednostce Małopolski. W południowej części Gór Świętokrzyskich skały charakteryzują się szerokim zakresem wartości epsilon neodymu (ϵ_{Nd}) od -11,5 do -5,5, podczas gdy w Łysogórach wartości ϵ_{Nd} zmieniają się w zakresie od -11,1 do -8,4. Chociaż zakresy te pokrywają się w dużej części, to nie można przeoczyć faktu, iż w jednostce Małopolski dominują wartości powyżej $\epsilon_{\text{Nd}(t)}$ -9,0, zaś w Łysogórach, poza jednym wyjątkiem, wszystkie wartości $\epsilon_{\text{Nd}(t)}$ są mniejsze niż -9,0. Wyraźna odmienność klastycznego materiału kambryjskiego Małopolski i Łysogóry wyrażona jest również różnymi wartościami wieku modelowego. W osadach Małopolski wahają się one od 1,71 do 2,18 Ga, przy czym przeważają wieki młodsze niż 1,9 Ga. W kambrze Łysogór sytuacja jest zgoła odwrotna. Wiek modelowy ($T_{2\text{DM}}$) są starsze niż 1,9 Ga. Dodatkowo, w sukcesji osadowej kambryjskiej jednostki Małopolski zaobserwować można pewien trend, a mianowicie w kambrze dolnym występuje najbardziej radiogeniczny detrytus (z najwyższymi wartościami epsilon neodymu powyżej -7,0), którego wiek modelowy jest młodszy niż 1,8 Ga, natomiast w kambrze środkowym materiał terygeniczny jest mniej radiogeniczny i wykazuje znacznie starsze wieki modelowe.

Dane izotopowe pokazały ponadto, że detrytus w ordowickich skałach klastycznych Małopolski pochodzi z podobnych źródeł co materiał detrytyczny kambryjskiego środkowego. Wartości ϵ_{Nd} wahają się w stosunkowo wąskim przedziale od -8,8 do -5,8, a ich wieki

modelowe oscylują w zakresie od 1,67 do 1,93 Ga. W Łysogórach zbyt mała ilość prób nie pozwala na wyciągnięcie jednoznacznych wniosków.

Skład izotopowy skał sylurskich jest zasadniczo różny od składu skał kambryjskich w obu jednostkach Gór Świętokrzyskich. W Łysogórach wartości ϵNd zmieniają w przedziale od -9,8 do -6,0. Bardzo podobny jest zakres ϵNd w skałach sylurskich Małopolski. Wyjątek stanowią jedynie dwie próby szarogłazów z Barda i Niestachowa, które zawierają znacznie bardziej radiogeniczny detrytus o wartościach ϵNd -4,5 i -4,4 i wiekach modelowych, odpowiednio 1,53 oraz 1,52.

System izotopowy Sr

Wartości stosunku izotopowego strontu ($^{87}Sr/^{86}Sr$) wahają się w skałach kambryjskich Małopolski w bardzo szerokim zakresie od 0,712 do 0,767. Przy tym zaznacza się wielomodalny rozkład tych wartości, wskazujący na pochodzenie z co najmniej trzech różnych źródeł, z charakterystycznym brakiem detrytus o stosunkach od 0,724 do 0,730. W przypadku izotopów strontu występuje wyraźna zależność pomiędzy składem izotopowym $^{87}Sr/^{86}Sr$ a wielkością ziarna. Wartości najmniej radiogeniczne cechują osady średnio- i gruboziarniste, takie jak piaskowce czy kwarcyty, natomiast mułowce mają generalnie bardziej radiogeniczny skład izotopowy. Podobne zależności obserwowane są również w skałach ordowickich i sylurskich. Detrytus w skałach ordowiku wykazuje dużą analogię do skał kambryjskich, cechując się podobnym zakresem zmienności izotopów strontu z głównym maksimum przypadającym na wartości od 0,711 do 0,724. Zupełnie odmienny charakter wykazuje natomiast materiał klastyczny osadów sylurskich o wartościach $^{87}Sr/^{86}Sr$ od 0,720 do 0,798.

Zależność składu izotopowego strontu od frakcji skały jest również zauważalna w skałach Łysogór. Sygnatury izotopowe strontu wahają się tutaj, w porównaniu do skał Małopolski, w zdecydowanie węższym przedziale od 0,722 do 0,754. Natomiast, skały sylurskie charakteryzują się stosunkami izotopów strontu w granicach od 0,715 do 0,749, w których można zaobserwować wyraźnie rozkład bimodalny.

Wnioski

Proweniencja materiału klastycznego

Wartości $\epsilon_{Nd(t)}$ osadów klastycznych zarówno w jednostce Łysogór, jak i w Małopolsce, wskazują na pochodzenie detrytusu wyłącznie z erozji skał skorupy kontynentalnej. Różnice w sygnaturach izotopowych neodymu oraz w wartościach wieków modelowych dowodzą, że w czasie kambru i ordowiku jednostki Małopolski i Łysogór zasilane były materiałem detrytycznym pochodzącym z różnych źródeł. Parametry składu izotopowego osadów kambryjskich Łysogór ($\epsilon_{Nd(t)}$ = od -11,1 do -8,4; T_{2DM} = od 1,90 do 2,12 Ga) wskazują, iż materiał klastyczny dostarczany był z obszaru Fennoskandii, gdzie erodowane były skały kompleksu „svecofeńskiego”. Najbardziej prawdopodobnym obszarem źródłowym była centralna i południowa Szwecja. Drugim potencjalnym obszarem alimentacyjnym mogła być północno-wschodnia Polska, Litwa oraz Białoruś, gdzie wieki modelowe podłoża krystalicznego wahają się w granicach od 1,99 do 2,32 Ga, lecz dominują tam skały o wieku modelowym w zakresie od 2,0 do 2,1 Ga.

W przeciwieństwie do Łysogór, osady klastyczne Małopolski poprzez swoją charakterystykę izotopową ($\epsilon_{Nd(t)}$ = -11,5 do -5,5; T_{2DM} = 1,71 do 2,18 Ga) wykazują wyraźne powinowactwo do źródeł gondwańskich. Jednakże w Małopolsce nie występują skały z wiekami modelowymi młodszymi od 1,7 Ga, które są charakterystyczne dla podłoża „grenwillskiego” w Ameryce Północnej oraz dla kilku kadomskich terranów: Zachodniej Awalonii, Florydy, Karoliny i wschodniej Awalonii. Zakres wieków modelowych skał kambryjskich Małopolski pokrywa się częściowo z wiekami rozpoznanymi w obrębie terranów Saksoturyngii, Tepla-Barandienu i Armoryki, które „zasilane” były z obszaru zachodniej Afryki. Materiał klastyczny pochodzący z kratonu Zachodniej Sahary nie zawiera jednak nigdy cyrkonów o wieku 1-1,7 Ga, które są obecne w skałach klastycznych Małopolski. Biorąc pod uwagę wszystkie dostępne dane (sygnatury izotopowe neodymu, wieki cyrkonów i łuszczaków) powstaje jasny wniosek, że materiał klastyczny, który docierał do jednostki Małopolski w czasie wczesnego do środkowego kambru, pochodził z zachodniej części kratonu Amazonii. Wzdłuż południowej krawędzi tego kratonu zachowała się do dzisiaj sekwencja skał wendyjsko-kambryjskich (w basenie Paragwaju), która ma identyczną charakterystykę izotopową, jak skały kambryjskie Małopolski. W czasie kambru środkowego obok materiału amazońskiego do Małopolski zaczął również docierać pierwszy detrytus z podłoża Bałtyki.

Zasadnicza zmiana proveniencji materiału w oby jednostkach nastąpiła jednak dopiero podczas syluru, kiedy w obu jednostkach obserwujemy bardzo podobny materiał. Miał on bardziej radiogeniczny skład izotopowy niż detrytus w skałach kambryjskich i był mieszaniną materiału z jakiegoś prekambryjskiego kompleksu krystalnego oraz neoproterozoicznego łuku wulkanicznego.

Implikacje paleogeograficzne

Sygnatury izotopowe neodymu kambryjskich skał klastycznych Małopolski i Łysogór stanowią potwierdzenie wcześniej postulowanego scenariusza, że oba bloki krystalne stanowiły niezależne jednostki w trakcie wczesnego paleozoiku. Rekonstrukcja ich paleogeograficznej pozycji nie jest jednak łatwym zadaniem, albowiem nie wszystkie typy danych (paleomagnetyka, wieki detrytycznych cyrkonów i łyszczków, sygnatury izotopowe neodymu, dane biogeograficzne) prowadzą to tych samych konkluzji. Pomimo tego, w kontekście przeprowadzonych badań izotopowych stało się oczywiste, iż jednostka Małopolski w czasie dolnego kambru musiała znajdować się w bezpośrednim sąsiedztwie kratonu Amazonii i jednocześnie blisko zachodniego końca pasa kadomskiego. Co ciekawe, już w czasie środkowego kambru Małopolska musiała zbliżać się sukcesywnie do krawędzi kontynentu Baltyki. Na taki scenariusz wskazuje pojawienie się detrytusu wieku svekofeńskiego w osadach najwyższej części środkowego kambru Małopolski oraz jednoczesny wzrost populacji bałtyckich ramienionogów bezzawiasowych. Końcowy etap dryftu Małopolski nastąpił najprawdopodobniej w późnym kambrze, a jego efektem były deformacje fałdowe, jakich doznały osady kambryjskie i prekambryjskie, które w starszej literaturze opisywane były jako „sandomierska faza orogeniczna”. Miejscem „dokowania” jednostki Małopolski do kontynentu Baltyki musiała być krawędź Teisseyra-Tornquista, w jej północnej, „fennoskandzkiej” części. Ponieważ w osadach dolnego paleozoiku Małopolski nie występuje detrytus wieku archaicznego wykluczone jest praktycznie, aby Małopolska mogła „dokować” do części sarmackiej kontynentu Baltyki, co wcześniej sugerowali Lewandowski (1993) oraz Nawrocki *i in.* (2004, 2007) na podstawie danych paleomagnetycznych.

Skład izotopowy neodymu kambryjskich skał Łysogór implikuje położenie tego bloku przy kontynencie Baltyki i dostawę materiału jedynie ze źródeł „svekofeńskich”. Biorąc jednak pod uwagę proveniencję ziaren detrytycznych muskowitów, które wskazują na obecność

detrytusu z bałtyckich oraz kadomskich źródeł, wydaje się, że Łysogóry musiały znajdować się podczas późnego kambru na przedpolu kontynentu Bałtyki ale równocześnie w niedużej odległości od skorupy kontynentalnej wieku kadomskiego. Na taki scenariusz wskazuje również mieszany, bałtycko-awaloński charakter fauny trylobitowej. Nie wydaje się jednak, aby Łysogóry były już w tym czasie połączone z litosferą Bałtyki. Ostateczna akrecja Łysogór do skorupy kontynentu Baltiki musiała nastąpić później, najprawdopodobniej dopiero w trakcie ordowiku. Współczesne, wzajemne położenie jednostek Łysogór i Małopolski zostało osiągnięte podczas późnego syluru. Wskazuje na to zarówno skład petrograficzny osadów jak i ich charakterystyka izotopowa.

INTRODUCTION

The Holy Cross Mountains, a small area located in the central part of Poland, comprise marginal parts of two continental lithospheric blocks, Łysogóry and Małopolska (Fig. 1), which are characterized by contrasting lithofacies, stratigraphy and tectonic development during early Paleozoic times (e.g., Lewandowski 1993; Belka *et al.* 2000; Dadlez 2001; Narkiewicz 2002). In addition, these crustal units are delimited by deep fractures recognizable in the relief of the Moho. Most often they were interpreted as terranes but their origin and accretionary history is still being discussed (for review see Belka *et al.* 2000; Belka *et al.* 2002; Pharaoh *et al.* 2006; Nawrocki *et al.* 2007; Narkiewicz *et al.* 2011; Żylińska 2013). Małopolska and Łysogóry are situated within the Trans European Suture Zone (TESZ), the most prominent crustal feature of Central Europe, which separates the old Precambrian basement of the East European Platform from the Variscan and Alpine tectonic belts of Western Europe (e.g., Pharaoh 1999; Winchester *et al.* 2002; Nawrocki & Poprawa 2006; Janutyte *et al.* 2014). Therefore, these units are crucial for understanding the geotectonic history of the European crust during the early Paleozoic. Of particular importance is also that the Holy Cross Mountains offer the largest exposure of Paleozoic rocks within the TESZ.

In the past, various scenarios were postulated to explain the individual sedimentary and geotectonic development of Małopolska and Łysogóry. On the one hand, both units were interpreted to be an integral part of the Baltica margin and located in the present-day position during the whole Paleozoic (e.g., Dadlez 2001; Jaworowski & Sikorska 2006), on the other hand, however, the majority of studies showed Małopolska and Łysogóry as individual terranes (Pożaryski 1991; Lewandowski 1993; Belka *et al.* 2000; Belka *et al.* 2002; Winchester *et al.* 2002; Nawrocki & Poprawa 2006; Winchester *et al.* 2006; Nawrocki *et al.* 2007; Narkiewicz *et al.* 2011) displaying linkages to both the Baltica paleocontinent and the Peri-Gondwanan plates. The principal evidence for the geotectonic status, origin and accretionary history of Małopolska and Łysogóry came from the composition of the Cambrian fauna, isotopic provenance data, paleomagnetic studies and from the deep seismic profiling of the crust. In the last years, Cambrian trilobite fauna has been substantially revised (Żylińska & Masiak 2007; Żylińska & Szczepanik 2009; Żylińska 2013). Thus, some earlier Cambrian paleogeographic reconstructions of Małopolska and Łysogóry, which were primarily based on old trilobite data, have become obsolete. Paleomagnetic data for the Cambrian (Nawrocki *et al.* 2007) are still unsatisfactory, as they only allow to indicate a rough paleogeographic



Fig.1. The simplified structural map of the Europe showing crustal units in the Variscan foreland of the southern Poland (from Belka *et al.* 2002, modified).

position of Małopolska in relation to Baltica. The isotopic ages of detrital micas and zircons extracted from lower Paleozoic rocks of Małopolska and Łysogóry (Belka *et al.* 2000; Valverde-Vaquero *et al.* 2000; Belka *et al.* 2002; Nawrocki *et al.* 2007) provided important constraints for identification of sources of the clastic material and led to recognition of linkages of Małopolska and Łysogóry to both Baltic and Gondwanan sources. However, because of sediment recycling and mixing the isotopic provenance data were not sufficiently unequivocal. Moreover, the previous studies were limited to coarse-grained clastics. The K-Ar measurements were performed on multigrain concentrates of micas, so that in many cases the K-Ar cooling ages may represent average ages of mixed mica populations and not ages of particular source areas. The usage of K-Ar method turned out to be limited in the Cambrian rocks of Małopolska because of very low content of detrital muscovites, which did not meet requirements for K-Ar isotope analysis. In consequence, the previous provenance studies have only partly unraveled the origin of Małopolska and Łysogóry. Another general problem is that

almost nothing is known on the provenance of fine-grained clastics and shales which constitute substantial parts of the lower Paleozoic successions in the Holy Cross Mountains and in the neighboring areas. So far, no studies on Sr and Nd isotope compositions of Paleozoic rocks have been carried out in the Holy Cross Mountains.

The present study represents first, systematic Nd and Sr isotope survey of siliciclastic rocks in the Małopolska and Łysogóry Terranes, with the aim to identify the sediment provenance of the lower Paleozoic rocks. The radiogenic Rb-Sr and Sm-Nd isotope systems are now well established as important tools for evaluating the provenance of terrigenous sedimentary rocks. The combined use of Sr and Nd isotopes can be a particularly useful in recognition of major terrestrial reservoirs (Dickin 2005) and provides essential constraints for more robust and comprehensive geotectonic reconstructions. The advantage of both methods is that each type of clastic rocks, including fine-grain material, can be analyzed. This is particularly important because it is generally impossible to determine petrographically the provenance of the fine-grained detritus. Another advantage is that small samples of clastics are sufficient for the analysis and that isotopic measurements can be performed on the whole rock and/or separately for various selected mineral grains extracted from the rock. The potential of Sm-Nd isotope system was already demonstrated in several provenance studies performed for clastic formations worldwide (e.g., DePaolo & Wasserburg 1976; McLennan *et al.* 1990; McDaniel *et al.* 1997; Swain 2005; Zhang *et al.* 2007; Dantas *et al.* 2009). In general, the Sm/Nd and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios of a clastic sedimentary rock, commonly expressed as ϵNd values, reflect the average of the source rocks in its provenance region. In addition, the Sm-Nd system provides estimates of the mean age of mantle extraction for rocks present in the source areas. The parameter known as the *Nd model age* (or crustal residence time) is based on the assumption that the major chemical fractionation of Sm and Nd took place when the source material was derived from depleted mantle (DM), i.e. prior to its incorporation into the crust (DePaolo 1981). The Sm-Nd system appears to remain usually very stable during erosion, sedimentation, high-grade metamorphism and even crustal melting events. Therefore, it can be expected that a systematic Nd isotope survey of the entire detrital material of the lower Paleozoic rocks of the Holy Cross Mountains will certainly allow verification of sources postulated earlier from detrital mica and zircon ages.

GEOLOGICAL BACKGROUND

The Małopolska and Łysogóry Terranes constitute elements of the Trans-European Suture Zone (TESZ). This complex, fundamental domain crosses Europe, from the North Sea through Denmark, northern Germany, Poland, Ukraine to the Black Sea, and reaches a length of more than 2000 km. It constitutes a collage of continental terranes situated between the Teisseyre-Tornquist Line (= the western edge of the East European Craton) and basement blocks underlying the Rhenohercynian and Moravo-Silesian belts (Fig. 1). The term TESZ was created by Berthelsen (1993) to underline the tectonic character of this domain. Because the crystalline basement is very deeply buried under extremely thick sedimentary cover the structural framework and the crustal evolution of the TESZ are still matter of debate (Dadlez *et al.* 1995; Pharaoh 1999; Nawrocki & Poprawa 2006; Narkiewicz *et al.* 2014). As far as known, most of the crustal terranes were amalgamated in early Paleozoic time. Information from the seismic experiments shown that the TESZ is not a steep structure as originally supposed, but it dips southwest around 20°, so that the Baltic crust underplates southwest the TESZ to the Elbe line in Germany. The TESZ is clearly recognizable not only in the deep lithosphere but also in the upper mantle (for overview see Janutyte *et al.* 2014).

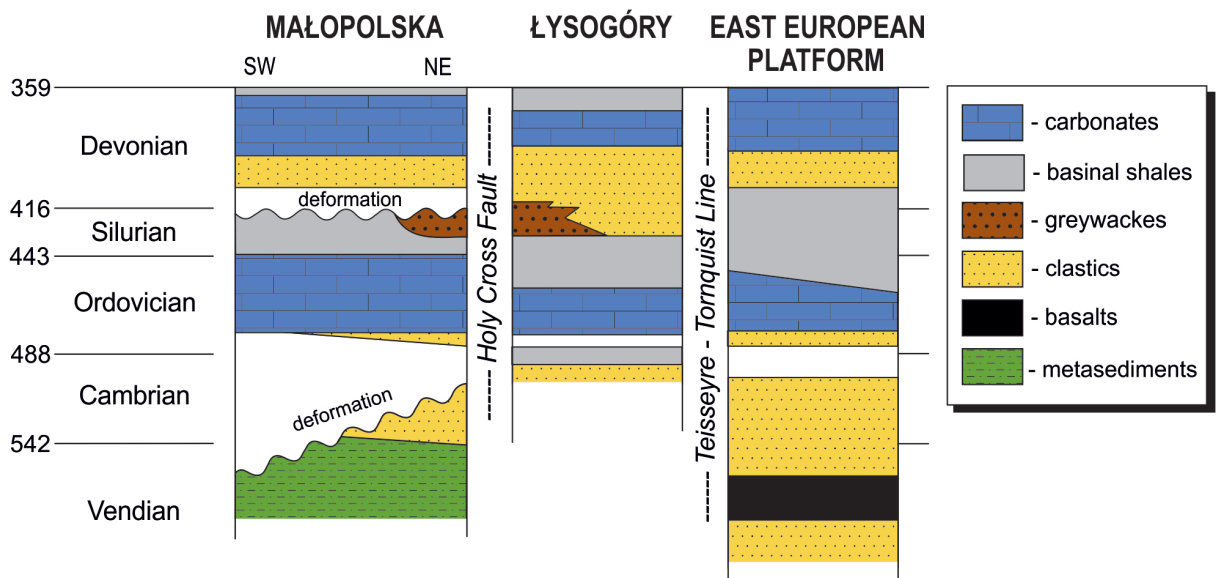


Fig. 2. Stratigraphic columns of the crustal units forming the Variscan foreland in the southern Poland. Note the difference in the occurrence of the stratigraphic gaps (expressed as a white spaces) and deformation phenomena (from Belka *et al.* 2002, modified).

The Moho depth increases across the TESZ from 30 km beneath the Variscan and Alpine Europe to 45 km beneath the craton of the EEP (Guterch & Grad 2006). The northern segment of the TESZ is characterized by a very thick Mesozoic and Cenozoic cover which prevents identification of individual terranes. Hence, the basic geological information (isotopic, paleomagnetic and structural data) from the Precambrian basement and the Paleozoic is restricted to a small number of boreholes only (see Frost 1982; Giese 1997). In contrast, in southern Poland the structure of the TESZ is relatively well understood. Extensive geological studies and hundreds of boreholes allowed to distinguished three fault-bounded crustal blocks (for overview see Narkiewicz *et al.* 2011). These are: starting from the edge of EEP, the Łysogóry Terrane (LT), the Małopolska Terrane (MT) and the Upper Silesian Terrane (Fig. 1). They consist of fragments of continental lithosphere and are bounded by deep fractures which are clearly visible in the relief of the Moho. The units are characterized by distinct tectonic and sedimentary development during the early Paleozoic (Fig. 2). Each of them represents, most likely, are terrane but the precise time of their amalgamation is not well defined. Similarly, their geotectonic provenance and the accretionary scenarios are still under discussion (e.g., Belka *et al.* 2002; Winchester *et al.* 2002). Among hypotheses discussed in the last decade on the origin of Małopolska and Łysogóry, one group assumes their peri-Gondwanan provenance, another one an origin from the Baltic crust (for a recent review see Narkiewicz *et al.* 2014).

Geology of the Łysogóry Terrane (LT)

The Łysogóry Terrane (LT) is a tectonic block composed of a thick continental crust. The Moho depth ranges from 44 km in the west to more than 50 in the east (e.g., Janik *et al.* 2005; Guterch & Grad 2006). The crystalline basement is deeply buried but has never been reached by boreholes, and thus, nothing is known about its age and structure. The term Łysogóry refers to a range of hills in the northern part of the Holy Cross Mountains where the oldest part of the Paleozoic succession of this unit is exposed. This is why the Łysogóry Unit is sometimes termed in the literature as the Northern Holy Cross Mountains. To the south, this tectonostratigraphic unit is separated from the MT by the WNW-ESE running Holy Cross Dislocation (Fig. 3), a fault zone recognizable in the Moho topography, which can be also traced in the field throughout the Holly Cross Mountains, but its route east of Opatów is unclear (Belka *et al.* 2002; Szczepanik *et al.* 2004). To the NW, the LT is delimited by the

Grójec Fault Zone (e.g., Winchester *et al.* 2002; Narkiewicz *et al.* 2011). Recent geophysical data (Narkiewicz *et al.* 2014) appear to suggest that this terrane may also include the Biłgoraj-Narol Block, which is bounded to the NE by the Teisseyre-Tornquist Line (TTL). If so, the LT is in contact with the crust of the East European Platform. The Paleozoic succession of Łysogóry exposed in the Holy Cross Mountains is estimated to be about 4600-5000 m thick and includes deposits ranging in age from Middle Cambrian to Late Devonian (Fig. 2). It starts with Middle Cambrian to early Ordovician clastics deposited in shallow-water environments. The Ordovician and much of the Silurian are represented by basinal graptolite shales. A very prominent part of the succession constitutes the Upper Ludlow to Pridoli, which is composed of fine-grained greywackes with interbeds of volcanoclastic rocks. The Lower Devonian is represented alluvial and marginal marine clastics which grade upwards into the Eifelian and Givetian shallow-water carbonate sequence (Szulczewski 1995). The Upper Devonian is poorly documented and consists predominantly of basinal shales with some carbonate gravity flow deposits. The succession investigated during the present study comprised the Cambrian to Silurian rocks. It is described in greater detail below.

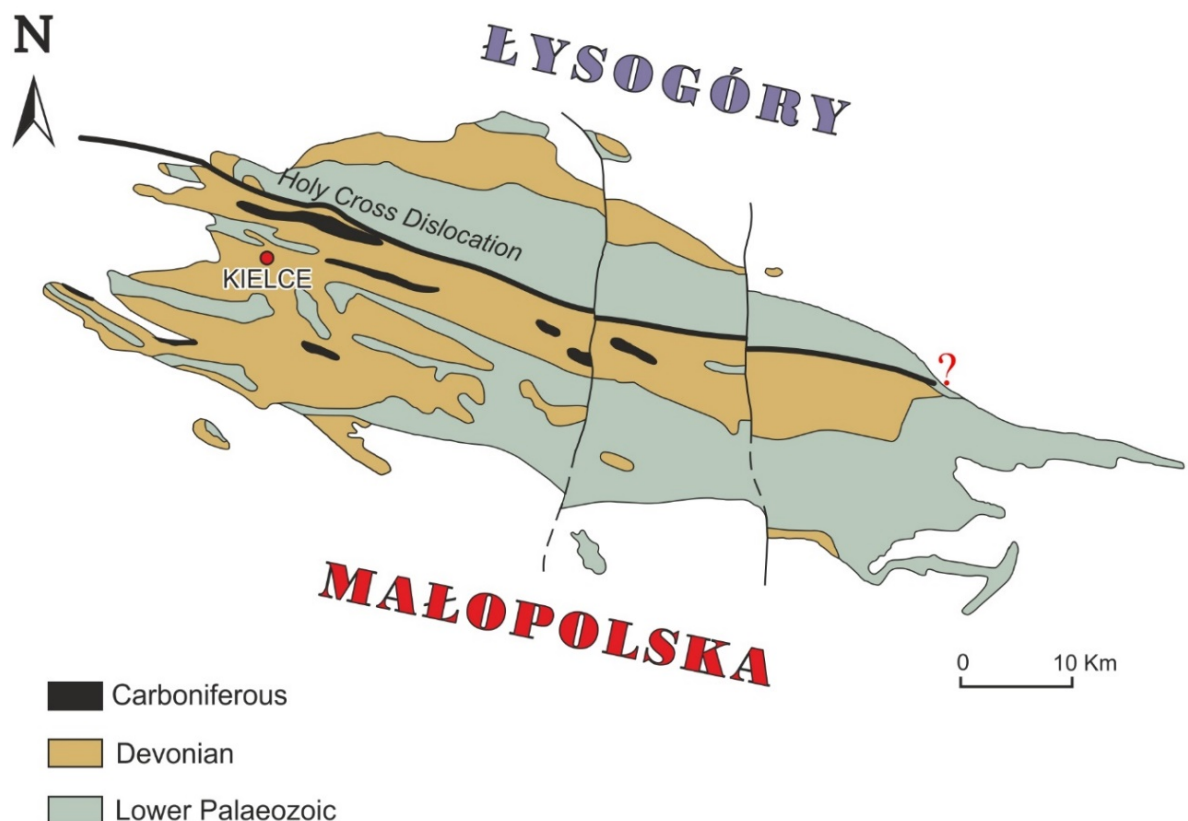


Fig. 3. Geological sketch-map of the Paleozoic rocks of the Holy Cross Mountains.

Cambrian

The Cambrian lithostratigraphic framework of Łysogóry consists of three main units: the Pieprzowe Mountains Formation, the Wiśniówka Formation and the Klonówka Formation which ranges into the lower Tremadocian (Fig. 4). The Pieprzowe Mountains Formation was introduced by Orłowski (1975). It is important to mention that this formation is the only one Cambrian unit that has been established both in Łysogóry and in Małopolska. Today, however, all occurrences of the formation are considered to belong to Łysogóry (e.g., Kowalczewski *et al.* 2006). Because of complex tectonic deformation the total thickness of the formation is not well recognized. Orłowski (1975, 1992) and Orłowski and Mizerski (1995) have estimated it as being about 400 m.

The Pieprzowe Mts. Formation is composed, first of all, of claystones, mudstones and sandstone-mudstone heteroliths. There are also some few conglomerate horizons. According to Przewłocki (2000), these sediments were deposited within a shallow-water shelf sea, between fair-weather and storm wave base and/or below the storm wave base. Sedimentary features attest a lower flow regime during the deposition of sandstone-mudstone heteroliths, unlike to sandstones which accumulated under an upper flow regime. A different interpretation provided Jaworowski and Sikorska (2006) suggesting that sandstones represent shallow-water deposits influenced by tidal currents and/or storms. Sand material in the heteroliths, however, was accumulated during storms, whereas fine-grained material (mudstones and claystones) deposited from suspension (Fig. 5).

The rocks of the Wiśniówka Sandstone Formation occur across the entire Łysogóry range, from Miedziana Góra in the west to Wąworków in the east, and they build the highest mountain chains (Masłowskie, Łysogórskie, Jeleniowskie). Nevertheless, natural outcrops of this formation are very rare. This is why our knowledge of the Wiśniówka Sandstone Formation is based on data collected in quarries at Wąworków and Wiśniówka. The former one, however, is abandoned for two decades, the latter is still in operation and is world-famous due to abundant trace fossils formed mainly by trilobites (e.g., Radwański & Roniewicz 1963; Orłowski *et al.* 1970; Seilacher 1970). There is a significant discrepancy in views as regards the thickness of the formation. Kowalczewski *et al.* (1986) recognized a strong tectonic folding and described the formation as being between 80 and 200 m thick. In contrary, Orłowski & Mizerski (1995), who favoured a monoclinical structure of the Cambrian succession in Łysogóry, have estimated it as being about 1400 m.

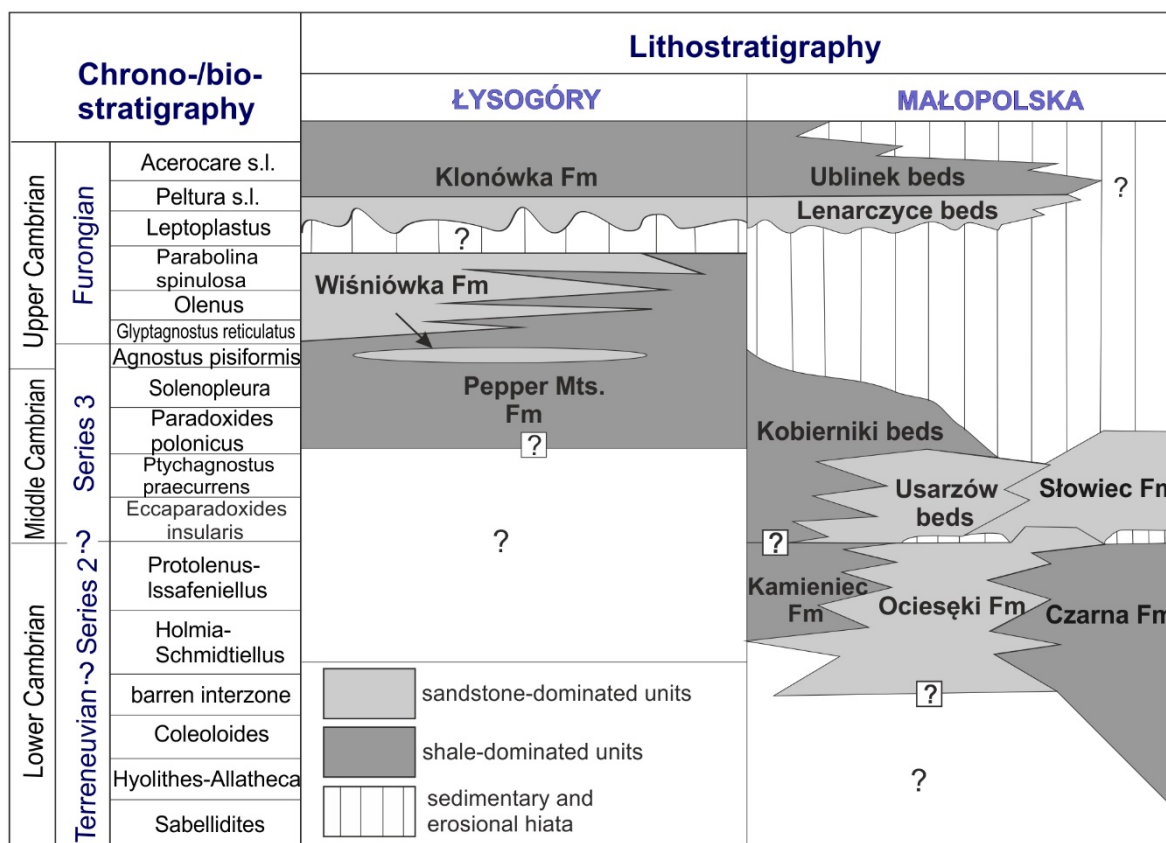


Fig. 4. Lithostratigraphic scheme of the Cambrian of Łysogóry and Małopolska (from Żylińska & Szczepanik 2009, modified). Compilation based on data of Orłowski (1975b, 1992), Kowalczewski (1990), Szczepanik *et al.* (2004 a,b), Żylińska (2002) and Kowalczewski *et al.* (2006). Chronostratigraphic subdivision after Babcock *et al.* (2005), Zhu *et al.* (2006), Babcock & Peng (2007).

Acritarchs and trilobite fauna provide evidence for the Late Cambrian age of the entire Wiśniówka Formation (Żylińska 2002). The formation is subdivided in two lithostratigraphic units: the lower one is termed the Łysogóry Quartzites and of the upper one is called the Wąworków Sandstones (for more details see: Kowalczewski 1995, 2000). Quartzite unit is composed of fine- to medium-grained quartzitic sandstones with rapidly alternating thin siltstone and claystone intercalations. The sandstone beds display a mature composition and are generally relatively pure quartz arenites. Heavy minerals, however, are present in large quantities. Among these, zircon and tourmaline are especially very frequent. Detrital muscovites, are very rich covering the upper surface of layers in the entire sequence. Sandstone bed thickness is from some centimeters up to 3 m and the thin beds clearly predominate. Thick beds are massive and devoid of any internal structure or grading as a rule. In contrast, very thin sandstone layers display often extensive bioturbations. A variety of sedi-

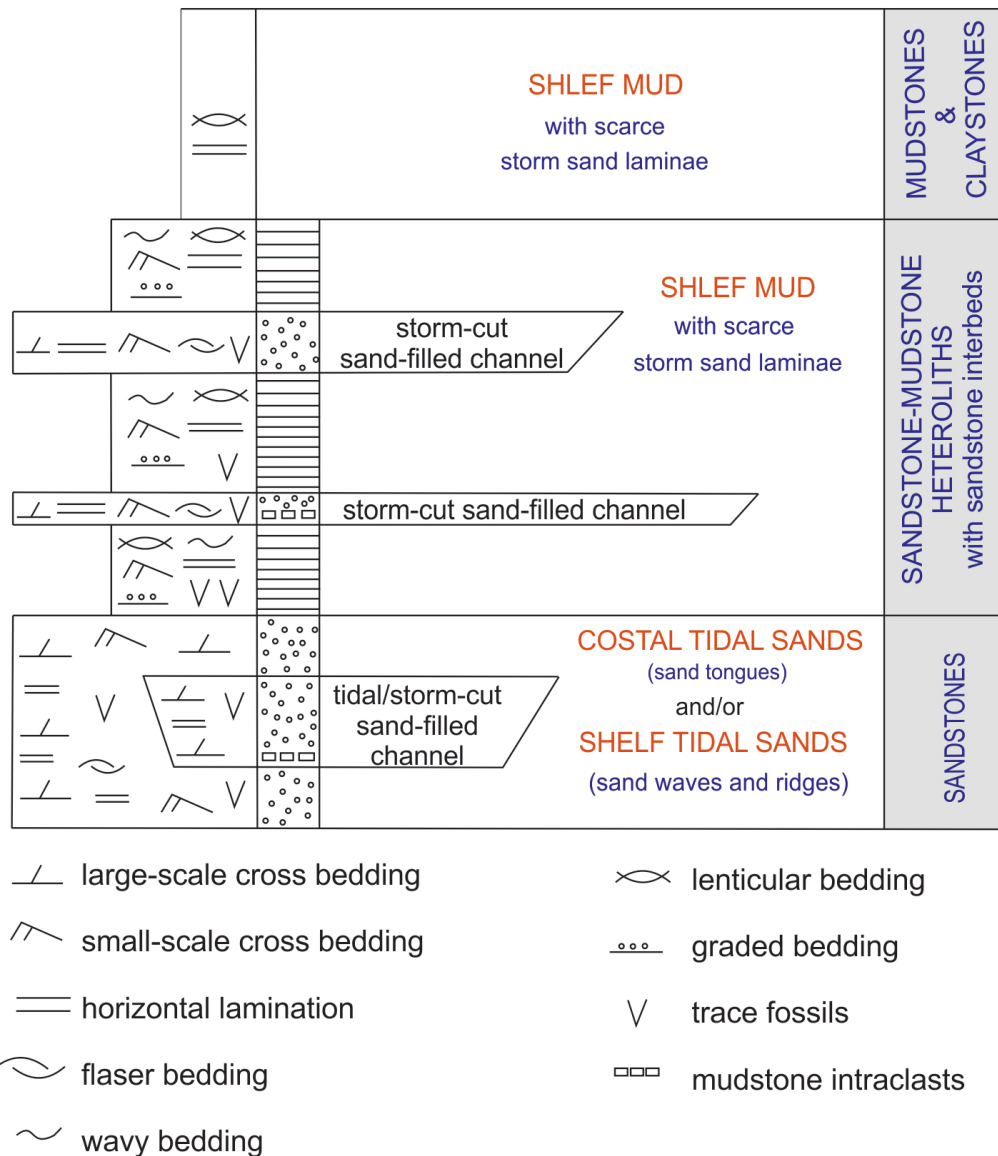


Fig. 5. Facies model for Cambrian deposits of Łysogóry and of the East European Platform (from Jaworski & Sikorska 2006, modified).

mentary structures occur abundantly both on the lower and upper surfaces of sandstones (Dżułyński & Żak 1960). Among these almost all kinds of current and wave ripples are the most frequent. Load structures, ripple-load-casts and gas bubble impressions are another distinct components. The upper part of the sequence, exposed in the road cuts leading to the quarry, consists predominantly of silty shales, in which illite is the chief constituent, while chlorite and kaolinite occur in minor quantities; sandstone beds are subordinate. Kowalczewski *et al.* (1986) reported thin bentonite and tuffite interbeds occurring in the shales. The most spectacular features on the Łysogóry Quartzites are biogenic structures formed by trilobites (e.g., Radwański & Roniewicz 1963; Orłowski *et al.* 1970). According to

Seilacher (2007 and pers. com.), the assemblage of trilobite traces is identical to those distributed throughout Gondwana and the Peri-Gondwanan microplates, but it is unknown from the paleocontinent of Baltica. Although trilobite traces are common, trilobite remains, however, are extremely rare. The Łysogóry Quartzites demonstrate many features that are characteristic of current and wave influenced shallow shoreface. Radwański and Roniewicz (1960) postulated a depth ranging from a score or so to some tens of meters. The shallow-water conditions can be inferred from the presence of the symmetrical and asymmetrical oscillation ripples, as well as from linguoid ripples associated by diagonal bedding. The dominant current direction was from SW to NE (Dżułyński & Żak 1960). Hence, the source area must have been situated somewhere to the south of the present-day position of the Wiśniówka Quarry. The upper unit, the Wąworków Sandstones, is approximately 300 m in thickness. It consists of quartzitic sandstones, mudstones, and claystones with sandstone-mudstone heteroliths containing interbeds of pyroclastics rocks (Kowalczewski 1976). According to Radwański and Roniewicz (1962), these sediments were deposited in the same environments as sediments of the Łysogóry Quartzites and the clastic material was also derived from southern sources.

The Klonówka Shale Formation (Fig. 4) was established and described by Orłowski (1975). The lowest part of this lithostratigraphic unit crops out in the Wiśniówka Duża Quarry, the middle part is exposed to near the Wiśniówka village, whilst the upper part has been recognized in several boreholes only (Tomczykowa 1968). According to Orłowski (1975), the thickness of the formation attains approximately 400 m and stratigraphically, it ranges from the uppermost Cambrian to lower Tremadocian. The unit is composed of sandstone-mudstone heteroliths which reveal a dominance of sandstone intercalations in its lower part (Orłowski 1968; Tomczykowa 1968; Kowalczewski 2000), whereas mudstone and claystones are more frequent in the upper. Sedimentological structures point to deposition within shelf environments (Kowalczewski 2000; Zylińska 2002), more distant from shoreline than the clastics of the Pieprzowe Mountains Formation.

Ordovician

The Ordovician rocks of Łysogóry are known mainly from boreholes. Currently, there are only two outcrops where small fragments of the Ordovician succession are accessible. The sequence is represented by monotonous shale series, sandstones and carbonates (for overview

see Trela 2006a, 2009). The maximum thickness is estimated to be about 230 m, but locally the sequence is remarkably condensed. The lowest Ordovician (Brzezinki Claystone Formation), which occurs in the sedimentary continuity with the underlying Upper Cambrian shales, is limited to a few meters of grey mudstones with claystone and sandstone interbeds (Trela 2006a). The age of this marine sequence has been defined by the presence of dendroid graptolites *Rhabdinopora*. The Middle Ordovician is represented by grey to greenish iron-rich limestones intercalated with limestones rich in numerous ooids, pebbles, intraclasts, quartz grains as well as fragments of various fossils (Tomczyk & Turnau-Morawska 1967). This condensed deposits (Pobroszyn Formation), 0.4 to 2.8 m in thickness only, are separated from the un- and overlying deposits by distinct discontinuities which reflect stratigraphic gaps. According to Trela (2008a), the sideritic carbonates accumulated in the shallow-water in an open-marine environment during a sea-level highstand. Their erosional upper surface, however, resulted from the emersion or non-deposition (Trela 2009). In consequence of the subsequent late Darriwilian transgression carbonate/shale succession rich in phosphorites (Bukowiany Formation) accumulated. The occurrence of phosphatized stromatolite mats attests an important role of benthic microbial communities in redistribution and concentration of phosphate (Trela 2008b). The Upper Ordovician is represented by monotonous series of dark mudrocks (Jeleniów and Wólka Formations) which are up to 190 m thick. The claystones of the Jeleniów Formation record the long-term dysoxic conditions lasting throughout the entire Caradoc (Trela 2007). An increase of the benthic oxygenation in the early Ashgill allowed intense bioturbation which is now well-expressed in the abundant ichnofabric of mudstones of the Wólka Formation. The top of the Ordovician sequence of Łysogóry is delineated by coarse-grained clastic deposits (Zalesie Formation), up to 6 m in thickness, which likely correspond to the late Ashgill glacio-eustatic sea-level drop (Trela 2007).

Silurian

The Silurian of Łysogóry is represented by predominantly fine-grained clastic deposits which attain a maximum thickness of about 2400 m and do not yield any significant stratigraphic gaps. The lithostratigraphy and facies pattern is very complex and this is why different lithostratigraphic schemes were proposed in the past (for review see Kozłowski 2008). The succession starts with a ca. 300 thick sequence of black and grey graptolitic shales (Ciekoty

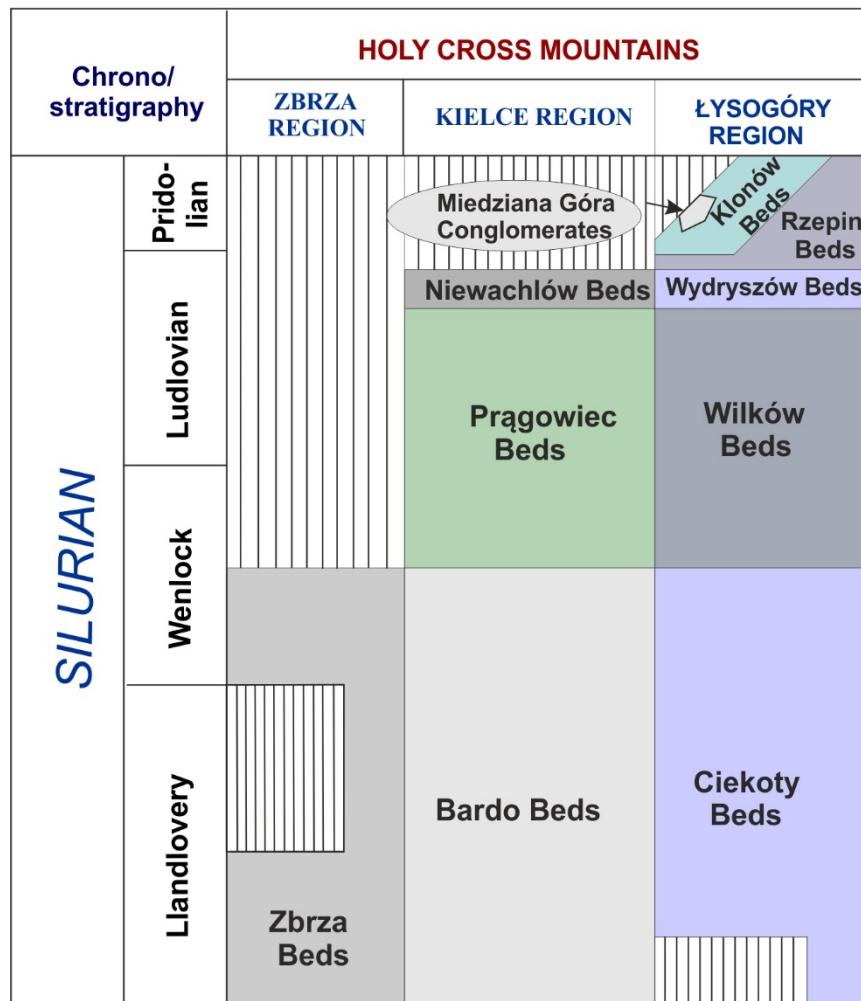


Fig. 6. Simplified lithostratigraphic scheme of the Silurian of the Holy Cross Mountains (from Polish Stratigraphic Table, 2008).

Beds) (Fig. 6) which comprises most of the Silurian time, from the Llandovery up to the early Ludlow (Gorstian). Overlying is a 500 m thick complex of fine-grained greywackes Trzcianka Formation (=Wydryszów Beds, Malec 2006) yielding commonly graded bedding, immature petrographic composition (Kozłowski *et al.* 2014) and a wide spectrum of sedimentary structures indicative of the sediment transport by turbidity currents. Records of rare graptolites indicate the late Ludlow (middle Ludfordian) age of greywackes. According to Kozłowski (2008), these deposits are interpreted to represent the underfilled flysch stage in the evolution of the basin. The succeeding unit Trochowiny Formation belongs also to the upper Ludlow. It is composed of 550 m thick shales interbedded with siltstone layers. The topmost part of the Ludlow is characterized by a remarkable facies change. Dark grey shales

are replaced by reddish-colored mudstones and sandstones of Rachtanka Formation (=Rzepin Beds, Malec 2006), which are up to 200 m thick most probably of continental origin. In the central part of Łysogóry these rocks interfinger with marine, shallow-water siliciclastics and carbonates (Winnica Formation). Fauna of trilobites and ostracods attest the latest Ludlow-earliest Pridoli age (Kozłowski 2008). The overlying Pridoli strata attain at least 800 m in thickness and are characterized by dramatic facies changes. While their lower and upper parts are represented by marine shales with abundant nektonic and planktonic fauna (Sarnia Zwola, Bronkowice and Rudki Formations), in the middle, there is a complex of red sandstones (Podchełmie Formation) which were deposited in a fluvial environment (Kowalczewski *et al.* 1998). The facies changes were very likely a response to eustatic sea-level fluctuation. Graptolite data indicate continuous marine deposition from the late Silurian into the Devonian in Łysogóry. According to (Narkiewicz 2002), the very thick Silurian sequence of Łysogóry documents a progressive trend of gradually increasing subsidence typical for a foredeep setting in front of the thrust belt.

Geology of the Małopolska Terrane (MT)

The Małopolska Terrane has well-defined tectonic boundaries in the north and south-west. To the north, it is separated by the Holy Cross Dislocation from the LT (Fig. 3) and to the south-west, the Cracow Fault from the Upper Silesian Terrane (UST). To the west, the MT hides under the overthrust of the Variscan Belt (Fig. 1). The extent of this unit in an easterly direction remains unclear. Belka *et al.* (2000) introduced the term the San Block for the eastern part of Małopolska, where there is a continuity in the Cambrian to Ordovician sedimentary succession, in contrast to the western part, the MT *sensu stricto*, where the Precambrian and Cambrian rocks have been tectonically deformed in pre-Ordovician time. The diagnostic features of the San Block can be only recognized in the area near the Polish-Ukrainian border. The location of the boundary between the San Block and the western Małopolska has not yet been identified. Most likely, it runs in the NE-SW direction, very close and parallel to the Vistula River (see Dadlez 2001). Because the Paleozoic rocks of Małopolska are exposed around Kielce in the Holy Cross Mountains, the unit was also frequently described in the literature as the “Kielce zone” or the “Kielce region”. The crystalline basement of Małopolska is unknown.

Precambrian

The Precambrian rocks have been recognized in the deep boreholes only. The succession is composed of polymictic clastics commonly characterized by a low degree of metamorphic overprint. According to Kowalski (1983), the total thickness of the succession exceeds 3000 m. In the central part of the MT, south of the Holy Cross Mountains, these are predominantly shales and siltstones with intercalations of coarse-grained clastics which exhibit a wide spectrum of clasts, from basic magmatic rocks and lithoclasts of sediments to fragments of metamorphic rocks (Kowalczewski 1990). The immature composition of detritus suggests a deposition within a forearc-trench system (Belka *et al.* 2000). In addition, the high amount of arkosic material together with volcanoclastic and plutoclastic detritus indicate the provenance from uplifted arc massif situated probably along a continental margin. Unfortunately, the age of the succession is not well defined. Studies of rare and poorly preserved acritarchs indicate the Ediacaran (=Vendian) age (Moryc & Jachowicz 2000). There is also a single numerical age of 549 ± 3 Ma for tuffs occurring in the topmost part of the succession (Compston *et al.* 1995). In the south-western part of MT, close to the contact with the UST, the Precambrian siliciclastic rocks are folded but unmetamorphosed. They are covered uncomfortably by Ordovician-Lower Silurian carbonates and Upper Silurian shales (Buła *et al.* 1997). Although Cambrian rocks of the MT are widely known from the many exposures and boreholes, nowhere the Cambrian base has been observed (Żelaźniewicz *et al.* 2009). In the easternmost Małopolska, Ediacaran clastics overlain by Cambrian rocks has been observed in the boreholes but always in tectonic contact.

Cambrian

The Cambrian sedimentary succession of Małopolska is known from many exposures in the southern part of the Holy Cross Mountains. It is composed of siliciclastic rocks and have a thickness which is estimated to be about 2000 m (e.g., Orłowski 1988). The succession starts with Lower Cambrian shales containing numerous intercalation of fine-grained siltstones and sandstone (Czarna Formation), (Fig. 4). However, the age of this unit is debated. According to (Orłowski 1989), it ranges from the *Hyolithes-Allathea* Zone up to the base of the *Holmia* Zone but (Kowalczewski 1995) placed it higher, into the *Holmia* Zone. In contrast, two other Early Cambrian units (Ociesęki and Kamieniec formations) are biostratigraphically well constrained thanks to the presence of abundant trilobite fauna (Żylińska 2013). The Ociesęki

Formation is represented by fine-grained sandstones with numerous intercalations of siltstones and shales in which bioturbations constitute a very characteristic feature. Sedimentary structures and ichnofossils indicate that deposition occurred in a relatively shallow-water marine environment (Orłowski 1989; Orłowski & Żylińska 2002). In the eastern part of the Holy Cross Mountains the Ociesęki Formation passes laterally into finer grained siliciclastic deposits of the Kamieniec Formation (Fig. 4). This unit is predominantly composed of monotonous shales and siltstones with only rare intercalations of fine-grained, non-bioturbated sandstones. These sediments were deposited below storm wave base, in a very deep basin characterized by poor water circulation and periodic oxygen deficiency (Mizerski *et al.* 1991).

In contrast to the Lower Cambrian, the Middle Cambrian deposits of Małopolska are primarily represented by coarse-grained siliciclastics. The lateral facies differentiation, however, shows some similarities to that of the Lower Cambrian one. In the western and central part of the southern Holy Cross Mountains, the Middle Cambrian is preserved in several patchy occurrences formed by coarse-grained, poorly-sorted sandstones (Słowiec Formation, Fig. 4) with trilobites and rare enigmatic echinoderms (Stasińska 1960; Żylińska & Masiak 2007; Żylińska & Szczepanik 2009). These sandstones were deposited in high-energy, shallow-water environment (Kowalczewski 1993). In the eastern part of the southern Holy Cross Mountains, however, the Middle Cambrian clastics have been assembled into the Usarzów Formation (Orłowski 1988). This unit comprises predominantly medium-grained, poorly-sorted sandstones with subordinate intercalations of shales, siltstones and coarse-grained sandstones. These rocks are considered as the most fossiliferous Cambrian clastics in the Holy Cross Mountains, being especially rich in trilobite fauna (Orłowski 1985). However, their sedimentary environment is not well defined. Kowalczewski (2000) suggested deposition within a shallow shelf area, most probably in deeper water than sediments of the Słowiec Formation.

A key point for the Cambrian stratigraphy of Małopolska seems to represent a sequence of the borehole Lenarczyce IG-1, which was drilled in the easternmost part of the southern Holy Cross Mountains. Szczepanik *et al.* (2004) documented there a 22 m thick complex of Middle Cambrian mudstones and siltstones (Kobierniki beds) overlain by the Late Cambrian (Furongian) clastic succession (89 m thick) yielding a distinct bipartite composition. While the lower part (Lenarczyce beds) is dominated by sandstones with numerous intercalations of mudstones, siltstones and conglomerates, the upper one (Ublinek beds) comprises

monotonous fine-grained dark shales with subordinate sandstones intercalations. It seems that each complex accumulated under different depositional conditions. Sedimentary structures recognized in the Lenarczyce beds point to shallow-water environment influenced by storms and tidal currents (Szczepanik *et al.* 2004). In contrast, the Ublinek beds were accumulated in more open-marine shelf environment. The whole Cambrian succession of the borehole Lenarczyce IG-1 is strongly deformed by folds and numerous shear zones. The deformation style is significantly different in the Middle and the Upper Cambrian rocks and there is a stratigraphic gap between them. Therefore, it is very likely that the Late Cambrian strata have been thrust over the Middle Cambrian Kobierniki beds.

Ordovician

Ordovician rocks of Małopolska are exposed only in a few places in the southern Holy Cross Mountains. The succession exhibits a clear tripartite subdivision with basal clastics, strongly condensed carbonate middle part and marly sediments at the top. Moreover, it is also characterized by lateral differences in thickness and facies, and also by the presence of stratigraphic gaps (for overview and details see: Trela 2005, 2006b). The succession starts with the glauconitic conglomerates (Kędziorka Formation) and/or sandstones (Międzygórz Formation) resting on various Cambrian clastics (Fig. 7). The contact is always erosional and associated with a stratigraphic gap. In addition, within the Bardo syncline, there is a distinct angular unconformity between the Ordovician and the underlying Early Cambrian strata. Here, the basal conglomerate is very thin (about 40 cm) and is covered by glauconitic mudstones with numerous chert nodules (Wysoczki Formation). These rocks contain significant amounts of pyroclastic material (Chlebowski 1971). The basal glauconitic clastics attain a maximum thickness of about 30 m and contain conodont and brachiopod faunas indicative of the Tremadocian and early Arenig age.

In the south-western part of the Holy Cross Mountains, which was deeper in comparison to the central and eastern areas, the basal clastics are overlain by claystones with graptolites (Brzeziny Formation), ranging from late Arenig to Llanvirn in age. They are approximately correlative with mudstones and sandstones (Szumsko and Bukówka formations, Fig.7), which accumulated in more shallow-water environments. The thick-bedded sandstones with some carbonate intercalations are characterized by the presence of numerous bioturbations and a distinct discontinuity at the top of the unit. According to Trela (2004), the sandstones

developed during a relative sea-level fall corresponding to a lowstand interval, well-documented in Baltoscandia, and the discontinuity underlines a stratigraphic gap comprising most of the Llanvirn time.

The Upper Ordovician exhibits a facies differentiation similar to that of the Arenig. In the south-west, up to 170 m thick, deep-water clay mudstones (Jeleniów and Wólka formations) were deposited, whereas in the central area sandy limestones of the Mójcza Formation (Fig. 7) accumulated. This carbonate unit is very fossiliferous and composed of a thick bedded organodetrital limestone with phosphate-coated grains and ferruginous ooids (Dzik *et al.* 1994; Dzik & Pisera 1994; Trela 2005). It represents an extreme case of stratigraphic condensation. At Mójcza, about 25 million years are represented by ten meters of rock succession, with only one biostratigraphically identifiable gap. In the eastern part of Małopolska, however, the sedimentary succession coeval with the Mójcza Formation is represented by medium- to thick-bedded dolomites (Mokradle Dolostones), locally with marly claystone intercalations.

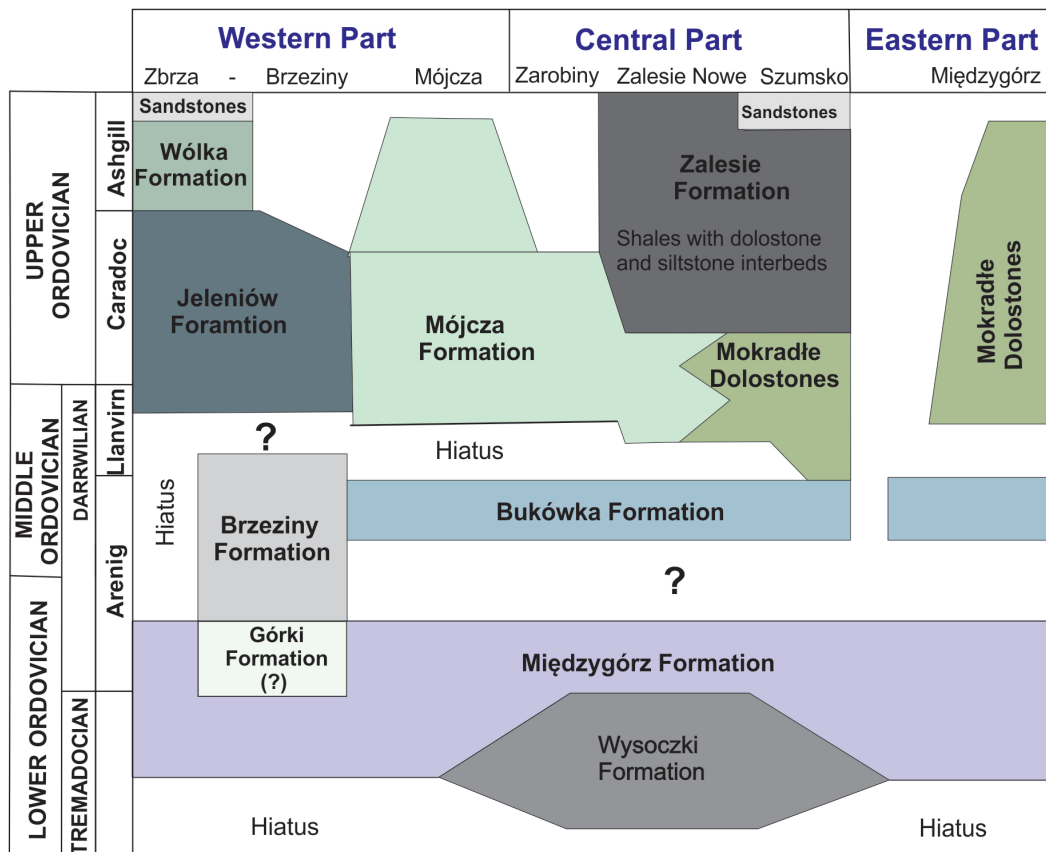


Fig. 7. Simplified Ordovician lithostratigraphy of Małopolska (from Bednarczyk (1981) and Dzik (1984), modified).

At Zalesie Nowe, the dolomites are overlain by a 7 m thick sequence of shales and dolomitic marls (Zalesie Formation). Fossils are abundant in shales but rare in dolomites. Brachiopods and graptolites define a Caradoc and Ashgill age of the unit (Bednarczyk 1981). Sedimentation across the Ordovician–Silurian boundary was continuous, and the Zalesie Formation grades upward into black Silurian shales.

Silurian

Silurian rocks of Małopolska crop out in numerous natural exposures in the southern part of the Holy Cross Mountains and were also recognized in many boreholes (for details see: Modliński & Szymański 2001). The succession in the Holy Cross Mountains, which is up to 1400-1500 m thick, is clearly tripartite. The lower and upper parts consist of shales, whereas the middle one is composed of medium- to coarse-grained clastics having predominantly a greywacke composition. The lower part of the sequence comprises Llandovery to lower Ludlow shales and claystones (Bardo and Prągowiec Beds, Fig. 6) up to 350 m in thickness, which bear typically very rich assemblages of graptolites (Tomczyk 1962; Tomczykowa & Tomczyk 1981). The shales are siliceous with lydite intercalations or dolomitic-calcareous with carbonate concretions. Besides graptolites other macrofossils occur also abundantly. These are nautiloids, tentaculoids, pelecypods, and various trilobites. The shales were deposited in a deep-water environment. The oriented pattern of graptolite rhabdosoms and the abundance of benthic organisms, however, provide evidence for oxic conditions and currents in the bottom part of the seawater column. A remarkable change in the lithological composition and sedimentary regime happened during mid-Ludlow time. It resulted in deposition of about 1000 m greywackes (Niewachłów Beds, Fig. 6), the most prominent lithological unit in the sequence. These rocks are chiefly lithic arenites composed of quartz, feldspars, and fragments of sedimentary and volcanic rocks. According to Przybyłowicz and Stupnicka (1989), the volcanic material was very probably derived from rhyolitic volcanoes situated in a short distance from the sedimentary basin. In contrary, Kozłowski *et al.* (2014) interpreted the greywackes as represented a sedimentary infill of a foreland basin and originated as a result of the collision of the Tornquist margin of Euramerica with a volcanic arc, located probably at the easternmost extent of the Avalonian Plate. The sedimentary structures of the greywackes, such as flute casts, tool marks, and grading, point to transport by local turbidity currents. The stratigraphic range of the unit is not well constrained. Some few graptolite shale intercalations has been dated to be late Ludlow in age (Malec 1991). The

greywackes are overlain by shales, assigned to the uppermost Ludlow and Pridoli, containing marine fauna in their lower portion (Lipniczek Beds). In contrast, the topmost part of the Silurian succession comprises fine-grained, red-colored shales with sandstone interbeds (Klonów Beds). Their continental vs marine origin was controversially discussed in the past (Tomczykowa & Tomczyk 1981; Malec 1993).

PALEOGEOGRAPHY

The European continent is composed of a large number of crustal fragments which primarily, in early Paleozoic time, have constituted parts of three paleocontinents: Baltica, Gondwana and Laurentia. These continents were formed as a result of disintegration of the Neoproterozoic continent Rodinia during two phases, at 750-725 Ma and 590-550 Ma (Fig. 8A), (Bond *et al.* 1984; Dalziel 1992; Powell *et al.* 1993). Their Vendian to Cambrian drift scenario is poorly documented and still remains a matter of discussion. But there is no doubt that Gondwana, Laurentia and Baltica were separated by oceans at least until late Ordovician time (Torsvik *et al.* 1996). In the context of the geotectonic provenance of Małopolska and Łysogóry linkages to Baltica and/or to the Peri-Gondwanan plates are considered (e.g., Belka *et al.* 2002; Nawrocki *et al.* 2007). This is why the most attention in this chapter is devoted to these continental domains. The paleocontinent Baltica comprised eastern Europe and most of the landmasses of the northern Europe. Its outline was roughly triangular with the eastern border along the Ural Mountains, the northwestern edge defined by the Iapetus suture along the British and Scandinavian Caledonides, and finally, the southwestern margin defined by the Teisseyre-Tornquist Line. The core of Baltica is formed by the East European Craton, which is composed of three major crustal segments: Fennoscandia, Sarmatia (Fig. 8B) and Volgo-Uralia (Claesson *et al.* 2001). Fennoscandia was docked to previously-established Volgo-Uralia and Sarmatia cratons at about 1.9 Ga and in a consequence the so-called Protobaltica continent was formed. According to the Meert & Torsvik (2003), Protobaltica constituted a unit within a supercontinent of Rodinia surrounded by crust of Amazonia, Laurentia and Siberia. Baltica was an independent plate from the Vendian (~570 to 550 Ma) until its soft collision with Avalonia at about 443 Ma. The position of Baltica during the early Paleozoic was fundamentally different from that of today and still there are some distinct ideas about the paleogeography. Late Precambrian and Paleozoic paleomagnetic data for Baltica were provi-

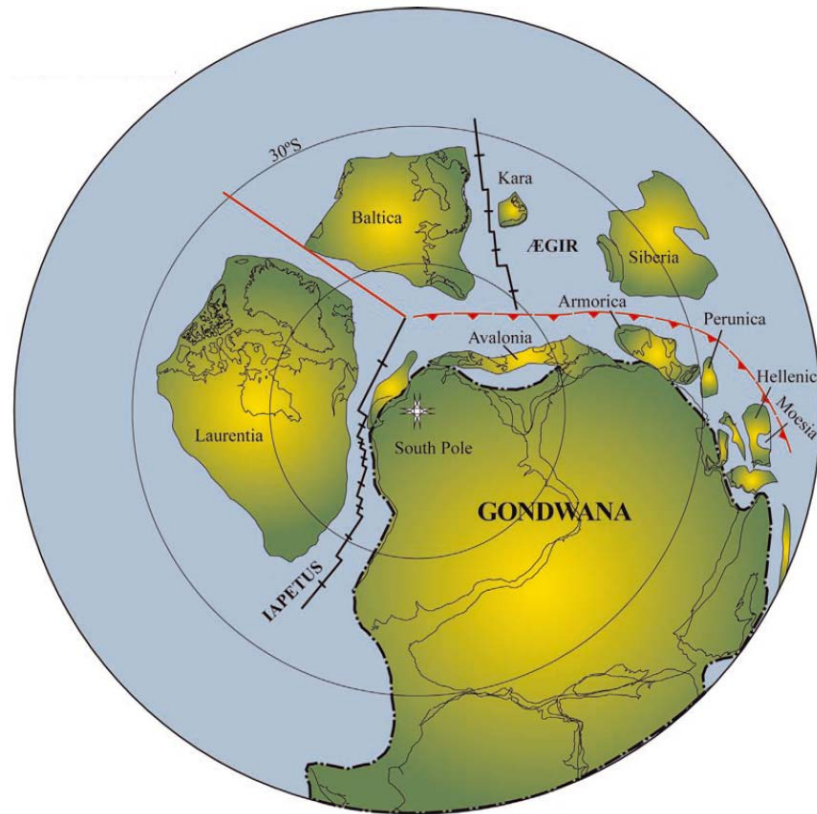


Fig. 8A. Reconstruction of Southern Hemisphere in the Late Vendian (550 Ma), modified from Hartz and Torsvik (2002) and Rehnström *et al.* (2002).

ded by Torsvik *et al.* (1996), Torsvik & Rehnström (2001), Nawrocki *et al.* (2004) and Cocks & Torsvik (2005). According to Torsvik & Rehnström (2001), for instance, Baltica stretched from about 30°S and to approximately 60°S and its present NE margin faced Gondwana. Belka *et al.* (2002), however, argued that the position of Baltica during the Cambrian, with its Teisseyre-Tornquist margin faced to the west not to the north is much more compatible with facies and faunal trends. Nawrocki *et al.* (2004a) showed in their reconstruction, however, the Baltic plate moving from moderate latitudes to the equator with a significant anticlockwise rotation of about 120°. This paleogeographical model explains the occurrence of tectonic units with the late Neoproterozoic deformation (Teisseyre-Tornquist Terrane Assemblage) near the present SW edge of Baltica. In this model the Teisseyre-Tornquist Terrane Assemblage was developed near the present southern margin of EEC and dextrally relocated along the TESZ edge of Baltica.

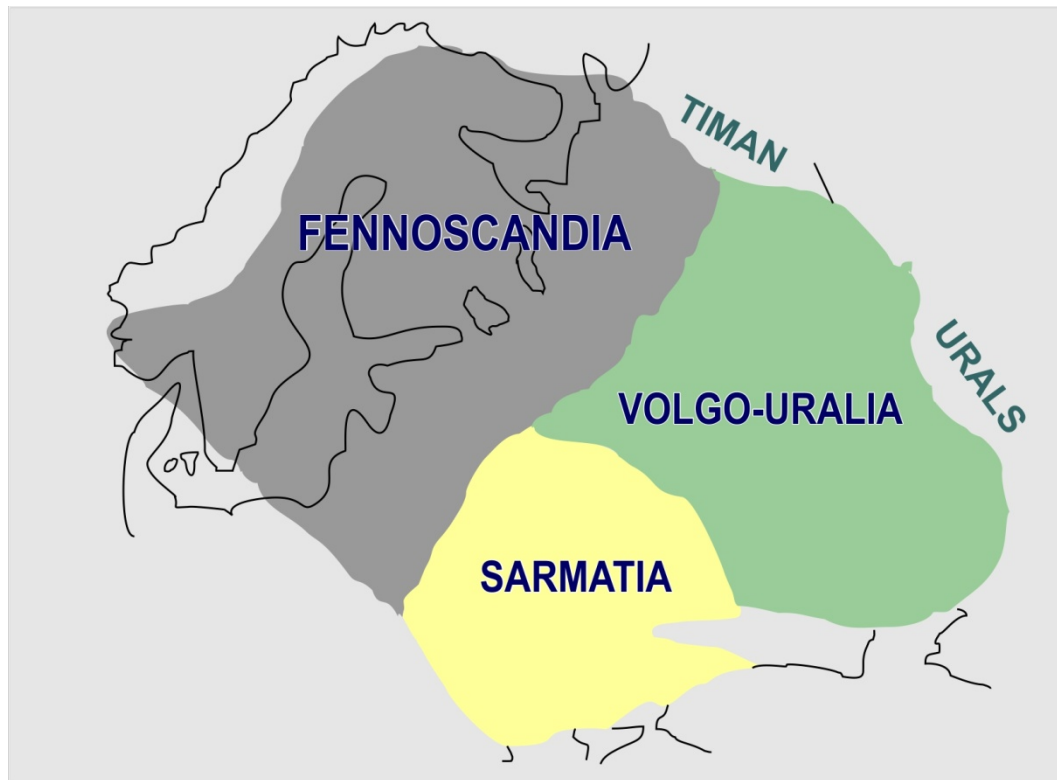


Fig. 8B. Schematic subdivision of the East European Craton to show the main structural segments: Fennoscandia, Sarmatia and Volgo-Uralia.

The first phase of the displacement took place as early as the latest Ediacaran. Cocks and Torsvik (2005) took into account all the previous paleomagnetic data and created a reconstruction in which claimed that Baltica underwent a very substantial rotation with the maximum rate of this relocation occurred in Late Cambrian and Early Ordovician times. This concept does not explain some tectonic problem such as the existence in the Trans-European Suture Zone along edge of Baltica as early as in the Cambrian and partly metamorphosed Ediacaran flysch in the basement of the Małopolska Massif. In general, the discrepancies in different paleogeographical models for Baltica position and drift cause main difficulties in reconstructing the geotectonic evolution of Małopolska and Łysogóry. During the early Paleozoic Gondwana created the largest continent, covering about 100 million km². Nowadays, its remnants make 64% of all land areas. Gondwana included South America, Africa, most of Antarctica and Australasia, as well as Madagascar and the Indian Subcontinent. Furthermore, Florida and most of Central America, southern Europe, and much of south-central and south-eastern Asia formed parts of Gondwana at different time span. Several smaller crustal fragments, termed as peri-Gondwanan blocks, which originally

constituted integral parts of Gondwana, were separated and rifted from it in the early Paleozoic.

The largest microplate forming originally the peri-Gondwanan terrane assemblage was Avalonia (Fig. 8A). It comprised the southern Ireland, Wales, New England, Belgium, the Netherlands and parts of northern Germany, in Europe, and eastern parts of the Appalachian orogen, in North America (Cocks *et al.* 1997). The primary edges of Avalonia are defined by sutures which originated due to closure of the Paleozoic oceans. The Iapetus suture forms the northern edge, the Tornquist suture the eastern edge, and the Rheic suture the southern one. Avalonia is considered as the first, large microplate that was detached from Gondwana, crossed the Iapetus Ocean and ultimately amalgamated with Laurentia and Baltica (Torsvik & Trench 1991a). There is plenty of evidence proving the Ordovician-Silurian drift of Avalonia. Paleomagnetic results from the 490 Ma volcanics in Wales and from 489 Ma igneous rocks in New England clearly indicate that Avalonia was situated between 65° and 62° S before separation from Gondwana (Trench *et al.* 1992; Thompson *et al.* 2010). The terrane broke off from Gondwana before the Llanvirnian (Early Ordovician). Palaeomagnetic data document that after its detachment Avalonia moved across Iapetus to 44°- 41° S (Trench *et al.* 1991; Hamilton and Murphy 2004).

The paleocontinent Laurentia, which included most of North America, Greenland, Svalbard and parts of northern Europe (northern Ireland, Scotland and some nappes in the Scandinavian Caledonides), was situated at high latitudes of the Southern hemisphere during the Early Cambrian and then drifted northwards to reach low latitudes in the Ordovician and Silurian times. Laurentia was separated from Avalonia, Baltica and Siberia by the Iapetus Ocean. Its maximum width occurred in the latest Cambrian, starting to close in the earliest Ordovician and continuing to close until the late Silurian when the continents Laurentia, Baltica and Avalonia were collided to create a new large continent Euramerica (Laurussia).

PREVIOUS PROVENANCE STUDIES

Biogeographic composition of the fauna

In terms of biogeography, the Cambrian trilobite and brachiopod faunas demonstrate a very specific composition in southern Poland. There occur taxa that are typical for Baltic or Peri-Gondwanan zoogeographical provinces, or for both (see for summary: Belka *et al.* 2002 and Fig. 9). In the past, the trilobites were considered as a main evidence that Małopolska

constituted an integral part of the paleocontinent of Baltica (Bergström 1984; Vidal *et al.* 1995). This interpretation was based on the presence of Baltic olenellid genera such as *Holmia*, *Kjerulfia* and *Schmidtiellus*, and ellipsocephalids (Orłowski 1985). Recently, the trilobite fauna collected in the Kielce region of Małopolska in the past decades has been revised (Żylińska & Szczepanik 2009; Żylińska 2013). The re-investigation revealed that the Early Cambrian trilobite fauna includes 12 endemic species belonging to the genera which are known from Baltica, Avalonia and West Gondwana. Representatives of holmiids display a correlation with Scandinavian successions and thus indicate a Baltic affinity. In turn, the ellipsocephalids associated with chengkouiid and atopid forms reveal strong similarity to faunas occurring in West Gondwana and Avalonia. However, the most numerous elements in the fauna, specimens of *Berabichia* and *Strenuella*, show linkages to Baltica, Avalonia and West Gondwana or only to West Gondwana and Avalonia (Żylińska 2013). The affinity to West Gondwana and Avalonia exhibits also the rich trilobite fauna present in the Lower-Middle Cambrian boundary interval of Małopolska (Żylińska & Masiak 2007; Żylińska & Szczepanik 2009). Results of the re-investigation of the Cambrian trilobite fauna from Małopolska clearly contradicts the conclusion of Cocks (2002), who postulated that these Early and Middle Cambrian trilobites “are of undoubted Baltic affinity”, although he has never examined directly the paleontological material and has based his interpretation exclusively on the data from the old literature.

Another group of the Cambrian fossils in Małopolska that show Avalonian affinities are inarticulate brachiopods (Jendryka-Fuglewicz 1992; 1998). In addition, the fauna is dominated by forms with calcitic shells what contrasts strongly with the Polish segment of the East European Platform, where brachiopod faunas are represented exclusively by forms with phosphatic shells. In the Lower Cambrian of Małopolska there is only one species among Avalonian forms, *Westonia bottnica*, which is known from successions of the East European Platform (Fig. 9). However, in the Middle Cambrian sequence a progressive migration of Baltic brachiopods to Małopolska can be observed. In the Ordovician rocks the linkages of the fauna to Gondwanan or Avalonian realms are not more recognizable. The fauna (brachiopods, conodonts) shows close Baltic similarities (Dzik *et al.* 1994), although Ordovician ostracods exhibit an increase of endemism (Olempska 1994). The trilobite record of Łysogóry starts with the Late Cambrian fauna which was collected and documented for over the last hundred years. But it was a very rich assemblage of trilobite traces that became

famous in the literature (e.g., Radwański & Roniewicz 1963; Orłowski *et al.* 1970) not the Cambrian trilobites themselves.

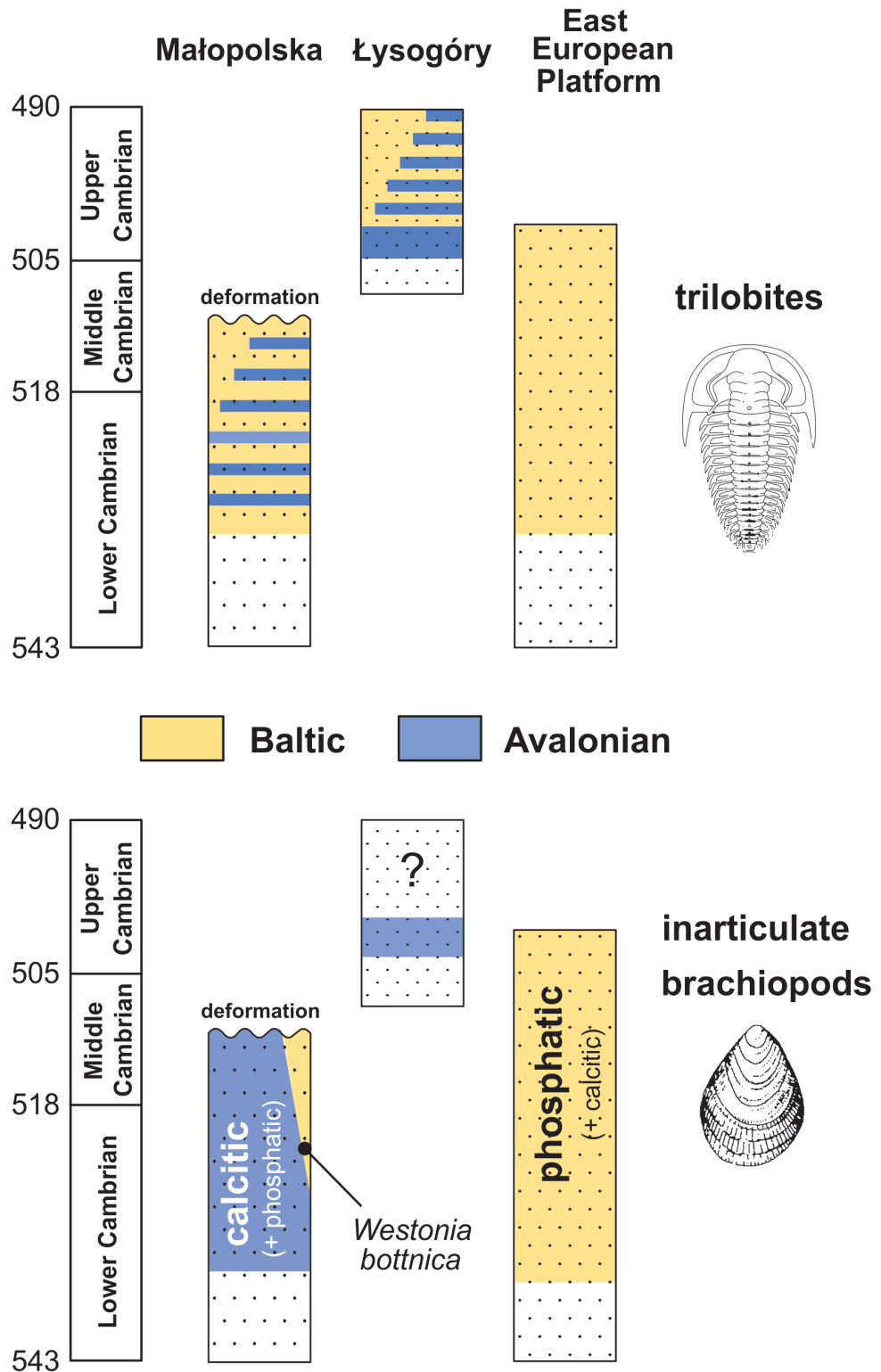


Fig. 9. Stratigraphic distribution and biogeographic affinity of the Cambrian trilobite and brachiopod faunas of Małopolska and Łysogóry in comparison to those of the Polish part of the East European Platform. Columns indicate the stratigraphic ranges of the Cambrian sequence in the crustal units. Shell mineralogy of articulate brachiopods is also indicated (from Belka *et al.* 2002, updated).

According to Seilacher (2007 and pers. comm.), this assemblage is identical to those distributed throughout Gondwana and the Peri-Gondwanan microplates, but it is unknown from the paleocontinent of Baltica. Żylińska (2001, 2002) re-investigated the Late Cambrian trilobite collections. She showed that a distinct change in the biogeographic composition of the fauna can be observed, from a low diversity fauna with a predominance of Avalonian forms in the early Late Cambrian, to more diversified assemblages, characterized by the constant increase to dominance of Baltic elements by the end of the Cambrian. However, there are also some few Gondwanan elements in the fauna, which are identical to those known from the Upper Cambrian of South America. In contrast to the Cambrian trilobites, the Ordovician biota of Łysogóry reveals a typical Baltic character with only a small exception of the occurrence of Gondwanan chitinozoan taxa in the Caradoc sequence (Wrona, pers. comm.).

Provenance of detrital micas

The pioneering K-Ar age determinations on detrital muscovites from the Lower Paleozoic and Devonian sedimentary rocks of the Holy Cross Mountains (Fig. 10A) were carried out by Belka *et al.* (2000). They noticed striking differences between Małopolska and Łysogóry not only in age of detrital mica grains but also in their concentration. In the Cambrian clastics of Łysogóry detrital muscovites were large and very abundant, whereas in the successions of Małopolska they were very rare and extremely small, and thus, mostly not suitable for K-Ar dating. Multigrain mica samples recovered from the Middle and Upper Cambrian of Łysogóry displayed a wide range of ages from 539 to 1745 Ma, suggesting a mixing of different mica populations. Belka *et al.* (2000) postulated a bimodal composition of the detritus due to a contribution from both Baltic (ages around 1.7 Ga) and Cadomian (ages from 540 to 600 Ma) sources. The provenance of 1.7 Ga old muscovites from the Baltic sources is well constrained since the Middle Cambrian sandstones in the marginal zone of the East European Platform (EEP) bear detrital muscovites with similar K-Ar cooling ages from c. 1775 to 1883 Ma. These ages point to a derivation from a Svecofennian basement of the EEP, most probably from the basement of the Mazury High. In contrast to Łysogóry, in the Cambrian clastics of Małopolska micas exclusively from Cadomian sources, with ages of 543 Ma and 547 Ma, have been recognized. Subsequent work of Belka *et al.* (2002) confirmed the Cadomian composition of detritus in both Cambrian and Ordovician clastics of Małopolska. The provenance data provided by Belka *et al.* (2000) revealed also that the remarkable difference in the derivation of clastic material between Łysogóry and Małopolska had to disappear

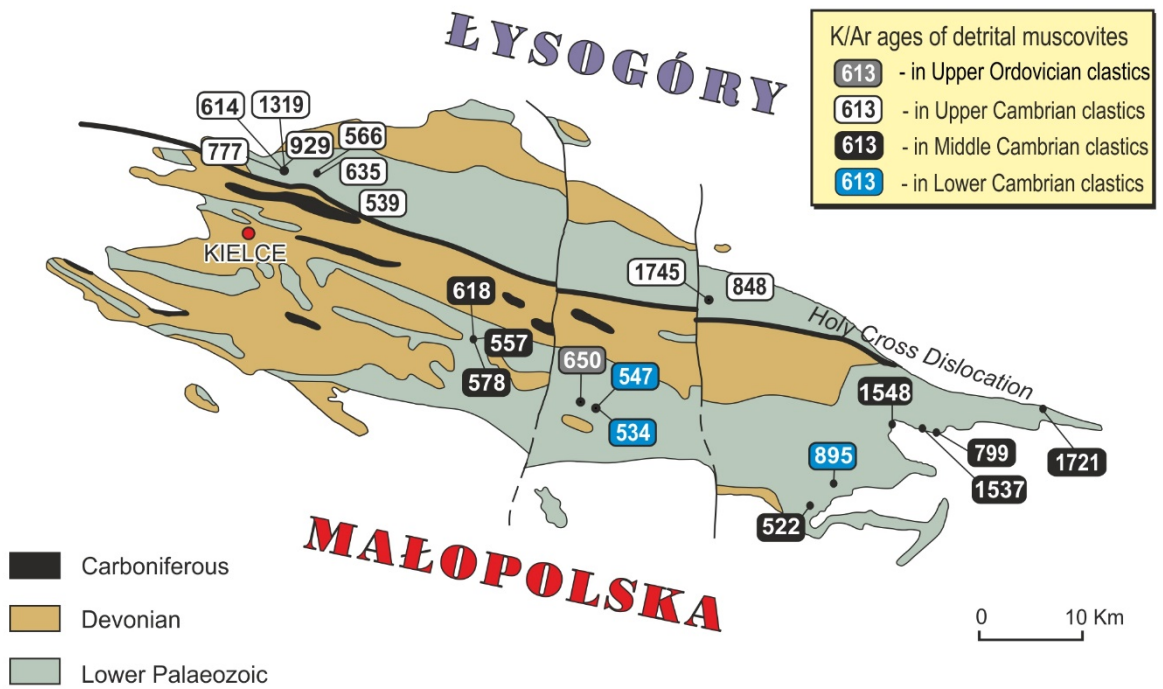


Fig. 10A. K-Ar cooling ages of detrital muscovite grains in the Cambrian and Ordovician clastics of Małopolska and Łysogóry (data from Belka *et al.* 2002 and Nawrocki *et al.* 2007).

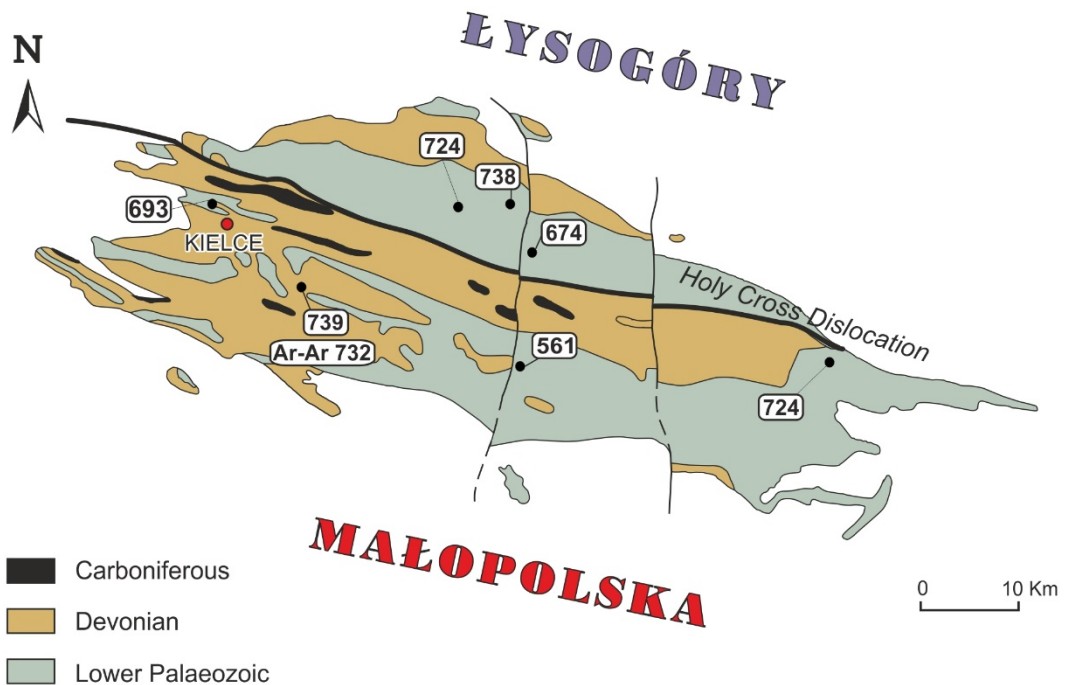


Fig. 10B. K-Ar cooling ages of detrital muscovites in the Upper Silurian clastics of Małopolska and Łysogóry, supplemented by the Ar-Ar age of the muscovite from Silurian greywackes (data from Nawrocki *et al.* 2007).

before Devonian time. Detrital micas recovered from the Lower Devonian sandstones of both terranes represent the same, unimodal mica population, derived from Paleozoic sources, most probably from the Scandinavian Caledonides.

K-Ar cooling ages of detrital micas were also studied by Nawrocki *et al.* (2007). They provided data from both the Cambrian and Silurian clastics. Samples collected in Łysogóry and in the central part of the southern Holy Cross Mountains confirmed the presence of a Cadomian detritus in the Cambrian of Łysogóry and Małopolska. However, in the eastern part of Małopolska, in the area where the contact of Małopolska with Łysogóry is not precisely defined, four muscovite samples recovered from the Lower and Middle Cambrian rocks gave ages of c. 0.8-0.9 Ga and 1.5 Ga. Nawrocki *et al.* (2007) interpreted them as indicating the sources with an igneous–metamorphic overprint of exactly the same age, i.e. about 0.8–0.9 Ga and 1.5 Ga. This interpretation, however, is very unlikely. It is very surprising that having the K-Ar age determinations performed on multigrain samples Nawrocki *et al.* (2007) did not take completely into account the possibility of detritus mixing from various sources. The muscovite concentrates with cooling ages of c. 1.5 Ga, for instance, were interpreted as probably derived from the contact zones of rapakivi-type granites of the East European Craton. The 1.5 Ga rapakivi granites and related rocks (Gothian rocks) are known to occur in the Precambrian basement of northeastern Poland (Doerr *et al.* 2002), i.e., situated today in the distance of about 400 km from the eastern part of Małopolska. The problem, however, is that the Cambrian clastics in the area between Małopolska and northeastern Poland are composed of older detrital material, with c. 1.7-1.8 Ga old muscovites (Belka *et al.* 2000; Valverde-Vaquero *et al.* 2000). This indicates that the Cambrian multigrain samples analyzed by Nawrocki *et al.* (2007), with ages of c. 0.8-0.9 Ga and 1.5 Ga, most probably represent mica grain populations originated from the mixing of Baltic and Cadomian materials.

Unlike the Cambrian, the Upper Silurian greywackes (Fig. 10B) contain virtually the same mica material in Łysogóry and Małopolska, with K-Ar cooling ages from 674 to 739 Ma. This detritus has been interpreted as derived from a crustal unit metamorphosed at c. 730 Ma, connected with a continental island arc (Kozłowski *et al.* 2004; Nawrocki *et al.* 2007; Kozłowski *et al.* 2014). In addition, a single sample from the Lower Silurian of southern Holy Cross Mountains gave a cooling age of 560 Ma, suggesting a Cadomian provenance.

Provenance of detrital zircons

To decipher the ancestry of Łysogóry and Małopolska terranes the Cambrian clastics were subject to studies using U-Pb geochronology of detrital zircons. It was Valverde-Vaquero *et al.* (2000), who provided first detrital zircon age populations from Wymyślona (Łysogóry) and compared them with zircons recovered from the Middle Cambrian sandstones of the Okuniew IG-1 borehole, drilled in the marginal part of the EEP. The zircons from the Upper Cambrian of Łysogóry (Fig. 11) showed a wide age spectrum with well-defined Archean (>2.5 Ga), Early Proterozoic (1.8–2.1 Ga) and Late Precambrian (600 Ma) ages. Valverde-Vaquero *et al.* (2000) pointed out that such age pattern is similar to that of the Gondwanan sources and that the remarkable absence of c. 1.0–1.2 Ga (Sveconorwegian) detritus excludes the derivation of clastic material from the Polish segment of the EEP. In Małopolska, detrital zircons were recovered from the Lower Cambrian sandstones of Kędziorka (Belka *et al.* 2002).

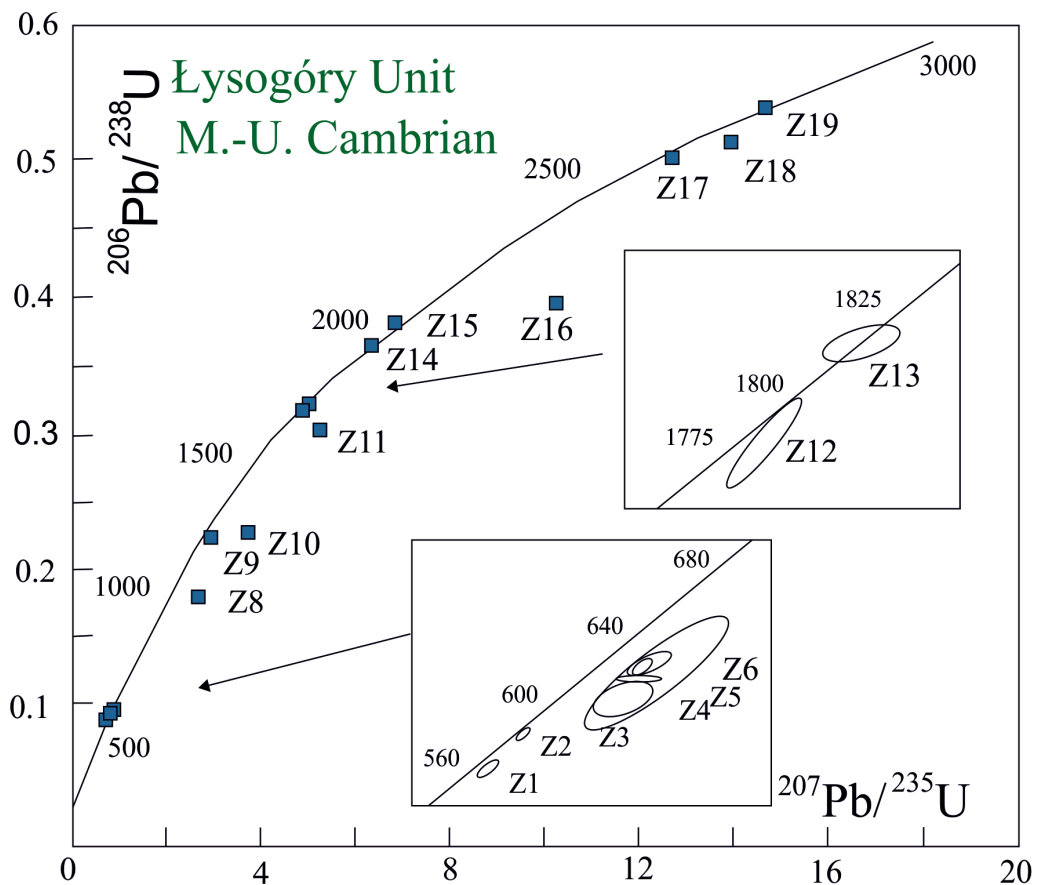


Fig. 11. The U-Pb concordia diagram for the detrital zircons from the Upper Cambrian sandstones at Wymyślona, Łysogóry (from Valverde-Vaquero *et al.* 2000).

Their main populations with ages of about 540 Ma and 2.0 Ga, and the presence of c. 1.2–1.37 Ga, 1.5 Ga, >2.5 Ga and 3.0 Ga old grains correlate well with detrital zircon populations known from Neoproterozoic rocks of West Avalonia and from the basement of the Amazonian Craton (see for comparison Nance & Murphy 1996). Friedl *et al.* (2000) recorded a similar geochronological fingerprint in the basement of the Upper Silesian Terrane. In summary, Belka *et al.* (2002) concluded that the clastic material of the Lower Cambrian at Kędziorka derived from a Cadomian source since the youngest population of zircons (540 Ma) coincided perfectly with K-Ar cooling ages of co-occurring detrital mica (535–545 Ma). In contrary, Nawrocki *et al.* (2007) provided a dataset of detrital zircon ages from the Upper Tremadocian sandstones of the Szumsko IG-1 borehole (located some few kilometers from Kędziorka) and suggested a derivation from a Fennoscandian (Baltic) source, based on the presence of c. 1.85 Ga old zircons. However, the occurrence of c. 1.85 Ga zircons alone does not represent conclusive evidence for the Baltic provenance because concordant zircons grains of this age were also recorded from West Avalonia (Nance *et al.* 2008). Moreover, the age spectrum of zircons known from clastics deposited on the Baltica margin typically comprises 1–1.2 Ga and 1.5 Ga populations (Valverde-Vaquero *et al.* 2000), which have not been observed in the Tremadocian sandstones at Szumsko.

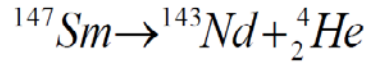
PALEOMAGNETIC DATA

Lewandowski (1993) was the first to provide paleomagnetic evidence for the mobility of Małopolska during pre-Carboniferous times, and hence an evidence for a possible terrane character of this block. He suggested a large-scale (c. 1000 km), dextral strike-slip displacement of Małopolska along the SW border of the East European Craton. This interpretation was based on Ordovician and Devonian paleopoles which deviate significantly from the apparent polar wander path (APWP) for the Baltica paleocontinent. The reliability of these data, however, was later questioned by Nawrocki (1995). For details of the discussion the reader is also referred to Lewandowski (1999). Further paleomagnetic studies carried out on Cambrian (Nawrocki *et al.* 2007), Ordovician (Schatz *et al.* 2006), and Silurian rocks (Nawrocki 2000) shown that Małopolska was fixed in its present-day position with respect to Baltica since the Middle Cambrian. According to Nawrocki *et al.* (2007), the tectonic mobility of Małopolska during latest Cambrian/earliest Ordovician time was presumably due to a large anticlockwise rotation of the whole of Baltica (Torsvik *et al.* 1996) and that this

movement resulted in deformation of Cambrian rocks of Małopolska. This means, however, that Małopolska had to be in direct contact to Baltica, thus it was not separated by the Łysogóry Terrane from the East European Craton.

NEODYMIUM AND SAMARIUM ISOTOPE GEOCHEMISTRY

Neodymium ($Z = 60$) and samarium ($Z = 62$) both belong to intermediate rare earth elements (REEs), in Group III B of the periodic table. They are very often known as lanthanides or lanthanoids after the element lanthanum, which is the first of the series. Usually, the low-atomic number members (elements from 57 to 62) of the REE series are called the light rare earth (LREE); in turn those with the higher atomic numbers (from 63 to 71) are named heavy rare earths (HREE). The size of the REE's ion is a little bit larger in the light REE and slightly smaller in the heavy REE. They occur in low concentration in rocks/minerals, usually less than 0.1%. Generally, they are reported in units of parts per million (ppm). Despite its low concentrations, the REE's are important research tools for geologists in studying igneous, sedimentary and metamorphic rocks and the processes that created them. Neodymium occurs in many silicate, phosphate and carbonate minerals by substitution for many ions. The REE's have a +3 valence with the exception of Eu, which is +2 under some conditions, and Ce, which is +4 also under some conditions. The main chemical difference between all rare earth elements is the ionic radius, which varies from 1.15 Å for La ($A=57$) to 0.93 Å for Lu ($A=71$) and is a key factor of their geochemical behavior. The ionic radii of Sm and Nd differ by only 0.04 Å (Nd= 1.08, Sm= 1.04). The atomic radii and relatively high charge causes that they do not easily fit into lattices structure of minerals. The REE's are considered as moderately incompatible, with exception of Nd that being a little bit more incompatible than Sm. Taking into consideration the rocks in the mantle, when they are starting to melt, the incompatible elements will be immediately "remove" from the solid and enter to the liquid phase. Thus, a low degree melt of the mantle rock causes high concentration of incompatible elements, whereas the melt process proceed, the concentration of incompatible elements will immediately decrease. Samarium has seven naturally-occurring isotopes of Sm of which three are radioactive: ^{147}Sm , ^{148}Sm , ^{149}Sm . The half-lives of ^{148}Sm and ^{149}Sm are extremely long hence, those isotopes are not used in radiometric dating. Nd is represented by five stable isotopes ^{142}Nd , ^{143}Nd , ^{145}Nd , ^{146}Nd , ^{148}Nd and two radioisotopes ^{144}Nd and ^{150}Nd . ^{143}Nd is produced by α -decay of ^{147}Sm with a half-life of $1.06 \cdot 10^{11}$ years.



Quantitative changes of these isotopes in the time can be representing by following linear equation:

$$\frac{^{143}\text{Nd}}{^{144}\text{Nd}} = \left(\frac{^{143}\text{Nd}}{^{144}\text{Nd}} \right)_0 + \frac{^{147}\text{Sm}}{^{144}\text{Nd}} (e^{\lambda t} - 1)$$

The Sm/Nd ratio in the mantle is higher than that of the crust and hence the $^{143}\text{Nd} / ^{144}\text{Nd}$ ratio is higher in the mantle than that in the crust. This subjection is strictly connected with the compatibility of the Sm and Nd. The samarium is more compatible than neodymium, therefore during partial melting Sm stays behind in the upper mantle, whereas Nd is taken up in the differentiation of crustal materials from the upper mantle. This process depletes the mantle more in neodymium than in samarium. The depleted mantle shows higher Sm/Nd ratio, whereas enriched domains conversely lower Sm/Nd ratios (Fig. 12).

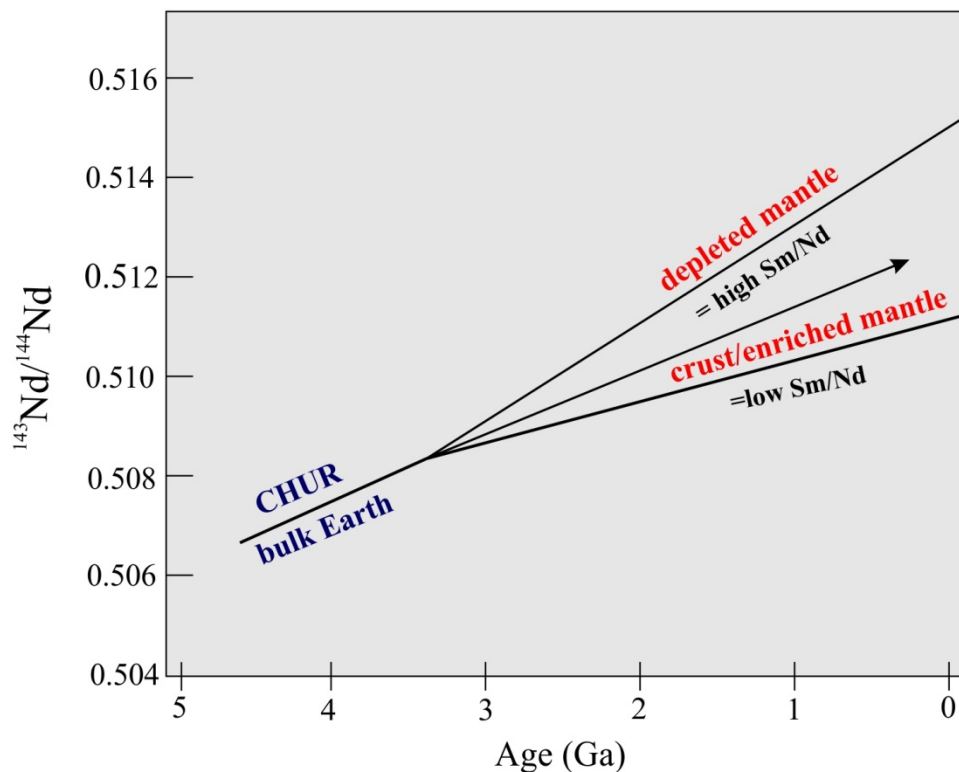


Fig. 12. Neodymium isotope evolution in the mantle and crust during earth's history (after, White 2009).

Therefore, the continental crust is characterized by low Sm/Nd and $^{143}\text{Nd}/^{144}\text{Nd}$ ratios. The largest variations of the Sm/Nd ratio are observed in mafic and ultramafic rock, in turn the smallest in acid rocks. The isotopic evolution of Nd in the Earth is presented as a model named CHUR, which is an abbreviation of “Chondritic Uniform Reservoir” (DePaolo and Wasserburg 1976). This model assumes that terrestrial Nd has evolved in a “uniform reservoir”, therefore the initial Nd isotopic system of terrestrial rocks is very similar to that defined by the CHUR evolution line. Chondritic meteorites are considered to represent the earliest (unsorted) material that formed in the solar system before planets were created. Chondritic meteorites were composed of relatively homogeneous trace elements signatures as a result their isotope evolution can be a model for the whole solar system and the “Bulk Earth”. All rocks of the continental crust reveal lower present-day $^{143}\text{Nd}/^{144}\text{Nd}$ ratios than those of the CHUR. The residual solids which remain behind after retraction of the magma show higher isotopic ratios than that recognized in chondritic reservoir. Because Sm and Nd belongs to the REE group with atomic numbers only two units apart, their chemical properties are very similar, thus they come under slight relative fractionation during crystal-liquid processes. This means that the isotope ratios in terrestrial rocks compared to the same ratio from CHUR shows no significant differences. Therefore, DePaolo and Wasserburg (1976a) introduced the “epsilon parameter” (ϵNd) whereby initial $^{143}\text{Nd}/^{144}\text{Nd}$ isotope ratio is represented in part per 10^4 deviation from the CHUR evolution line. Mathematically, the equation is presented as:

$$\epsilon\text{Nd} (t) = \left[\frac{(^{143}\text{Nd}/^{144}\text{Nd})_{\text{sample}} (t)}{(^{143}\text{Nd}/^{144}\text{Nd})_{\text{CHUR}} (t)} - 1 \right] \times 10^4$$

where t is the time at which ϵNd is calculated. The ϵ notation allows for easier comparison of initial Nd isotope ratios of crustal bodies of different ages. The present-day CHUR value is ~ 0.512638 and $^{147}\text{Sm}/^{144}\text{Nd} \sim 0.1966$ (Wasserburg *et al.* 1981). A positive epsilon values means that the rocks were originated from residual solids in the reservoir after magma had been retracted at an earlier time. This part of the reservoir is considered to be “depleted” in the “large ion lithophile” (LIL) elements which are concentrated in the liquid phase during partial melting. A negative epsilon values indicates that the rocks were probably derived from/or assimilated from old crustal rocks whose Sm/Nd has been lowered primary, when they were separated from the mantle. Finally, when the rocks take the value $\epsilon\text{Nd} = 0$, indisputably, they are derived directly from chondritic reservoir (Fig. 13).

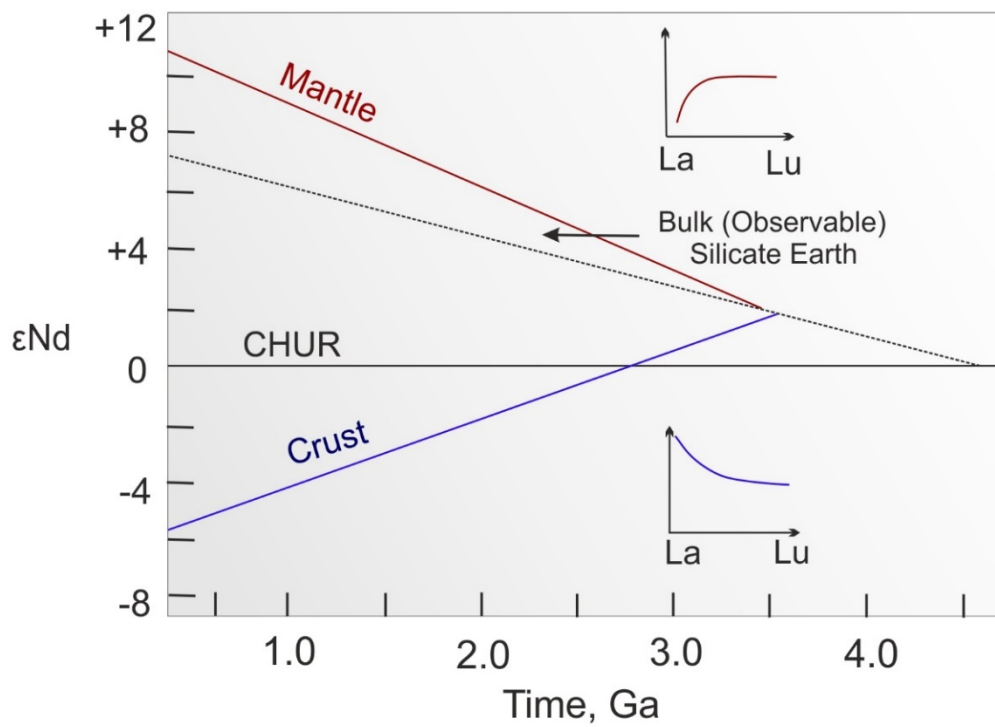


Fig. 13. Evolution of bulk earth, crust and mantle when $^{143}\text{Nd}/^{144}\text{Nd}$ is transformed to ϵNd (from White 2009).

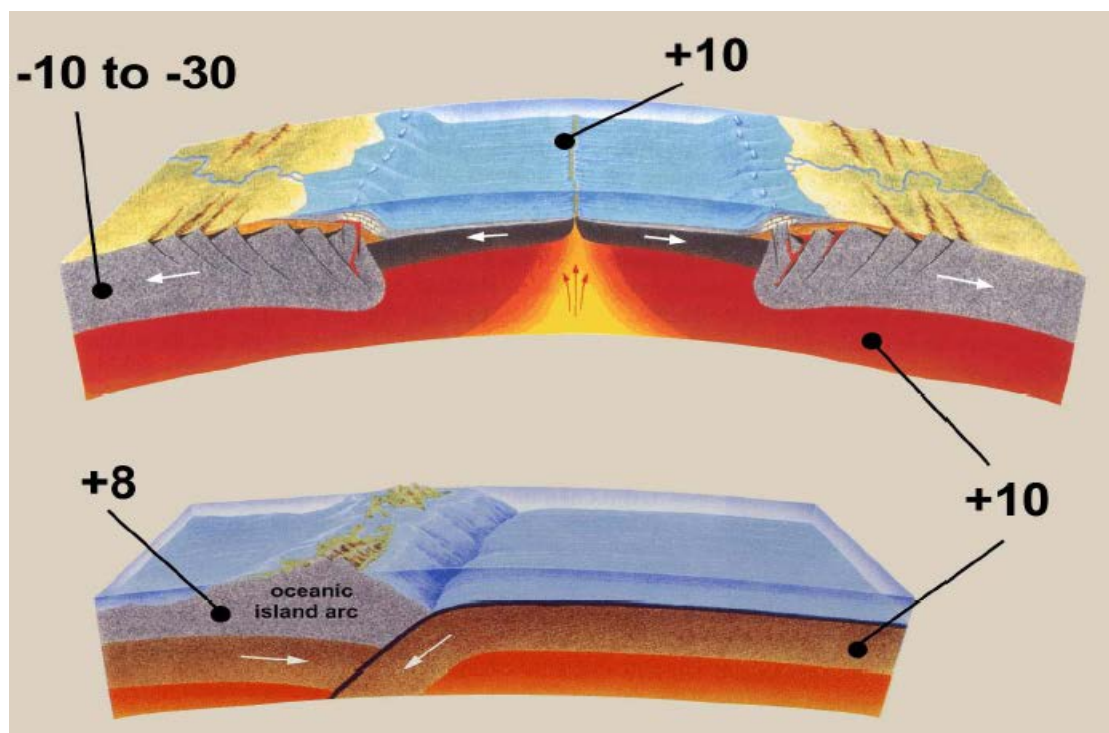


Fig. 14. Neodymium isotope composition of the upper mantle and continental crust, (taken from Dopieralska 2003, based on data of DePaolo 1988).

Different types of crustal rocks reveal varied values of the ϵNd from +10 to -30. For instance, the old continental crust reveals negative ϵNd values ranging from -10 to -30 (DePaolo 1988). As James (1982) and Farmer & DePaolo (1984) demonstrated the volcanic rocks appearing within active continental margin like those in Andes and in California are characterized by much higher ϵNd values than the continental crust. They exhibit values ranging from -13 to +6. The rocks of the oceanic crust are derived from depleted upper mantle with originally high Sm/Nd ratios thus, display positive ϵNd values (Fig. 14). Contemporary MORB shows a uniform ϵNd values of +10, however, the ϵNd values for Phanerozoic ophiolite complexes are about 2 units lower, usually about +8 (DePaolo 1988).

Assuming that crustal rocks are created by partial melting and that the $^{147}\text{Sm}/^{144}\text{Nd}$ ratios of crustal rocks have not changed since time when they were forming, it becomes possible to estimate the “crustal residence time”, i.e. the time in which the rocks spend in the crust. This so-called “model age” is calculated from the basic evolution of the Sm-Nd system. The model ages can be estimated for any individual rock from a single pair of parent-daughter isotopic ratios. It is also necessary to make an assumption about the isotopic composition of the mantle source from which the rocks were primarily derived. In the case of neodymium there are two models of the mantle reservoir: CHUR and DM (Depleted Mantle). The first model assumes that the Earth’s original mantle had the same isotopic signature as the average chondritic meteorite during a creation of the Earth. The second one, calculated to CHUR, describes the time when the rock is extracted from the mantle and gained a different Sm/Nd ratio. It is likewise the time when the sample had the same $^{143}\text{Nd}/^{144}\text{Nd}$ ratio as CHUR. Mathematically, the equation is as follows:

$$\tau_{CHUR} = \frac{1}{\lambda} \ln \left(\frac{^{143}\text{Nd} / ^{144}\text{Nd}_{sam} - ^{143}\text{Nd} / ^{144}\text{Nd}_{chon}}{^{147}\text{Sm} / ^{144}\text{Nd}_{sam} - ^{147}\text{Sm} / ^{144}\text{Nd}_{chon}} + 1 \right)$$

Detailed examination of the initial $^{143}\text{Nd}/^{144}\text{Nd}$ ratio from Precambrian rocks indicates that the mantle, which is the supplier of the continental crust, has evolved along steeper slope to higher $^{143}\text{Nd}/^{144}\text{Nd}$ values than CHUR. Thus, model ages for continental crust are usually estimated with the reference to the depleted mantle rather than CHUR. Mathematically, the depleted mantle model age is obtained by replacement the appropriate DM values in place of $(^{143}\text{Nd}/^{144}\text{Nd})_{CHUR}$ and $(^{147}\text{Sm}/^{144}\text{Nd})_{CHUR}$ in equation. Both concepts presume that the model age is calculated by extrapolating the $^{143}\text{Nd}/^{144}\text{Nd}$ ratio back to the intersection with the depleted mantle/or CHUR growth curve (Fig. 15).

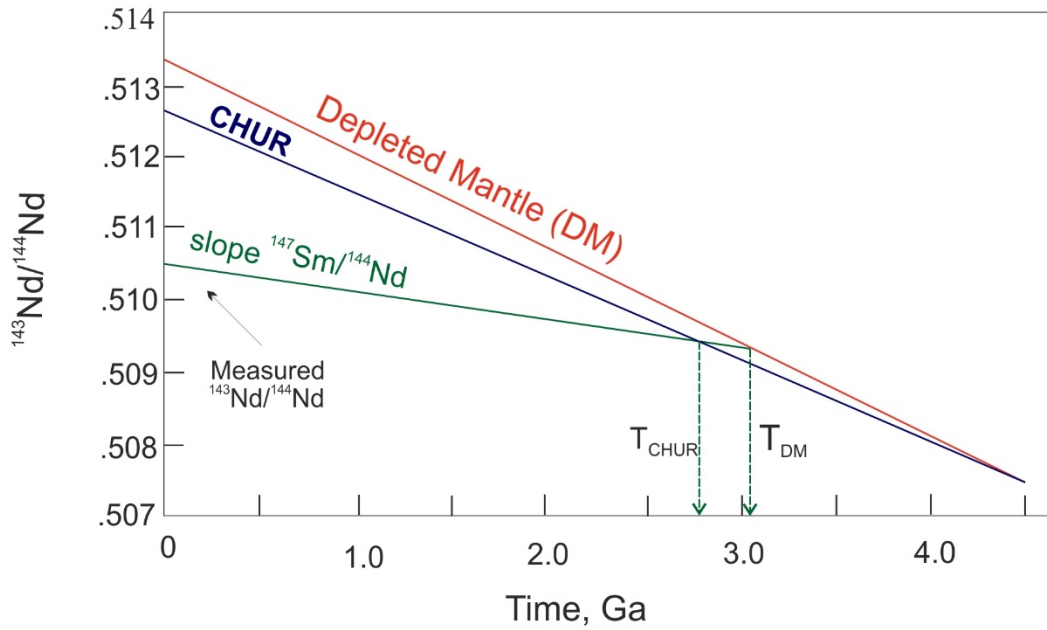


Fig. 15. Graphic diagram of Nd model ages; the $^{143}\text{Nd}/^{144}\text{Nd}$ is extrapolated backward until it meets with a chondritic or depleted mantle growth line (from White 2008).

Sometimes, the Sm-Nd system can be modified during erosion and sedimentation (e.g., Bock *et al.* 1994), so that a second phase of Sm/Nd fractionation take part (first occurred at the mantle melting). If so, better constraints on the model age of the source can be obtained by applying a “two stage” model age calculation (Keto and Jacobsen 1987). The two-stage model ages ($T_{2\text{DM}}$) can be obtained from the single stage T_{DM} model age by equation:

$$T_{2\text{DM}}^{\text{Nd}} = T_{\text{DM}}^{\text{Nd}} - (T_{\text{DM}}^{\text{Nd}} - T_{\text{START}}) \left[\frac{f_{\text{CC}} - f_{\text{meas}}}{f_{\text{CC}} - f_{\text{DM}}} \right]$$

where $f_{\text{DM}} = 0.08592$ is the $f_{\text{Sm}/\text{Nd}}$ value of the DM reservoir (Keto & Jacobsen 1987). A very significant feature of REEs is their stable behavior in the sedimentary environment and also during high-grade metamorphism (Green *et al.* 1969). During the forming of fine-grained sediments from crystalline rocks there is only a little change in the relative abundance of these elements. Thus, the Sm/Nd ratios in fine-grained sediments do not generally differ much from the ratio in the precursor crystalline rock. It is widely accepted that Sm-Nd isotopic system is powerful instrument to interpret the geotectonic background of both magmatic and sedimentary rocks.

STRONTIUM ISOTOPE GEOCHEMISTRY

Strontium is a metal which belongs to alkaline earth elements. It has four naturally occurring isotopes ^{84}Sr (0.56%), ^{86}Sr (9.86%), ^{87}Sr (7.0%) and ^{88}Sr (82.58%) but only one of them ^{87}Sr is radiogenic and is produced by radioactive decay of ^{87}Rb , which has a half-life of 4.88×10^{10} years. Strontium substitutes Ca^{2+} and is present in minerals like plagioclase, apatite and calcite. Rubidium has two naturally occurring isotopes ^{87}Rb and ^{85}Rb . In turn, Rb substitutes K in potassium-bearing minerals like muscovite, biotite in potassium feldspar and in some clay and evaporate minerals. Generally, there are two sources of ^{87}Sr in any given material, the first one was formed during primordial nucleosynthesis along with ^{84}Sr , ^{86}Sr and ^{88}Sr and the second one is created permanently by radioactive decay of ^{87}Rb . The parameter commonly used in geological studies is the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio, which is derived from the following relationship:

$$\frac{{}^{87}\text{Sr}}{{}^{86}\text{Sr}} = \frac{{}^{87}\text{Sr}}{{}^{86}\text{Sr}_0} + \frac{{}^{87}\text{Rb}}{{}^{86}\text{Sr}}_{\lambda t}$$

The Rb-Sr isotope system is in many respects analogous to the Sm-Nd isotope one. The variation of the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio in rocks of the Earth's crust arises from properties of both rubidium and strontium. Both elements are incompatible in silicate magmatic system during the early stage of fractional crystallization, but Rb is more incompatible than Sr. Thus, the Rb/Sr ratio increase with the degree of differentiation, the highest ratio is recognized in pegmatites, whereas the lowest is observed in carbonatites and Ca-rich plutonic rocks. Strontium first separates from the stop and is preferentially incorporated into plagioclase, leaving Rb in liquid phase. This leads to growth of Rb/Sr ratio in residual magma. In consequence, the rocks originating during progressive differentiation processes have higher Rb/Sr ratio (Fig. 16). The initial $^{87}\text{Sr}/^{86}\text{Sr}$ ratio of the Earth is calculated from the initial ratio in meteorites based on assumption that the whole solar system had a uniform $^{87}\text{Sr}/^{86}\text{Sr}$ ratio at the time it was formed. A new created crust shows a higher Rb/Sr ratio than the mantle because, as was mention above, Rb is more incompatible element than Sr. The residual mantle is characterized by lower Rb/Sr ratio than the bulk Earth. The mafic and ultramafic rocks, which are created from the depleted mantle, are characterized by low $^{87}\text{Sr}/^{86}\text{Sr}$ isotope ratio. In contrast, acid rocks of the continental crust reveal high $^{87}\text{Sr}/^{86}\text{Sr}$ isotope ratios because they are formed during fractional crystallization of magma which is enriched in Rb. During the sedimentary processes, which take place at low temperatures, the behavior of the

Rb and Sr ions is quite different. This is related to the fact that Sr is more susceptible to weathering than Rb (Faure & Mensing 2005).

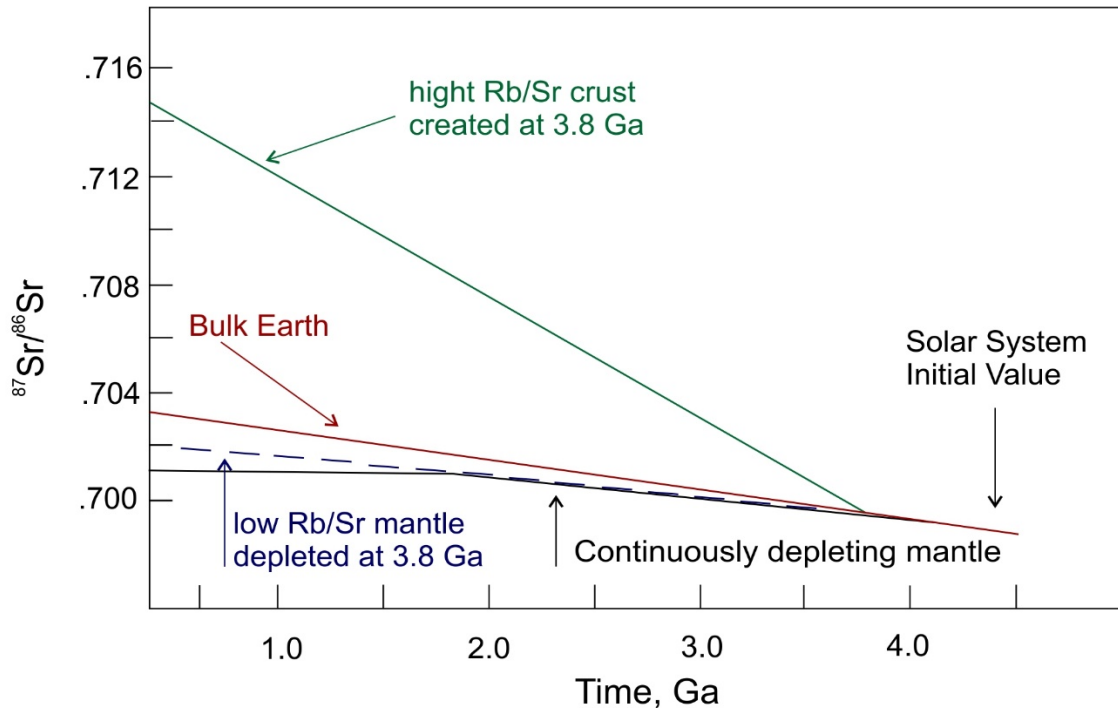


Fig. 16. Strontium isotope evolution of the bulk earth; example of high Sr/Rb crust formed at 3.8 Ga, with evolution of the residual mantle and evolution of the mantle being continuously depleted (from White 2009).

Rubidium is also more readily absorbed by clay minerals, while Sr is removed from crystal lattice of the minerals and incorporated to the interstitial fluids. As a result, the $^{87}\text{Sr}/^{86}\text{Sr}$ ratio in fine grained sediments enriched in clay minerals became very homogenous. The rubidium-strontium isotope system remains one of the most widely used isotopic whole-rock methods but it can also be used to determine the isotopic composition of individual minerals.

MATERIAL AND METHODS

Field work

A total of 87 samples of rocks were collected in 34 localities distributed across the whole area of the Holy Cross Mountains (Fig.17 and Tab. 2). Rock samples, each about 250 g, included various coarse- and fine-grained clastics, Early Cambrian to Late Silurian in age. A summary



Fig. 17. Simplified geological map of the Holy Cross Mountains to show Paleozoic rocks and location of sampled exposures.

of the samples analyzed in this study is given in Tab. 2 and in Appendix. Wherever it was possible samples were collected from formations and localities sampled for detrital zircons and micas during the previous provenance studies.

Analytical procedure

The Nd and Sr isotopic analyses were carried out at the Isotope Laboratory of the Adam Mickiewicz University in Poznań. Prior to isotopic analysis, all samples were cleaned using deionized water. Once dried, the samples were ground to a fine powder with an agate mortar and pestle. From 50 to 95 mg of powdered sample were carefully weighed, spiked with a mixed ^{149}Sm – ^{150}Nd spike, and dissolved with concentrated HF-16N HNO_3 in closed Teflon vials on a hot plate for a few days. After this stage, the solution was evaporated to dryness and re-dissolved in 3 ml 6 N HCL. Finally, the samples were treated with 300 μl of 2N HNO_3 . In order to separate any insoluble particles from the solution, samples were centrifuged for 15 minutes with 15 000 RPM (rotation per minute). Sr and REE were separated using standard cation exchange methods in 50 μl columns filled with the Sr and TRU Resins, respectively (Fig. 18).

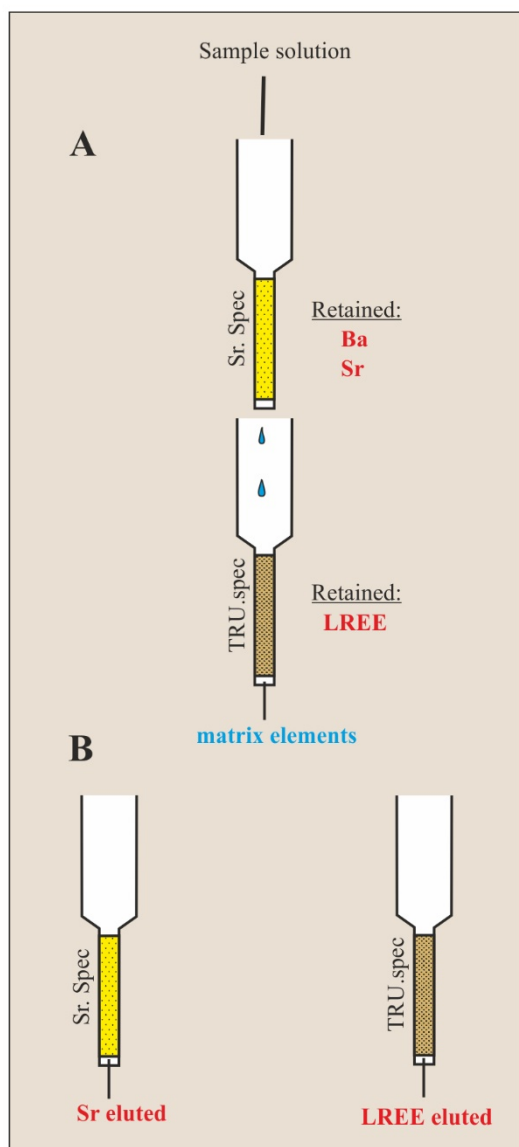


Fig. 18. Scheme of extraction chromatographic procedure to separate Sr and light REE using Sr. Spec. and TRU. A-Loading of sample and extraction step. B- elution step (from Dopieralska 2003).

For separation of Nd and Sm large, 2 ml columns filled with the Ln Resin were used. The columns with Sr and TRU Resins were washed with 0.5 ml of deionized H₂O, preconditioned with 0.3 ml of 2 N HNO₃ and coupled in a sequence. Subsequently, centrifuged sample solutions were loaded into the upper, Sr columns. Afterwards, the resin was rinsed successively with 0,25 ml and twice with 0,2 ml portions of 2N HNO₃, each time waiting before the whole acid added in the former step soaked into the resin. After this step, both columns were decoupled and Sr is eluted after 1 ml of deionized H₂O and collected into Teflon vials. During the next step, the TRU columns were coupled with Ln columns, and REE's were eluted from the TRU Resin with 1,3 ml H₂O. The columns were then decoupled and the Ln Resin was rinsed three times with 0.25 ml of 0.25 N HCL. The following 11 ml fraction of 0,25 N HCL, enriched in La, Ce and Pr were discarded, and subsequently Nd was

collected into clean Teflon vials with 6 ml of the same acid. In the next step, the columns were rinsed with 2 ml of 0,25 N HCL, the eluent was switched to 0,75 N HCL and Sm was collected in a 3 ml fraction. The collected Sm and Nd fractions were evaporated to dryness on a hot plate. Subsequently, Sm and Nd were dissolved in 4 µl of the 1N HCL and 0,35N H₃PO₄ mixture, and loaded on the Re double filament. Sm and Nd measurements were carried out in dynamic collection mode on a Finnigan MAT 261 mass spectrometer. Total procedural blanks for Nd were <35 pg. Nd isotopic ratios were normalized to ¹⁴⁶Nd/¹⁴⁴Nd = 0,7219, and Sm ratios to ¹⁴⁷Sm/¹⁵²Sm = 0,56081. Repeated measurements of the AMES standard yielded ¹⁴³Nd/¹⁴⁴Nd = 0,512127 ± 7 (2σ, n = 24). The data presented here were corrected to the

accepted value of AMES (0,512140). Nd isotopic analysis are reported in the standard epsilon notation calculated using the $^{143}\text{Nd}/^{144}\text{Nd}$ ratio of CHUR that corresponds to a present-day value of 0,512638 (Hamilton *et al.* 1983). All $\epsilon_{\text{Nd}(t)}$ values were recalculated according to the measured $^{147}\text{Sm}/^{144}\text{Nd}$ ratios for the time of deposition.

RESULTS

Strontium isotopes

The results are summarized in Tab. 2 and in Figs. from 19 to 21. The Cambrian clastics of Małopolska are characterized by an extremely wide range of Sr isotope ratios, from 0.712 to 0.767 (Fig. 19). The data, however, show a polymodal distribution with at least three distinct detritus populations and a characteristic break in the distribution between 0.724 and 0.730. Moreover, there is also a general tendency that $^{87}\text{Sr}/^{86}\text{Sr}$ ratios generally increase with decreasing grain size. The most non-radiogenic population is represented by medium- and coarse-grained sandstones (or quartzites) whereas mudstones tend to have much more radiogenic composition (Fig. 21).

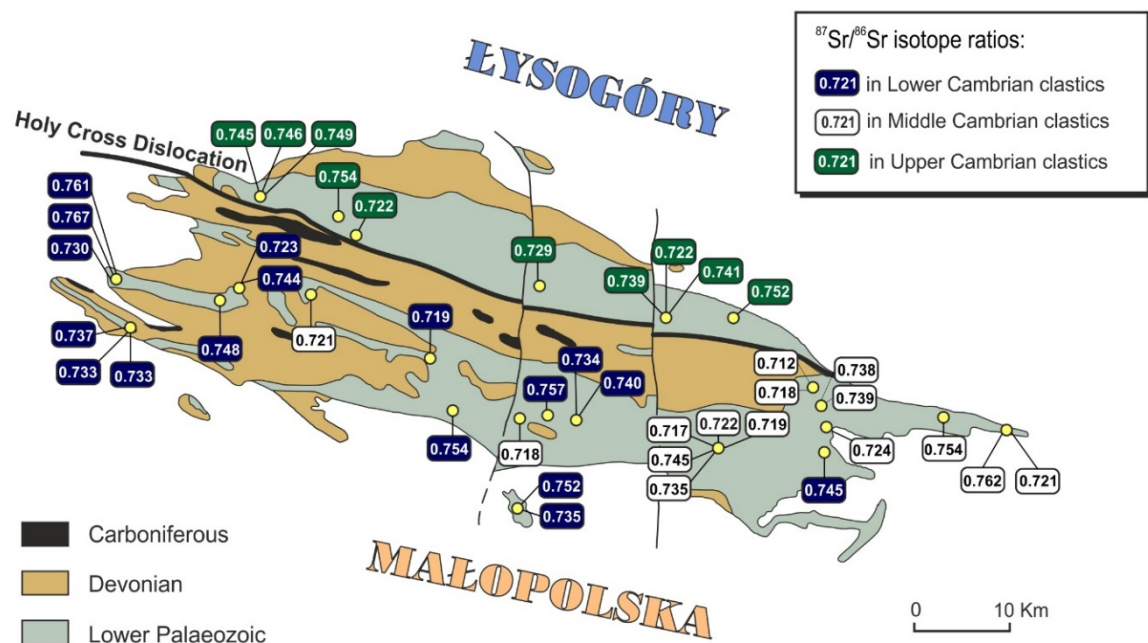


Fig. 19. The isotopic composition of strontium ($^{87}\text{Sr}/^{86}\text{Sr}$) of the Cambrian clastics of Małopolska and Łysogóry.

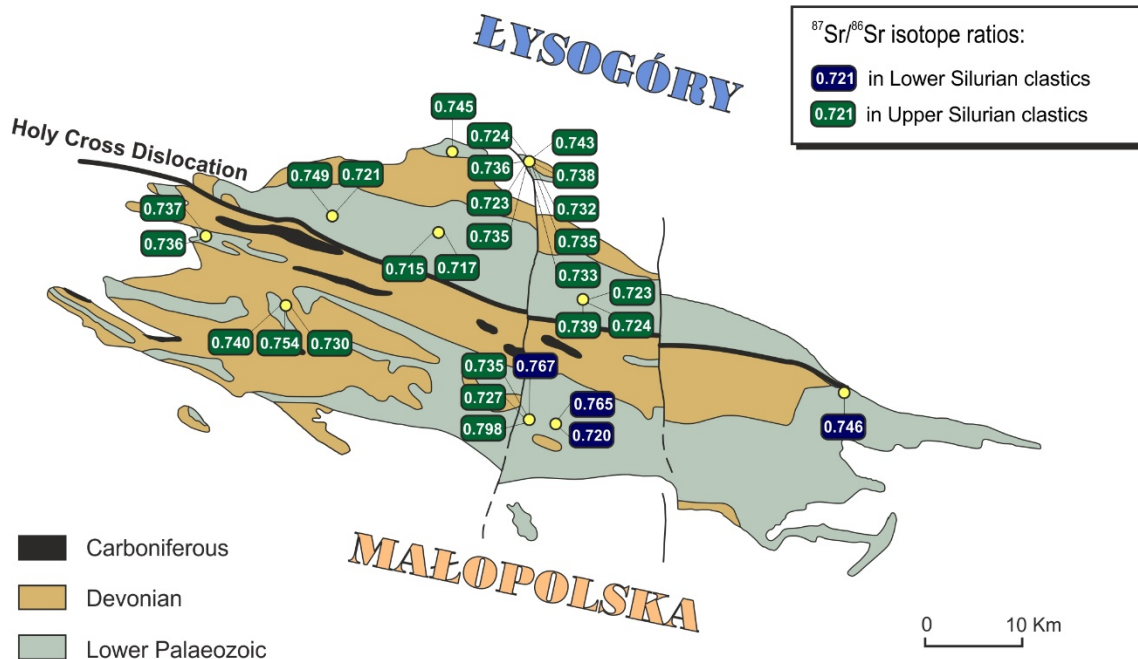


Fig. 20. The isotopic composition of strontium ($^{87}\text{Sr}/^{86}\text{Sr}$) of the Silurian clastics of Małopolska and Łysogóry.

This relationship can also be observed in the Ordovician and Silurian clastics of Małopolska although the data sets are significantly smaller. Ordovician clastics show some similarities to those of the Cambrian. Their $^{87}\text{Sr}/^{86}\text{Sr}$ values vary within a similar range and the dominant population is represented by the most non-radiogenic detritus, with ratios between 0.711 and 0.724. In contrast, the Silurian clastics of Małopolska appear to exhibit a different Sr isotope composition. The rocks contain more radiogenic detritus, with $^{87}\text{Sr}/^{86}\text{Sr}$ values from 0.720 to 0.798 (Fig. 20). Although the number of samples from Łysogóry was relatively small, the Sr isotope data of clastics irrespective of their age are also here clearly influenced by variations in grain size. It is almost a rule that mudstones are more radiogenic than sandstones and greywackes (Fig. 21). The Sr isotope composition of clastic rocks in Łysogóry varies within much narrower ranges when compared to Małopolska. The Cambrian clastics have $^{87}\text{Sr}/^{86}\text{Sr}$ values that range from 0.722 to 0.754 (Fig. 20). Admittedly, the Sr isotope signatures of Silurian rocks fall within a similar range, from 0.715 to 0.749, but they appear to show a bimodal distribution.

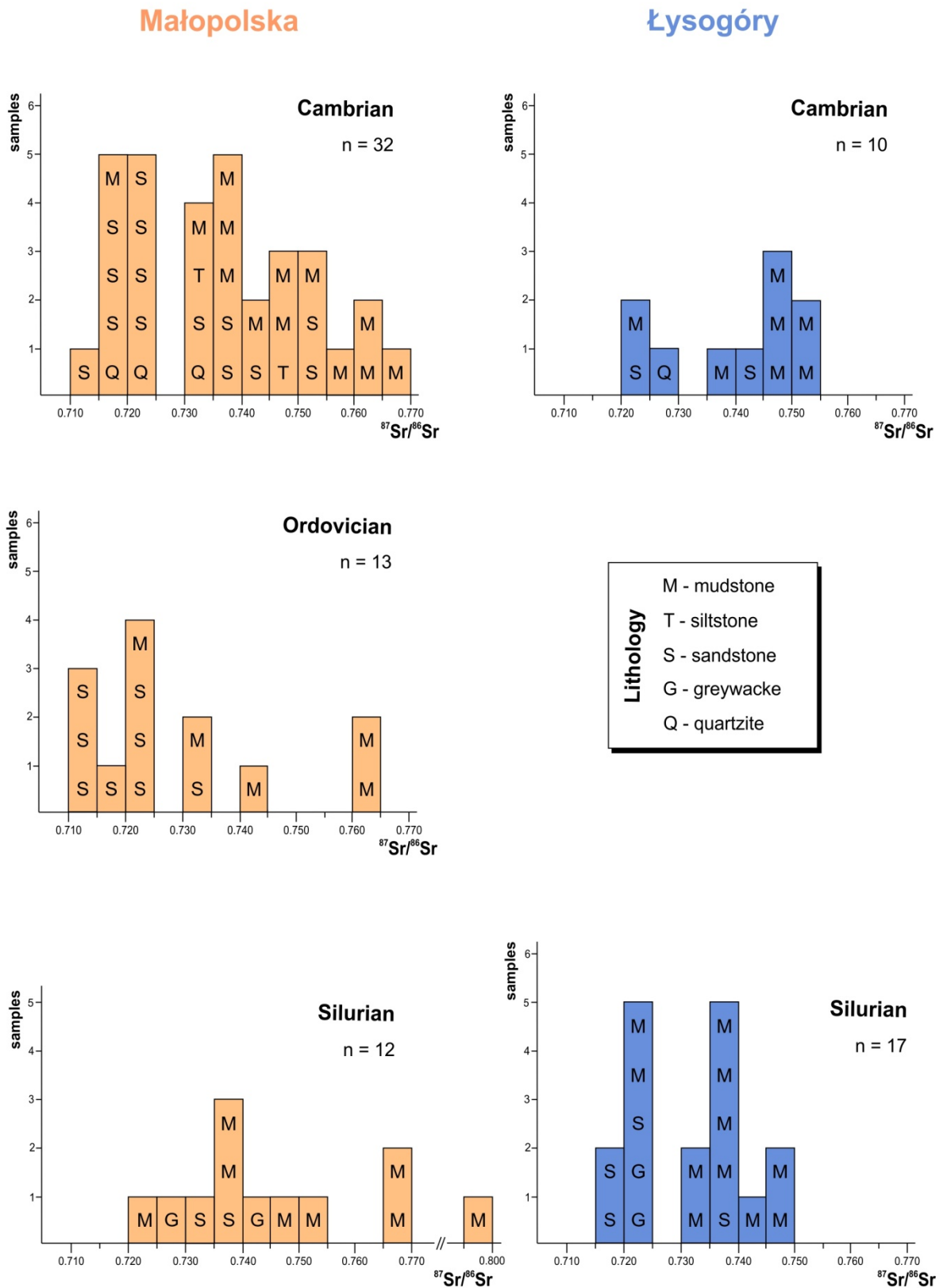


Fig. 21. Histograms showing the relationship between grain sizes and strontium isotope ratios in the Lower Paleozoic clastic sediments of Małopolska and Łysogóry.

Neodymium isotopes

The results are summarized in Tab. 2 and in Figs. from 22 to 25. Unlike the Sr isotope ratios, the Nd isotope composition of investigated clastics does not show any dependence on grain size (Fig. 22B). The major feature of the collected Nd data is that the Cambrian clastic rocks of Małopolska differ noticeably in their Nd isotope composition from those of Łysogóry. Clastics from Małopolska are characterized by a wide range of the $\epsilon_{Nd(t)}$ values, from -11.5 to -5.5 , whereas in Łysogóry the $\epsilon_{Nd(t)}$ values vary within a fairly narrow range, from -11.1 to -8.4 (Fig. 22A). Although these ranges are overlapping a remarkable difference is visible because rocks with more radiogenic composition $\epsilon_{Nd(t)}$ values generally higher than -9.0 dominate in Małopolska (Fig. 22A). In contrast, the Cambrian clastics of Łysogóry exhibit more unradiogenic $\epsilon_{Nd(t)}$ values which, with exception of a single sample, are lower than -9.0 . A clear disparity in clastic material of the Cambrian of Małopolska and of Łysogóry can also be observed in its model age values (T_{2DM}). The clastics of Małopolska are characterized by a wide range of model ages of from 1.71 Ga to 2.18 Ga, but ages younger than 1.9 Ga predominate. These data are contrasted by T_{2DM} model ages from the Cambrian clastics of Łysogóry which all are older than 1.9 Ga (Fig. 23A).

A certain temporal trend can be observed in the wide spectrum of the Nd isotope composition of Cambrian rocks of Małopolska. The most radiogenic detritus, which is the most frequent one, with the $\epsilon_{Nd(t)}$ values higher than -7 and T_{2DM} model ages younger than 1.8 Ga, occurs predominantly in the Lower Cambrian deposits of the western part of Małopolska. The Middle Cambrian rocks, however, contain predominantly clastic material which is less radiogenic and yielding older model ages (Figs. 22A, 23A). Compared to the Cambrian the number of isotopic measurements performed on Ordovician rocks was small. Nevertheless, the Nd isotope data show that clastic material contained in the Ordovician rocks of Małopolska appears to derive from similar sources to that of the Middle Cambrian. The samples exhibit $\epsilon_{Nd(t)}$ values which vary within a relatively narrow range, from -5.8 to -8.8 . Their T_{2DM} model ages range between 1.67 Ga to 1.93 Ga (Fig. 24A, 25A). In Łysogóry, however, samples from one section only were available. Two samples have $\epsilon_{Nd(t)}$ values around -10 and T_{2DM} model ages around 2 Ga. This clastic material corresponds well with isotopic characteristics of the Upper Cambrian clastics in the region. One sandstone sample, however, has a more radiogenic Nd isotope composition ($\epsilon_{Nd(t)} = -6.6$) and much younger

model age of 1.74 Ga, and thus, it resembles the clastic material present in Małopolska within the Cambrian and the Ordovician sequences.

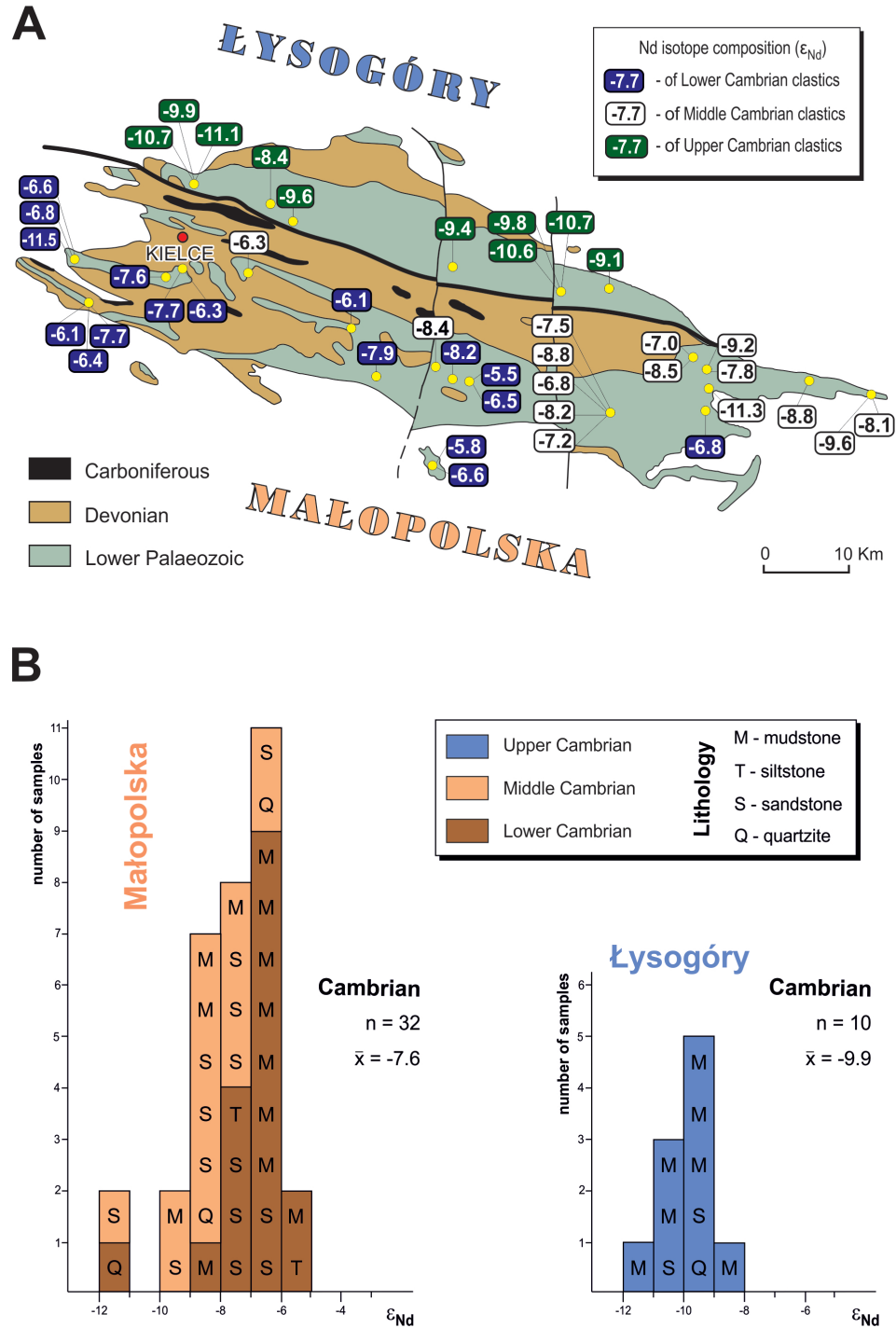


Fig. 22.A. The neodymium isotope composition ($\epsilon_{Nd(t)}$) of the Cambrian clastics of Małopolska and Łysogóry. **B.** Dataset from Fig. 22A presented as frequency distribution of the Nd isotope signatures with their relation to lithology and stratigraphy.

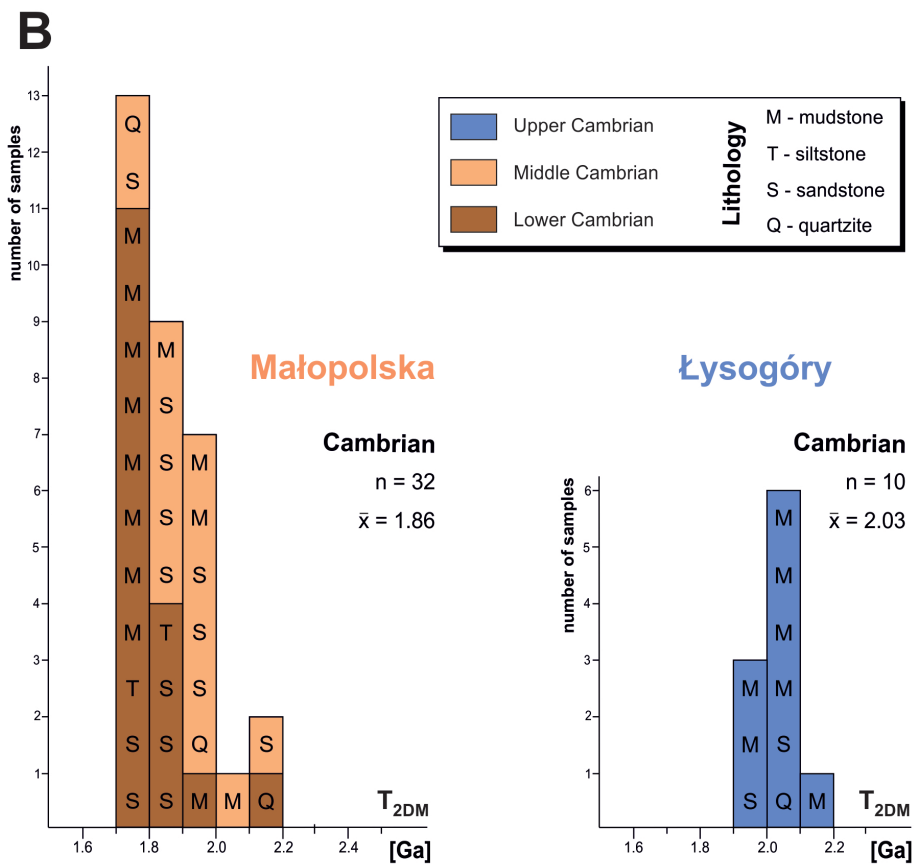
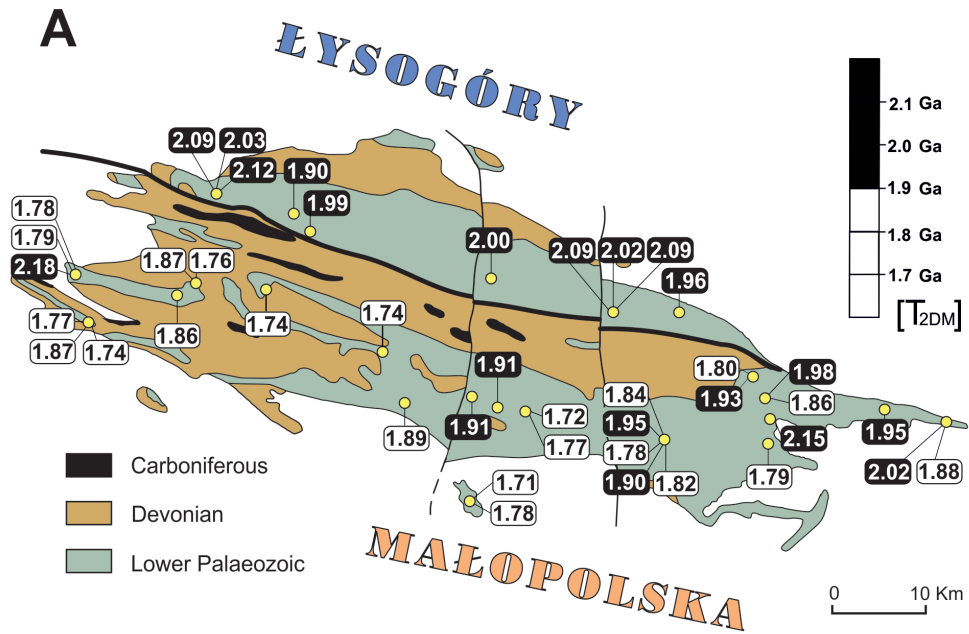


Fig. 23.A The neodymium model age values (T_{2DM}) of the Cambrian clastics of Małopolska and Łysogóry. **B.** Dataset from Fig. 23A presented as frequency distribution of Nd model ages with their relation to lithology and stratigraphy.

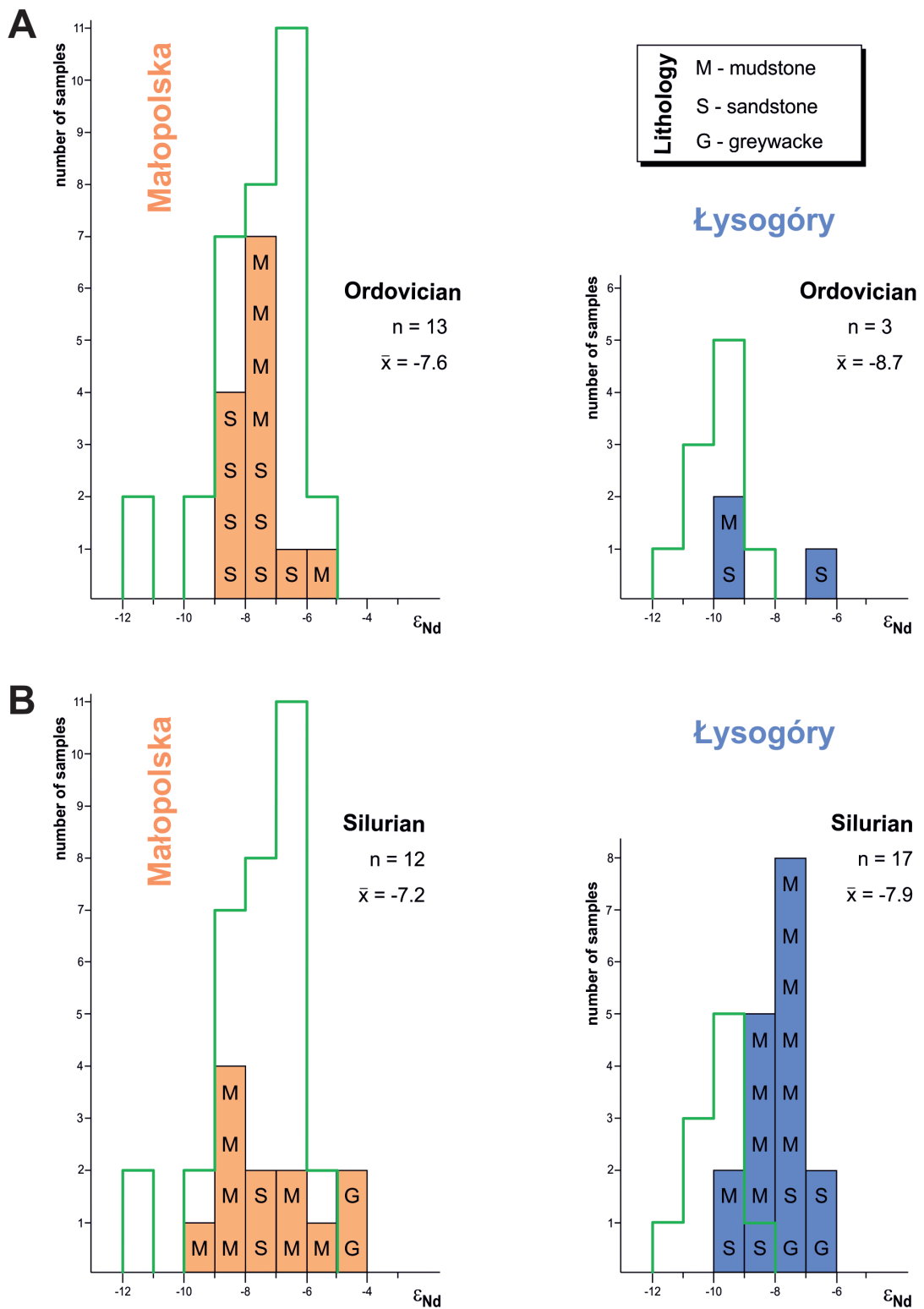


Fig. 24. Histograms to show the frequency distribution of $\epsilon_{Nd(t)}$ values of the Ordovician (A) and Silurian (B) clastic deposits of Łysogóry and Małopolska in relation to $\epsilon_{Nd(t)}$ values of Cambrian clastics (green contour).

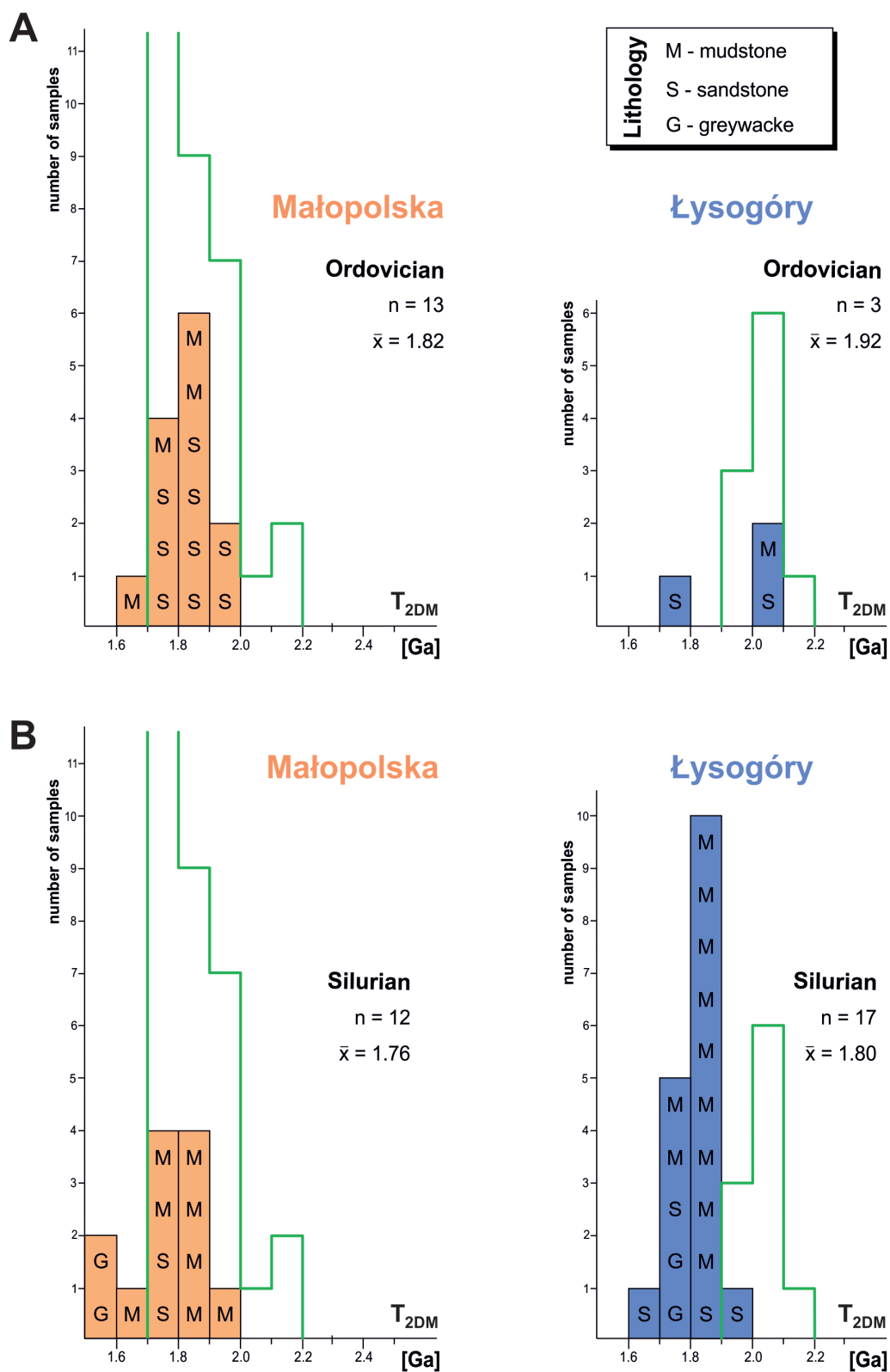


Fig. 25. Histograms to show the frequency distribution of Nd model ages values (T_{2DM}) of the Ordovician (A) and Silurian (B) clastic deposits of Łysogóry and Małopolska in relation to model ages of Cambrian clastics (green contour).

The Nd isotope compositions of the Silurian clastics of Łysogóry differ significantly from those of the Cambrian rocks of this terrane. The clastic material with $\epsilon_{\text{Nd}(t)}$ values ranging from -9.8 to -6.0 is more radiogenic (Fig. 24B). In addition to this, its $T_{2\text{DM}}$ model ages, with a mean value of 1.80 Ga, are clearly younger. The majority of Silurian samples in Małopolska exhibit also similar isotopic characteristics (Fig. 24B, 25B). The exception, however, constitute greywackes sampled at Bardo (BA) and Niestachów (NI), which contain radiogenic detritus with $\epsilon_{\text{Nd}(t)}$ values of 4.5 and 4.4, and $T_{2\text{DM}}$ model ages of 1.53 and 1.52 Ga, respectively.

DISCUSSION

Provenance of clastic material

Strontium isotopes have been successfully applied to characterize sediment provenance and to reconstruct history of sediment supply to basins (e.g., Weldeab *et al.* 2002; Farmer *et al.* 2003; Mahoney 2005; Hemming *et al.* 2007; Ahmad *et al.* 2009; Révillon *et al.* 2011; Dou *et al.* 2012). Although $^{87}\text{Sr}/^{86}\text{Sr}$ ratios of detritus appear to fingerprint generally well the isotopic composition of its source rocks they can be influenced by grain-sized sorting and diagenesis of the sediment (e.g., Dasch 1969; Biscaye & Dasch 1971; Blum & Erel 2003; Meyer *et al.* 2011). Moreover, several studies showed that temporal shifts in Sr isotope composition of sediments were caused by changes in the degree of physical versus chemical weathering on land (e.g., Jung *et al.* 2004; Wang *et al.* 2007; Cole *et al.* 2009).

Variations of Sr isotope signatures observed in Lower Paleozoic clastics of Małopolska and Łysogóry suggest a complex provenance of clastic material. Wide ranges of $^{87}\text{Sr}/^{86}\text{Sr}$ values can be attributed to mixing of material from many different sources. In addition, there is no doubt that clastic material underwent different modes of transports and weathering, which were characteristic by specific grain-size distribution. This is indicated by a relationship between Sr isotope ratios and grain-size. Sr isotope signatures of detritus higher than 0.710 are indicative of the provenance exclusively from continental crust domains in which rocks of the upper crust (granites and gneisses) were essential constituents exposed to erosion and weathering. The comparison of the Sr isotope composition of Cambrian rocks in Małopolska and Łysogóry suggests that these crustal units were supplied with a clastic material from different areas. A similar scenario still existed during the Silurian, despite the fact that the

petrographic composition of clastic material, compared to the Cambrian time, has changed dramatically in both units (Kozłowski *et al.* 2014).

Contrary to behavior of strontium in sedimentary environments, the Sm-Nd isotope system usually remains undisturbed during sedimentary and diagenetic processes (e.g., DePaolo 1988; Goldstein & Jacobsen 1988; Blum & Erel 2003). In addition, the Nd isotope signatures are not dependent on grain size of detritus (Meyer *et al.* 2011). All this makes that the Nd isotope composition of clastics is regarded as a very good indicator of sediment provenance (e.g., McCulloch & Wasserburg 1978; McLennan *et al.* 1990; McLennan & Hemming 1992; McDaniel *et al.* 1997; Fagel 2004). The method is especially useful for fine-grained siliciclastic rocks and dust (e.g., Taylor *et al.* 1983; Allegre & Rousseau 1984; Goldstein *et al.* 1984; DePaolo 1988; Innocent *et al.* 2000; Cole *et al.* 2009; Allan *et al.* 2013; Konieczna *et al.* 2015) because it is generally impossible to determine the provenance of fine-grained detritus petrographically. The Nd isotope signatures, commonly expressed as $\epsilon_{Nd(t)}$ values, reflect the average value of the crust from which the sediments are derived. They also allow to decipher a geotectonic regime under which rocks in the source area were formed (e.g., McLennan *et al.* 1990; Thorogood 1990; Murphy & Nance 2002; Wade *et al.* 2005; Zhang *et al.* 2007). Another advantage of Nd isotopes is that large fragments of the continental crust can be characterized by consistent model ages. Hence, Nd model ages measured on clastic rocks constitute an additional, useful parameter in provenance studies which may help in identification of detritus source areas (e.g., Goldstein & Jacobsen 1988; McLennan *et al.* 2000; Dantas *et al.* 2009; Drost *et al.* 2011; Linnemann *et al.* 2012).

The $\epsilon_{Nd(t)}$ values of the investigated clastic rocks of both terranes, Małopolska and Łysogóry, prove the provenance of terrigenous detritus from continental crustal domains. However, the obvious differences in Nd isotope signatures and model ages provide a strong indication that Małopolska and Łysogóry were supplied with clastic material from different source areas during Cambrian and Ordovician times. In addition, the data demonstrate also that clastic material was much better mixed prior to deposition in Łysogóry than it was the case in Małopolska. This conclusion is consistent with earlier petrographic examinations of Cambrian rocks (e.g., Czerwiński 1959; Michniak 1969), which revealed the presence of highly mature detritus in the Cambrian clastics of Łysogóry and rather an immature composition of Cambrian rocks in Małopolska. The Nd isotope composition of Cambrian clastics of Łysogóry ($\epsilon_{Nd(t)}$ values = -11.1 to -8.4; Nd model ages T_{2DM} = 1.90 to 2.12 Ga) allows, in

fact, only one conclusion: the clastic material must have been derived from the Svecofennian basement of Fennoscandia. The central and southern Sweden appears to represent one of the most likely source area. Wilson *et al.* (1985) reported here the Svecokarelian and post-Svecokarelian granites yielding a similar range of Nd model ages to that of the Łysogóry clastics (Fig. 26). Consequently, there is also a perfect overlap with the Nd isotope characteristics of the Mesoproterozoic clastics which were deposited on the Svecofennian basement in Sweden (Miller *et al.* 1986). A second potential source area, with Svecofennian rocks exposed to erosion during Cambrian times, could constitute NE Poland, Lithuania and Belarus where model ages of the crystalline basement rocks range from 1.99 to 2.32 Ga, and among which ages between 2.0 and 2.1 Ga clearly predominate (Claesson *et al.* 1995; Claesson & Ryka 1999; Claesson *et al.* 2001).

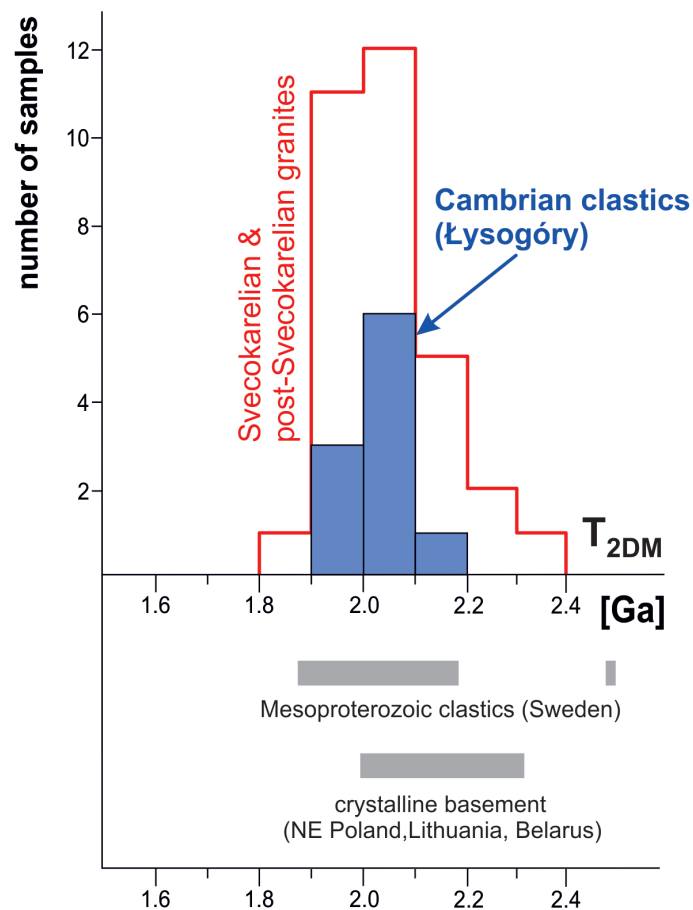


Fig. 26. Nd model ages (T_{2DM}) of Cambrian clastic rocks of Łysogóra to compare with model ages of Svecofennian crystalline and sedimentary rocks in central Sweden, NE Poland, Lithuania and Belarus (data from Wilson *et al.* 1985; Miller *et al.* 1986; Claesson *et al.* 1995; Claesson & Ryka 1999; Claesson *et al.* 2001).

In contrast to Łysogóry, the Baltic provenance of detritus of the Cambrian and Ordovician clastics of Małopolska is very unlikely. This is because that rocks having Nd model ages of 1.7-1.8 Ga, which are the dominant ages among Małopolska clastics, are very rare in the Precambrian basement of the East European Platform (= paleocontinent Baltica). Although there are many small, mafic intrusions (dolerites) along the Protogine Zone in southwestern Sweden (Johansson & Johansson 1990), these rocks cannot be considered as a potential detritus source because they are characterized by much more negative $\epsilon_{Nd(t)}$ values than those of Małopolska clastics. Claesson *et al.* (1995) reported Nd model ages between 1.7 and 2.3 Ga from an anorthosite and gabbro complex at Suwałki (NE Poland). However, they did not consider them as true protolith ages but as a result of REE fractionation during the formation of anorthosite-gabbro rock association.

Taking into consideration the wide ranges of $\epsilon_{Nd(t)}$ values and model ages of the Cambrian clastics of Małopolska ($\epsilon_{Nd(t)} = -11.5$ to -5.5 ; $T_{2DM} = 1.71$ to 2.18 Ga), it appears that Gondwanan crust is the best candidate for such a source reservoir. At the same time one must also take into account that Nd isotope characteristics of rocks may not represent the real variation in $\epsilon_{Nd(t)}$ and crustal residence time of a single source but can reflect probably a mixture of material derived from different sources. On the other hand, it does not appear that the Nd isotope characteristics of the investigated clastics results from a mixing of old Paleoproterozoic and young Neoproterozoic detritus because the distribution of T_{2DM} values is clearly asymmetric (Fig. 27) and there are no records of model ages younger than 1.7 Ga in Małopolska. Such model ages are characteristic of the basement rocks of the Grenville Belt in North America and several Avalonian-Cadomian terranes, West Avalonia, Florida, Carolina, and East Avalonia, for instance (Nance & Murphy 1996). Although the Nd model ages of the Małopolska clastics partly overlap with model ages of basement rocks of Armorica (Cadomia) and Lower Paleozoic sediments of Saxo-Thuringia (Linnemann *et al.* 2004) and Tepla-Barrandian (Drost *et al.* 2007), the provenance of detritus from these terranes is rather very unlikely. These crustal units show a clear link to the West African craton, indicated by a characteristic pattern of zircon ages (Nance & Murphy 1996; Linnemann *et al.* 2004; Drost *et al.* 2007; Nance *et al.* 2008; Drost *et al.* 2011). Because of lack of any magmatic activity within the West African craton, between Panafrican (0.5-0.75 Ga) and Eburnean (1.9-2.2Ga) orogenies, detrital zircons with Mesoproterozoic ages (1.0-1.7 Ga) are absent in the Ediacaran and Cambrian sedimentary sequences of terranes yielding a West African provenance. This is, however, not the case in Małopolska. As already mentioned above (in the chapter *Previous*

Provenance Studies), detrital zircons with ages of ca.1.2–1.37 Ga and 1.5 Ga were identified in the Lower Cambrian sandstones of Małopolska (Belka *et al.* 2002). Therefore, the West African proximity of Małopolska during the Cambrian seems to be excluded.

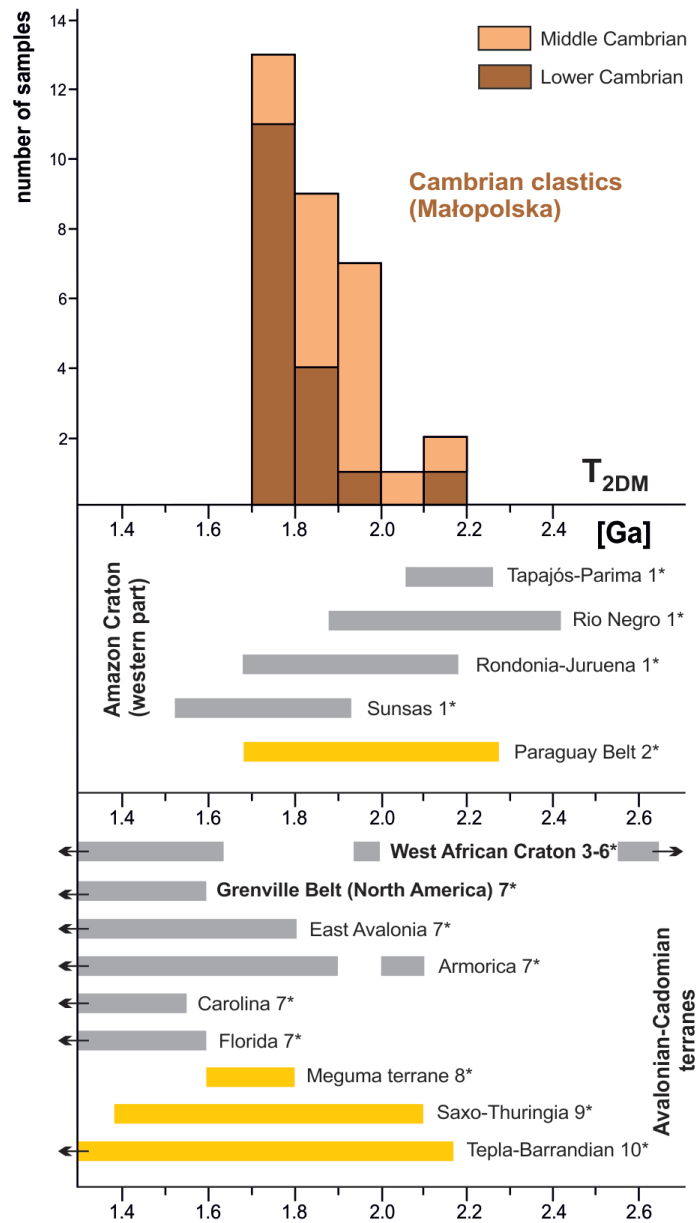


Fig. 27. Nd model ages (T_{2DM}) of Cambrian clastic rocks of Małopolska to compare with model ages of basement crystalline rocks (grey) and sediments (yellow) of the Amazon Craton, Grenville Belt and Avalonian-Cadomian terranes. Note the perfect overlap of ages from Małopolska clastics with those of the Paraguay Belt. Data sources for Nd model ages: 1*- Santos *et al.* (2000). 2*- Dantas *et al.* (2009) – T_{DM} ages recalculated into T_{2DM} ages. 3*- Allegre and Othman (1980). 4*- Toteu *et al.* (1994). 5*- Potrel *et al.* (1998). 6*- Gasquet *et al.* (2005). 7*- Nance and Murphy (1996). 8*- Clarke and Halliday (1985). 9*- Linnemann *et al.* (2004). 10*- Drost *et al.* (2007).

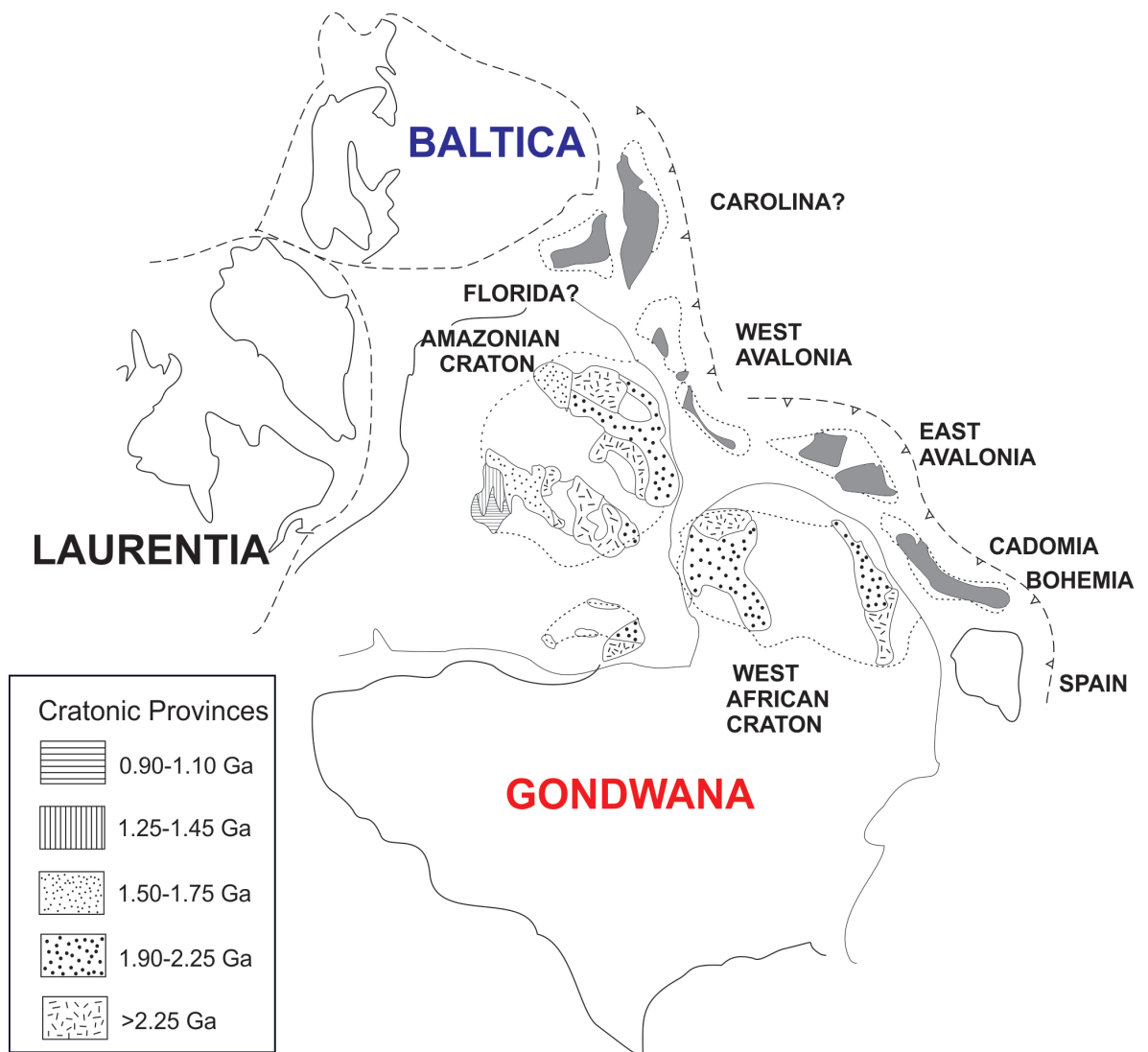


Fig. 28. Neoproterozoic continental reconstruction of Dalziel 1992; with main cratonic provinces from Teixeira *et al.* 1989 (after Nance & Murphy 1996).

Taken together, the Nd isotope data and detrital zircon ages strongly suggest that Małopolska must have been supplied with clastic material from Amazonian sources during Early to Middle Cambrian times. The range of Nd model ages of Cambrian clastics of Małopolska closely matches age provinces of the western part of the Amazon craton (Fig. 27, 28). It is also important to note that the ranges of Nd model ages (T_{2DM}) and $\epsilon_{Nd(t)}$ values of the Małopolska clastics are identical to those of the Neoproterozoic sedimentary succession of the Paraguay Belt, which developed along the western margin of the Amazonian craton (Dantas

et al. 2009). A peri-Amazonian location for Małopolska is also suggested by detrital zircons with ages of 0.54 Ga, ca. 1.2-1.37 Ga, 1.5 Ga, 2.0 Ga, >2.5Ga and >3.0 Ga (Belka *et al.* 2002).

Nd isotope data shed new light on some interpretations of detrital mica ages obtained in the study area in the past and they highlight complexity of various types of provenance data. Belka *et al.* (2000) reported detrital muscovite grains showing K-Ar cooling ages of about 535-545 Ma in the Lower Cambrian sandstones exposed at Kędziorka (Małopolska) and muscovites with a similar age (c. 539 Ma) in the Upper Cambrian quartzites at Wymyślona (Łysogóry). This material has been interpreted to document clastic input of the Cadomian provenance, both in Małopolska and Łysogóry. However, the rocks exposed at Kędziorka and Wymyślona included entirely different populations of detrital zircon grains (Belka *et al.* 2002). It has been shown that zircons recovered from sandstones at Kędziorka correlate well with detrital zircon ages known from Neoproterozoic rocks of West Avalonia and with basement isotopic signatures of the Amazonian craton. In contrary, the provenance of detrital zircon population from Wymyślona was considered to be unclear, because its geochronological signatures of c. 0.6 Ga, 1.8-2.1 Ga and >2.5 Ga are known from both Gondwanan and Baltic sources (e.g., Nance & Murphy 1996; Valverde-Vaquero *et al.* 2000). The present study revealed that rocks at both localities exhibit different Nd isotope compositions. Detritus of the Lower Cambrian sandstones at Kędziorka with $\epsilon_{Nd(t)}$ values of -6.5 to -5.5 is more radiogenic than that at Wymyślona ($\epsilon_{Nd(t)} = -9.6$). Consequently, the Nd model ages T_{2DM} are also different, 1.72 and 1.77 Ga at Kędziorka and 1.99 Ga at Wymyślona. In conclusion, it can be assumed that although detrital mica grains having the same age (540 Ma) occur in Małopolska and Łysogóry, they are most probably delivered from two different Cadomian sources.

Although the number of the Nd isotope data obtained from Ordovician clastics is small, it allows to formulate a preliminary conclusion that at least Małopolska was supplied with clastic material from the same sources as during the Cambrian. It seems that a significant change in material provenance occurred in the Silurian because the Nd isotope characteristics of rocks is for the first time similar within coeval rocks in Małopolska and Łysogóry. Recently, Kozłowski *et al.* (2014) compared the petrological composition of Silurian greywackes in whole area of the Holy Cross Mountains. Based on differences in the petrographic and geochemical composition they concluded that during the late Silurian Małopolska had to be located in a more proximal position to the source area than Łysogóry, at

the same time stating that both regions were supplied from a common source. They postulated a source area, located to the west of the present-day position of the Holy Cross Mountains, in which the basement of the East European Platform collided with a volcanic arc developed on the eastern margin of an Avalonian terrane. If so, a young juvenile material should constitute a substantial portion of the investigated greywackes. The negative $\epsilon_{\text{Nd}(t)}$ values of these rocks, however, indicate clearly that most of the material must have been derived from old continental crust sources.

Paleogeographic implications

Nd isotope data obtained from clastic rocks of Łysogóry and Małopolska provide additional evidence for the previously postulated development of both units as independent crustal blocks during the Early Paleozoic. Nevertheless, the reconstruction of their paleogeographical location is not a simple task, because various types of provenance data do not lead to exactly the same conclusions. Moreover, there are no unambiguous paleomagnetic data for the Cambrian time because in the past there were often wildly contradictory results from similar-age rocks (cf. Meert 2014; Klein *et al.* 2015; Levashova *et al.* 2015; Nawrocki 2015). In particular, the position of Baltica during the Cambrian is poorly constrained.

The crust of Małopolska exhibits clearly Avalonian affinities (Narkiewicz *et al.* 2011). There is a broad consistency of provenance data regarding the proximity of Małopolska to Amazonia during Early and Middle Cambrian times. The Amazonian provenance of clastic material is well supported by both Nd isotope data (this study) and zircon ages (Belka *et al.* 2002). Detrital muscovites clearly indicate a derivation from crust with a Cadomian imprint (Belka *et al.* 2000; Nawrocki *et al.* 2007). It is now apparent that the Cambrian trilobite fauna of Małopolska was endemic, but showing linkages to Baltica, Avalonia and West Gondwana (Żylińska & Szczepanik 2009; Żylińska 2013). Some uncertainty appears to exist only regarding the occurrence of mica grains from Baltic sources in the area around Sandomierz, where the contact of Małopolska with Łysogóry is not precisely defined. However, it is important to note that the clastics including the Baltic-derived muscovites are characterized by strong negative $\epsilon_{\text{Nd}(t)}$ values and old Nd model ages, around 2.0 Ga, supporting also a Baltic provenance. Because the Baltic-derived detritus occurs only in the youngest part of the Cambrian succession of Małopolska (Cambrian Series 3), its appearance can easily be explained by the progress of the Małopolska-Baltica convergence.

In summary, all provenance and biogeographic data indicate that during the Early Cambrian Małopolska must have been situated in the immediate vicinity of the Amazonian craton and near the western end of the Cadomian belt (Fig. 29). It was not until the late Middle Cambrian that Małopolska collided with the Baltica margin. By Middle Cambrian times, as Małopolska progressively approached Baltica, increasing numbers of Baltic brachiopod taxa were able to reach Małopolska (Belka *et al.* 2002) and also clastic material from Svecofennian sources of Baltica was deposited within this crustal unit. Finally, continuing convergence resulted in oblique collision leading to the fold deformations of Cambrian and Precambrian rocks. This event, locally termed the Sandomierz “orogenic phase”, was not accompanied by any magmatic activity.

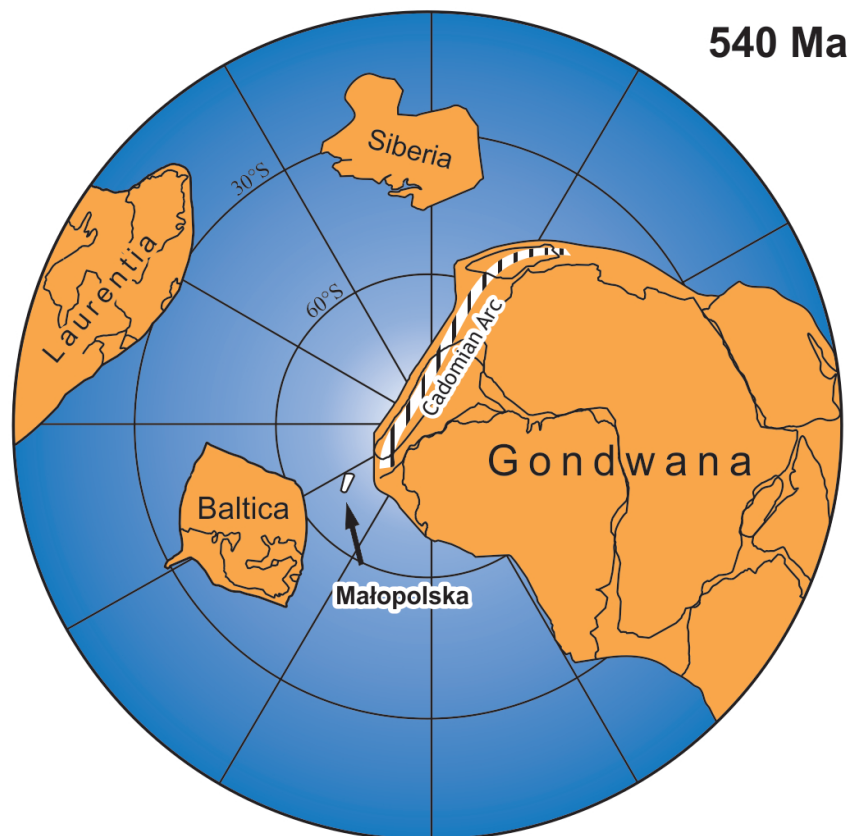


Fig. 29. Early Cambrian (c. 540 Ma) paleogeographic reconstruction to show the position of Małopolska in relation to principal paleocontinents. The location of Laurentia, Siberia, Baltica and Gondwana are taken from the reconstruction of Torsvik & Cocks (2013) with exception of the orientation of Baltica, which is now faced to the west. This orientation is favored because it is compatible with faunal and provenance data.

Detrital muscovites with K-Ar cooling ages of about 1.7 Ga, identified in the Middle Cambrian clastics (Belka *et al.* 2000), indicate that Małopolska must have docked to the Fennoscandian segment of Baltica, along the Tornquist margin, being afterwards probably always located near its present position. Among all available provenance data there is no evidence to constrain the docking of Małopolska at the southern, Sarmatian margin of Baltica, as previously suggested (Lewandowski 1993; Nawrocki *et al.* 2004; Nawrocki *et al.* 2007). A close position to Sarmatia would result in delivery of Archean detritus which is absent in the Lower Paleozoic clastics of Małopolska. Because the pre-Furongian rocks of the Łysogóry Terrane are unknown so far from exposures and boreholes the geotectonic provenance of this crustal block remains enigmatic. Recent deep-seismic sounding investigations showed that the crystalline crust of Łysogóry, compared to that of the East European Platform, is thinner, mostly lacks the middle layer, and has a thickened uppermost low-velocity layer. Nevertheless, according to Narkiewicz *et al.* (2011), this crustal unit represents probably a proximal terrane displaced dextrally along the Baltica margin in pre-Devonian times, with a magnitude on the order of a few hundred kilometers at most. The Nd isotope data seem to confirm that the Łysogóry Terrane was located at the Baltica margin during Late Cambrian time and was sourced with clastic material from the Svecofennian crust. By contrast, detrital zircons show similarities to Gondwanan sources. Valverde-Vaquero *et al.* (2000) argued that the absence of Sveconorwegian detritus (1.0-1.2 Ga) in Łysogóry questions the derivation of clastic material from the Polish segment of the EEP. This conclusion was probably not correct because Ediacaran sandstones drilled in south-eastern Poland, on the margin of the EEP, included a zircon population with dominant 1.5-1.6 Ga components but without Sveconorwegian grains (Żelaźniewicz *et al.* 2009). Although the Nd isotope data from the Upper Cambrian of Łysogóry provide clear evidence for detritus supply from the Baltic sources only, a strong variation in age of detrital micas points to the presence of clastic material from both Baltic and Cadomian basement rocks (Belka *et al.* 2000). Mixed, Baltic-Cadomian composition of clastic sediments and a similar mixed character of the trilobite fauna (Zylińska 2001, 2002) supports a paleogeographic scenario in which Łysogóry occupied a position between Baltica and the Cadomian belt during the Late Cambrian. It seems, however, that the terrane was not accreted to the Baltic crust at that time. The temporal change in the composition of the trilobite fauna, from Avalonian- to Baltic-dominated assemblages, suggests that Łysogóry could approach the Baltica margin before the end of the Cambrian.

Previous provenance investigations performed on detrital muscovites revealed that the final amalgamation of Małopolska and Łysogóry was largely terminated before late Early Devonian times (Belka *et al.* 2000). Similar Nd isotope signatures recognized in the Silurian clastics in Małopolska and Łysogóry document that both units were supplied with detritus from the same source area but they do not allow to constrain the progress of the amalgamation process. However, the fact that the sedimentary infill in the marginal, contact zones of the Małopolska and Łysogóry terranes was almost identical at that time attests the final phase of amalgamation during the late Silurian (cf. Kozłowski *et al.* 2014).

Conclusions

From the Sr and Nd isotope investigations of Lower Paleozoic clastics of the Małopolska and Łysogóry terranes, coupled with provenance and biogeographic data, the following main conclusions can be drawn:

- Cambrian clastic rocks of Małopolska differ in their Sr and Nd isotope composition from those of Łysogóry. The former are characterized by a wide range of the $\epsilon_{Nd(t)}$ values, from -11.5 to -5.5 , and Sr isotope ratios from 0.712 to 0.767 , whereas the $\epsilon_{Nd(t)}$ values of the latter vary within a fairly narrow range of the $\epsilon_{Nd(t)}$, from -11.1 to -8.4 , and Sr isotope ratios from 0.722 to 0.754 ;
- Cambrian clastics of Małopolska are characterized by Nd model ages (T_{2DM}) from 1.71 Ga to 2.18 Ga, but ages younger than 1.9 Ga clearly predominate. This pattern is contrasted by T_{2DM} model ages of the Cambrian clastics of Łysogóry, which all are older than 1.9 Ga;
- Contrary to the Cambrian rocks, the Silurian clastics exhibit similar Sr and Nd isotope characteristics within both terranes, Małopolska and Łysogóry;
- Sr and Nd isotope composition of Cambrian rocks in Małopolska and Łysogóry suggests that these crustal units were supplied with clastic material from different source areas of continental crustal domains. Detritus recognized in Łysogóry material must have been mostly derived from the Svecofennian basement of Fennoscandia, but it has also a small addition from a Cadomian source. In contrast, Nd isotope data and detrital zircon ages strongly suggest that Małopolska must have been supplied with clastic material from Amazonian sources

during Cambrian times, being also progressively sourced from the Baltic Svecofennian crust in the late Middle Cambrian;

- All provenance and biogeographic data indicate that Małopolska must have been situated in the immediate vicinity of the Amazonian craton and near the western end of the Cadomian belt during the Early Cambrian, and it was not until the late Middle Cambrian that Małopolska collided with the Baltica margin;
- The Łysogóry Terrane occupied a position between Baltica and the Cadomian belt during the Late Cambrian but it was probably not accreted to the Baltic crust at that time.
- The phase of final amalgamation of Małopolska and Łysogóry took place during late Silurian times.

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APPENDIX

Tab. 1. Location code of analyzed samples.

Locality	Locality Code		Locality	Locality Code
Łysogóry			Małopolska	
Bronkowice	BR		Bardo	BA
Czarna Woda	CW		Belnia	BE
Jurkowice	JR		Biesak	BI
Paprocice	PA		Brzechów	BRZ
Pobroszyn	PO		Chęciny	DCH; DC
Podmąchocice	PD		Gruchawka	GR
Rzepin	RZ		Góry Pieprzowe	GP
Wąworków	WWW		Jugoszów	JUG
Wilków	LUB		Kędziorka	KE; KED
Winnica	WN		Konary	KO
Wiśniówka	WNW		Kopiec	KOP
Wymyślona	WYMYSL		Kotuszów	KT
			Krobielice	KRO
			Lenarczyce	LE
			Lipnik	LI
			Międzygórz	MDZ
			Mójcza	MJ
			Niestachów	NI
			Ociesęki	OC
			Słowik	SL
			Usarzów	US
			Widełki	WI
			Zalesie	ZE
			Żurawiki	ZU

Tab .2. Table showing the Nd, Sm, Sr isotopic compositions of analyzed samples.

Sample	Weight (mg)	Sm (ppm)	Nd (ppm)	$^{147}\text{Sm}/^{144}\text{Nd}$	$^{143}\text{Nd}/^{144}\text{Nd}$	ϵNd	$^{143}\text{Nd}/^{144}\text{Nd}$ (T)	ϵNd (T)	$T_{2\text{DM}}$ model age (in Ga)	Stratigraphic age	$^{87}\text{Sr}/^{86}\text{Sr}$	Lithology
Malopolska												
Cambrian												
BE-01	60.31	8.81	43.81	0.1219	0.512044 ± 6	-11.6	0.511628 ± 6	-6.6	1.78	Lower Cambrian	0.760707 ± 10	mudstone
BE-02	67.21	5.28	29.19	0.1094	0.511994 ± 12	-12.6	0.511621 ± 12	-6.8	1.79	Lower Cambrian	0.766852 ± 11	mudstone
BE-03	58.75	2.58	15.18	0.1031	0.511730 ± 10	-17.7	0.511379 ± 10	-11.5	2.18	Lower Cambrian	0.730592 ± 10	quartzite
BI-02	54.97	1.99	11.85	0.1015	0.511918 ± 13	-14.0	0.511586 ± 13	-7.7	1.87	Lower Cambrian	0.723140 ± 10	sandstone
BI-03	60.58	3.54	22.45	0.0953	0.511970 ± 11	-13.1	0.511643 ± 11	-6.3	1.76	Lower Cambrian	0.743917 ± 11	sandstone
DC-01	108.98	2.62	13.56	0.1167	0.512052 ± 19	-11.4	0.511654 ± 19	-6.1	1.74	Lower Cambrian	0.737112 ± 14	mudstone
DC-02	103.63	3.51	14.81	0.1434	0.512061 ± 15	-11.2	0.511573 ± 15	-7.7	1.87	Lower Cambrian	0.733051 ± 10	sandstone
DCH-01	64.53	5.40	27.43	0.1190	0.512044 ± 10	-11.6	0.511638 ± 10	-6.4	1.77	Lower Cambrian	0.732808 ± 10	mudstone
KE-01	100.68	3.13	16.93	0.1168	0.512068 ± 12	-11.1	0.511687 ± 12	-5.5	1.72	Lower Cambrian	0.734215 ± 77	siltstone
KE-02	57.97	4.87	26.02	0.1130	0.512019 ± 11	-12.1	0.511634 ± 11	-6.5	1.77	Lower Cambrian	0.740013 ± 11	mudstone
KRO-01	46.54	5.14	28.70	0.1082	0.511991 ± 16	-12.6	0.511622 ± 16	-6.8	1.79	Lower Cambrian	0.744644 ± 19	mudstone
KT-01	67.95	6.05	31.47	0.1163	0.512068 ± 33	-11.1	0.511672 ± 33	-5.8	1.71	Lower Cambrian	0.751922 ± 13	mudstone
KT-02	107.12	4.28	20.61	0.1254	0.512056 ± 106	-11.3	0.511629 ± 106	-6.6	1.78	Lower Cambrian	0.734793 ± 12	mudstone
OC-01	56.74	5.25	29.45	0.1075	0.511928 ± 7	-13.8	0.511562 ± 7	-7.9	1.89	Lower Cambrian	0.754217 ± 10	sandstone
SL-01	95.95	2.81	14.44	0.1175	0.511981 ± 12	-12.8	0.511581 ± 12	-7.6	1.86	Lower Cambrian	0.747647 ± 19	siltstone
WI-01	62.50	1.84	9.09	0.1223	0.512071 ± 26	-11.0	0.511655 ± 26	-6.1	1.74	Lower Cambrian	0.718641 ± 10	sandstone
ZE-02	53.96	5.12	32.62	0.0949	0.511871 ± 10	-15.0	0.511547 ± 10	-8.2	1.91	Lower Cambrian	0.756679 ± 10	mudstone
BA-05	56.89	2.00	11.10	0.1087	0.511920 ± 11	-14.0	0.511558 ± 11	-8.4	1.91	Middle Cambrian	0.718318 ± 15	sandstone
BRZ-01	55.53	2.64	16.54	0.0965	0.511984 ± 9	-12.7	0.511665 ± 9	-6.3	1.74	Middle Cambrian	0.721379 ± 9	sandstone

Sample	Weight (mg)	Sm (ppm)	Nd (ppm)	¹⁴⁷ Sm/ ¹⁴⁴ Nd	¹⁴³ Nd/ ¹⁴⁴ Nd	εNd	¹⁴³ Nd/ ¹⁴⁴ Nd (T)	εNd (T)	T _{2DM} .model age (in Ga)	Stratigraphic age	⁸⁷ Sr/ ⁸⁶ Sr	Lithology
GP-01	52.99	5.97	35.19	0.1029	0.511834 ± 10	-15.7	0.511494 ± 10	-9.6	2.02	Middle Cambrian	0.762014 ± 11	mudstone
GP-02	54.28	7.39	9.20	0.1326	0.512016 ± 31	-12.1	0.511571 ± 31	-8.1	1.88	Middle Cambrian	0.721034 ± 18	sandstone
JUG-01	49.50	3.75	20.84	0.1089	0.511767 ± 11	-17.0	0.511407 ± 11	-11.3	2.15	Middle Cambrian	0.724380 ± 13	sandstone
KO-01	41.21	3.12	18.25	0.1034	0.511948 ± 13	-13.5	0.511605 ± 13	-7.5	1.84	Middle Cambrian	0.735026 ± 8	mudstone
KO-02	54.16	5.08	29.10	0.1056	0.511884 ± 22	-14.7	0.511535 ± 22	-8.8	1.95	Middle Cambrian	0.744934 ± 30	mudstone
KO-03	70.77	0.97	5.91	0.0993	0.511972 ± 9	-13.0	0.511644 ± 9	-6.8	1.78	Middle Cambrian	0.716850 ± 12	quartzite
KO-04	45.15	2.42	13.78	0.1060	0.511918 ± 14	-14.0	0.511567 ± 14	-8.2	1.90	Middle Cambrian	0.721719 ± 22	quartzite
KO-05	42.85	1.32	6.98	0.1148	0.511996 ± 20	-12.5	0.511616 ± 20	-7.2	1.82	Middle Cambrian	0.719453 ± 14	sandstone
LE-01	49.54	7.08	42.48	0.1007	0.511871 ± 9	-14.6	0.511530 ± 9	-8.8	1.95	Middle Cambrian	0.754434 ± 21	sandstone
US-01	48.28	4.25	23.81	0.1081	0.511874 ± 9	14.1	0.511517 ± 9	-9.2	1.98	Middle Cambrian	0.738007 ± 21	sandstone
US-02	52.11	3.56	19.25	0.1115	0.511961 ± 7	-13.2	0.511592 ± 7	-7.8	1.86	Middle Cambrian	0.739156 ± 13	sandstone
ZU-01	63.02	1.04	4.76	0.1318	0.512066 ± 32	-11.1	0.511630 ± 32	-7.0	1.80	Middle Cambrian	0.711627 ± 9	sandstone
ZU-02	62.63	1.03	5.59	0.1112	0.511918 ± 17	-14.0	0.511550 ± 17	-8.5	1.93	Middle Cambrian	0.717741 ± 9	mudstone
Ordovician												
BI-01	57.60	1.66	12.27	0.0817	0.511853 ± 13	-15.3	0.511586 ± 13	-8.3	1.89	Lower Ordovician	0.723140 ± 10	sandstone
KED-01	51.09	9.23	43.64	0.1278	0.512040 ± 10	-11.7	0.511638 ± 10	-7.5	1.82	Lower Ordovician	0.732041 ± 9	mudstone
KED-02	55.70	4.20	19.31	0.1315	0.512048 ± 9	-11.5	0.511635 ± 9	-7.5	1.73	Lower Ordovician	0.731357 ± 16	sandstone
MDZ-02	51.42	13.62	59.51	0.1383	0.512116 ± 8	-10.2	0.511681 ± 8	-6.7	1.75	Lower Ordovician	0.711462 ± 10	sandstone
MDZ-03	54.60	11.03	49.28	0.1351	0.512006 ± 8	-12.3	0.511582 ± 8	-8.6	1.90	Lower Ordovician	0.714493 ± 10	sandstone
MDZ-04	50.69	2.07	8.88	0.1408	0.512010 ± 11	-12.2	0.511567 ± 11	-8.8	1.93	Lower Ordovician	0.715622 ± 7	sandstone
MDZ-05	62.14	6.80	27.08	0.1519	0.512072 ± 12	-11.0	0.511595 ± 12	-8.3	1.88	Lower Ordovician	0.712837 ± 10	sandstone
ZE-03	61.10	24.15	109.38	0.1335	0.512060 ± 6	-11.3	0.511640 ± 6	-7.4	1.81	Lower Ordovician	0.719701 ± 9	mudstone
ZE-04	52.38	5.47	26.06	0.1269	0.512056 ± 12	-11.3	0.511657 ± 12	-7.2	1.79	Lower Ordovician	0.724340 ± 10	sandstone
MJ-01	49.95	1.93	12.94	0.0901	0.511933 ± 9	-13.7	0.511649 ± 9	-7.4	1.80	Lower Ordovician	0.721304 ± 11	sandstone
KED-03	55.71	4.87	27.96	0.1052	0.511955 ± 10	-13.5	0.511616 ± 10	-7.9	1.84	Middle Ordovician	0.762947 ± 10	mudstone

Sample	Weight (mg)	Sm (ppm)	Nd (ppm)	$^{147}\text{Sm}/^{144}\text{Nd}$	$^{143}\text{Nd}/^{144}\text{Nd}$	ϵNd	$^{143}\text{Nd}/^{144}\text{Nd}$ (T)	ϵNd (T)	$T_{2\text{DM}}$.model age (in Ga)	Stratigraphic age	$^{87}\text{Sr}/^{86}\text{Sr}$	Lithology
ZE-05	52.04	0.66	4.26	0.0944	0.512028 ± 14	-11.9	0.511740 ± 14	-5.8	1.67	Middle Ordovician	0.762164 ± 14	mudstone
ZE-06	51.91	3.25	31.55	0.1006	0.511966 ± 8	-13.1	0.511660 ± 8	-7.4	1.79	Middle Ordovician	0.742018 ± 11	mudstone
Silurian												
BA-01	59.34	5.22	30.46	0.1035	0.511884 ± 12	-14.7	0.511585 ± 12	-9.5	1.95	Lower Silurian	0.766788 ± 9	mudstone
MDZ-01	48.15	6.44	34.94	0.1115	0.511955 ± 13	-13.3	0.511634 ± 13	-8.5	1.87	Lower Silurian	0.746269 ± 10	mudstone
ZE-01	62.39	5.18	29.33	0.1068	0.511939 ± 12	-13.6	0.511631 ± 12	-8.6	1.88	Lower Silurian	0.765499 ± 9	mudstone
ZE-07	62.54	1.73	8.34	0.1256	0.512105 ± 15	-10.4	0.511740 ± 15	-6.4	1.70	Lower Silurian	0.720448 ± 13	mudstone/chert
BA-02	57.00	5.77	31.85	0.1096	0.511966 ± 11	-13.1	0.511665 ± 11	-8.4	1.85	Upper Silurian	0.734762 ± 10	mudstone
BA-03	55.87	4.40	22.77	0.1169	0.512187 ± 10	-8.8	0.511865 ± 10	-4.5	1.53	Upper Silurian	0.726600 ± 11	greywacke
BA-04	74.85	4.56	27.17	0.1015	0.512073 ± 10	-11.0	0.511794 ± 10	-5.9	1.64	Upper Silurian	0.7977606 ± 16	mudstone
GR-01	50.59	6.14	33.45	0.1109	0.512010 ± 48	-12.2	0.511705 ± 48	-7.6	1.78	Upper Silurian	0.737568 ± 10	sandstone
GR-02	56.04	6.17	33.78	0.1105	0.511979 ± 12	-12.8	0.511675 ± 12	-8.2	1.83	Upper Silurian	0.735626 ± 7	mudstone
NI-01	58.18	5.07	22.47	0.1363	0.512247 ± 11	-7.6	0.511872 ± 11	-4.4	1.52	Upper Silurian	0.740516 ± 10	greywacke
NI-02	49.25	7.88	39.44	0.1208	0.512079 ± 7	-10.9	0.511747 ± 7	-6.8	1.72	Upper Silurian	0.754574 ± 13	mudstone
NI-03	55.51	1.84	11.99	0.0928	0.511952 ± 7	-13.4	0.511697 ± 7	-7.8	1.79	Upper Silurian	0.729817 ± 20	sandstone
Lysogóry												
Cambrian												
JR-01	105.25	8.35	41.99	0.1202	0.511840 ± 19	-15.5	0.511453 ± 19	-10.7	2.09	Upper Cambrian	0.738927 ± 33	mudstone
JR-02	105.14	0.72	4.09	0.1070	0.511844 ± 29	-15.5	0.511498 ± 29	-9.8	2.02	Upper Cambrian	0.721785 ± 25	mudstone
JR-03	102.98	11.97	64.87	0.1116	0.511818 ± 12	-16.0	0.511456 ± 12	-10.6	2.09	Upper Cambrian	0.741173 ± 14	sandstone
PA-01	55.72	1.10	6.20	0.1077	0.511863 ± 8	-15.1	0.511514 ± 8	-9.4	2.00	Upper Cambrian	0.729172 ± 13	quartzite
PD-01	62.71	7.47	39.56	0.1142	0.511938 ± 16	-13.6	0.511560 ± 16	-8.4	1.90	Upper Cambrian	0.753928 ± 12	mudstone
WNW-01	97.63	6.17	27.78	0.1342	0.511890 ± 111	-14.6	0.511453 ± 111	-10.7	2.09	Upper Cambrian	0.745408 ± 27	mudstone
WNW-02	63.38	6.82	36.39	0.1133	0.511862 ± 10	-15.1	0.511495 ± 10	-9.9	2.03	Upper Cambrian	0.746477 ± 12	mudstone
WNW-03	96.69	4.21	20.81	0.1223	0.511828 ± 10	-15.8	0.511432 ± 10	-11.1	2.12	Upper Cambrian	0.749403 ± 25	mudstone

Sample	Weight (mg)	Sm (ppm)	Nd (ppm)	$^{147}\text{Sm}/^{144}\text{Nd}$	$^{143}\text{Nd}/^{144}\text{Nd}$	ϵNd	$^{143}\text{Nd}/^{144}\text{Nd}$ (T)	ϵNd (T)	$T_{2\text{DM}}$.model age (in Ga)	Stratigraphic age	$^{87}\text{Sr}/^{86}\text{Sr}$	Lithology
WW-01	56.95	2.18	12.01	0.1098	0.511890 ± 10	-14.6	0.511534 ± 10	-9.1	1.96	Upper Cambrian	0.752348 ± 10	mudstone
WY-01	93.34	2.73	14.62	0.1127	0.511880 ± 13	-14.8	0.511511 ± 13	-9.6	1.99	Upper Cambrian	0.721768 ± 16	sandstone
Ordovician												
PO-01	53.65	5.8	33.59	0.1045	0.511843 ± 32	-15.5	0.511515 ± 32	-9.9	2.01	Lower Ordovician	0.754171 ± 9	mudstone
PO-02	55.7	13.21	56.23	0.1419	0.512134 ± 10	-9.8	0.511688 ± 10	-6.6	1.74	Lower Ordovician	0.709791 ± 12	sandstone
PO-03	59.23	5.54	28.76	0.1164	0.511889 ± 10	-14.6	0.511523 ± 10	-9.7	2.00	Lower Ordovician	0.724984 ± 10	sandstone
Silurian												
BR-01	51.34	7.15	49.55	0.0873	0.511953 ± 9	-13.4	0.511713 ± 9	-7.5	1.77	Upper Silurian	0.745555 ± 10	mudstone
CW-01	58.96	0.78	4.71	0.1000	0.511918 ± 18	14.0	0.511643 ± 18	-8.9	1.88	Upper Silurian	0.715426 ± 54	sandstone
CW-02	52.85	1.65	9.25	0.1076	0.511919 ± 13	-14.0	0.511590 ± 13	-9.3	1.91	Upper Silurian	0.717170 ± 10	sandstone
LUB-01	53.01	3.35	19.64	0.1030	0.511988 ± 9	-12.7	0.511705 ± 9	-7.7	1.79	Upper Silurian	0.720685 ± 10	greywacke
LUB-02	65.55	7.22	45.74	0.0955	0.511989 ± 12	-12.6	0.511727 ± 12	-7.2	1.75	Upper Silurian	0.748760 ± 10	mudstone
RZ-01	60.31	5.09	25.62	0.1201	0.512059 ± 11	-11.3	0.511728 ± 11	-7.2	1.75	Upper Silurian	0.723763 ± 10	sandstone
RZ-02	52.36	3.07	20.97	0.1146	0.512012 ± 11	-12.2	0.511697 ± 11	-7.8	1.80	Upper Silurian	0.736182 ± 10	mudstone
RZ-03	50.5	4.36	20.63	0.1277	0.512092 ± 10	-10.6	0.511741 ± 10	-6.9	1.73	Upper Silurian	0.723055 ± 10	greywacke
RZ-04	51.39	4.42	25.20	0.1061	0.512082 ± 10	-10.8	0.511790 ± 10	-6.0	1.65	Upper Silurian	0.735530 ± 10	sandstone
RZ-05	47.96	7.31	40.83	0.1083	0.511990 ± 11	-12.6	0.511692 ± 11	-7.9	1.81	Upper Silurian	0.743041 ± 11	mudstone
RZ-06	57.99	6.47	35.82	0.1091	0.511969 ± 11	-13.0	0.511668 ± 11	-8.4	1.84	Upper Silurian	0.738477 ± 10	mudstone
RZ-07	59.02	4.36	23.82	0.1107	0.511944 ± 13	-13.7	0.511640 ± 13	-9.8	1.89	Upper Silurian	0.731672 ± 10	mudstone
RZ-08	61.76	5.06	30.40	0.1007	0.511960 ± 12	-13.2	0.511683 ± 12	-8.1	1.82	Upper Silurian	0.735110 ± 10	mudstone
RZ-09	46.79	6.06	34.12	0.1075	0.511964 ± 11	-13.1	0.511669 ± 11	-8.4	1.84	Upper Silurian	0.733313 ± 9	mudstone
WN-01	51.72	5.98	35.36	0.1022	0.511965 ± 9	-13.1	0.511684 ± 9	-8.1	1.82	Upper Silurian	0.738712 ± 13	mudstone
WN-02	48.30	26.19	4.94	0.1140	0.512009 ± 9	-12.3	0.511695 ± 9	-7.8	1.80	Upper Silurian	0.723708 ± 10	mudstone
WN-03	56.86	5.09	27.29	0.1127	0.512006 ± 11	-12.3	0.511696 ± 11	-7.8	1.80	Upper Silurian	0.723300 ± 10	mudstone

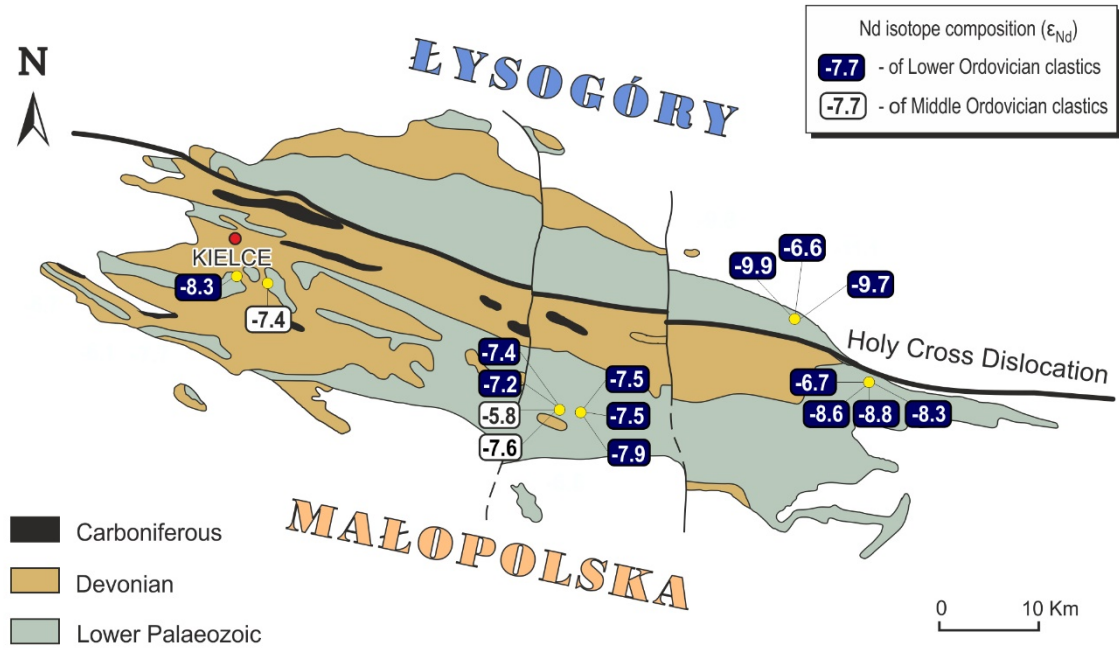


Fig. 1'. The neodymium isotope composition ($\epsilon_{Nd(t)}$) of the Ordovician of Małopolska and Łysogóry.

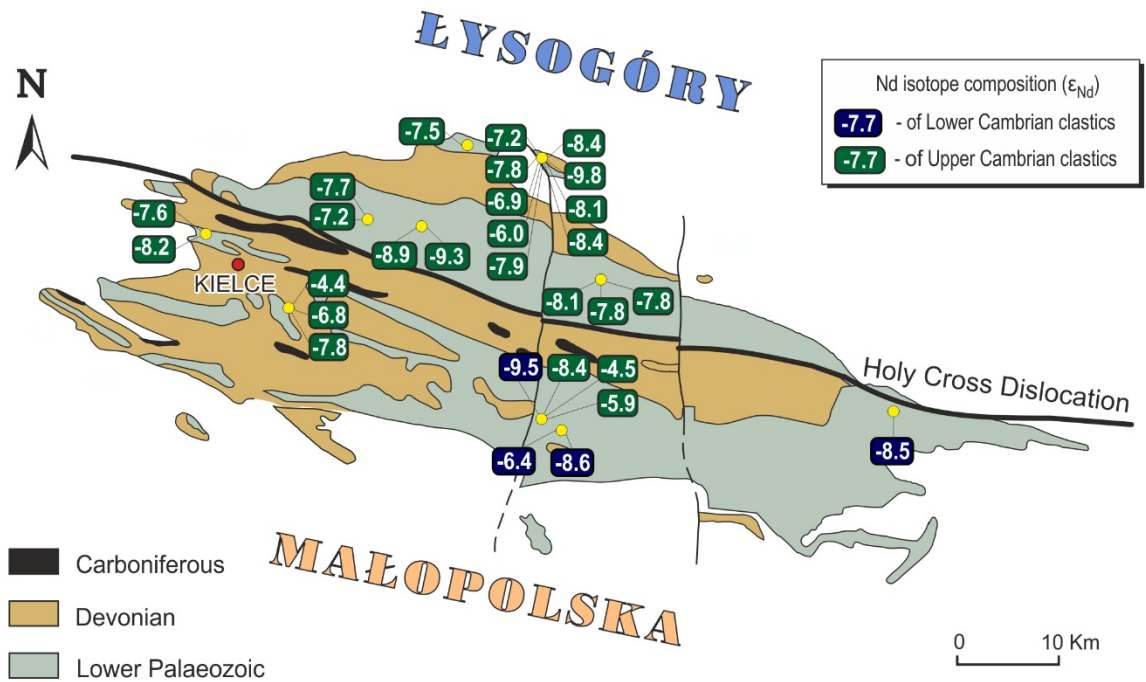


Fig. 2'. The neodymium isotope composition ($\epsilon_{Nd(t)}$) of the Silurian clastics of Małopolska and Łysogóry.

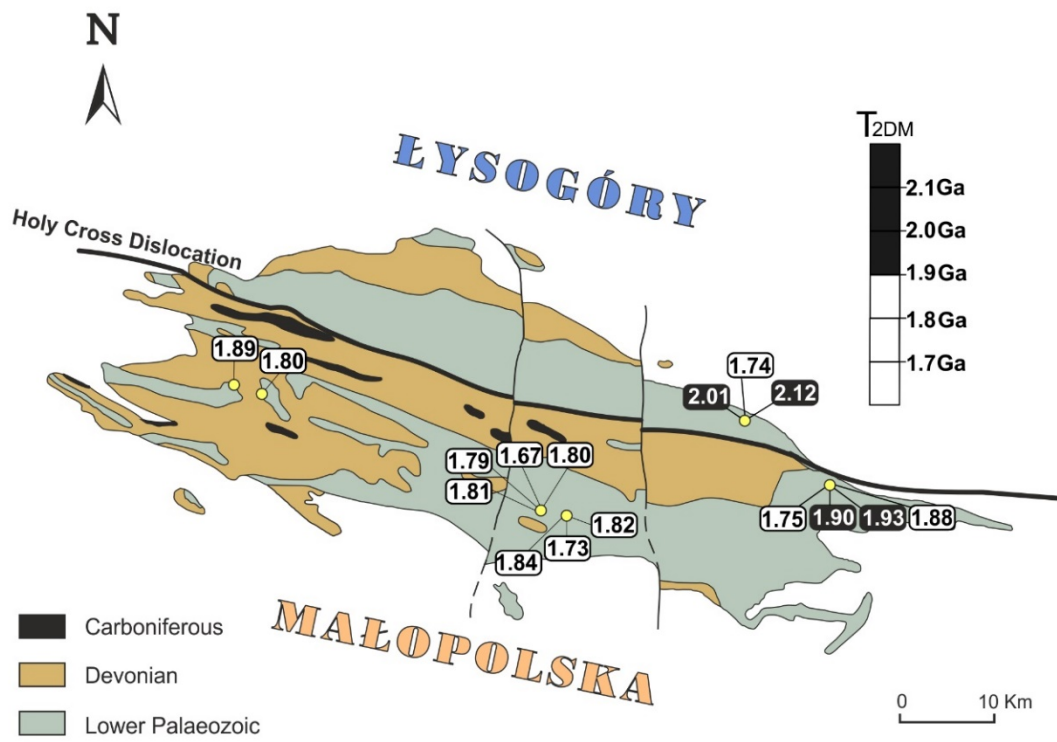


Fig. 3'. The neodymium model age values (T_{2DM}) of the Ordovician clastics of Małopolska and Łysogóry.

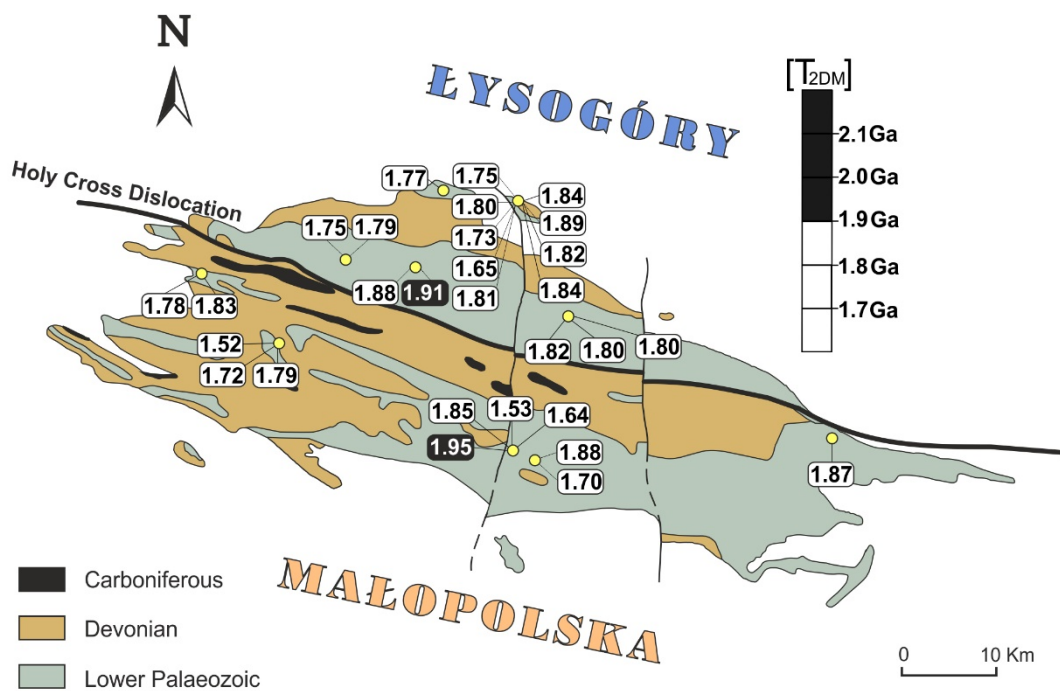


Fig. 4'. The neodymium model age values (T_{2DM}) of the Silurian clastics of Małopolska and Łysogóry.

LOCALITES DATA:

ŁYSOGÓRY

<i>Locality:</i>	Bronkowice
<i>Geographical coordinates:</i>	50°58'35.29"N; 21°00'54.28"E
<i>Character of locality:</i>	natural exposure along the road
<i>Stratigraphy:</i>	Upper Silurian
<i>Samples code:</i>	BR
<i>Lithology:</i>	mudstones
<i>Literature data</i>	Kozłowski 2008



Fig. 5'. Satellite picture with the setting of the Bronkowice village, (taken from *Google Earth*).

<i>Locality:</i>	Czarna Woda
<i>Geographical coordinates:</i>	50°54'35.45''N; 20°56'56.75''E
<i>Character of locality:</i>	rock blocks in the river bed
<i>Stratigraphy:</i>	Upper Silurian
<i>Samples code:</i>	CW
<i>Lithology:</i>	sandstones



Fig. 6'. Satellite picture with the location of the sampled point in the river bed of Czarna Woda (taken from *Google Earth*).

<i>Locality:</i>	Jurkowice
<i>Geographical coordinates:</i>	50°47'30.66''N; 21°22'24.85''E
<i>Character of locality:</i>	natural outcrops
<i>Stratigraphy:</i>	Upper Cambrian
<i>Samples code:</i>	JR
<i>Lithology:</i>	mudstones, sandstones
<i>Literature data</i>	Tomczykowa 1993; Romanek & Rup 1989.



Fig. 7'. Satellite picture with the location of the sampled outcrop at Jurkowice (taken from Google Earth).

<i>Locality:</i>	Paprocice
<i>Geographical coordinates:</i>	50°49′50.49″N; 21°04′55.99″E
<i>Character of locality:</i>	natural exposure along railway
<i>Stratigraphy:</i>	Upper Cambrian
<i>Samples Code:</i>	PA
<i>Lithology:</i>	quartzites



Fig. 8’. Satellite picture with the location of the sampled outcrop at Paprocice (taken from *Google Earth*).

<i>Locality:</i>	Pobroszyn
<i>Geographical coordinates:</i>	50°46'53.58"N; 21°27'51.63"E
<i>Character of locality:</i>	slope with natural outcrops
<i>Stratigraphy:</i>	Lower Ordovician
<i>Samples Code:</i>	PO
<i>Lithology:</i>	sandstones, mudstones
<i>Literature data:</i>	Kowalczewski <i>et al.</i> 1976; Trela <i>et al.</i> 2001; Trela 2008a; Bednarczyk & Stempień-Sałek 2011.



Fig. 9'. Satellite picture with the location of the sampled section at Pobroszyn (taken from *Google Earth*).

<i>Locality:</i>	Podmachocice
<i>Geographical coordinates:</i>	50°54'17.89''N; 20°47'27.81''E
<i>Character of locality:</i>	natural outcrops
<i>Stratigraphy:</i>	Upper Cambrian
<i>Samples code:</i>	PD
<i>Lithology:</i>	mudstones

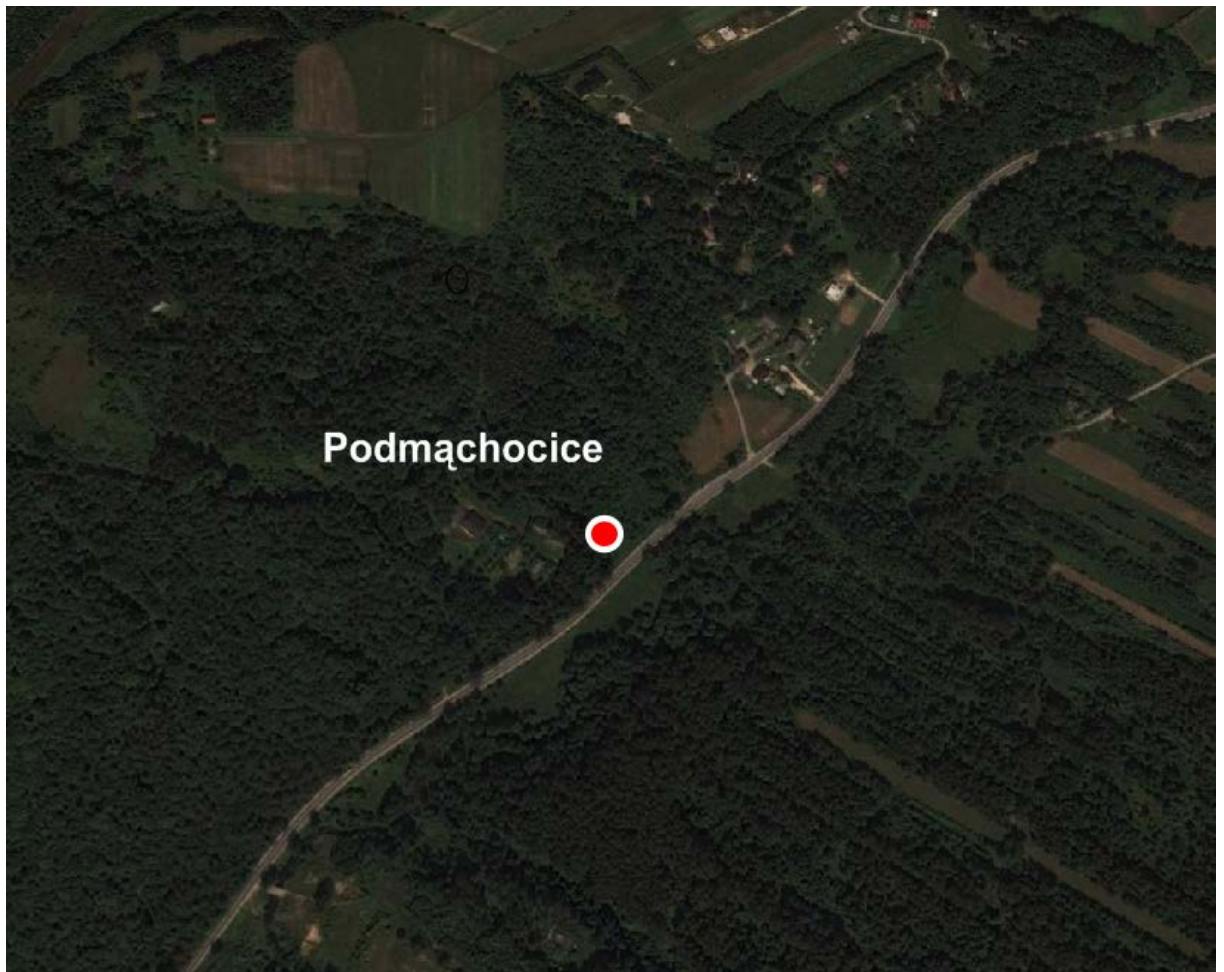


Fig. 10'. Satellite picture with with the location of the sampled outcrop at Podmachocice (taken from *Google Earth*).

<i>Locality:</i>	Rzepin Pierwszy/Rzepin Kolonia
<i>Geographical coordinates:</i>	50°58'06.87"N; 21°04'24.11"E 50°57'33.09"N; 21°05'21.05"E
<i>Character of locality:</i>	slope with natural outcrops/ abandoned quarry
<i>Stratigraphy:</i>	Upper Silurian
<i>Samples Code:</i>	RZ
<i>Lithology:</i>	sandstones, mudstones, greywackes
<i>Literature data</i>	Czarnocki 1950; Tomczykowa & Tomczyk 1981; Bednarczyk <i>et al.</i> 1983; Malec 1993, 2000; Kozłowski 2008.



Fig. 11'. Satellite picture with the setting of the Rzepin Pierwszy (taken from *Google Earth*).



Fig. 12'. Satelite picture with the location of the sampled exposure close to Rzepin Kolonia village (taken from *Google Earth*).



Fig. 13'. Natural outcrop of the Upper Silurian sediments in Rzepin Kolonia.



Fig. 14'. Exposure of the Upper Silurian fine grained sandstones in Rzepin Kolonia.

<i>Locality:</i>	Wąworków
<i>Geographical coordinates:</i>	50°47'18.24''N; 21°26'51.07''E
<i>Character of locality:</i>	abandoned quarry
<i>Stratigraphy:</i>	Upper Cambrian
<i>Samples code:</i>	WWW
<i>Lithology:</i>	mudstones
<i>Literature data</i>	Kowalczewski <i>et al.</i> 1976; Żylińska 2002.



Fig. 15'. Satellite picture with the location of the Wąworków Quarry (taken from *Google Earth*).

<i>Locality:</i>	Wilków - Parcele
<i>Geographical coordinates:</i>	50°54′34.34″N; 20°49′29.42″E
<i>Character of locality:</i>	natural exposure along the road
<i>Stratigraphy:</i>	Upper Silurian
<i>Samples Code:</i>	LUB
<i>Lithology:</i>	greywackes, mudstones
<i>Literature data</i>	Cieśla <i>et al.</i> 1962; Deczkowski & Tomczyk 1969a; Malec 2000d.



Fig. 16’. Satelite picture with the location of the sampled exposure close to Lubrzanka river near the Wilków village (taken from *Google Earth*).

<i>Locality:</i>	Winnica
<i>Geographical coordinates:</i>	50°52'21.29"N; 21°06'14.68"E
<i>Character of locality:</i>	natural outcrops in the valley
<i>Stratigraphy:</i>	Upper Silurian
<i>Samples code:</i>	WN
<i>Lithology:</i>	mudstones
<i>Literature data</i>	Kozłowski 2000, 2008; Bodzioch <i>et al.</i> 2003.



Fig. 17'. Satellite picture with the location of the sampled exposure in the riverbed of Dobruchna at Winnica (taken from *Google Earth*).



Fig. 18'. Natural outcrop of the Upper Silurian shales in the riverbed of Dobruchna at Winnica village.

<i>Locality:</i>	Wiśniówka Duża
<i>Geographical coordinates:</i>	50°55′57.55″N; 20°42′00.01″E
<i>Character of locality:</i>	active quarry
<i>Stratigraphy:</i>	Upper Cambrian
<i>Samples Code:</i>	WNW
<i>Lithology:</i>	mudstones
<i>Literature data:</i>	Kowalczewski <i>et al.</i> 1986; Studencki 1994; Orłowski & Mizerski 1995; Kowalczewski & Dadlez 1996; Żylińska & Szczepaniak 2002.



Fig. 19’. Satelite picture with the location of the sampling place in the Wiśniówka Duża Quarry (taken from *Google Earth*).

<i>Locality:</i>	Wymyślona
<i>Geographical coordinates:</i>	50°53'47.55''N; 20°49'43.79''E
<i>Character of locality:</i>	natural outcrops
<i>Stratigraphy:</i>	Upper Cambrian
<i>Samples code:</i>	WY
<i>Lithology:</i>	sandstones
<i>Literature data</i>	Valverde-Vaquero <i>et al.</i> 2000.



Fig. 20'. Satellite picture with with the location of the sampled outcrop at Wymyślona (taken from *Google Earth*).

MAŁOPOLSKA

<i>Locality:</i>	Bardo-Prągowiec Ravine
<i>Geographical coordinates:</i>	50°44′46.66″N; 21°02′07.82″E
<i>Character of locality:</i>	natural outcrops
<i>Stratigraphy:</i>	Middle Cambrian, Lower/Upper Silurian
<i>Samples code:</i>	BA
<i>Lithology:</i>	sandstones, mudstones, greywackes
<i>Literature data:</i>	Bednarczyk & Tomczyk 1981; Masiak <i>et al.</i> 2003; Trela & Salwa 2007.



Fig. 21'. Satellite picture with the location of the Prągowiec Ravine (taken from Google Earth).

<i>Locality:</i>	Belnia Hill
<i>Geographical coordinates:</i>	50°50′58.54″N; 20°29′33.20″E
<i>Character of locality:</i>	natural outcrop along the road
<i>Stratigraphy:</i>	Lower Cambrian
<i>Samples code:</i>	BE
<i>Lithology:</i>	mudstones, quartzites

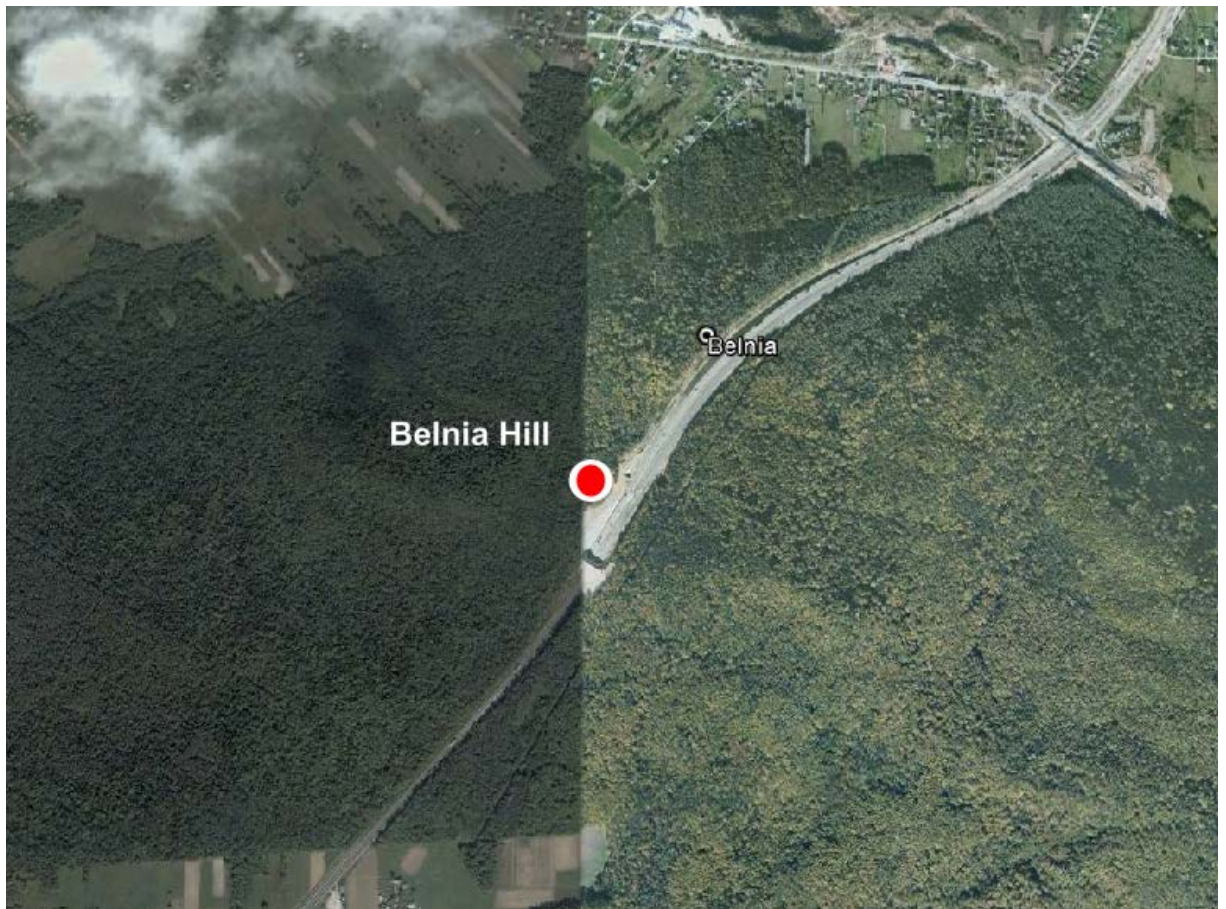


Fig. 22'. Satellite picture with the location of the sampled outcrop of the Belnia hill (taken from Google Earth).

<i>Locality:</i>	Biesak-Białogon (Nature Reserve)
<i>Geographical coordinates:</i>	50°50′39.59″N; 20°34′22.19″E
<i>Character of locality:</i>	abandoned quarry
<i>Stratigraphy:</i>	Lower Cambrian, Lower Ordovician
<i>Samples code:</i>	BI
<i>Lithology:</i>	sandstones
<i>Literature data:</i>	Mizerski 1996; Orłowski 1988.



Fig. 23'. Satellite picture with the location of the Biesak-Białogon Quarry (taken from *Google Earth*).

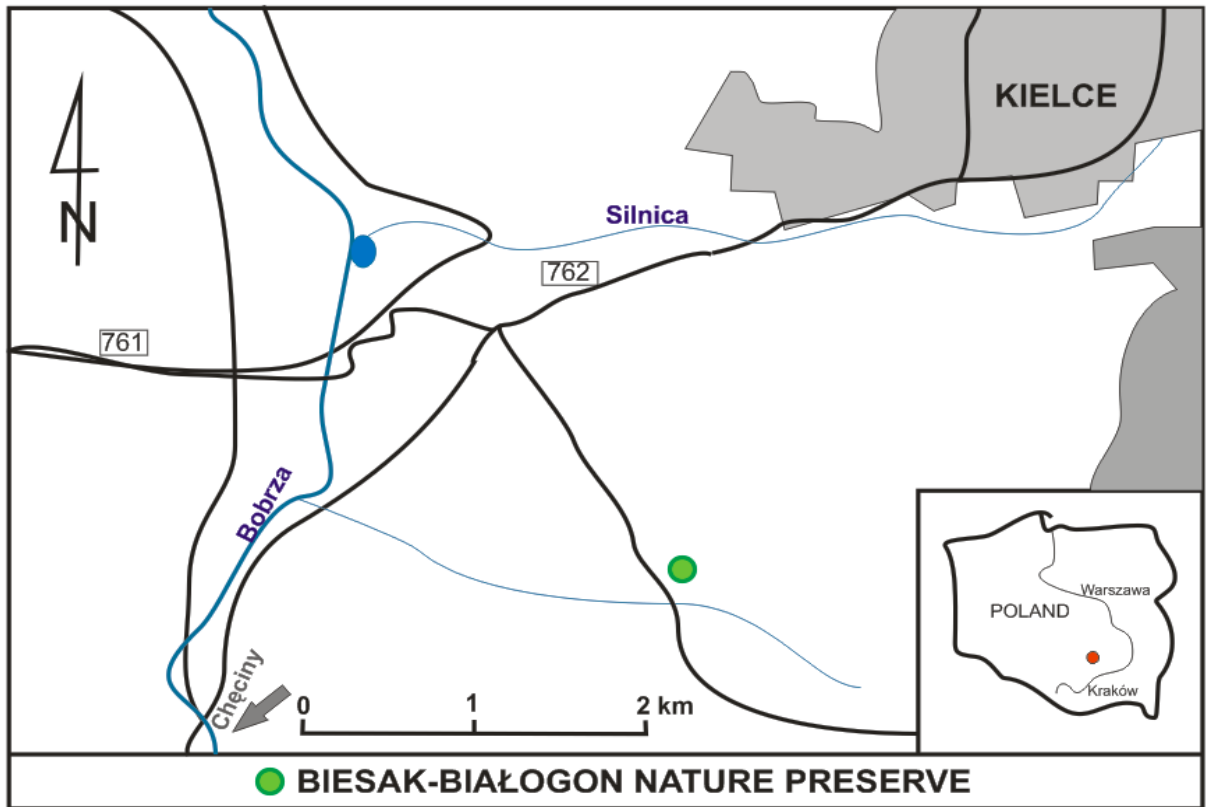


Fig. 24'. Map showing the location of the Biesak-Białogon Nature Reserve.



Fig. 25'. Natural outcrop of the Ordovician sandstone in the Biesak-Białogon Quarry.

<i>Locality:</i>	Brzechów
<i>Geographical coordinates:</i>	50°50′02.29″N; 20°46′28.00″E
<i>Character of locality:</i>	natural exposure
<i>Stratigraphy:</i>	Middle Cambrian
<i>Samples code:</i>	BRZ
<i>Lithology:</i>	sandstones
<i>Literature data</i>	Bednarczyk 1970; Masiak & Żylińska 1994; Żylińska & Masiak 2007.



Fig. 26'. Satellite picture with the location of the sampled exposure in the Brzechów village (taken from *Google Earth*).

<i>Locality:</i>	Dolina Chęcińska
<i>Geographical coordinates:</i>	DC1- 50°48'36.23''N; 20°28'17.43''E DC2- 50°48'27.53''N; 20°27'35.24''E DCH- 50°48'21.05''N; 20°28'59.34''E
<i>Character of locality:</i>	natural outcrops
<i>Stratigraphy:</i>	Lower Cambrian
<i>Samples code:</i>	DC, DCH
<i>Lithology:</i>	mudstones, sandstones
<i>Literature data:</i>	Stupnicka 1986



Fig. 27'. Satellite picture with the location of sampled outcrops of Lower Cambrian rocks in the Chęciny Valley (taken from *Google Earth*).

<i>Locality:</i>	Kielce - Gruchawka
<i>Geographical coordinates:</i>	50°54'42.68''N; 20°37'18.85''E
<i>Character of locality:</i>	exposures along a railway track
<i>Stratigraphy:</i>	Upper Silurian
<i>Samples code:</i>	GR
<i>Lithology:</i>	mudstones, sandstones,
<i>Literature data:</i>	Malec 1993; Tomczykowa 1993; Malec 2001.

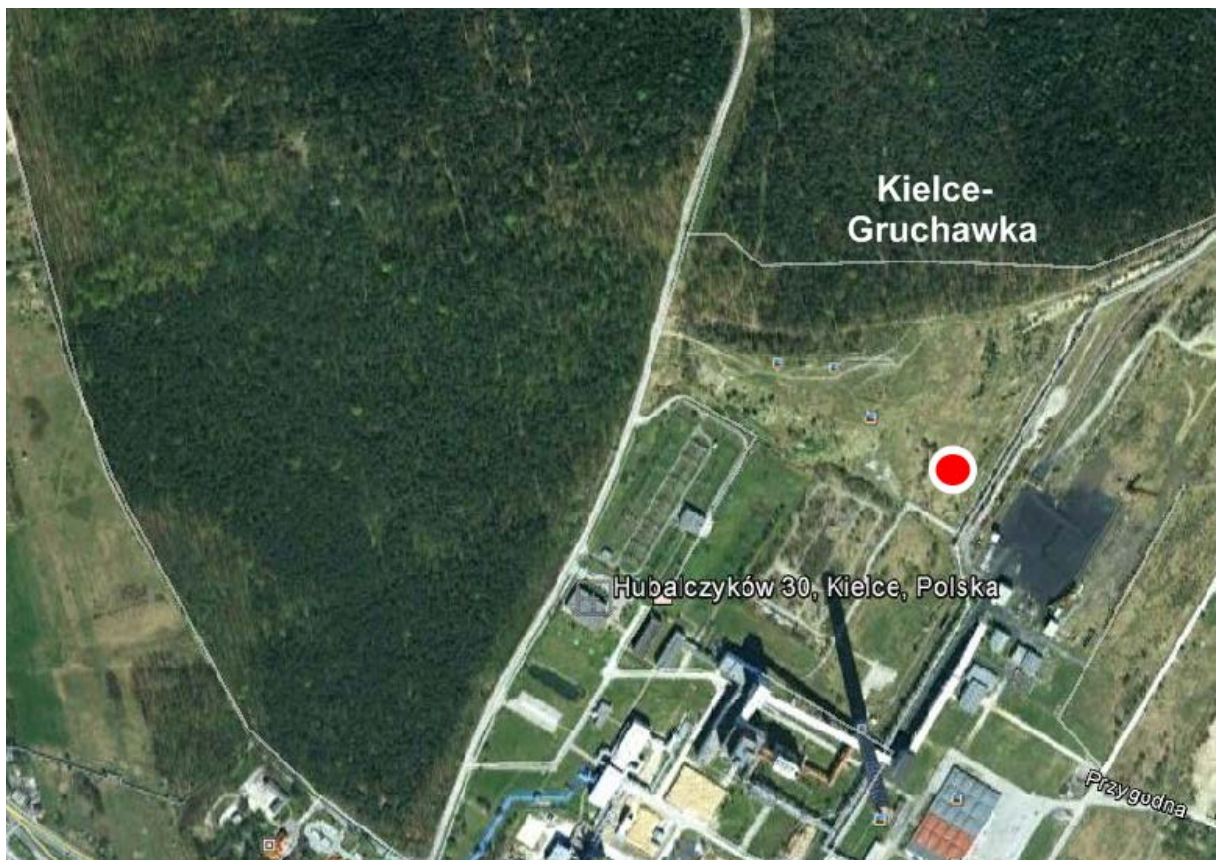


Fig. 28'. Satellite picture with the location of exposures at Kielce-Gruchawka.

<i>Locality:</i>	Pieprzowe MTS Nature Reserve
<i>Geographical coordinates:</i>	50°41'13.58"N; 21°48'17.11"E
<i>Character of locality:</i>	Slope with natural outcrops
<i>Stratigraphy:</i>	Middle Cambrian
<i>Samples code:</i>	GP
<i>Lithology:</i>	mudstones, sandstones
<i>Literature data</i>	Orłowski 1975, 1992; Orłowski & Mizerski 1995; Kowalczewski 1990; Żylińska & Szczepanik 2002.



Fig. 29'. Satellite picture with the position of the sampled outcrop in the Pepper Mountains (taken from *Google Earth*).

<i>Locality:</i>	Jugoszów
<i>Geographical coordinates:</i>	50°41'26.00"N; 21°32'34.92"E
<i>Character of locality:</i>	slope of natural outcrops
<i>Stratigraphy:</i>	Middle Cambrian
<i>Samples code:</i>	JU
<i>Lithology:</i>	sandstones
<i>Literature data</i>	Orłowski 1964, 1988; Żylińska & Szczepanik 2009.



Fig. 30'. Satellite picture with the location of the sampled exposure close to Jugoszów (taken from *Google Earth*).

<i>Locality:</i>	Kędziorka
<i>Geographical coordinates:</i>	50°42'43.87''N; 21°07'23.97''E
<i>Character of locality:</i>	natural outcrops in the Chojnów Ravine
<i>Stratigraphy:</i>	Lower Cambrian, Lower/Middle Ordovician
<i>Samples code:</i>	KE, KED
<i>Lithology:</i>	siltstones, mudstones, sandstones
<i>Literature data</i>	Chlebowski 1971; Tomczyk 1962; Bednarczyk 1981; Nawrocki 2000.



Fig. 31'. Satelite picture with the location of the sampled outcrops in the Chojnów Ravine (taken from *Google Earth*).



Fig. 32'. Natural outcrop of the Lower Ordovician sandstones in the Chojnów Ravine.

<i>Locality:</i>	Konary
<i>Geographical coordinates:</i>	50°41'03.92"N; 21°21'44.82"E 50°41'08.83"N; 21°22'18.66"E
<i>Character of locality:</i>	natural exposures
<i>Stratigraphy:</i>	Middle Cambrian
<i>Samples code:</i>	KO
<i>Lithology:</i>	mudstones, quartzites, sandstones
<i>Literature data</i>	Orłowski 1971



Fig. 33'. Satellite picture with the positions of sampled outcrops at Konary (taken from Google Earth).

<i>Locality:</i>	Kotuszów
<i>Geographical coordinates:</i>	50°36'22.13"N; 21°04'20.19"E
<i>Character of locality:</i>	natural outcrop
<i>Stratigraphy:</i>	Lower Cambrian
<i>Code Samples:</i>	KT
<i>Lithology:</i>	mudstones
<i>Literature data:</i>	Kowalczewski <i>et al.</i> 1987.



Fig. 34'. Satellite picture with the location of the sampled outcrop at Kotuszów (taken from *Google Earth*).



Fig. 35'. Exposure of the Lower Cambrian mudstones at Kotuszów.

<i>Locality:</i>	Krobielice
<i>Geographical coordinates:</i>	50°40′52.34″N; 21°32′21.91″E
<i>Character of locality:</i>	natural exposure along the valley
<i>Stratigraphy:</i>	Lower Cambrian
<i>Samples code:</i>	KRO
<i>Lithology:</i>	mudstones



Fig. 36’. Satellite picture with the location of the sampled exposure at Krobielice (taken from *Google Earth*).

<i>Locality:</i>	Lenarczyce
<i>Geographical coordinates:</i>	50°42′06.08″N; 21°40′25.45″E
<i>Character of locality:</i>	natural outcrop
<i>Stratigraphy:</i>	Middle Cambrian
<i>Samples code:</i>	LE
<i>Lithology:</i>	sandstones
<i>Literature data</i>	Szczepanik <i>et al.</i> 2004a, 2004b; Żylińska & Szczepanik 2009.

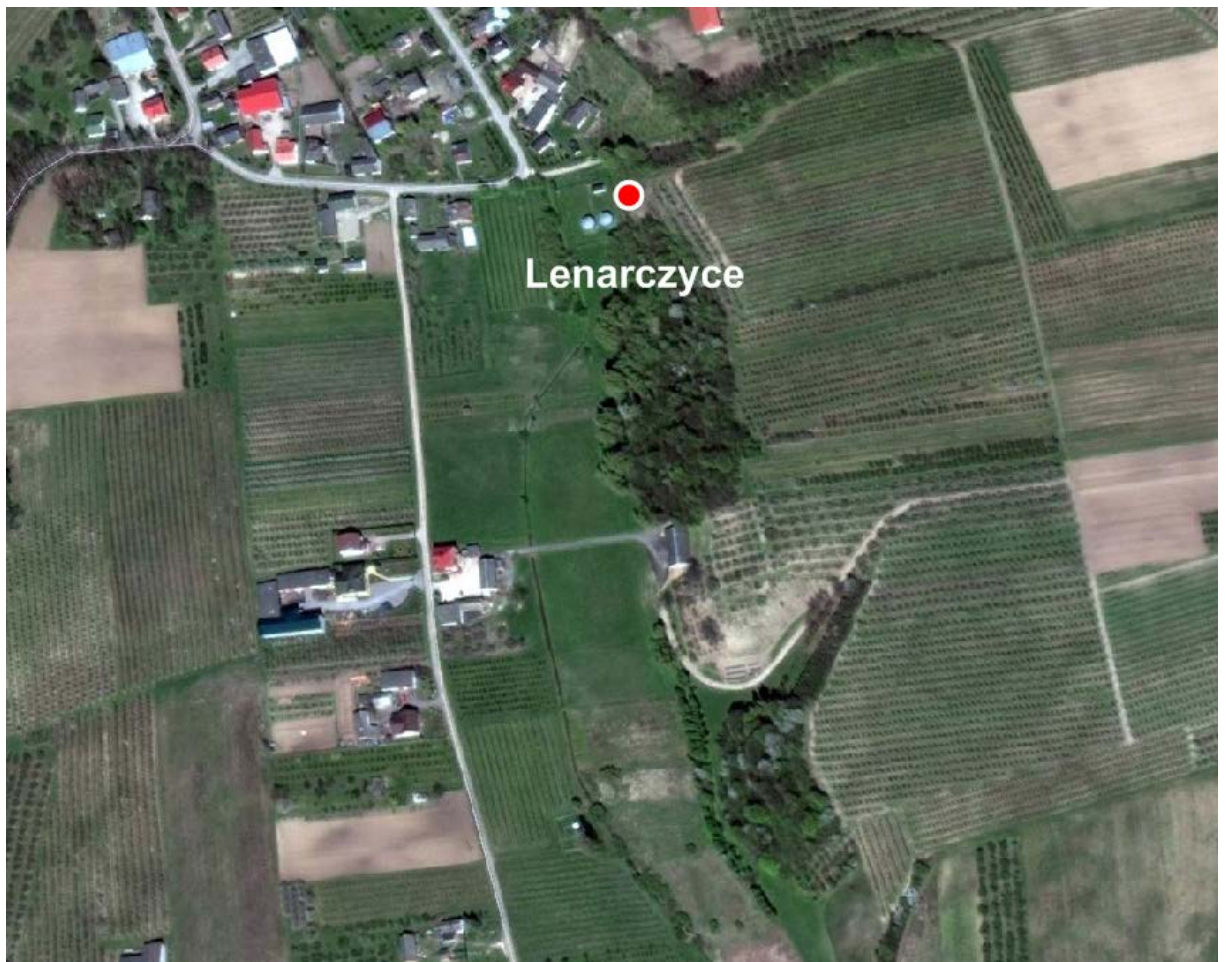


Fig. 37'. Satellite picture with the location of the sampled exposure at Lenarczyce (taken from *Google Earth*).

<i>Locality:</i>	Międzygórz
<i>Geographical coordinates:</i>	50°44′09.34″N; 21°33′45.35″E
<i>Character of locality:</i>	abandoned quarry
<i>Stratigraphy:</i>	Lower Ordovician, Lower Silurian
<i>Samples code:</i>	MDZ
<i>Lithology:</i>	mudstones, sandstones
<i>Literature data</i>	Bednarczyk 1981; Bednarczyk & Stupnicka 2000; Dzik & Pisera 1994.



Fig. 38'. Satelite picture with the location of the Międzygórz Quarry (taken from *Google Earth*).

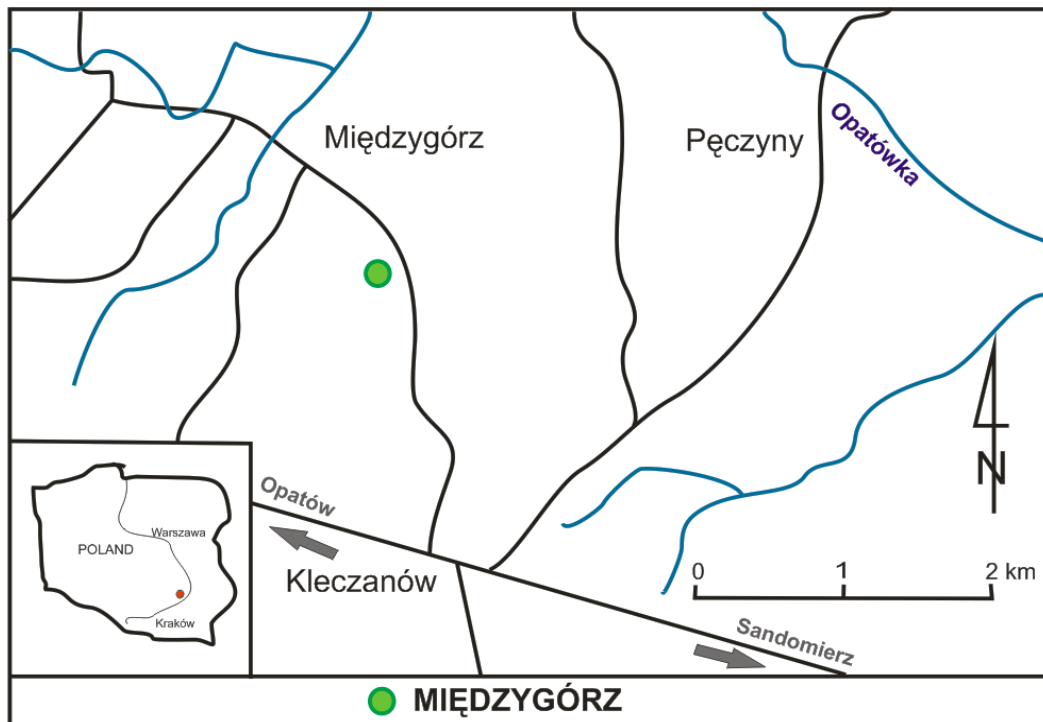


Fig. 39'. Map showing the location of the Międzygórz Quarry.



Fig. 40'. SE face of the Międzygórz Quarry with outcrop of the Lower Ordovician rocks.

<i>Locality:</i>	Mójcza
<i>Geographical coordinates:</i>	50°50′29.44″N; 20°41′09.11″E
<i>Character of locality:</i>	abandoned quarry
<i>Stratigraphy:</i>	Lower Ordovician
<i>Sample codes:</i>	MJ
<i>Lithology:</i>	sandstones
<i>Literature data:</i>	Dzik & Pisera 1994; Trela 1998; Olempska 1994.

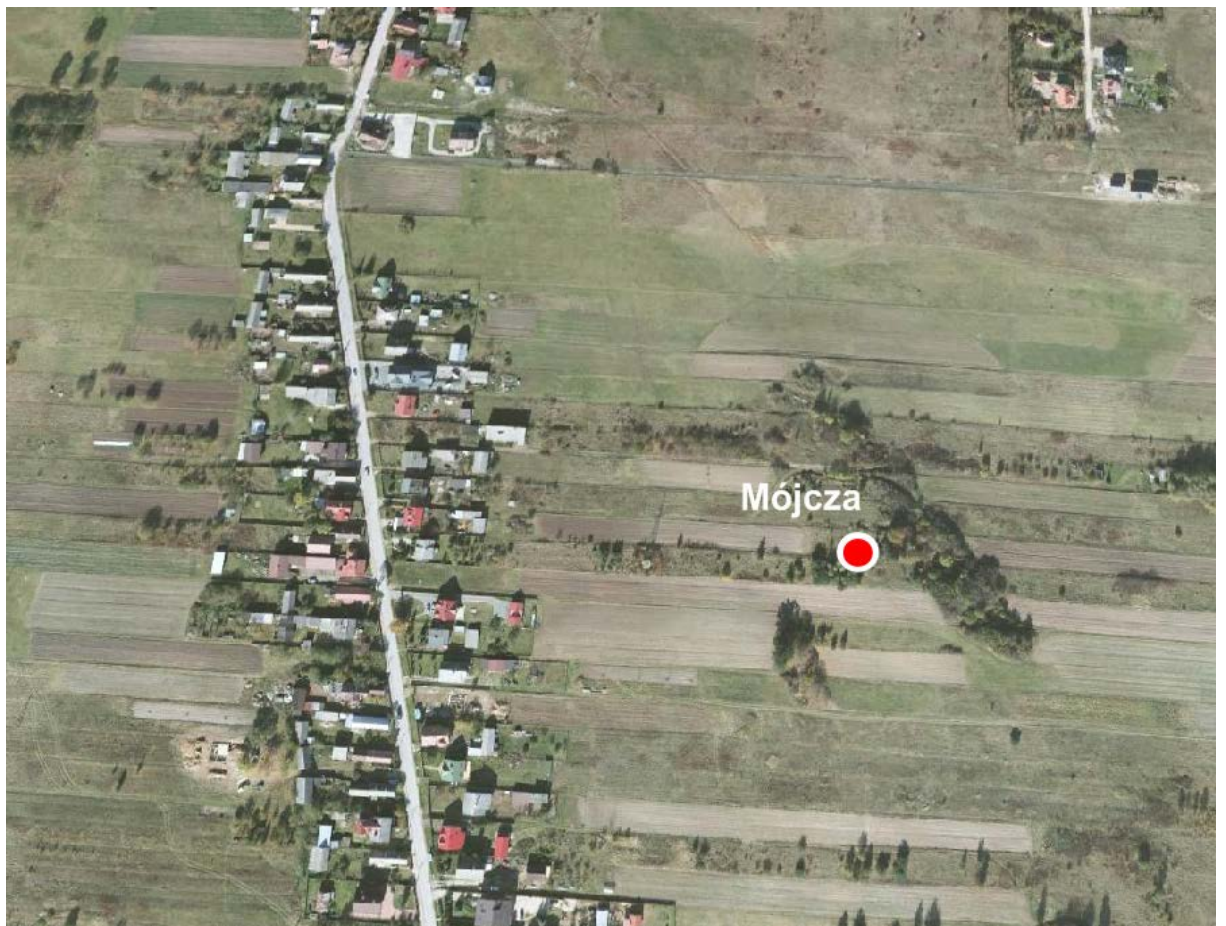


Fig. 41'. Satellite picture with the location of the abandoned quarry at Mójcza (taken from *Google Earth*).

<i>Locality:</i>	Niestachów
<i>Geographical coordinates:</i>	50°50′40.60″N; 20°44′31.36″E
<i>Character of locality:</i>	abandoned quarry
<i>Stratigraphy:</i>	Upper Silurian
<i>Samples code:</i>	NI
<i>Lithology:</i>	greywackes, mudstones, sandstones
<i>Literature data:</i>	Przybyłowicz & Stupnicka 1989; Malec <i>et al.</i> 2016.



Fig. 42'. Satellite picture with the position of the sampled outcrop at Niestachów (taken from Google Earth).

<i>Locality:</i>	Ociesęki
<i>Geographical coordinates:</i>	50°43'45.79"N; 20°58'14.70"E
<i>Character of locality:</i>	natural outcrops on the Sterczyna Hill
<i>Stratigraphy:</i>	Lower Cambrian
<i>Samples code:</i>	OC
<i>Lithology:</i>	sandstones
<i>Literature data</i>	Kowalczewski 1990; Orłowski & Żylińska 2002; Żylińska 2013.



Fig. 43'. Satellite picture with the location of the Sterczyna Hill (taken from *Google Earth*).



Fig. 44'. Outcrop of the Lower Cambrian sandstones on the Sterczyna Hill at Ociesęki.

<i>Locality:</i>	Słowik
<i>Geographical coordinates:</i>	50°49′51.12″N; 20°32′05.04″E
<i>Character of locality:</i>	natural outcrop
<i>Stratigraphy:</i>	Lower Cambrian
<i>Samples code:</i>	SL
<i>Lithology:</i>	siltstones



Fig. 46’. Satelite picture with the location of the sampled exposure at Słowik near Kielce (taken from *Google Earth*).

<i>Locality:</i>	Usarzów
<i>Geographical coordinates:</i>	50°42'24.10"N; 21°31'28.78"E
<i>Character of locality:</i>	abandoned quarry
<i>Stratigraphy:</i>	Middle Cambrian
<i>Samples code:</i>	US
<i>Lithology:</i>	sandstones
<i>Literature data:</i>	Orłowski 1988; Dzik & Orłowski 1995; Żylińska & Szczepanik 2009.



Fig. 47'. Satellite picture with the location of the sampled exposure at Usarzów (taken from *Google Earth*).

<i>Locality:</i>	Widelki
<i>Geographical coordinates:</i>	50°46′32.97″N; 20°56′43.55″E
<i>Character of locality:</i>	natural outcrops
<i>Stratigraphy:</i>	Lower Cambrian
<i>Samples code:</i>	WI
<i>Lithology:</i>	sandstones
<i>Literature data</i>	Żylińska & Masiak 2007; Żylińska & Szczepanik 2009.



Fig. 48’. Satellite picture with the location of the sampled outcrop at Widelki, (taken from Google Earth).

<i>Locality:</i>	Zalesie
<i>Geographical coordinates:</i>	50°43'24.64"N; 21°04'32.60"E
<i>Character of locality:</i>	geological natural outcrops
<i>Stratigraphy:</i>	Lower Cambrian, Lower/Middle Ordovician, Lower Silurian
<i>Collected Samples:</i>	ZE
<i>Lithology:</i>	mudstones
<i>Literature data:</i>	Bednarczyk 1981; Dzik, Pisera 1994; Trela 2006a; Masiak <i>et al.</i> 2003.



Fig. 49'. Satellite picture with the location of the sampled exposure at Zalesie (taken from Google Earth).

<i>Locality:</i>	Żurawiki
<i>Geographical coordinates:</i>	50°44'50.75"N; 21°29'28.91"E
<i>Character of locality:</i>	slope of natural outcrops
<i>Stratigraphy:</i>	Middle Cambrian
<i>Samples code:</i>	ZU
<i>Lithology:</i>	sandstones, mudstones



Fig. 50'. Satellite picture with the location of the sampled outcrop at Żurawniki (taken from *Google Earth*).