JACEK GRABOWSKI

PALAEO- AND ROCK MAGNETISM OF MESOZOIC CARBONATE ROCKS IN THE SUB-TATRIC SERIES (CENTRAL WEST CARPATHIANS) — PALAEOTECTONIC IMPLICATIONS

Polish Geological Institute Special Papers, 5

WARSZAWA 2000

CONTENTS

1. Introduction . . . 6
2. Geological framework . . . 7
2.1. The concept of Adria/Apulia. . . . 7
2.2. Geological structure of the Western Carpathians . . . 8
2.2.1. The Outer Zone . . . 9
2.2.2. Pieniny Klippen Belt . . . 12
2.2.3. Central and Inner West Carpathians . . . 13
2.3. The Tatra Mts. as a part of the Central West Carpathians . . . 14
2.3.1. Tatricum. . . . 15
2.3.2. Fatricum. . . . 15
2.3.3. Hronicum . . . 17
2.3.4. Palaeogene cover . . . 18
2.4. Thermal history of the Tatra Mts. in the Mesozoic and Tertiary . . . 19
2.5. Mesozoic and Tertiary Apparent Polar Wander Paths for Europe and Africa and expected palaeomagnetic directions for the Tatra Mts. . . . 19
3. Sampling and laboratory methods . . . 23
3.1. Sampling . . . 23
3.2. Laboratory methods . . . 24
4. Experimental results . . . 24
4.1. Skupniów Upłaz and Siklawica . . . 24
4.1.1. Locality descriptions . . . 24
4.1.2. Petrography and rock magnetism . . . 24
4.1.3. Demagnetization results . . . 29
4.2. Wierch Spalenisko . . . 31
4.2.1. Locality description . . . 31
4.2.2. Petrography and rock magnetism . . . 31
4.2.3. Demagnetization results . . . 34
4.3. Dolina Długa — Liassic . . . 36
4.3.1. Locality description . . . 36
4.3.2. Petrography and rock magnetism . . . 36
4.3.3. Demagnetization results . . . 38
4.4. Dolina Kryta . . . 39
4.4.1. Locality description . . . 39
4.4.2. Petrography and rock magnetism . . . 40
4.4.3. Demagnetization results . . . 42
4.5. Dolina Długa — Malm/Neocomian . . . 43
4.5.1. Locality description . . . 43
4.5.2. Petrography and rock magnetism . . . 43
4.5.3. Demagnetization results . . . 46
4.6. Sarnia Skała. . . . 48
4.6.1. Locality description . . . 48
4.6.2. Petrography and rock magnetism . . . 49
4.6.3. Demagnetization results . . . 51

4.7. Dolina Strążyska 52
4.7.1. Locality description 52
4.7.2. Petrography and rock magnetism 52
4.7.3. Demagnetization results 55
4.8. Wielkie Koryciska 57
4.8.1. Locality description 57
4.8.2. Petrography and rock magnetism 57
4.8.3. Demagnetization results 59
5. Interpretation and discussion of palaeomagnetic data 60
5.1. Summary of the rock magnetic properties 60
5.2. Summary of the susceptibility and AMS data 61
5.3. Tilt tests and the age of magnetization 62
5.3.1. Normal polarity component A 63
5.3.2. Mixed polarity component 67
5.4. Links between magnetic mineralogy and remagnetization 68
6. Geological implications 69
6.1. Syntectonic remagnetizations of Late Cretaceous age 69
6.2. Primary(?) Early Cretaceous magnetization 71
6.3. Comparisons to Mesozoic palaeopoles from the Tatra Mts. to Mesozoic palaeopoles from adjacent regions 71
6.3.1. Central West Carpathians of Slovakia 71
6.3.2 Outer West Carpathians 73
6.3.3. Eastern Alps 73
6.3.4. Transdanubian Central Mts. 74
6.3.5. Inner Western Carpathians of Slovakia 74
6.4. Mesozoic palaeomagnetic data from the Tatra Mts. in the light of Tertiary kinematics of the Carpathian-Pannonian domain 75
6.4.1. Review of Tertiary palaeomagnetic data from the Western Carpathians and Northern Pannonia. 75
6.4.2. Geological interpretation of Tertiary palaeomagnetic data. 76
6.4.3. Mesozoic poles from the Tatra Mts. corrected for Tertiary movements 77
7. Conclusions 80
7.1. Palaeomagnetism of Mesozoic rocks in the Tatra Mts. 80
7.2. Geological implications of palaeomagnetic results 80
7.3. Considerations for future investigations 81
References 82

GRABOWSKI J., 2000 — **Palaeo- and rock magnetism of Mesozoic carbonate rocks in the Sub-Tatric series (Central West Carpathians) — palaeotectonic implications**. *Polish Geological Institute Special Papers*, 5: 1–88.

Jacek Grabowski, Państwowy Instytut Geologiczny, ul. Rakowiecka 4, 00-975 Warszawa, Poland. E-mail: jgra@pgi.waw.pl

Abstract. The paper presents new palaeomagnetic and rock magnetic data from the Mesozoic carbonates of the Fatric (KriŽna) and Hronic (Choč) units from the Polish part of the Tatra Mts. 55 independently oriented hand samples were collected in the Western Tatra Mts. from the KriŽna nappe (8 localities; Middle Triassic–Lower Cretaceous) and Choč nappe (1 locality; Middle Triassic). The results are interpreted together with data already published from the High Tatric units and compared to the palaeomagnetic database of Mesozoic and Tertiary results from the Eastern Alpine-Carpathian-Pannonian area.

The rock magnetic investigations include IRM experiments, hysteresis measurements, thermomagnetic analysis and low temperature susceptibility measurements. All palaeomagnetically investigated sedimentary rocks in the Tatra Mts. were remagnetized. The age of remagnetization was interpreted as 113–88 Ma during Cretaceous Quiet Zone of normal polarity, synchronously with the Late Cretaceous thrusting in the Central West Carpathians (CWC). Fold test in some KriŽna (Bobrowiec and Suchy Wierch) units revealed that the remagnetization took place before the internal deformations of these units took place. Remagnetization is related mostly to pseudo-single domain (PSD) magnetite. In one locality ("Biancone" limestones of Tithonian/Berriasian age) the mixed polarity component was noted. The component passed the reversal test and was preliminarily interpreted as primary. However, different hysteresis parameters and maximum unblocking temperatures in the normally and reversely magnetized samples indicate complex rock magnetic properties and further investigations should be performed to prove its primary nature.

Identification of syntectonic Late Cretaceous remagnetization let to determine the dip of strata in investigated tectonic units during thrusting. The most numerous and reliable data were obtained in this and earlier studies from the High Tatric parautochthon and Bobrowiec, Hawrań and Suchy Wierch units belonging to the KriŽna nappe. Parautochthon was remagnetized in roughly horizontal position (±5°) while the mentioned KriŽna units were dipping at least 10–20° to the S to SW during magnetization. This implies that horizontal compression might be an important factor of their emplacement. However in the two localities from some other KriŽna units palaeomagnetic directions indicate that rocks were magnetized dipping 20–60° to the north thus the attitude of strata during overthrusting was complex. Palaeolatitude of the Tatra Mts. in the Late Cretaceous amounts to 30–36° N. Possibly primary component isolated in the Tithonian/Berriasian limestones indicate palaeolatitude 21–23° N which is closer to the African/Adriatic than European plate. Palaeodeclinations of Mesozoic components reveal 20–50° clockwise rotation of parautochthonous unit and KriŽna nappe in relation to the European platform. These are most likely resultant values of ca. 60° counter-clockwise rotation after Oligocene and 80–110° clockwise rotation between Cenomanian–Turonian and Eocene. After subtracting the effect of Tertiary rotation, the Mesozoic palaeopoles from the Tatra Mts. are matched with pre-Gosau palaeopoles from the Northern Calcareous Alps (NCA). These two rotational events are most probably characteristic also for the CWC in Slovakia, however their magnitude is variable due to local tectonic effects. Existing palaeomagnetic data point to palaeotectonic affinity of the CWC and NCA in the Mesozoic. On the other hand, the CWC reveal different rotation pattern than the areas belonging to the Adriatic plate (Southern Alps, Inner West Carpathians (IWC) and Northern Pannonia). It seems that different azimuth of the Cretaceous palaeodeclinations between the CWC (predominantly clockwise rotations) and IWC, and Outer West Carpathians (exclusively counter-clockwise rotations) point to Cretaceous rotational movements along the Pieniny Klippen Belt and Meliata suture zones.

Key words: palaeomagnetism, rock magnetism, palaeotectonics, Mesozoic, Tertiary, Carpathians, the Tatra Mts.

Abstrakt. W pracy przedstawiono nowe i częściowo opublikowane dane paleo- i petromagnetyczne z mezozoicznych skał węglanowych płaszczowin reglowych dolnej (kriżniańska — Fatricum) i górnej (choczańska — Hronicum) w polskiej części Tatr. Próbki do badań paleomagnetycznych pobrano z 8 stanowisk (środkowy trias–dolna kreda) w płaszczowinie reglowej dolnej i 1 stanowiska (środkowy trias) w płaszczowinie reglowej górnej. Ogółem pobrano 55 niezależnie zorientowanych próbek ręcznych. Wyniki badań zinterpretowano razem z opublikowanymi już danymi z tatrzańskich serii wierchowych i odniesiono do paleomagnetycznej bazy danych z obszaru wschodnich Alp, Karpat i rejonu pannońskiego.

Przeprowadzone eksperymenty petromagnetyczne obejmowały badania izotermicznej pozostałości magnetycznej (IRM), parametrów pętli histerezy, analizy termomagnetyczne i pomiary podatności magnetycznej w niskich temperaturach. Wszystkie badane skały uległy przemagnesowaniu. Wiek przemagnesowania zinterpretowano na przedział 113–88 mln lat temu, najprawdopodobniej podczas ruchów płaszczowinowych na obszarze Centralnych Karpat Zachodnich, w czasie trwania długiej zony o normalnej polarności („Cretaceous Quiet Zone”). W jednym stanowisku (wapienie „biancone” wieku tyton–berias) stwierdzono obecność składowych namagnesowania o mieszanej polarności, które dają pozytywny wynik testu inwersji i zostały wstępnie zinterpretowane jako składowe pierwotne. Jednak zróżnicowane właściwości petromagnetyczne normalnie i odwrotnie namagnesowanych próbek (maksymalne temperatury odblokowujące i parametry pętli histerezy) sprawiają, że kwestia pierwotności namagnesowania pozostaje otwarta.

Wyróżnienie syntektonicznego przemagnesowania późnokredowego pozwoliło na określenie położenia warstw w poszczególnych jednostkach tektonicznych podczas ruchów płaszczowinowych. Najliczniejsze i najbardziej wiarygodne dane uzyskano z jednostki wierchowej parautochtonicznej oraz z jednostek kriżniańskich Bobrowca, Suchego Wierchu i Hawrania. Parautochton uległ przemagnesowaniu w pozycji subhoryzontalnej (±5°), podczas gdy wymienione jednostki reglowe uzyskały namagnesowanie zapadając 10–20° w kierunku S do SW. Wniosek ten może mieć znaczenie przy określaniu mechanizmu transportu płaszczowin reglowych: w tym wypadku wydaje się, że główną rolę odgrywała pozioma kompresja. Jednak w dwóch stanowiskach z innych jednostek kriźniańskich wtórne kierunki przemagnesowania wskazują, że skały zostały przemagnesowane zapadając 20–60° na N.

Paleoszerokość Tatr w późnej kredzie wynosiła 30–36°, natomiast na przełomie jury i kredy 21–23°. Paleodeklinacje kierunków mezozoicznych z Tatr wykazują 20–50° prawoskrętnej rotacji w stosunku do platformy europejskiej. Jest to najprawdopodobniej wypadkowa wartość dwóch rotacji o różnym wieku: lewoskrętnej o kąt 60° po oligocenie i prawoskrętnej o kąt 80–110° między cenomanem/turonem a eocenem. Po odjęciu efektów rotacji trzeciorzędowych paleobieguny z Tatr stają się bliskie przedsenońskim paleobiegunom z północnych Alp Wapiennych. Wymienione dwie rotacje najprawdopodobniej objęły również cały blok Centralnych Karpat Zachodnich na Słowacji, jednak ich amplituda mogła być zróżnicowana wskutek efektów lokalnych. Dane paleomagnetyczne wskazują na bliskość Centralnych Karpat Zachodnich i północnych Alp Wapiennych w mezozoiku, natomiast wykazują różnice w stosunku do obszarów zaliczanych do płyty adriatyckiej (Alpy Południowe, Wewnętrzne Karpaty Zachodnie, Północna Pannonia). Zróżnicowane zwroty pokredowych rotacji tektonicznych w Centralnych Karpatach Zachodnich i obszarach położonych bezpośrednio na N i S od nich (tzn. w Wewnętrznych Karpatach Zachodnich i Karpatach Zewnętrznych) wskazują na możliwość rotacji tektonicznych wzdłuż linii pienińskiego pasa skałkowego i szwu oceanu Meliata.

Słowa kluczowe: paleomagnetyzm, petromagnetyzm, paleotektonika, mezozoik, trzeciorzęd, Karpaty, Tatry.

1. INTRODUCTION

Palaeomagnetism has been established a useful tool for investigation of tectonics of mobile belts (Van der Voo, 1993; Morris, Tarling, 1996). Alpine and Carpathian orogens have been extensively studied palaeomagnetically since the 1970s (for review see Heller *et al*., 1989; Marton, Mauritsch, 1990) but the amount of palaeomagnetic data is still insufficient to solve some key problems e.g. to reconstruct the Mesozoic and Tertiary history of the terranes in the Eastern Alpine-Carpathian area. This study was performed in the Tatra Mts. which are situated in the northern part of the Central West Carpathians in Poland and Slovakia. The investigation aims to obtain reference Mesozoic and Tertiary palaeomagnetic data for the Tatra Mts. and to use them for palaeogeographic and palaeotectonic reconstructions comparing with existing palaeomagnetic data base for central Mediterranean and European and African continents. Intriguing question is whether palaeomagnetism supports the Apulian (Adriatic) origin of the Tatra Mts. and the Central West Carpathian block.

First palaeomagnetic data obtained from Middle and Upper Jurassic sediments of the KriŽna nappe in the Western Tatra Mts. were published by Kruczyk *et al*. (1980), Kądziałko-Hofmokl *et al*. (1985) and Kądziałko-Hofmokl, Kruczyk (1987). The pre-folding palaeomagnetic directions were interpreted as primary and indicated that KriŽna unit in the Tatra Mts. did not rotate significantly relatively to the Carpathian foreland since ca. 160 Ma. This conclusion was very confusing for geologists because their tectonic models required large mobility of the Central West Carpathians in both Mesozoic and Tertiary (e.g. Birkenmajer, 1985; Balla, 1987; Sotak, 1992; Csontos *et al*., 1992). Further palaeomagnetic work on Jurassic sediments of the KriŽna unit in Slovakia (Kruczyk *et al*., 1992) revealed local tectonic rotations between studied localities which were interpreted as related to oroclinal bending of the area. However these results were not compared to existing palaeomagnetic data base of Central Mediterranean (Marton, Mauritsch, 1990) and were treated as "anomalous" since bulk of Mesozoic and Tertiary palaeomagnetic data from Outer West Carpathians, Inner West Carpathians and Northern Pannonia revealed consistent counter-clockwise rotation of palaeomagnetic declinations (Krs *et al*., 1982; Marton, Marton, 1983; Marton *et al*., 1988). Very recent study of Marton *et al*. (1999) confirmed that post-Oligocene counter-clockwise rotation embraced also the Central Carpathian Palaeogene Basin. In 1990s the author started systematic palaeomagnetic observations in various tectonic units of the Tatra Mts. (Grabowski, 1995a, 1995b, 1997a). It was indicated that Mesozoic rocks of High Tatric units were remagnetized in the Middle Cretaceous prior to the folding. The proximity of the Tatra Mts. to the European Platform during remagnetization event was confirmed. Net 30° clockwise rotation of the Tatra Massif in relation to the European Platform after Middle Cretaceous was implied (Grabowski, 1997a). It was also suggested that the same remagnetization might occur in the Sub-Tatric units, thus the results of Kądziałko-Hofmokl *et al*. (1985) should be reinterpreted. In this study the new palaeomagnetic data from the Middle Triassic–Lower Cretaceous rocks of Sub-Tatric series are presented and the hypothesis of Cretaceous remagnetization in the Tatra Mts. is evaluated. The problem of variegated sense of rotations (counter-clockwise in the Tertiary, clockwise in the Mesozoic) in the Central Carpathian area is also highlighted (Grabowski, Nemčok, 1999).

The paper comprises shortened version of the author's Ph.D. dissertation. Preliminary results of the study were presented at the PANCARDI workshop in Zakopane (Grabowski, 1997b). Some ideas presented in the paper were published in 1999 (Grabowski *et al*., 1999; Grabowski, 1999; Grabowski, Nemčok, 1999).

2. GEOLOGICAL FRAMEWORK

2.1. THE CONCEPT OF ADRIA/APULIA

Interaction of minor continental plates between Africa and Eurasia is of central importance in understanding of orogenic processes within the Mediterranean mobile belt. The number of these microplates and the boundaries between them are often disputed. The presence of Apulian microplate is accepted in all palaeotectonic reconstructions and it certainly played a key

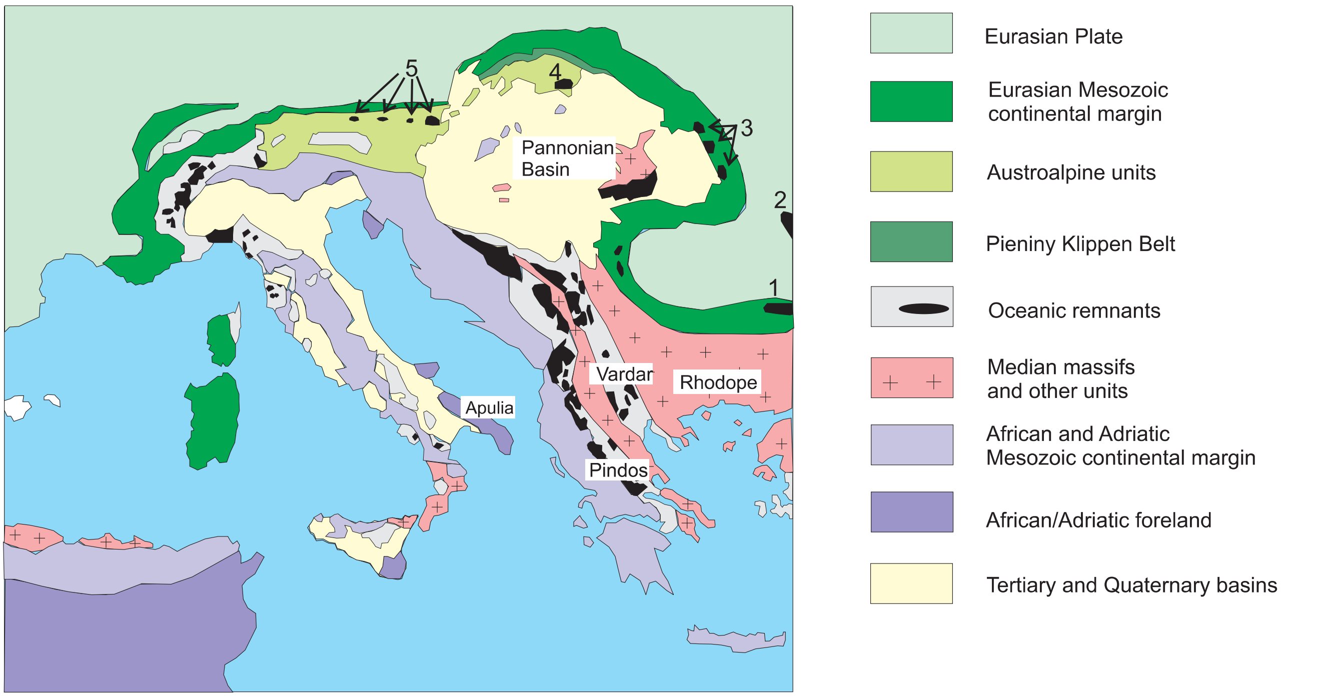


Fig. 1. Structural elements of central part of Alpine Belt (modified after Channell and Kozur, 1997)

1 — Kotel zone, *2* — Dobrudzha, *3* — Transylvanian nappes, *4* — Meliaticum of the Inner West Carpathians, *5* — Meliaticum of the Eastern Alps

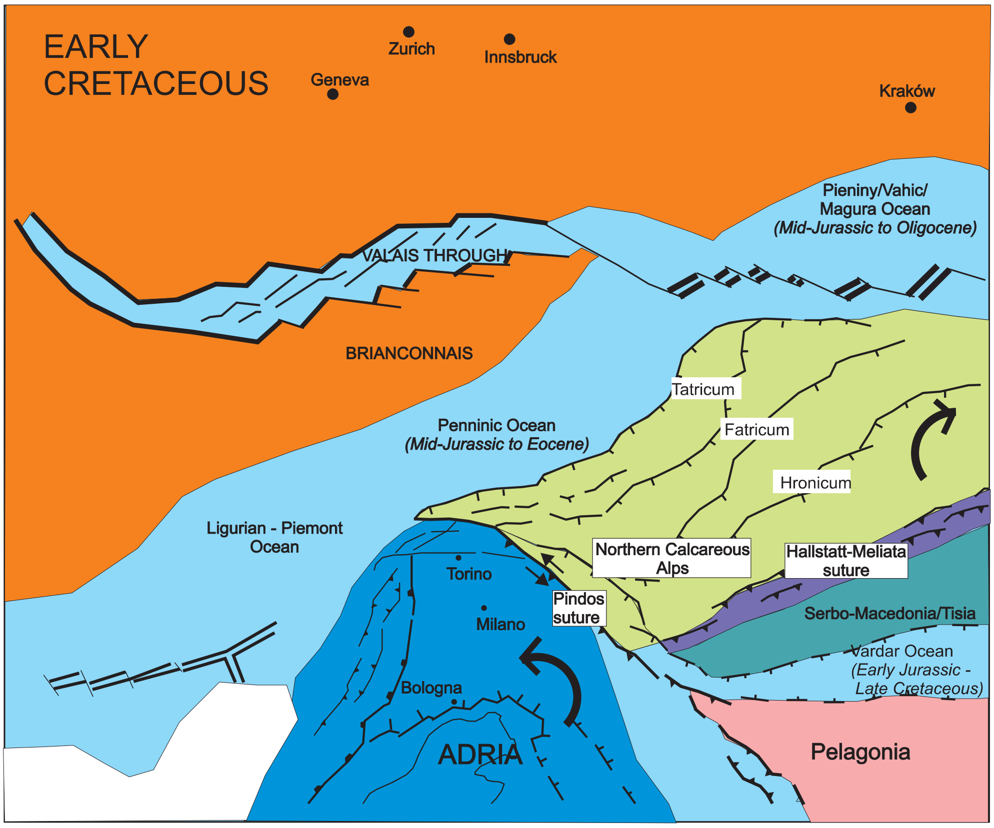


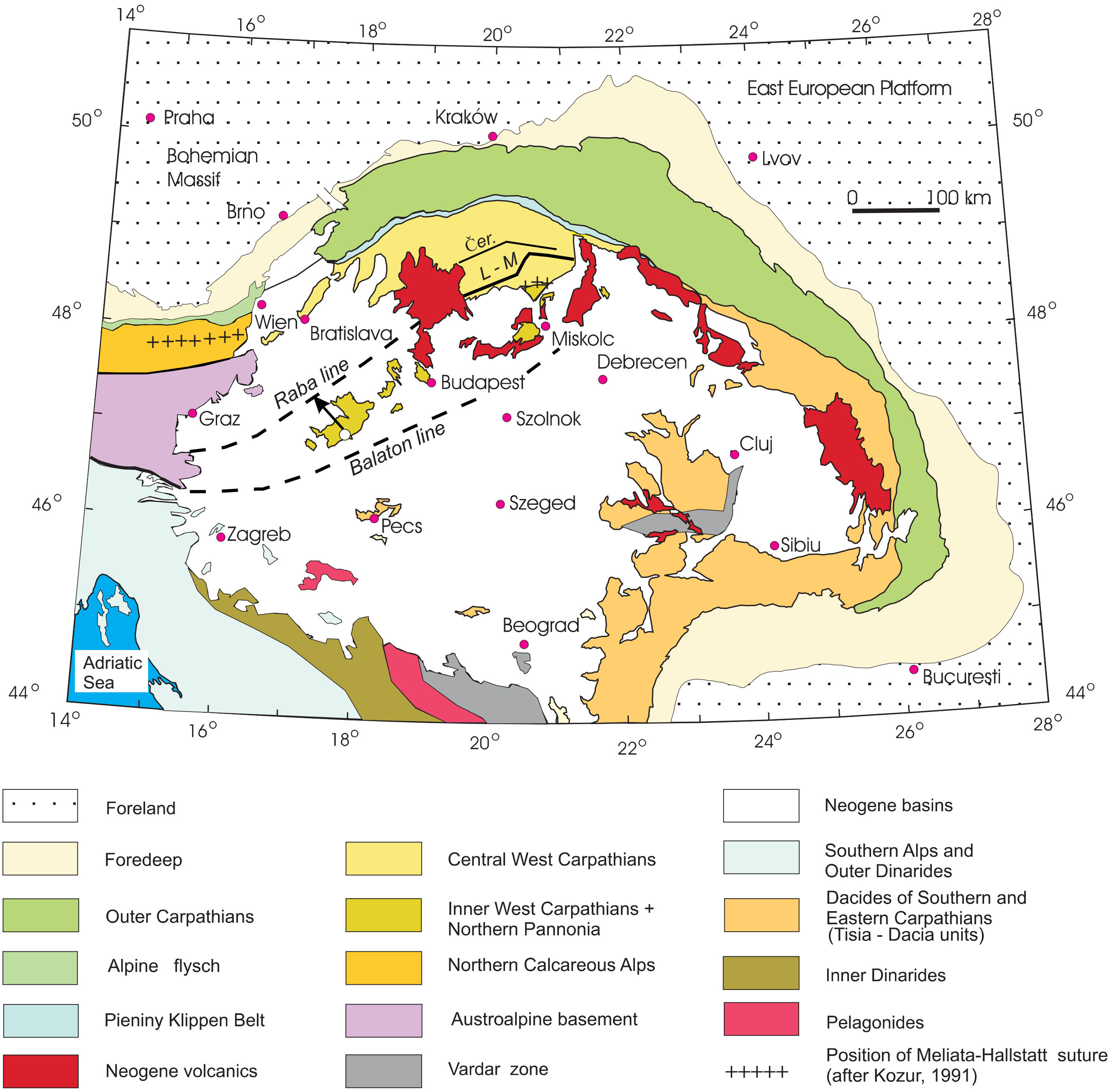
Fig. 2. Schematic Early Cretaceous palaeogeography of the Mediterranean region (modified after Channel, 1996 and Channell, Kozur, 1997)

Arrows — predominant senses of Mesozoic rotations relative to stable European plate

role in the origin of the Alpine-Carpathian system. Sometimes the term Adriatic microplate (or Adria) is also used. The concept of Apulia as a promontory of the African plate was first put forward by Argand (1924) and supported by subsequent investigations of geologists (i.e. Frisch, 1979; Burchfiel, 1980; Gealey, 1988; Yilmaz *et al*., 1996) and palaeomagnetists (Channell *et al*., 1979; Lowrie, 1986; Channell, 1996). The present state of knowledge about Apulia was summarized by Channell (1996). The autochthonous remnants of the microplate are preserved in several places of the circum--Adriatic area (Fig. 1). They include Apulia (*sensu stricto*) and Gargano regions of the Southern Italy, a part of the Southern Alps and Istria. Palaeomagnetic data from the Mesozoic and Lower Tertiary rocks of these regions clearly indicate their African palaeotectonic affinity (Channell, 1996). Also apparent polar wander paths (APWPs) of the Umbria Marche (Appenines), Southern Alps and Transdanubian Central Mts. (Hungary) are of African shape, although these areas were rotated respective to the autochthonous formations. It is a matter of debate whether Apulia was attached to the African plate and, if not, when did the separation take place (see Channell, 1996 for discussion). The northern boundary of Apulia is also not clear. Frisch (1979) included the Southern Alpine as well as Austroalpine units to the Apulian plate and this point of view is generally accepted by geologists (i.e Dercourt *et al*., 1986; Ratschbacher *et al*., 1991; Ziegler *et al*., 1996). In this concept the western and northern boundaries of Apulia would trace along the Ligurian–Piemont–Southern Penninic Ocean. However Mesozoic palaeomagnetic data from the Southern Alps and Northern Calcareous Alps differ considerably (Marton, Mauritsch, 1990; Mauritsch, Marton, 1995) and it is very likely that the northern Austroalpine units constituted a separate microplate, at least in the Jurassic (Channell *et al*., 1992; Channell, 1996) (Fig. 2). From the east the Adriatic promontory was bounded by the Pindos or Vardar Oceans (different concepts, see Stampfli and Marchant, 1995; Yilmaz *et al*., 1996). Evaluation of palaeomagnetic constraints for Apulian origin of the Central West Carpathians is one of important aims of the present study.

2.2. GEOLOGICAL STRUCTURE OF THE WESTERN CARPATHIANS

The tectonic division of the Western Carpathians applied in this paper is a compilation from the papers of Polish, Slovakian and Hungarian geologists. It is based mainly upon the papers of Biely (1990), Csaszar *et al*. (1990), Kozur, Mock (1996), Häusler *et al*. (1993), Plašienka *et al*. (1997a, 1997b) and Unrug (1982). Classical division comprise: Foreland, Carpathian Foredeep, Outer Carpathians (Carpathian Flysch Belt) the Pieniny Klippen Belt and Central + Inner West Carpathians (Fig. 3). First three units, sometimes combined as the Outer Zone, are described only briefly.



Čer. — Čertovica line, L-M — Lubenik–Margecany line

Fig. 3. Tectonic map of the Carpathians and adjacent areas

2.2.1. THE OUTER ZONE

The Outer Zone consists of the Foreland, Carpathian Foredeep and Carpathian Flysch Belt. The Foreland is heterogeneous and is formed by the Bohemian Massif in Czechia, Epi-Variscan Platform in the Moravia and Silesia-Cracow region (partially covered with Mesozoic and Tertiary sediments) and the East European Platform with its sedimentary cover in the eastern Poland and Ukraine. The Carpathian Foredeep started to form in the Early Miocene before the final overthrusting of the Outer Carpathian nappes. It is filled with mostly clastic sediments and evaporites of Miocene age of few hundreds up to 3500 m thickness (Oszczypko, 1997). Carpathian Flysch Belt consists of a stack of sedimentary nappes involving Lower Cretaceous to Miocene deposits. They are thrust over the Carpathian Foredeep. The flysch is predominant component of the sedimentary series although some alcaline volcanites (teschenites — in the Silesian unit), carbonates and radiolarites also occur especially in the Lower Cretaceous. The most important flysch nappes are: Skole, Sub-Silesian, Silesian and the most internal Magura Nappe. The outermost Stebnik Nappe which occurs in the eastern part of the Western Carpathians only, consists of flysch and Miocene rocks of the Carpathian Foredeep folded together and thrusted over the autochtonous Miocene of the Foredeep. The

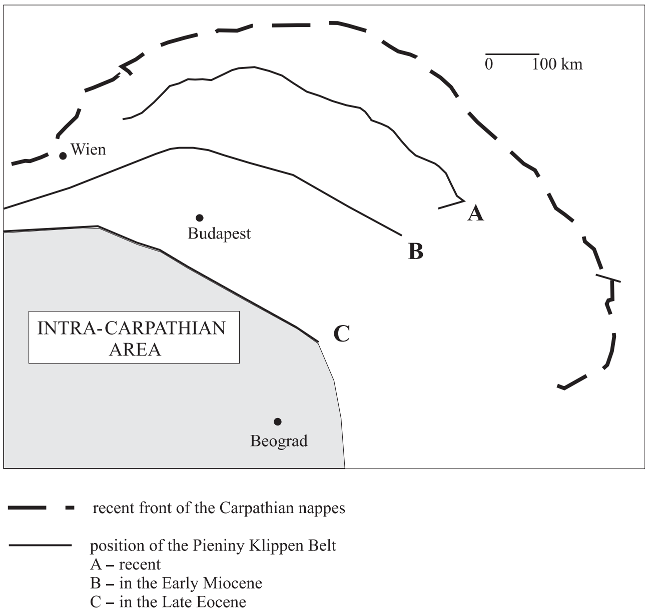


Fig. 4. Position of the Pieniny Klippen Belt in time (after Oszczypko, Ślączka, 1985 and pers. commun., *fide* Csontos *et al.*, 1992)

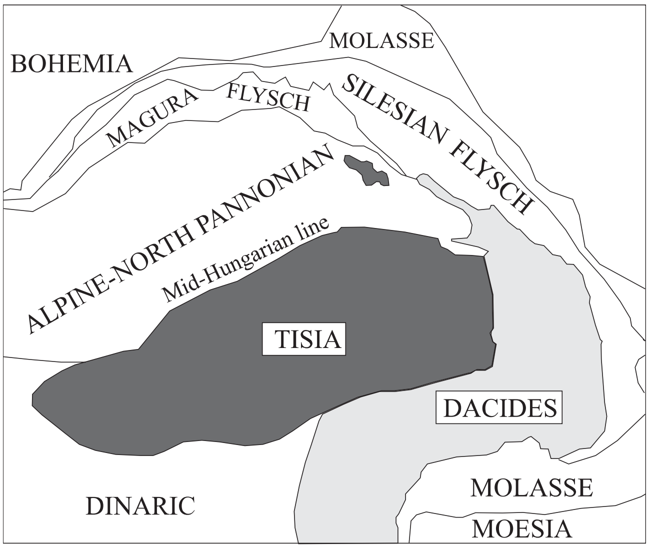


Fig. 5. Palaeogene terranes in the Carpathians (after Csontos *et al.*, 1992)

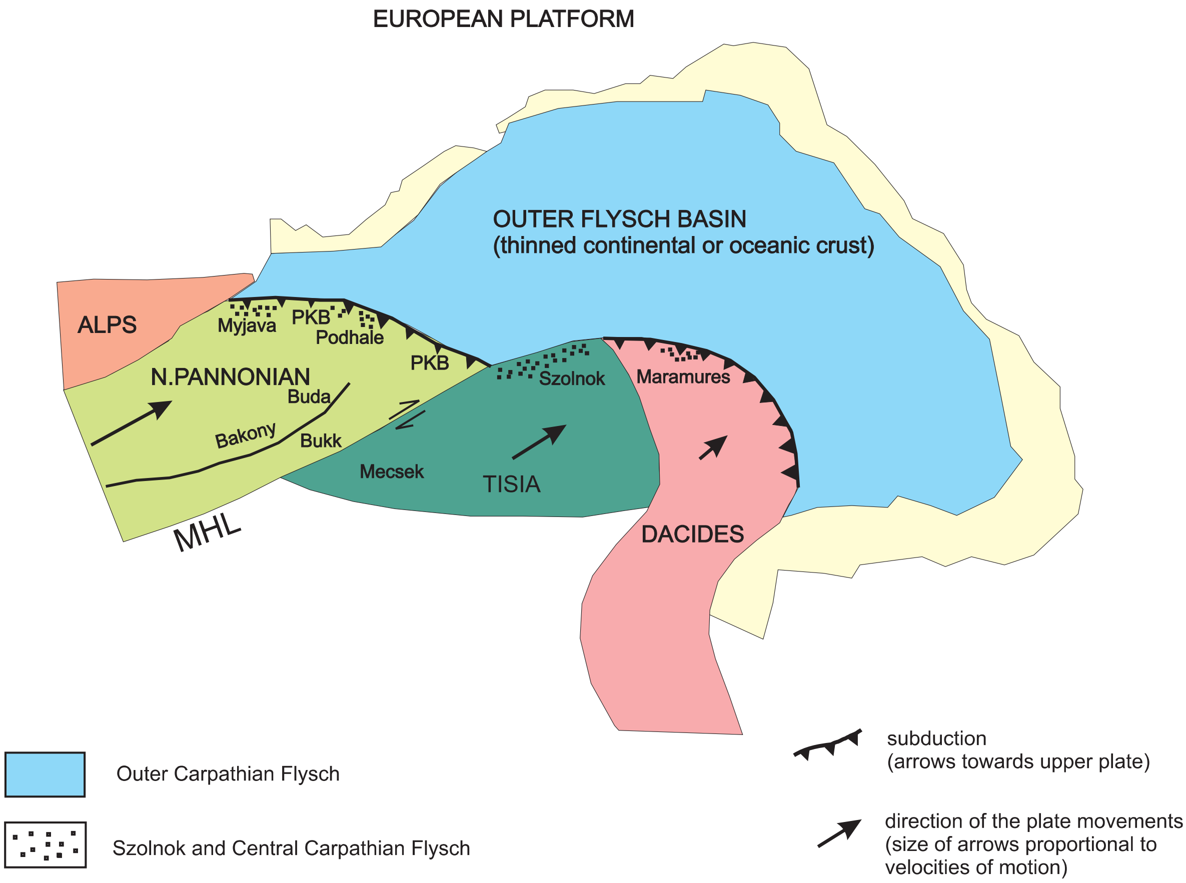


Fig. 6. Proposed relative position of the terranes at the end of Eocene (after Csontos *et al.*, 1992)

MHL — Mid-Hungarian Lineament

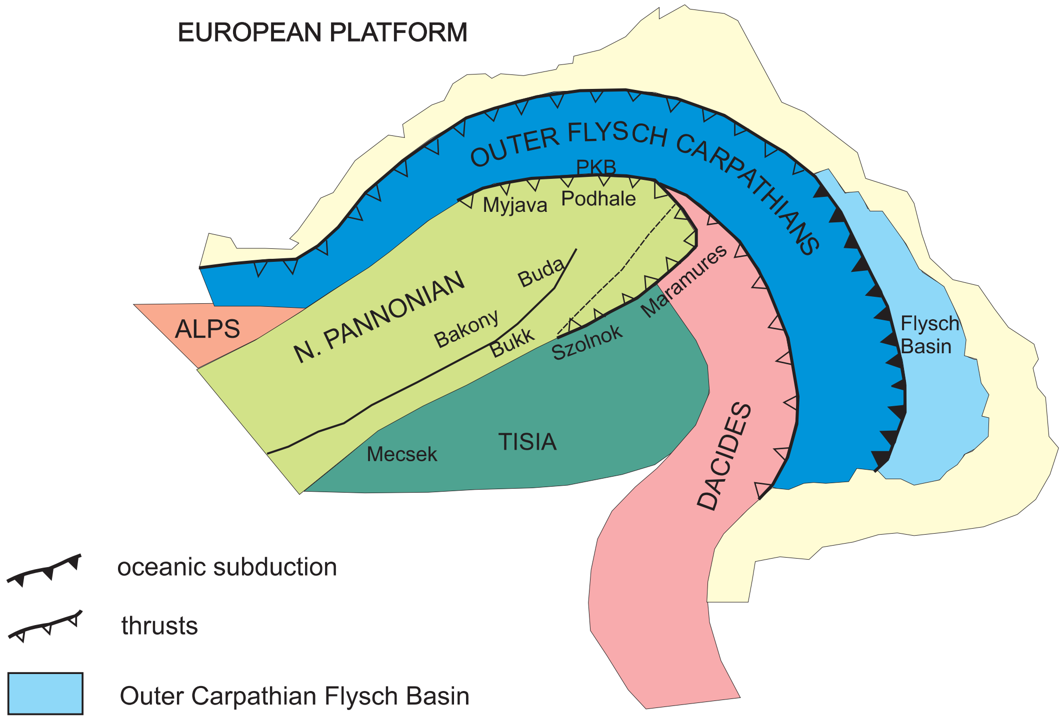


Fig. 7. Proposed position of the terranes in the Early Miocene (after Csontos *et al.*, 1992)

total width of the Outer Flysch Carpathians before the Neogene tectogenesis is estimated as 500–600 km (Unrug, 1979).

The Neogene development of the Outer Carpathians was related to theconvergence of European and African plates. The collision of the Northern Pannonian terrane with European foreland plate took place in the Miocene. It is recently interpreted as the result of tectonic escape of the Northern Pannonian terrane from the Alpine collision zone (Balla, 1987;

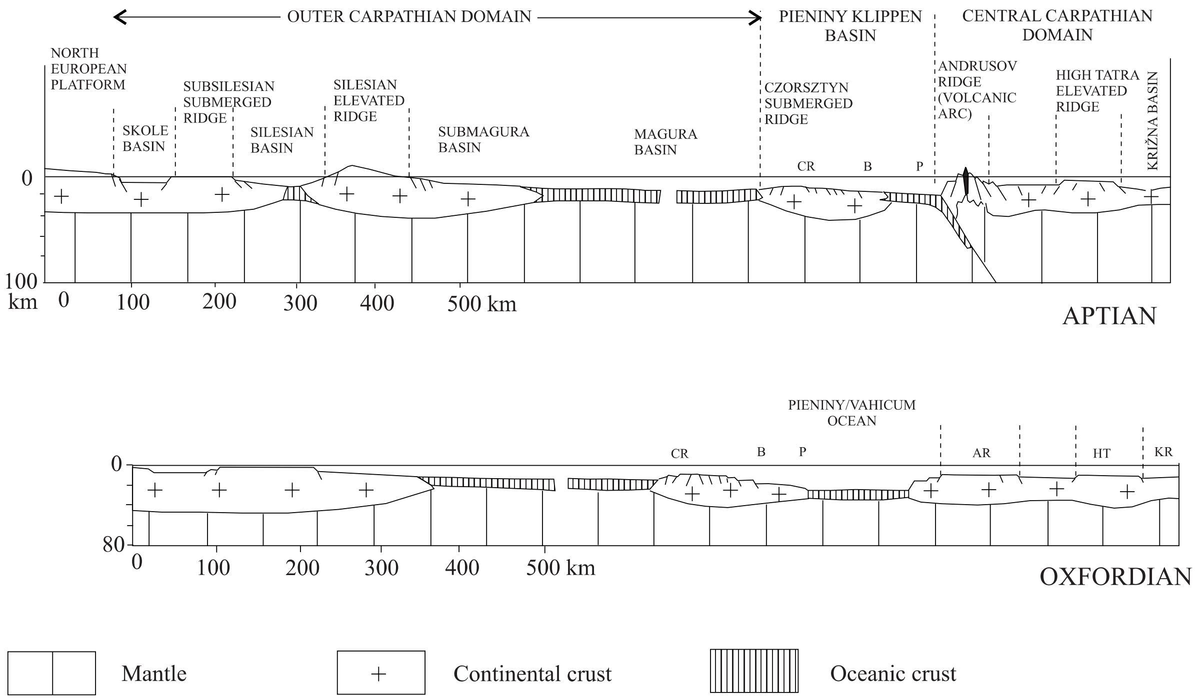


Fig. 8. Palinspastic reconstructions of the Pieniny Klippen Basin with adjacent Outer and Central West Carpathians in Oxfordian and Aptian (after Birkenmajer, 1988, modified)

CR — Czorsztyn Ridge, B — Branisko succession, P — Pieniny succession, AR — Andrusov Ridge, HT — High Tatra Ridge, KR — KriŽna Basin

Csontos *et al*., 1992). The movement would take place due to subduction of the oceanic or thinned continental crust of the European plate. One possible scenario is presented in the Figures 4–7. In this model the Outer Carpathians are interpreted as an accretionary wedge that evolved on the northern margin of the North Pannonian terrane. The main driving forces of the plate tectonic mechanisms in the area were the Neogene subduction roll-back beneath the Carpathians (e.g. Royden, Burchfiel, 1989) and lateral extrusion of the continental masses from the Alpine collision zone, resulting in east-northeastward motion of tectonic wedges (Ratschbacher *et al*., 1991).

It is still not clear if the flysch sediments of the Outer Carpathians were deposited on the continental or oceanic crust. Magura ocean was introduced by Birkenmajer (1976, 1986) as the substratum of the Magura unit (Fig. 8). This idea was supported by arguments of Ney (1976), Sikora (1976) and Tokarski (1980). The ocean opened in the Early/Middle Jurassic, its partial subduction took place in the Late Cretaceous (Laramian phase) while total closure was completed by the Early Miocene. According to Oszczypko (1992) the Magura and Pieniny sedimentary areas were a uniform palaeogeographic zone up to Albian, bordered to the north by the Silesian submerged ridge. The short-lived Silesian ocean (opening: Malm/Neocomian, closure: Late Cretaceous/Early Tertiary) occurred north of the Silesian ridge (Birkenmajer, *op. cit*). However, according to heavy mineral data of Winkler and Ślączka (1992), the Outer Carpathian, Silesian and Dukla basins as well as the northern part of the Magura basin were supplied by continental basement rocks. They occuppied the position of a deep continental basement floored foreland basin. Chemical composition of the Neocomian teschenites of the Silesian unit point to a short-term rifting of the continental crust (Narębski, 1990; Hovorka, 1996) discarding the possibility of oceanic type of crust. On the other hand, the existence of the Neogene subduction zone is very likely because of the presence of well preserved volcanic arc in the Slovakia (Burchfiel, 1980). Andesitic stratovolcanoes developed along the Western Carpathian arc between 16 and 10 Ma (Downes, 1996; Kovač *et al*., 1997; Nemčok *et al*., 1998 and references herein). Csontos *et al*. (1992) argue for oceanic substratum of Magura flysch nappes. According to these authors consumption of 500 km continental substratum would have to imply a "Himalaya-type" of orogeny followed by elevation of the high mountain.

2.2.2. PIENINY KLIPPEN BELT

Pieniny Klippen Belt is a narrow zone, from a few hundred meters to about 20 km wide, that stretches along the strike of the Western Carpathian arc from the Vienna Basin to Poiana Botizei (Romania) (Fig. 3). It is a heterogeneous structure that consists of variegated Mesozoic sedimentary series (so called Klippen successions). They consist of the Lower Jurassic to Upper Cretaceous limestones and shales with their uppermost Cretaceous/Palaeogene sedimentary cover. They were thrusted and folded together during several tectonic phases in the Late Cretaceous/Palaeogene and Miocene (for detailed review see Birkenmajer, 1986). The most important Klippen successions are (from north to south): the Czorsztyn succession, Pieniny, Branisko and Haligovce successions. The Czorsztyn succession was deposited on the aseismic-type ridge (Fig. 8). It is characterized by condensed calcareous deposition with numerous stratigraphical breaks. It is regarded as a boundary between the Outer Carpathian and Central Carpathian realm in the Mesozoic. The Pieniny and Branisko successions originated in the deeper zone than the Czorsztyn succession what is inferred from the radiolarites occurring in the Middle and Upper Jurassic. The Pieniny Klippen Belt area was bounded from the south by the Andrusov ridge which was reconstructed basing on the analysis of exotic pebbles (Birkenmajer, 1988). The position of the Andrusov ridge as the southern boundary of the Pieniny basin was recently questioned by Plašienka (1995) who placed the exotic ridge in the Veporic or even more southern areas of the Central West Carpathians. This concept is still disputed, also among the Slovakian and Hungarian geologists (e.g. Kozur, Mock, 1996). The intrusion of andesites took place in the Neogene (Birkenmajer, 1984; Birkenmajer *et al*., 1987). They cut the Cretaceous and Palaeogene cover units. The Pieniny Klippen Belt is bounded in the south and north along most of its length by strike-slip faults of Early Miocene age. According to Birkenmajer (1985) the Pieniny Klippen Belt acted as a mega-shear zone in the Early Neogene due to transpression between the Central Carpathian block and the Outer Carpathian flysch belt.

Existence of the oceanic domain between the Central West Carpathians and Czorsztyn Ridge of the Pieniny Klippen Belt was postulated by Birkenmajer (1976, 1986). It is regarded as the prolongation of the Alboran-Ligurian-South Penninic Ocean (Mahel, 1981; Plašienka, 1995; Tomek, 1993; Dumont

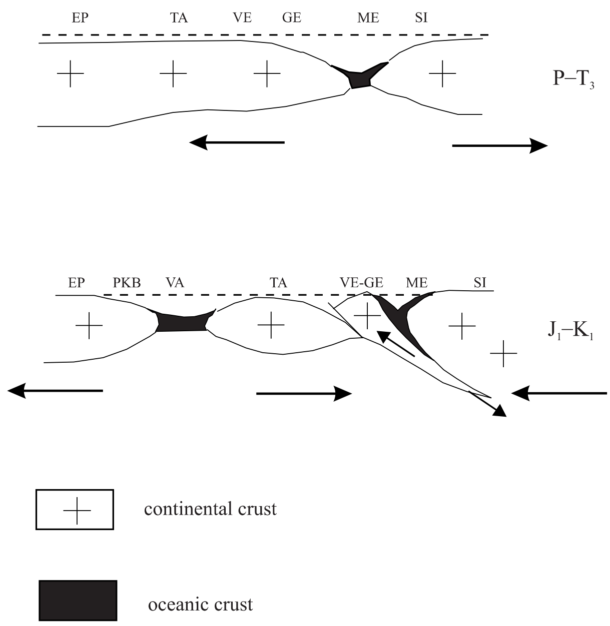


Fig. 9. Palaeotectonic-kinematic model of the Central and Inner West Carpathians during Permian to Early Cretaceous period (after Putiš, 1992, modified)

EP — European Platform, TA — Tatricum, VE — Veporicum, GE — Gemericum, ME — Meliaticum, SI — Silicicum, PKB — Pieniny Klippen Belt, VA — Vahicum

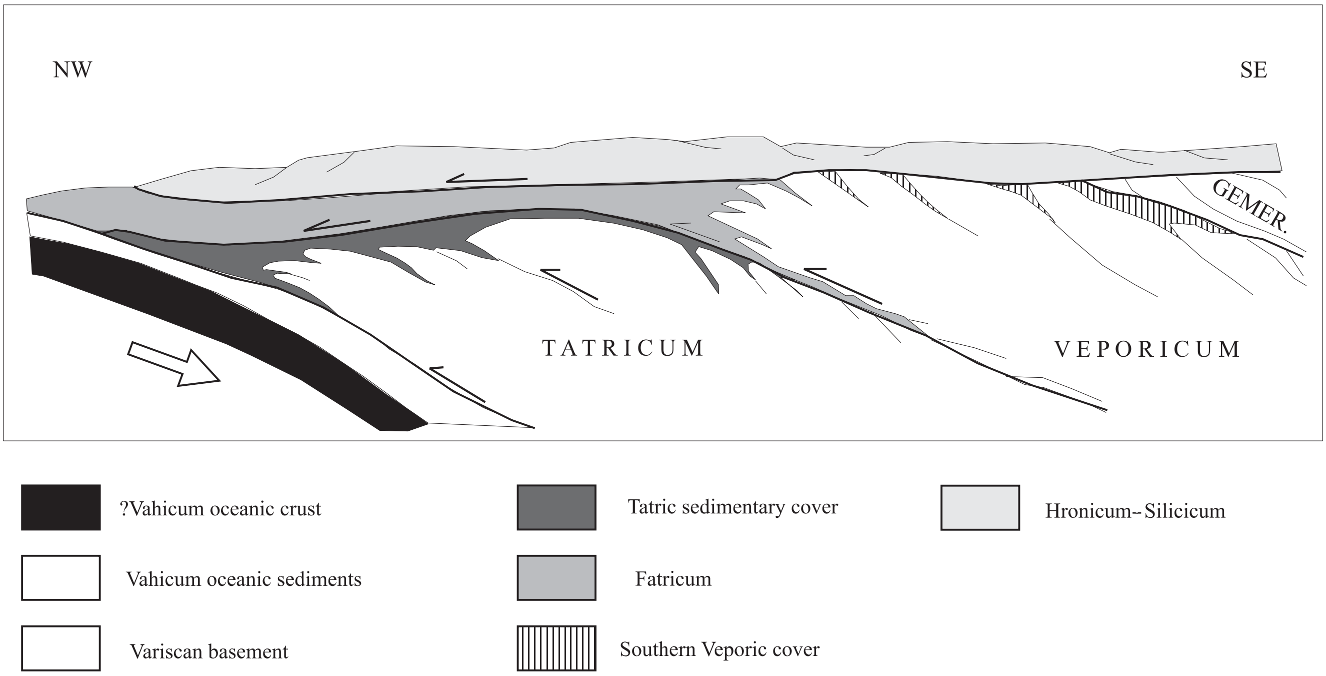


Fig. 10. Hypothetical cross section of the principal units of the Central West Carpathians in pre-Senonian time (after Häusler *et al.*, 1993)

et al., 1996). The terms Vahicum or Pieniny Ocean are commonly used. The ocean opened not later than in the Middle Jurassic (Fig. 9) and closed in the Latest Cretaceous to Early Tertiary (Fig. 8). The subduction took place under the Andrusov Ridge. The blueschist facies metamorphism of Middle Jurassic $^{40}Ar/^{39}Ar$ age (155 Ma), which is an evidence for subduction, was reported by some authors (e.g. Faryad, 1997 and references herein) from the Cretaceous–Palaeogene conglomerates of the Klape Unit in Slovakia. The remnants of Vahicum/Pieniny Ocean are now mostly buried under the overthrust of the Tatric units (Fig. 10). They were described by Plašienka *et al.* (1995) in the Povazsky Inovec (Belice Unit) and in the substratum of Transcarpathian Depression in Eastern Slovakia (Iňačovce–Kričevo Unit). Analysis of the heavy minerals from the Peri-Klippen series (Winkler, Ślączka, 1994) confirm the source of ophiolitic detritus situated between the Pieniny Klippen Belt and Central West Carpathian domain. Alternative location of the Vahicum was suggested by Kozur, Mock (1996). They argue that Vahicum was not a part of the Pieniny Ocean. This interpretation is based upon the assignment of Belice and Iňačovce–Kričevo units to the northernmost part of Veporicum. Thus they place the Vahicum suture in the Central West Carpathians south of the Fatricum.

2.2.3. CENTRAL AND INNER WEST CARPATHIANS

The Central West Carpathians are situated south of the Pieniny Klippen Belt. Because the area of investigation, the Tatra Mts., is a part of the Central West Carpathians more detailed description of this zone will be presented.

According to Plašienka (in Häusler *et al.*, 1993; Plašienka *et al.*, 1997a) the Central West Carpathians extend as far to the south as to the Meliata suture in the Northern Hungary, thus they would comprise also Gemericum with its sedimentary cover. The Inner West Carpathians units, that originated to the south of the Meliata suture, are bounded by the Mid-Hungarian lineament, which is "a dextral fault which juxtaposes Western Carpathian units against the Tisia terrane" (Plašienka *et al.*, 1997c). A different definition of the Inner West Carpathians is found in some other papers (i.e. Mock, 1978; Unrug, 1982; Tomek, 1993; Kozur, Mock, 1996), where northern boundary of the Inner West Carpathians is traced along the Lubenik–Margecany line (Fig. 3). Tectonic units of two types are recognized in the Central and Inner West Carpathians (Biely, 1990). **First type** comprises units with crystalline pre-Mesozoic rocks and their Upper Palaeozoic and Mesozoic sedimentary cover. These are from north to south: Tatricum, Veporicum and Gemericum (Fig. 10). Veporicum is thrusted over the Tatricum along the Čertovica line, while in the south it is buried under Gemericum along the Lubenik–Margecany fault. The **second type** of tectonic units comprises relatively thin near surface nappes consisting mainly of Mesozoic and less frequently Upper Palaeozoic rocks that are thrusted over Tatricum, Veporicum and Gemericum. To this type belong Fatricum, Hronicum, Silicicum and Meliaticum. Fatricum comprises mainly the KriŽna nappe. The Sturec and Choč nappes occur within Hronicum while the StraŽov and Nedzov nappes are characteristic for Silicicum (Biely, 1990). There is still controversy among Slovakian geologists concerning definition of Silicicum. Occurrence of Silicicum in the Central West Carpathians north of the Lubenik–Margecany line was disputed by Plašienka *et al.* 1997b, following the ideas of Mahel (1968). Plašienka *et al.* (*op. cit.*) include the StraŽov and Nedzov nappes also to the Hronicum. Original substratum of Fatricum is most probably buried below the Čertovica overthrust. Situation of the "root" zone of the Hronic-Silicic units is not clear but most authors agree that they could originate to the present south of Veporicum or even south of Gemericum (Fig. 10). A term Meliaticum designates a complex of fine grained carbonate and cherty sediments resting upon the basic and ultrabasic rocks of Ladinian to Lower Carnian age (Kozur, 1991). It is interpreted as a remnant of an oceanic domain. The ocean opened in the Middle Triassic in the beginning of the Mesozoic break-up of the Pangea. The ophiolitic sequences are described from the Northern Hungary (Kozur, 1991; Csaszar *et al.*, 1990). The closure of the Meliata ocean was completed by the end of Jurassic (Fig. 9). Its SE continuation in the Dinaric-Hellenic region (Vardar, Pindos) is sometimes suggested (i.e. Kovacs, 1992; Ziegler *et al.*, 1996). Alternative concept is the prolongation of the Meliata–Hallstatt zone into the Palaeotethys (Cimmerian) Ocean (Kozur, 1991). To the west it is linked with the "Hallstatt Ocean" in the southern margins of the Northern Calcareous Alps, see Fig. 1 (i.e. Channell *et al.*, 1992; Neubauer, 1996). Its presence was supported by Wagreich *et al.* (1995) from the analysis of heavy minerals in the Northern Calcareous Alps.

Chronology of the Cretaceous tectonic events in the Central and Inner West Carpathians was summarized by Jacko and Sasvari (1990), Putiš (1992), Plašienka (1996) and Plašienka *et al.* (1997a, b). Glaucophanized basalts from the base of Meliata unit in the eastern Slovakia yielded the Late Jurassic (160–150 Ma) age of high pressure–low temperature metamorphism which is possibly related to the closure of the Meliata oceanic trough. Thrusting of the Gemerides onto Veporides took place just afterwards in the Early Cretaceous (130–110 Ma). Radiometric Rb–Sr dates between 150 and 120 Ma were obtained from the Gemeric granites and mylonites (Kovach *et al.*, 1986). The Veporic crystalline complexes were subjected to low-grade metamorphism in the Early Cretaceous (Kovač *et al.*, 1994). Thrust of the Veporides over the Tatricum occurred in the early Late Cretaceous (90–80 Ma). Around 90 Ma emplacement of the Fatric and Hronic nappe systems took place. In the Coniacian the extensional Gosau-type basins originated.

Pre-Cainozoic basement of the Central West Carpathians occurs now in the form of tectonic horsts (the Tatra Mts., Nizke Tatry, Mala and Velka Fatra and others) surrounded by Palaeogene (i.e Central Carpathian Palaeogene Basin) and Neogene basins (i.e Vienna Basin, Pannonian Basin and others, see Kovač *et al.*, 1993; Kovač *et al.*, 1997). These features originated due to variegated transtensional and transpressional stress regimes in the Tertiary (Kovač *et al.*, 1993; Kovač *et al.*, 1994; Nemčok *et al.*, 1996). The uplift of the Variscan basement rocks was taking place from the Late Cretaceous–Early Palaeogene (Vepor-Gemeric domain) to Late Neogene (northern Tatricum). In the Neogene extensive volcanism developed in the Central West Carpathians which is interpreted as a result of the collision of the European part of the Eurasian Platform with the Northern Pannonian Plate.

2.3. THE TATRA MTS. AS A PART OF THE CENTRAL WEST CARPATHIANS

The Tatra Mts. is the northernmost occurrence of the "core mountains" which comprise the several Central West Carpathians massifs with Tatric basement which were uplifted in the Neogene. Several structural units are distinguished (Fig. 11): the pre-Mesozoic crystalline basement (Tatricum basement), Mesozoic parautochthonous and over-

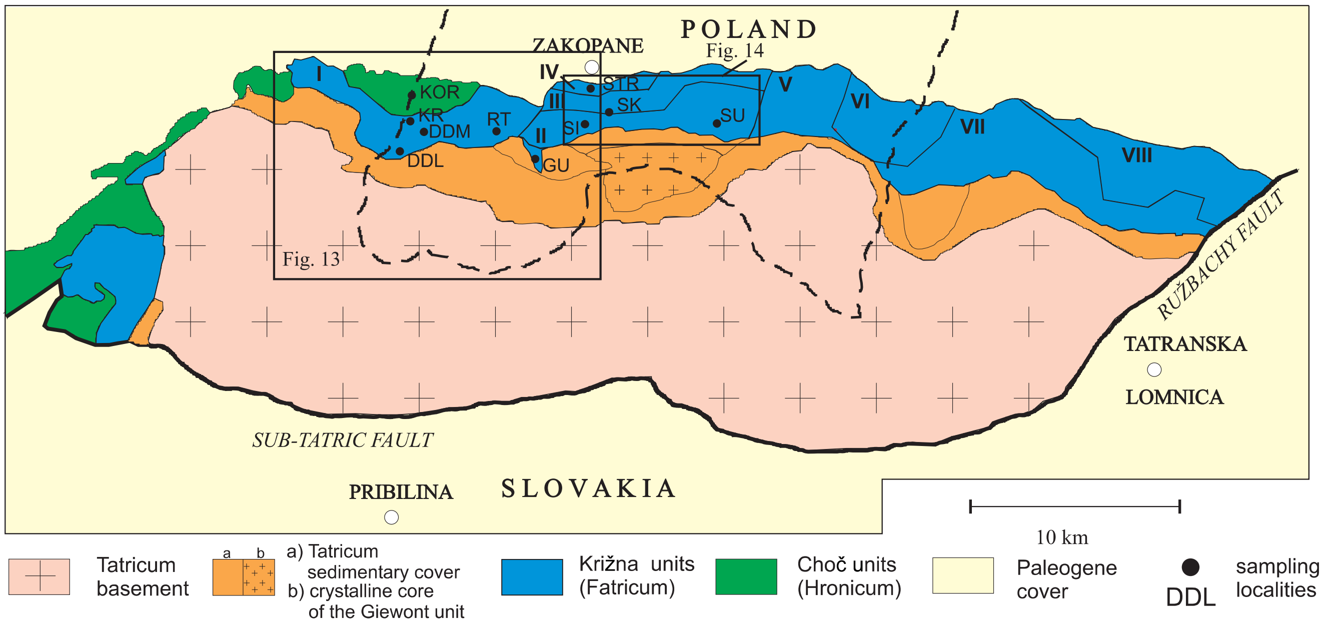


Fig. 11. Tectonic sketch map of the Tatra Mts. with palaeomagnetic sampling localities

Fatric units: I — Bobrowiec, II — Suchy Wierch, III — Mała Świnica, IV — Samkowa Czuba, V — Kopy Sołtysie, VI — Gęsia Szyja and Skałki, VII — Havran, VIII — Bujači; GU — Gładkie Upłaźniańskie slice; rectangles indicate areas pictured in the Figures 13 and 14 thrusted units (Tatricum sedimentary cover + Fatric and Hronic nappes), Palaeogene sedimentary cover (filling of the Central Carpathian Palaeogene Basin).

2.3.1. TATRICUM

The pre-Mesozoic basement of the Tatra Mts. consists of variegated metamorphic rocks and granitoids. Radiometric data indicate the Variscan age of igneous rocks. The age of cooling of the High Tatra granite is estimated between 290 and 347 Ma (Burchart, 1968 — Rb/Sr; Maluski *et al.*, 1993 — Ar/Ar). Rb/Sr age of the Western Tatra pegmatites is 345 ±9.5 Ma according to Gawęda (1995). Metamorphic rocks of the “Goryczkowa crystalline island” (crystalline core of the overthrusted Giewont unit) revealed the Rb/Sr age 420 Ma (Burchart, 1968). Tatra crystalline massif is regarded as an allochthonous body, as inferred from the geophysical evidences (Lefeld, Jankowski, 1985). The very high electric conductivity layer at depths 6–15 km present under the Polish part of the Central Carpathians was interpreted as complex of sedimentary rocks saturated with water, what would imply existence of a deep thrust under the Tatra crystalline massif (Lefeld, Jankowski, *op. cit.*).

The Mesozoic of the Tatricum in the Tatra Mts. bears a local name “High Tatric (wierchowa) unit”. It is divided into parautochthonous unit, that is a roughly *in situ* sedimentary cover of the crystalline rocks, and several detached units (Fig. 11). Among the latter the most important are: Czerwone Wierchy unit, Giewont unit and Široka unit. Parautochthonous unit forms more or less a continuous belt between the western and eastern margin of the Tatra Massif. This unit comprises the sequence of Mesozoic sediments from the Lower Triassic (in one place also the Upper Permian) up to the Middle Cretaceous. They were deposited upon the topographic swell and therefore numerous stratigraphic breaks occur (Kotański, 1961), especially at the Triassic/Jurassic boundary.

The overthrusted High Tatric units occur only locally in the depressions of the crystalline basement (Kotański, 1961). Giewont and Široka units have their own crystalline cores. The complicated tectonic structures of the High Tatric units reveal the northern vergency.

2.3.2. FATRICUM

The Fatricum in the Tatra Mts. is represented by the KriŽna nappe. Local term “lower Sub-Tatric unit” (reglowa dolna) is also used. Synthetic stratigraphical profile of the KriŽna nappe is presented in the Figure 12. It differs from the Tatricum mainly by presence of open marine Jurassic–Cretaceous sequence. Lower Scythian is developed as quartzitic sandstones of fluvial origin. Upper Scythian consists of marine sandstones and shales followed by dolomites and limestones. In the Middle Triassic variegated dolomites prevail, limestones occur subordinately mainly in the Anisian. Similarly as in the High Tatric units Middle Triassic reaches huge thickness of about 1000 m. Carpathian Keuper contains dolomites, sandstones and red shales up to 130 m thick. They are followed by dark Rhaetian limestones of shallow marine origin. Jurassic starts with the Kopieniec Formation (Hettangian–Sinemurian) developed as fine grained clastics with intercalations of marly limestones. It is overlaid by complex of biodetritic and sandy limestones often spotted (Fleckenmergel) of Sinemurian–Pliensbachian age (Sołtysia Marlstone Formation). Spongiolites, crinoidal and nodular limestones occur in the Pliensbachian–Aalenian (Huciska Limestone Formation). Radiolarites form two stratigraphical horizons: the lower of Bajocian/Bathonian (Sokolica Radiolarite Formation) and the upper of Oxfordian age (Czajakowa Radiolarite Formation). They are separated by thin (4–6 m) intercalation of the Bathonian–Callovian nodular limestone that is correlated with the Niedzica Limestone Formation. The uppermost level of nodular limestone belongs to Kimmeridgian/Lower Tithonian (Czorsztyn Limestone Formation). It is followed by pale pelagic limestones of Tithonian-/Berriasian (Pieniny Limestone Formation). The youngest formation in the Western Tatra Mts. is the Kościeliska Marl Formation consisting of marls and marly limestones. Their age is estimated as Upper Berriasian up to Aptian.

KriŽna nappe in the Tatra Mts. is a highly differentiated structure that consists of numerous slices and partial nappes. In the western part of the Tatra Mts. a Bobrowiec unit (Bac, 1971) is the most important (Fig. 13). It is a northward dipping monocline that comprises the complete profile from the Middle Triassic up to Lower Cretaceous. It consists of two tectonic elements. The lower — Głębowiec slice is composed mainly of the Middle Triassic dolomites. The upper — Parządczak slice contains the upper part of succession from Keuper up to Barremian–Aptian. The Bobrowiec unit is thrust from NW over the middle part of the Sub-Tatric zone. This part of the

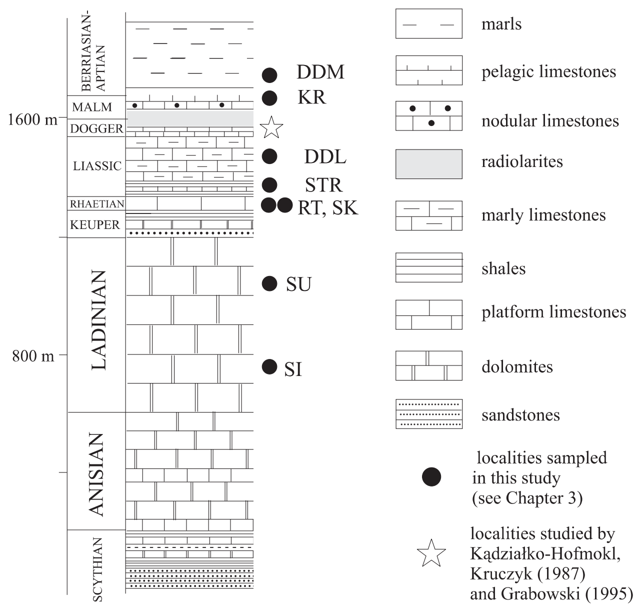


Fig. 12. Synthetic stratigraphic profile of the Križna (Lower Sub-Tatric) nappe, compiled after Kotański (1976) and Lefeld *et al.* (1985)

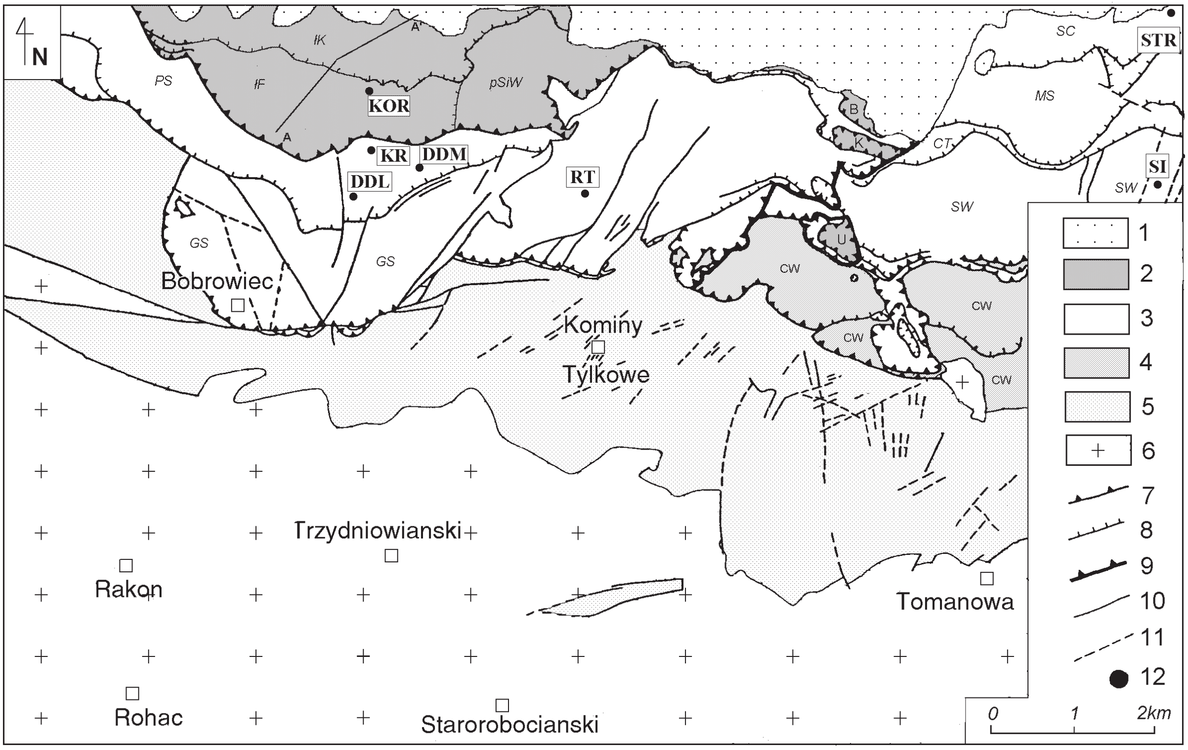


Fig. 13. Tectonic sketch map of the Bobrowiec unit (after Bac, 1971, modified)

1 — **Eocene**; *2* — **elements of the Choč nappe**: pSiW — Siwa Woda unit, łF — Furkaska unit, łK — Koryciska unit, B — Brama Kantaka slice, K — Kończysta slice, U — Upłaz slice; *3* — **elements of the Križna nappe**: Bobrowiec unit: GS — Głębowiec slice, PS — Parządczak slice; Zakopane part of the Sub-Tatric unit: SW — Suchy Wierch unit, CT — Czarna Turnia slice, MS — Mała Świnica unit, SC — Samkowa Czuba unit; *4* — **overthrusted High Tatric units:** CW — Czerwone Wierchy unit; *5* — **sedimentary parautochthonous cover**; *6* — **crystalline rocks**; *7* — main overthrusts; *8* — subordinate overthrusts; *9* — eastern boundary of the Bobrowiec unit; *10* — faults; *11* — faults recorded by fotointerpretation; *12* — palaeomagnetic sampling localities; A–A' — line of the geological cross-section shown in the Figure 15

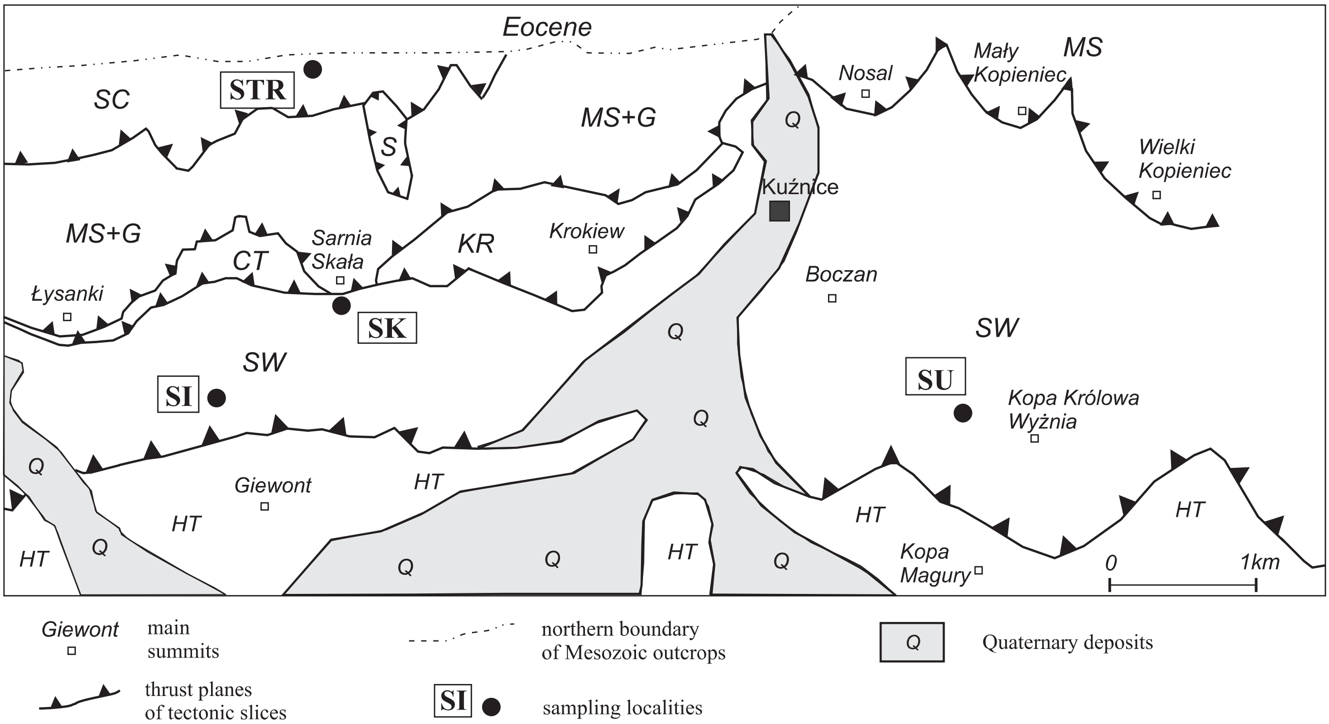


Fig. 14. Tectonic sketch map of the Zakopane part of the Sub-Tatric zone (after Guzik and Kotański, 1963, simplified) with palaeomagnetic sampling localities

Sub-Tatric units: SW — Suchy Wierch unit, CT — Czarna Turnia slice, KR — Krokiew slice, MS + G — Mała Świnica unit and Grześkówki slice, SC — Samkowa Czuba unit, S — Spadowiec unit; HT — High Tatric units

Sub-Tatric zone (in the vicinity of Zakopane) is composed of several small-scale tectonic units (Guzik, Kotański, 1963) which comprise only the Lower Triassic–Lower Jurassic rocks. One of the largest is the Suchy Wierch unit (Fig. 14). It rests directly on the High Tatric substratum. The strata dip monoclinally to the north. The dip is very steep (70–80°) just north of the Mt. Giewont and decreases to the west and east. The second important unit is the Mała Świnica unit which is a tectonic element higher than Suchy Wierch unit. Both Suchy Wierch and Mała Świnica units are spread from west to east over the entire middle part of the Sub-Tatric zone. Other tectonic units distinguished by Guzik and Kotański (1963) (Krokiew, Czarna Turnia, Grzeskówki, Samkowa Czuba and Spadowiec slices) occur only locally (Fig. 14)[1]. In the eastern part of the Sub-Tatric zone again different assemblage of second order tectonic units occurs. Between Sucha Woda and Białka valleys Kopy Sołtysie and Gęsia Szyja units are distinguished[2]. Kopy Sołtysie unit consists mainly of the Lower Jurassic–Cretaceous limestones. The unit is in the overturned position what is unusual in the Sub-Tatric zone. The overlying Gęsia Szyja unit strikes N–S, concordantly with the prominent Białka dislocation (Sokołowski, 1978). In the Belanske Tatry a fairly large Sub-Tatric units (Havran and Bujači units) occur with some subordinate tectonic slices. These units again dip monoclinally to the north (Sokołowski, 1948).

2.3.3. HRONICUM

The higher Sub-Tatric nappes crop out in the western part of the Tatra Mts. and are built mainly of the Middle–Upper Triassic and Lower Jurassic carbonates. Their structural position is still a matter of debate. According to Nemčok ed. (1995) only the Choč nappe (Hronicum) is present in the Tatra Mts. It consists of two sub-nappes: lower — Siwa Woda and upper — Furkaska–Koryciska (Fig. 13). The schematic profile of the Furkaska–Koryciska unit is presented in the cross-section at the Figure 15. It is commonly compared with the Eastern Alpine Triassic. The sequence starts with the Anisian Ramsau dolomites. Above them the nodular Reifling limestones occur (Upper Anisian/Lower Ladinian?). They are overlain by the Partnach Beds which are marls intercalated with limestones. The upper part of the Furkaska–Koryciska unit consists mainly of the Ladinian/Carnian Wetterstein dolomite which exceeds 600 m in thickness. The Siwa Woda unit contains the complementary profile of the Upper Triassic: the Hauptdolomite of Upper Carnian/Norian age and Rhaetian of the Norovica Formation.

Michalik and Gaździcki (1980), and Passendorfer (1984) argue for occurrence of Choč nappe in the Tatra Mts. without higher nappe system. According to Bac *et al*. (1979), after the

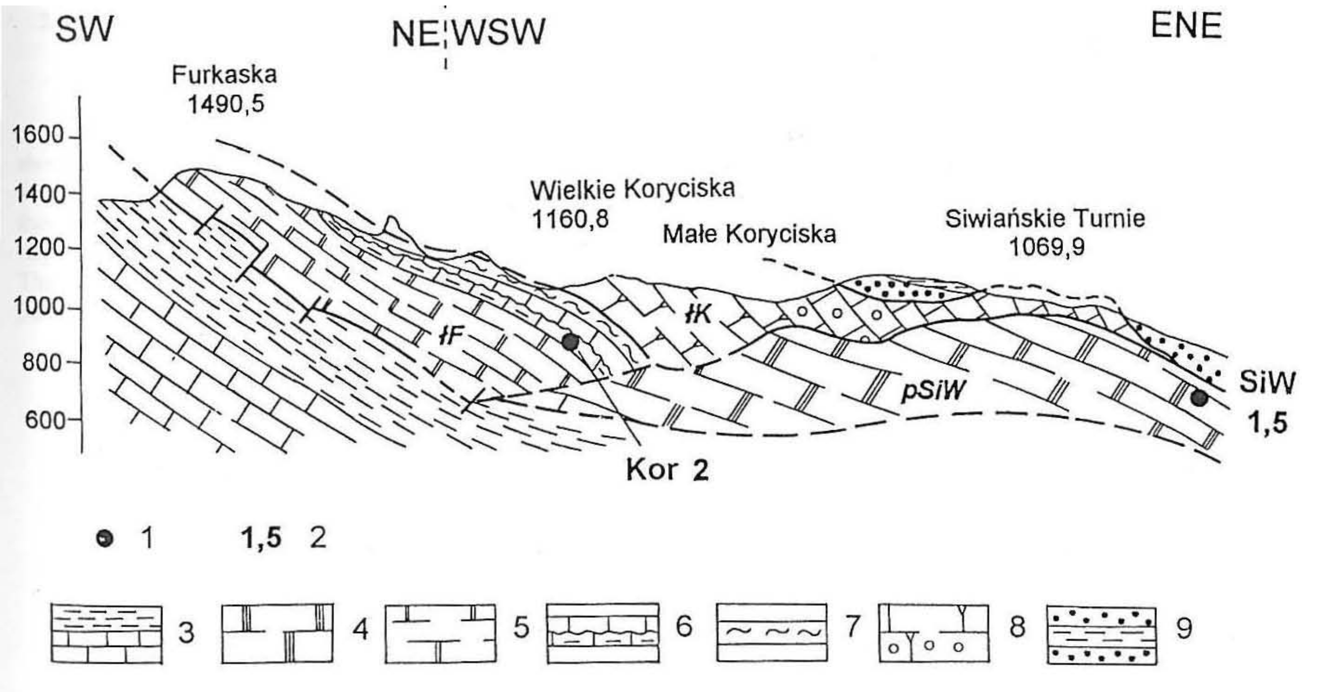


Fig. 15. Geological cross-section of the Furkaska–Koryciska unit (modified after Kotański, Iwanow, 1997)

1 — localities sampled for CAI studies, *2* — CAI indexes (after Grabowski *et al.*, 1999), *3* — Jurassic and Cretaceous of the Križna nappe, *4* — Hauptdolomite and Rhaetian Kössen beds in the Siwa Woda unit (pSiw), *5–7* — Furkaska–Koryciska unit (łF and łK): *5* — Anisian dolomites (Ramsau dolomites), *6* — Upper Anisian Reifling limestones, *7* — Anisian/Ladinian Partnach Beds, *8* — Ladinian/Carnian Wetterstein dolomites, *9* — Eocene

[1] Recently Bac-Moszaszwili (1998) introduced some modifications to the model of Guzik and Kotański (1963), changing the boundaries and names of some tectonic units in the Zakopane part of the Sub-Tatric unit. However, the modifications do not concern the tectonic units sampled in this study.

[2] Tectonic structure of the eastern part of the Sub-Tatric unit has been revised by Lefeld (1999).

concept of Kotański (1973), only Siwa Woda unit belongs to the Choč nappe and the Furkaska–Koryciska unit belongs to the StraŽov nappe being a remnant of the Silicicum. In this study both Siwa Woda and Furkaska–Koryciska units are assigned to the Choč nappe, after the most recent paper of Bac-Moszaszwili (1998).

Few isolated tectonic slices (Brama Kantaka, Kończysta and Upłaz, see Fig. 13) built predominantly of Liassic limestones (Uchman, 1993) occur to the east of the Furkaska–Koryciska and Siwa Woda units. According to Kotański (1976), the Upłaz scale belong to the Veporic unit as can be inferred from the transgression of Liassic onto the Anisian. Brama Kantaka and Kończysta slices are interpreted as remnants of the Choč nappe.

2.3.4. PALAEOGENE COVER

From the north and south the Tatra Mts. are surrounded by the Central Carpathian Palaeogene Basin (CCPB) (Fig. 16). Geological structure of the CCPB was recently reviewed by Nemčok *et al.*, (1996). Sedimentation started in the Middle Eocene with marine transgressive deposits: conglomerates and nummulitic limestones that crop out on the northern margin of the Tatra Massif. Higher members of the Central Carpathian Palaeogene are variegated terrigenous (flysch) rocks that fill the Podhale and Liptov depressions. The maximum thickness of Podhale Palaeogene is 2200 m but in the Levoča Mts. thicknesses up to 4000 m are noted (Nemčok *et al.*, 1996). In the Neogene, between 10 and 20 Ma according to the fission track data of Burchart (1972), the uplift of the Tatra Massif from under the Palaeogene cover took place. The present day Tatra Massif reveals an asymmetric structure. From the south it is cut by the distinct Sub-Tatric fault (Fig. 11). Crystalline rocks contact here directly with the Palaeogene of the Liptov depression. Mesozoic tectonic units are preserved on the northern slopes of the Tatra Massif. They plunge under the Palaeogene of the Podhale basin. The presence of the fault between the Nummulitic Eocene and Eocene flysch rocks on the northern Tatra margin is very likely (Bac-Moszaszwili, 1995). The RuŽbachy fault constitutes the eastern termination of the Tatra Massif. In

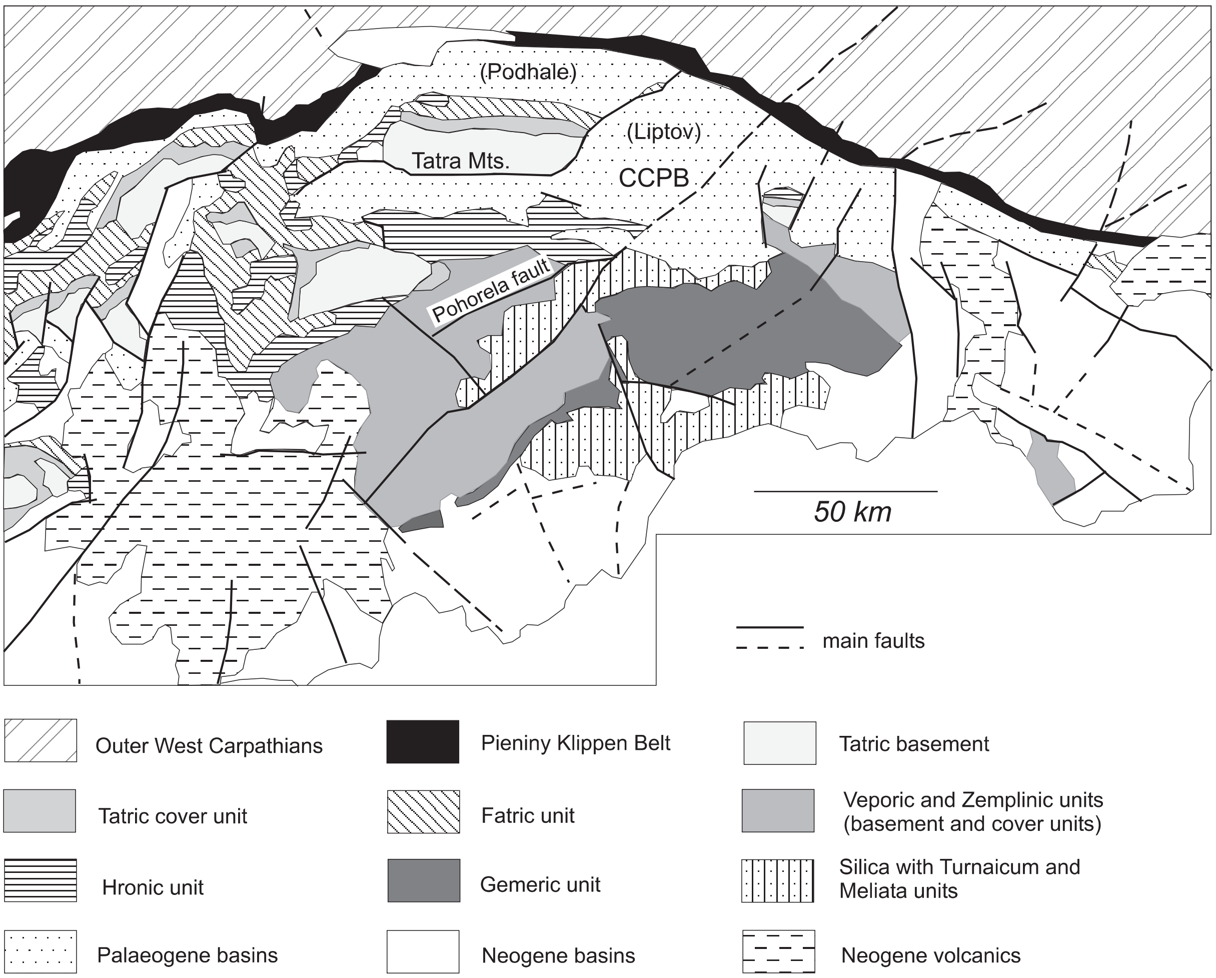


Fig. 16. Tectonic sketch map of the Central and Inner West Carpathians of Poland and Slovakia

the west the crystalline Tatricum is hidden under the nappe structures that are the prolongation of the Choč Hills.

According to Piotrowski (1978) the crystalline basement together with Mesozoic and Palaeogene were tilted by about 20° to the north as the result of the Neogene uplift. This model was accepted by Bac-Moszaszwili *et al*. (1984) and Sperner (1996). However according to Bac-Moszaszwili (1995) the Neogene tilt could be diversified in particular parts of the Tatra Massif. Also Wieczorek and Barbacki (1997) pointed out that Mesozoic units situated under the Podhale Flysch and in the central part of the Tatra Massif preserved the geometry of tectonic structures since the Late Cretaceous without Neogene disturbations. In this model the effects of Neogene tectonics would be the most conspicuous at the northern margin of the Tatra Massif as postulated already by Guzik and Kotański (1963). The uncertainties in estimating the effects of Neogene movements for the Tatra Mts. are a serious drawback in palaeoreconstructions based on palaeomagnetism.

2.4. THERMAL HISTORY OF THE TATRA MTS. IN THE MESOZOIC AND TERTIARY

Thermal history of the Mesozoic units of the Tatra Mts. has not been investigated in detail. Recently obtained values of the CAI indexes for the highest Sub-Tatric units (i.e. Siwa Woda and Furkaska–Koryciska units) are quite low: 1.5–2 (Grabowski *et al*., 1999). This corresponds to temperature range 50–80°C during 10 Ma. Few palaeotemperature estimations in Mesozoic and Tertiary exist also for the crystalline substratum of the Tatra Mts. The maximum temperature of crystalline rocks could not exceed 300–350°C (Janak, 1994) because $^{40}Ar/^{39}Ar$ dating indicates Variscan cooling ages. According to Kovač *et al*. (1994) northern part of the Tatricum was buried in the post-Permian times to a maximum depth of 12 km what corresponds to the temperature ca. 250°C. Also Lefeld (1997) accepts that palaeotemperature at the top of crystalline core during Late Cretaceous thrusting was not higher than 200°C thus the Cretaceous orogenesis was rather "cold". Results of the fission-track apatite studies reveal that temperature of the crystalline core fell down below 100–120°C during the Miocene uplift of the Tatra Mts. 10–15 Ma ago (Burchart, 1972; Kovač *et al*., 1994).

2.5. MESOZOIC AND TERTIARY APPARENT POLAR WANDER PATHS FOR EUROPE AND AFRICA AND EXPECTED PALAEOMAGNETIC DIRECTIONS FOR THE TATRA MTS.

"Expected" palaeomagnetic directions for the Tatra Mts. should situate the area in the western part of the Tethys between the European and African plates. Palaeomagnetic data from these two major plates should serve as a framework for the interpretation of palaeomagnetic directions from the Tatra Mts. Therefore it is important to evaluate critically the reference palaeomagnetic results from Europe and Africa.

The most extensive data set of palaeomagnetic poles for African and European plates as well as Tethyan terranes was recently published by Besse and Courtillot (1991), and Van der Voo (1993) (Tabs. 1 and 2). New results concerning European apparent polar wander path (APWP) for Triassic–Early Jurassic were supplied by Edel and Duringer (1997) (Tab. 3). Besse and Courtillot (1991), and Van der Voo (1993) selected the

Table 1

Reference palaeomagnetic directions for the Tatra Mts. (20° E, 49° N) calculated from the European and African palaeopoles, after Van der Voo (1993)

Age in Ma	European coordinates				African coordinates			
	D	I	PLAT	α_{95}	D	I	PLAT	α_{95}
37–66 (Palaeogene)	5.9	57.2	37.8	4	3	59.4	40.3	11
67–97 (Late Cretaceous)	15.8	54.7	35.2	6	341.2	50.1	30.9	6
98–144 (Early Cretaceous)	2.7	48.1	29.1	15	325.9	45.9	27.3	7
145–176 (Late Jurassic)	4.0	43.3	25.2	15	332.4	41.8	24.1	13
177–195 (Early Jurassic)	25.5	59.5	40.3	10	343.3	55.3	35.8	5
215–196 (Late Triassic/ Early Jurassic)	–	–	–	–	354.8	49.6	30.4	–
232–216 (Late Triassic)	39.1	44.2	25.9	14	330.8	41.5	23.9	–
245–233 (Early Triassic)	30.0	35.4	19.6	4	–	–	–	–

D — declination, I — inclination, PLAT — palaeolatitude, α_{95} — Fisher statistics parameter

Table 2

Reference palaeomagnetic directions for the Tatra Mts. (20° E, 49° N) calculated from the European and African palaeopoles, after Besse and Courtillot (1991)

Age in Ma	European coordinates				African coordinates			
	D	I	PLAT	α_{95}	D	I	PLAT	α_{95}
0–10 (Miocene/Recent)	7	62.5	43.9	3	5.5	62.9	44.3	3
10–30 (Oligocene/Miocene)	11.1	59.1	39.9	8	4.6	59.8	40.6	5
30–50 (Eocene/Oligocene)	14	61	42.1	7	–	–	–	–
40–60 (Palaeocene/Eocene)	12.8	56.6	37.2	4	2.1	51.4	32.1	5
50–70 (Palaeocene/Eocene)	–	–	–	–	353.5	47.5	28.6	8
60–80 (Late Cretaceous/Palaeogene)	7.8	55.7	36.2	10	349.5	43.5	25.4	6
70–90 (Late Cretaceous)	–	–	–	–	345.6	47.8	28.8	5
80–100 (Late Cretaceous)	5.6	55.1	35.6	4	342.8	49.2	30.1	4
100–120 (Early Cretaceous)	357	49.3	30.1	9	321.5	43.6	25.5	12
140–160 (Late Jurassic)	17	55.4	36	7	336	51	31.7	12
160–180 (Middle Jurassic)	34.8	57.7	38.4	6	–	–	–	–
180–200 (Early Jurassic)	25.4	62.8	44.2	10	342.2	50.8	31.5	8

Explanations as in the Table 1

most reliable results from major lithospheric plates. Then they used the kinematics for the Atlantic bordering plates given by various authors (i.e. Bullard *et al*., 1965; Savostin *et al*., 1986; Rowley, Lottes, 1988 and others) to transfer the data to a single continent reference frame to obtain a "master APWP" for each continent. The sense of this operation was to combine palaeomagnetic data from various plates and to use the data also for the plates where palaeomagnetic results are not numerous. Also the coherence of "master APWP" would be the test if palaeomagnetic data are in agreement with kinematic parameters. Unfortunately Besse and Courtillot (1991), and Van der Voo (1993) utilised the kinematics elaborated by different authors and their "master APWPs" are not directly comparable. Therefore for the purpose of that dissertation I used their "pure" African and European data in their own coordinates. This approach would eliminate the uncertainties of palaeoreconstructions which arises from different kinematic models. An exception was made for the African APWP for Jurasssic and Cretaceous processed by Channell (1996). He found only one pole from Africa as reliable. All other entries used for his African APWP are North American poles rotated to the African coordinates (Tab. 4). They are used here for comparison with the "pure" African data of Besse and Courtillot (1991), and Van der Voo (1993). As Apulian/Adriatic microplate was involved in the Alpine collision between Europe and Africa the data from that microplate were also considered as possible reference directions for the Tatra Mts. (Tab. 5).

Table 3

Reference Middle Triassic/Early Jurassic palaeomagnetic directions for the Tatra Mts. (20° E, 49° N) calculated from Western European palaeopoles, after Edel and Duringer (1997)

Age in Ma	D	I	PLAT	α_{95}
187 (Toarcian/Aalenian)	23.9	62.9	44.3	4
195 (Pliensbachian)	44.6	68.9	52.3	8
200 (Sinemurian)	50.7	61.9	43.1	11
205 (Hettangian)	51.5	54.3	34.8	7
225 (Carnian/Norian)	41.6	43.4	25.3	5
235 (Anisian/Ladinian)	33.4	36.7	20.5	5

Explanations as in the Table 1

Table 4

Reference Jurassic/Cretaceous palaeomagnetic directions for the Tatra Mts. calculated from the "African" palaeopoles, after Channell (1996)

Age in Ma	D	I	PLAT	α_{95}
80 (Campanian)	348.8	55.8	36.4	10
90 (Turonian)	341.4	44.5	26.2	4
120 (Barremian)	323.9	39.4	22.3	–
143 (Tithonian/Berriasian)	309.3	24.6	12.9	4
145 (Tithonian/Berriasian)	302.8	31.3	16.9	9
166 (Callovian)	333.5	57.4	38	–
175 (Bajocian/Bathonian)	340.2	54.7	35.3	3
175 (Bajocian/Bathonian)	327.8	44.2	25.9	6

Explanations as in the Table 1

Table 5

Reference paleomagnetic directions for the Tatra Mts. calculated from the Adrian paleopoles, after Channell (1996), Channell *et al.* (1992), Channell and Doglioni (1994). The data from Southern Alps were utilized, except the Pliensbachian direction which is calculated from the NW Umbrian pole rotated, after Channell (1996)

Age in Ma	D	I	PLAT	α_{95}
76 (Senonian)	350.3	40.5	23.1	4
124 (Hauterivian/Barremian)	326.2	39	22	5
152 (Kimmeridgian/Tithonian)	314.4	26.7	14.1	5
163 (Callovian/Oxfordian)	313.5	37.9	21.3	8
196 (Pliensbachian)	333.7	52.8	33.3	4.5
242 (Scythian)	336	10.9	5.5	7

Explanations as in the Table 1

Tables 1–5 present the expected declinations, inclinations and palaeolatitudes calculated for the Tatra Mts. as if they were a part of the European, African or Apulian/Adrian plates according to mean palaeopoles. The same data are plotted in the Figs. 17–19 as the reference curves of palaeodeclination, palaeoinclination and palaeolatitude in the function of time, calculated from palaeopoles for the present day position of the Tatra Mts.

Mesozoic and Tertiary palaeoinclinations from the Tatra Mts. should indicate the position of the area as intermediate between the African and European plates. Thus they should not be significantly higher than the "European" inclinations and lower than the "African". In the case that the observed palaeoinclinations did not fit this framework another options should be considered. This refers to the inclination shallowing which might be encountered in sedimentary rocks. It could originate as the effect of post-depositional compaction or gravitational forces acting during deposition (Garces *et al.*, 1996). Depositional conditions should also be taken into account. When deposition takes place on a topographic slope correction of primary magnetization vector to the horizontal position introduces erroneous interpretation of inclination parameter (Kruczyk, Kądziałko-Hofmokl, 1988b). Also intensive tectonic deformations might disturb the alignment of ferromagnetic particles in the host rocks (Kligfield *et al.*, 1983). Special attention should be paid for a possible synfolding magnetization. Investigated Mesozoic sedimentary rocks in the Tatra Mts. occur in the thrust sheets. Tectonic movements which could affect their geometry ceased in the area of the Tatra Mts. as late as in the late Miocene. The intermediate stages of folding and thrusting are usually poorly known. Therefore a synfolding magnetization could reveal anomalously shallow or steep inclinations what should be born in mind by interpretation of the characteristic directions.

The framework of expected palaeodeclinations can not be established for the Tatra Mts. unambiguously. Local rotations around vertical as well as inclined axes should be expected. This is the consequence of tectonic structure of the Mesozoic of the Tatra Mts. with numerous independent nappes and tectonic slices. The amount of rotations could be very large. Recent study of Houša *et al.* (1996) indicates that the local rotation of Mesozoic sedimentary sequences near Žilina (western part of the Pieniny Klippen Belt in Slovakia) was almost 130°. These local rotations most likely occurred in the Late Cretaceous dur-

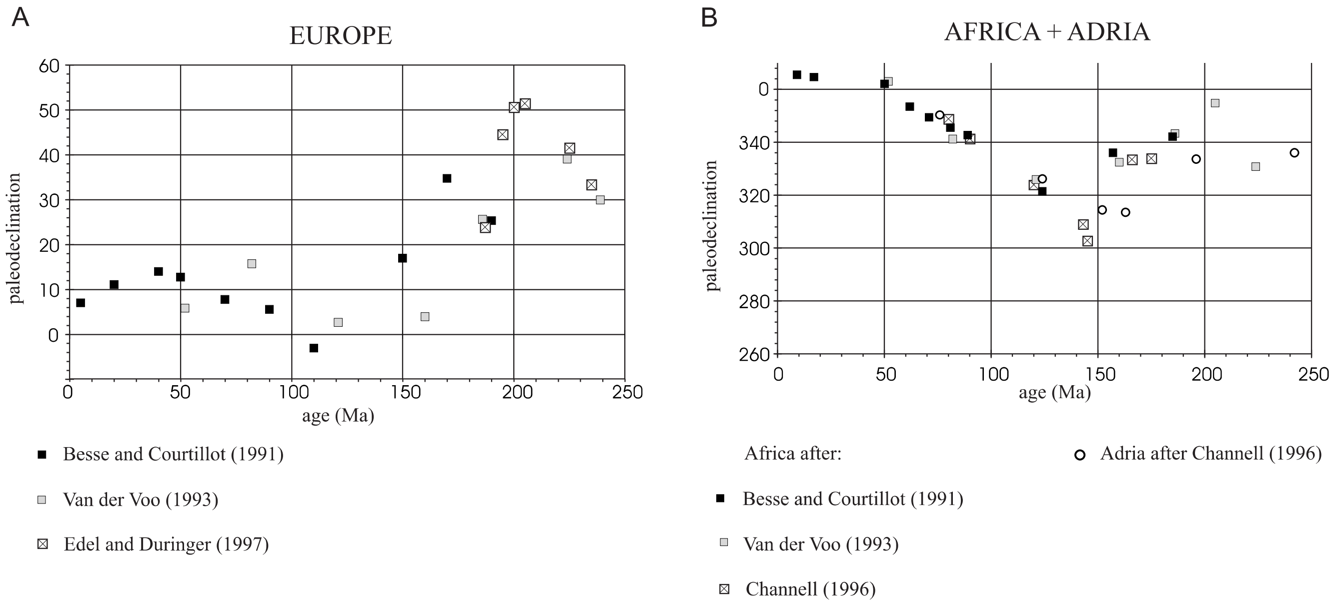


Fig. 17. Expected paleomagnetic declinations for European and African + Adriatic plates since 250 Ma, calculated for geographical position of the Tatra Mts. (20° E, 49° N)

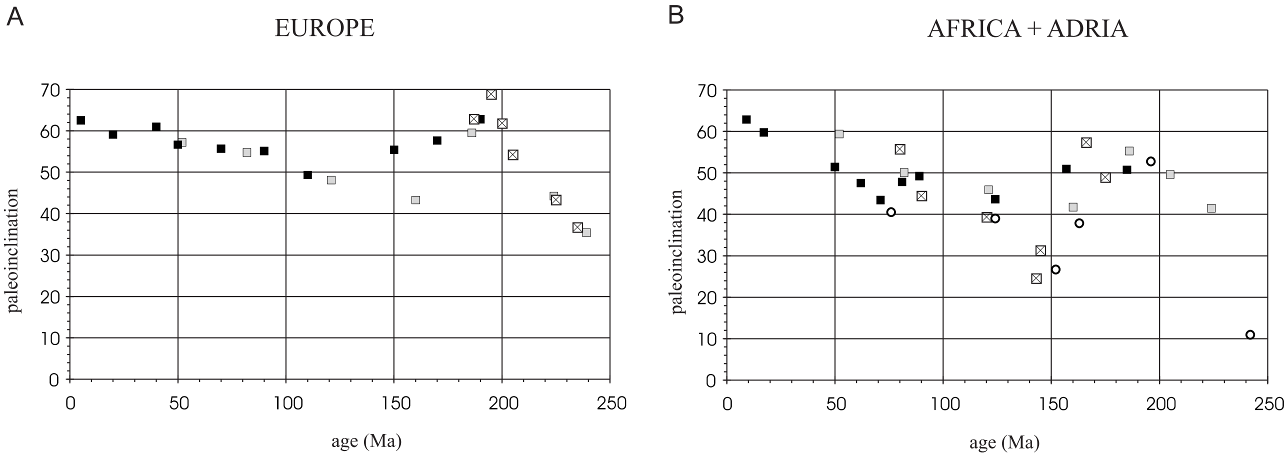


Fig. 18. Expected paleomagnetic inclinations for European and African + Adriatic plates since 250 Ma, calculated for geographical position of the Tatra Mts. (20° E, 49° N)

Explanations of squares as in Figure 17

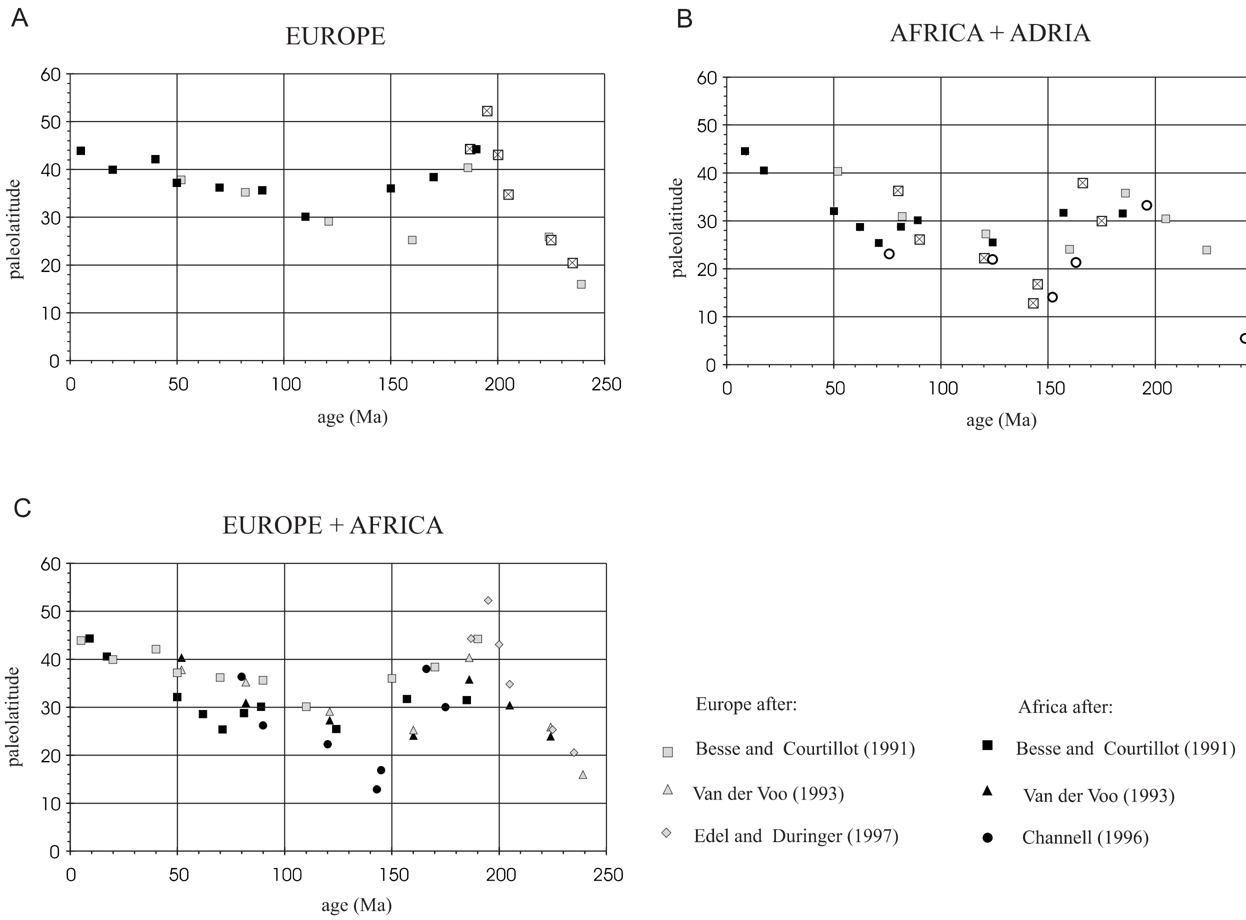


Fig. 19. Expected paleolatitudes for European and African + Adriatic plates since 250 Ma, calculated for geographical position of the Tatra Mts. (20° E, 49° N)

Explanations of A and B as in Figure 17

ing the nappe thrusting. Post-Eocene/pre-Neogene rotations of particular nappes and tectonic slices are not unlikely (Bac-Moszaszwili, 1995) especially at the northern margin of the Tatra Massif where backthrusting of the Eocene and Sub-Tatric units were observed (Piotrowski, 1978; Bac-Moszaszwili, 1993). There is no proof for tectonic rotations which postdate the Neogene uplift.

3. SAMPLING AND LABORATORY METHODS

3.1. SAMPLING

Field work was carried out in the summer seasons 1996 and 1997. 55 independently oriented hand samples were collected in the Western Tatra Mts. from the KriŽna nappe (8 localities) and Furkaska–Koryciska unit (1 locality). The new palaeomagnetic results presented in this study were interpreted together with data already published from the High Tatric units and Nummulitic Eocene (Grabowski, 1997a). The final interpretation of palaeomagnetic data from the Tatra Mts. led in the chapter 4 was based on almost 90 hand samples collected by the author plus results published by Kądziałko-Hofmokl, Kruczyk (1987) and Kruczyk *et al*. (1992).

The oldest sediments sampled in the KriŽna nappe were Ladinian dolomites that belong to the Suchy Wierch unit. They were collected at Skupniów Upłaz ridge (SU) and Siklawica waterfall (SI) in the Dolina Strążyska (Tab. 6; Figs. 13 and 14). It was possible to perform the fold test between these two localities because the tectonic position of beds in SU and SI is different (see Tab. 6). Most localities within the KriŽna nappe were situated in the Bobrowiec unit (Tab. 6; Fig. 13) where sedimentary sequence was sampled from the top of Keuper up to the Lower Cretaceous. Entire profile of the Rhaetian together with the overlying Hettangian and underlying Keuper strata was investigated in the Dolina Lejowa at the Wierch Spalenisko (RT). Lower Jurassic–Lower Cretaceous limestones were sampled in the western slopes of the Dolina Chochołowska in the Dolina Długa and Kryta (DDL, KR and DDM). Additional localities of the Rhaetian–Hettangian limestones were studied in the vicinity of Sarnia Skała (SK) and at the mouth of the Dolina Strążyska (STR) (Tab. 6; Figs. 13 and 14). Locality STR is situated at the top of the Samkowa Czuba unit. Locality SK belongs to the unnamed tectonic slice which is incorporated in the so called Czerwona Przełęcz zone being the upper part of the Suchy Wierch unit (Guzik, Kotański, 1963). Stratigraphical succession of localities sampled in the KriŽna nappe is presented in the Figure 12.

The locality belonging to the Furkaska–Koryciska unit (Choč nappe) was sampled in the Wielkie Koryciska (KOR). It comprises the Anisian/Ladinian “Reifling type” limestones. (Tab. 6; Figs. 13 and 15).

Detailed description of each sampling locality with results of petrological observations is enclosed in the chapter 4.

Table 6

Description of sampling localities

Locality	Code	Rock type	Age	Tectonic unit	Dip azimuth(°)	Dip (°)	N
Skupniów Upłaz	SU	platy dolomites	Upper Ladinian	Suchy Wierch unit	20–50	34–44	9
Siklawica	SI	platy dolomites	Ladinian	Suchy Wierch unit	350–15	52–70	5
Wierch Spalenisko	RT	dolomites, dark limestones	Keuper–Hettangian	Bobrowiec unit	30–46	30–45	9
Dolina Długa (Liassic)	DDL	dark micritic limestones	Sinemurian–Pliensbachian	Bobrowiec unit	356–25	23–68	6
Dolina Kryta	KR	marly limestones	Tithonian	Bobrowiec unit	10–20	38–47	6
Dolina Długa (Malm/Neocomian)	DDM	light micritic limestones	Tithonian–Berriasian	Bobrowiec unit	0–50	30–65	8
Sarnia Skała	SK	dark limestones	Rhaetian	*	0	66	3
Dolina Strążyska	STR	dark limestones	Hettangian	Samkowa Czuba unit	14–18	106–120	4
Wielkie Koryciska	KOR	dark nodular limestones	Anisian–Ladinian	Furkaska–Koryciska unit	345–20	40–45	5

N — number of hand samples taken from the locality

* unnamed tectonic slice between the Suchy Wierch and Mała Świnica units

3.2. LABORATORY METHODS

Cylindrical specimens of 25 mm diameter and 22 mm height were drilled from the hand samples. Usually 3–7 specimens were obtained from each hand sample. Natural remanent magnetization (NRM) was measured by means of JR-5 spinner magnetometer (AGICO, Brno; noise level 10^{-5} A/m) in the palaeomagnetic laboratory of Polish Geological Institute (PGI) in Warsaw, cryogenic SQUID magnetometers (2G Enterprises, USA; noise level 10^{-6} A/m)) in the palaeomagnetic laboratories of Institute of Geophysics (Polish Academy of Sciences — PAS, Warsaw) and ETH in Zurich. Alternating field (AF) demagnetization was performed with the instrument made by 2G Enterprises, attached to the cryogenic magnetometer in the laboratories of PAS and ETH (max. demagnetization field 160 mT). Thermal demagnetization was carried out in the PGI and PAS laboratories using the non-magnetic oven MMTD (Magnetic Measurements, UK; maximum demagnetization temperature 700°C) and in the ETH laboratory with Schonsted oven. NRM measurements and demagnetization experiments were carried in the magnetically shielded space. In the PGI and PAS laboratories it was a low-field cage (Magnetic Measurements, UK) reducing the ambient geomagnetic field by about 95%. In the ETH laboratory the procedure took place in the steel room which almost completely cancels the influence of geomagnetic field. Magnetic susceptibility was monitored with KLY-2 bridge (AGICO, Brno; sensitivity 10^{-8} SI units) after each thermal demagnetization step. The same instrument was used for measurements of anisotropy of magnetic susceptibility (AMS). AMS parameters were determined using the ANISO software supplied by AGICO (Jelinek, 1977). Characteristic remanence magnetization (ChRM) directions were calculated basing on the principal component analysis (Kirschvink, 1980) using the PALMAG package of Lewandowski *et al*. (1997). Palaeopoles and Apparent Polar Wander Paths were plotted and constructed with the GMAP for Windows software of Torsvik and Smethurst (1994). Fold tests were performed using the method of McFadden (1990).

Several methods were used for studying rock magnetic properties. Investigation of the isothermal remanent magnetization (IRM) (stepwise acquisition and 3-axes thermal demagnetization — Lowrie, 1990) were performed at PGI using the MMPM1 pulse magnetizer produced by Magnetic Measurements (UK) (maximum field applied 3T). Thermomagnetic analyses were performed in palaeomagnetic laboratory “Fort Hoofdijk” in Utrecht using the horizontal Curie balance (Mullender *et al*., 1993). Hysteresis behaviour was studied with an alternating gradient force magnetometer (AGFM) produced by Princeton Measurements Corporation (USA) also in Utrecht. The frequency dependence and low temperature susceptibility measurements were carried on in School of Environmental Sciences (University of East Anglia — Norwich), using Bartington susceptibility bridge. Additional petrographic observations were carried out in reflected and transmitted light microscopy. Scanning electron microscope (SEM) investigations were performed using JEOL JSM 35 device equipped with Link Isis Energy Dispersive Spectrometer (EDS). All the SEM samples were carbon coated for EDS analysis.

4. EXPERIMENTAL RESULTS

4.1. SKUPNIÓW UPŁAZ AND SIKLAWICA

4.1.1. LOCALITY DESCRIPTIONS

Nine hand samples of the platy dolomites (Upper Ladinian according to Kotański, 1963) were taken from the Skupniów Upaz (SU) ridge. Entire profile about 70 m thick was sampled there above the upper part of Długi Żleb in the Jaworzynka Valley. The outcrops are of very good quality because the slopes of the Jaworzynka Valley are almost devoid of trees and bushes. The strata dip 34–44° to the NE (Tab. 6). Each hand sample was collected from separate dolomite bed.

Five hand samples of the platy dolomites (Lower Ladinian according to Kotański, 1971) were taken from the southern end of the Strążyska Valley in the vicinity of the Siklawica waterfall (SI). According to Bac-Moszaszwili (1995) the Siklawica cascade could be developed along the fault of post-Eocene age. One sample (SI2) was taken directly from the waterfall quest, four others from the little outcrop on the western slope of the valley. The fresh rocks are dark, almost black. They disintegrate very easily due to numerous joints. The strata dip steeply (70°) to the N (Tab. 6).

4.1.2. PETROGRAPHY AND ROCK MAGNETISM

Thin sections were prepared from samples SU3, SU4, SI2 and SI5. The rocks are dolomicrites, mostly without bioclasts, in sample SI2 cyanobacterial structures are observed. Non-transparent minerals consist mostly of very fine grains of unidentified iron oxides or hydroxides which are of post-pyrite origin. Unaltered pyrite is rare. Opaque minerals are quite uniformly distributed between the dolomite grains and do not occur within fine dolomitic veins. Ferrugineous substance is present in the stylolitic seams which are apparently older than veins.

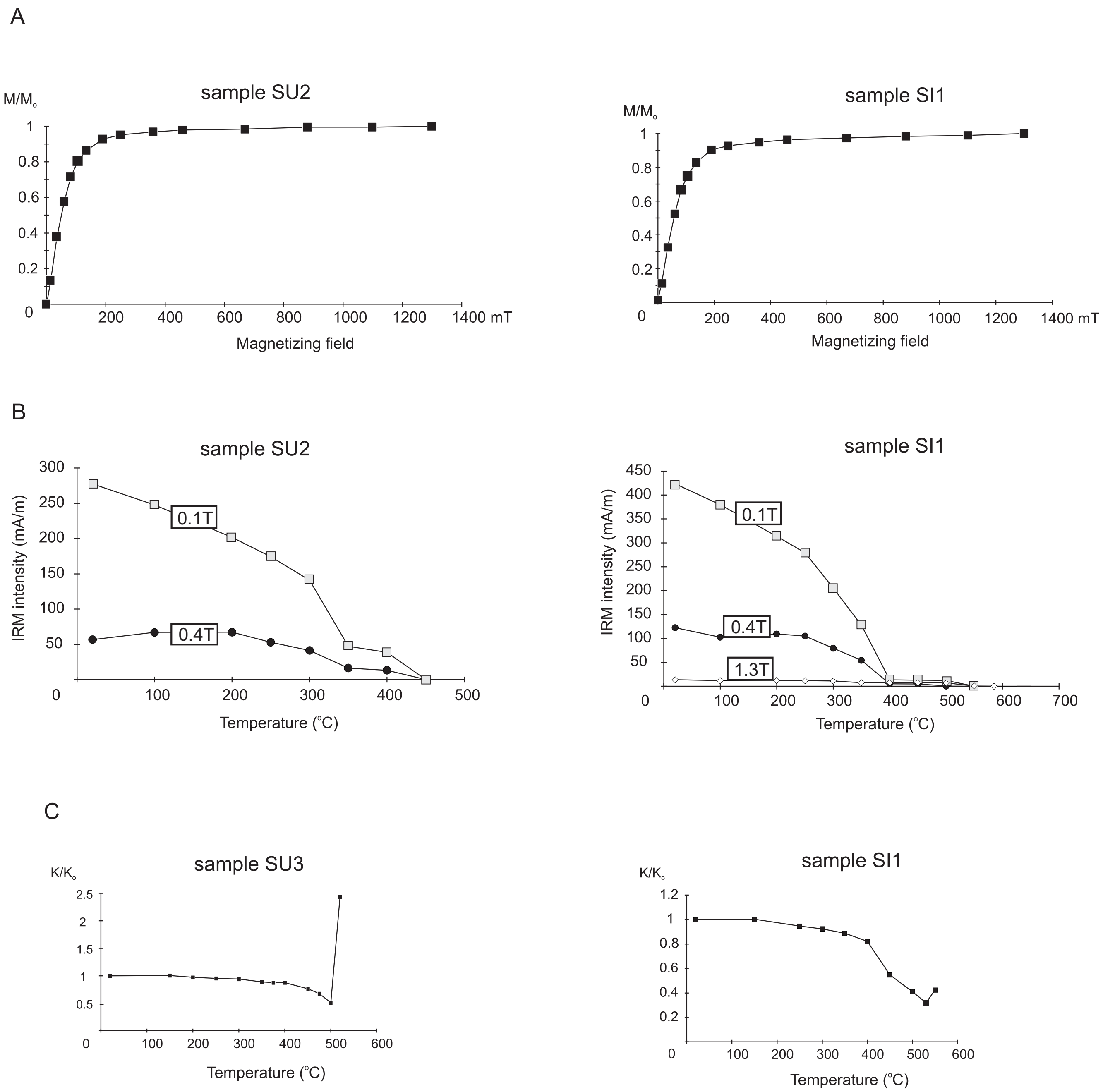


Fig. 20. Rock magnetic properties of the Middle Triassic dolomites from the localities SU and SI

A — IRM acquisition curve; **B** — thermal demagnetization of the 3-axes IRM acquired in the fields 0.1T, 0.4T and 1.3T; **C** — susceptibility changes during thermal treatment

IRM aquisition experiments were carried on for two specimens of samples SU1 and SU2 and for two specimens of the sample SI1. They revealed, that low coercivity minerals predominate in both localities. The samples are almost saturated in the fields as low as 250 mT (Fig. 20A). High coercivity phase (i.e. goethite and hematite) is apparently absent. Thermal demagnetization of the 3-axes IRM (Fig. 20B) reveals the presence of 2 magnetic phases: first with maximum unblocking temperature 300–350°C, the second with maximum unblocking temperature 450–500°C.

One thermomagnetic analysis was performed for the sample SU4 (Fig. 21A). Curie temperature slightly exceeds 600°C. As hematite is absent in this locality, the temperature should be regarded as characteristic for oxidized magnetite (maghemite) (Mullender *et al.*, 1993). The presence of maghemite is additionally supported by decrease of magnetic susceptibility after 350°C (Fig. 20C). Above this temperature maghemite transforms to hematite (Opdyke, Channell, 1996). There is no indications of iron sulphides on the thermomagnetic curve: either pyrrhotite (Curie temperature 300–350°C) or pyrite (production

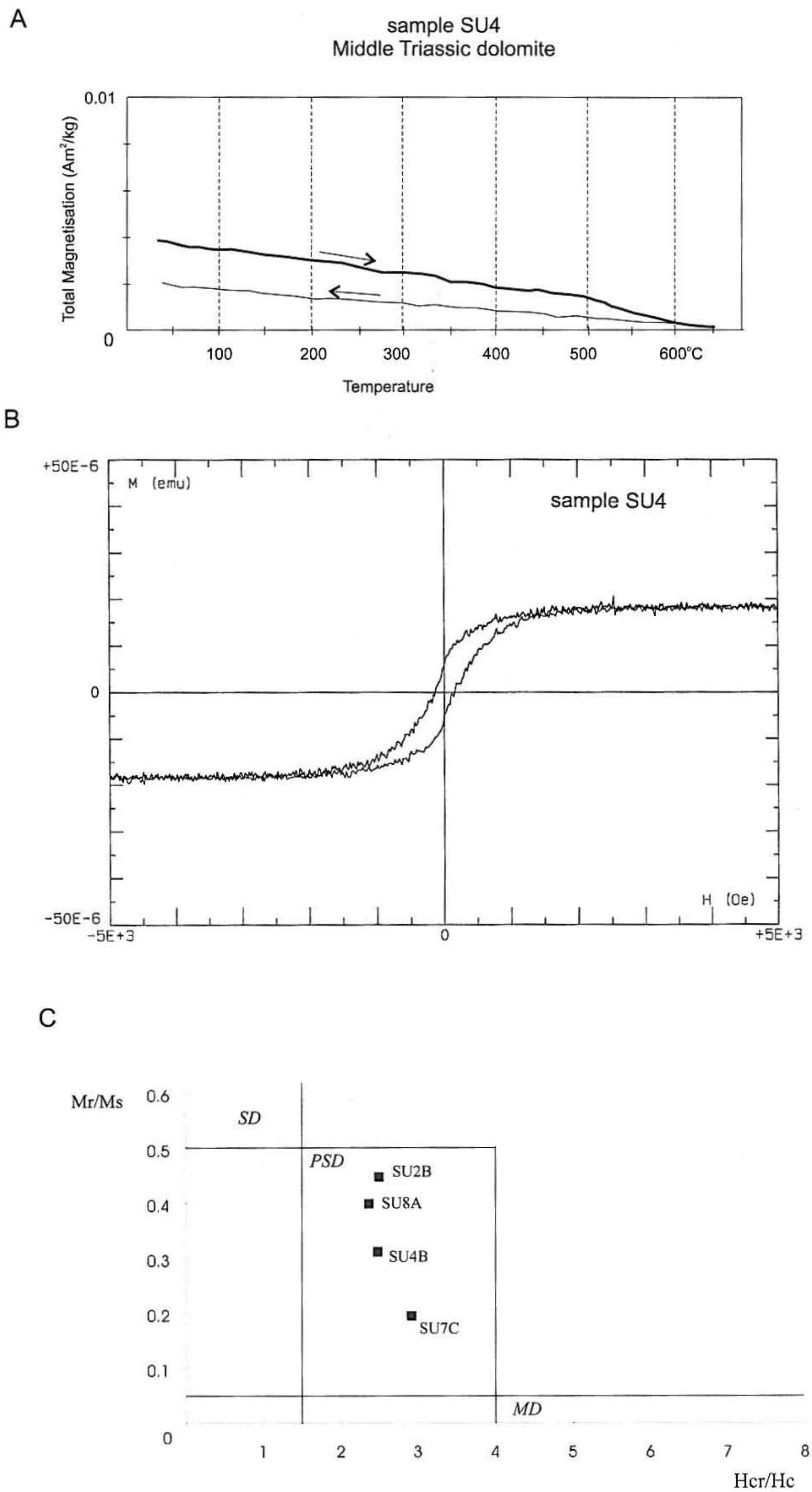


Fig. 21. Rock magnetic properties at the locality SU (continuation)

A — thermomagnetic analysis; bold line — heating curve, thin line — cooling curve; **B** — hysteresis loop; **C** — hysteresis ratios plotted on a Day *et al.* (1977) diagram

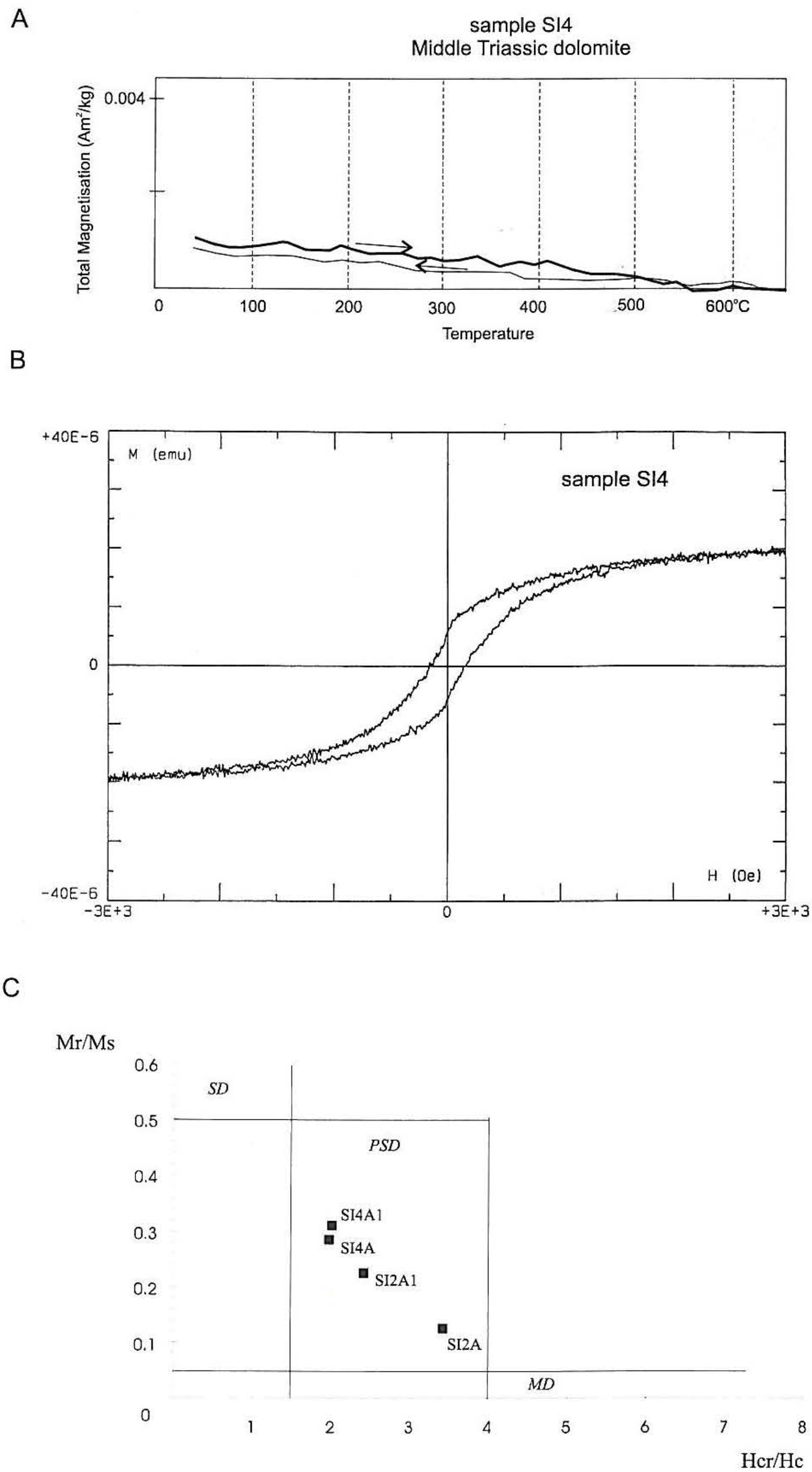


Fig. 22. Rock magnetic properties at the locality SI (continuation)

A — thermomagnetic analysis; bold line — heating curve, thin line — cooling curve; **B** — hysteresis loop; **C** — hysteresis ratios plotted on a Day *et al.* (1977) diagram; explanations as in the **Figure 21**

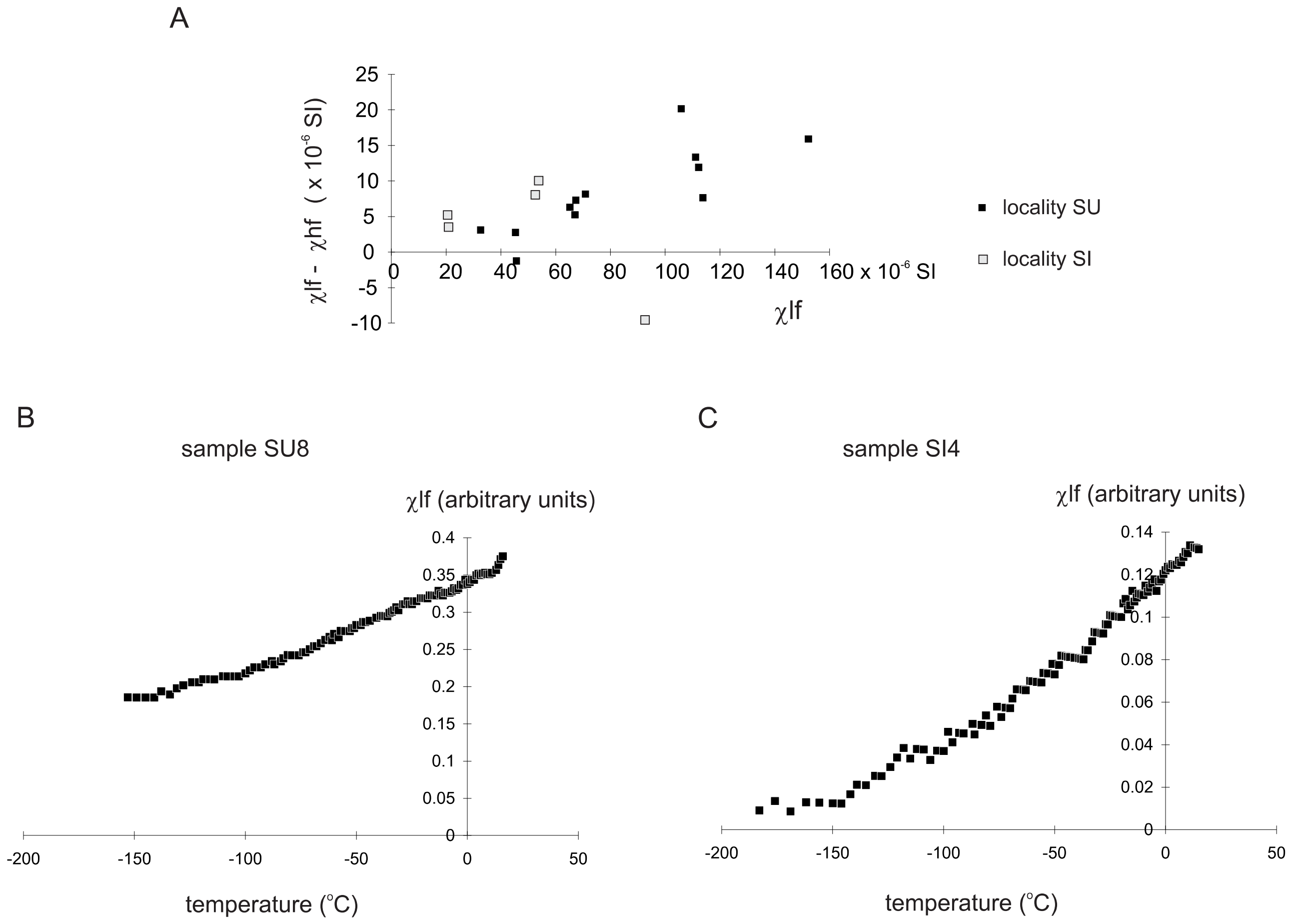


Fig. 23. A — Susceptibility differences χlf–χhf measured at low (0.47 kHz: χlf and high (4.7 kHz: χhf) frequency plotted as a function of low frequency susceptibility χlf.; B, C — low temperature susceptibility measurements

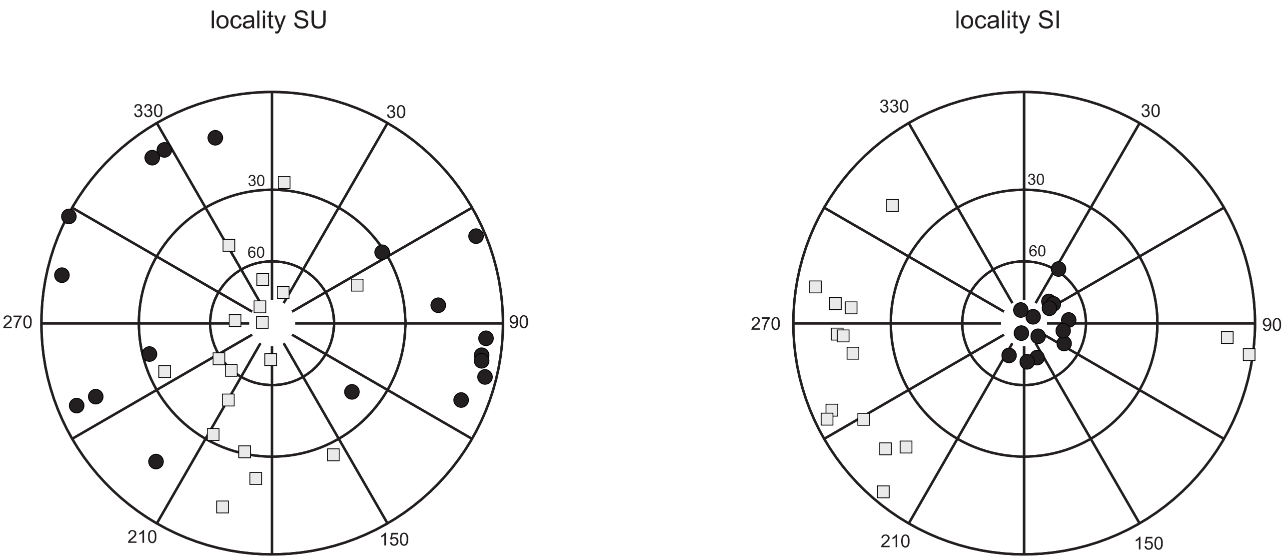


Fig. 24. Lower hemisphere projection of the maximum (squares) and minimum (dots) susceptibility axes at the localities SU and SI (after tectonic correction)

of secondary magnetite would be expected at ca. 420°C). During heating magnetite is altered to a mineral with lower value of saturation magnetisation (cooling curve in the Figure 21A is systematically below the heating curve) which is possibly hematite. Thermomagnetic curve for the sample SI4 from SI locality (Fig. 22A) is noisy and does not permit definite interpretation.

Magnetic hysteresis measurements were performed for rock chips of 8 samples. The hystereses are gently wasp-waisted (Figs. 21B and 22B) what, in the absence of hematite and pyrrhotite, might results from variegated coercivities of magnetite/maghemite grains (Tauxe *et al*., 1996). The hysteresis ratios Mr/Ms and Hcr/Hc point to a pseudo-single domain (PSD) range of magnetite (Fig. 21C and 22C).

There are evidences for presence of significant amounts of the superparamagnetic (SP) grains. There is a quasi-linear dependance between the value of susceptibility measured at low frequency (χlf) and the difference between susceptibilities measured at low and high frequencies (χlf–χhf) (Fig. 23A). The frequency dependent susceptibility difference originates from ultra fine SP magnetite particles (Forster *et al*., 1994). The intercept of the regression line seems positive due to contribution of magnetic material with no frequency dependence (most probably PSD magnetite as revealed from the hysteresis experiments). The presence of SP grains is further confirmed by the low temperature susceptibility measurements (Fig. 23B and 23C). Susceptibility increase from the liquid nitrogen up to room temperature is caused by shift of relaxation times, from SD to SP behaviour (Gialanella *et al*., 1994).

Magnetic fabric in the dolomites of the Suchy Wierch unit is well defined but differs considerably between localities (Fig. 24). In the locality SI the K3 axes group generally close to the pole of the bedding plane. This feature is characteristic for weakly deformed sediments in the compaction stage. K1 axes are dispersed within the bedding plane with SW to NW azimuths. On the other hand principal susceptibility axes in the locality SU are quite dispersed. K1 axes form a girdle in the N–S direction while most of K3 axes group within the bedding plane, preferably in W–E direction. If magnetic fabric in both localities is related to SP magnetite grains (see above), then it must be implied that dolomites in the locality SU suffered more intensive internal deformation than those in the locality SI.

4.1.3. DEMAGNETIZATION RESULTS

NRM intensities ranged between 3.43 and 12.5 x 10^{-3} A/m (mean 7.92 x 10^{-3} A/m) for the locality SU and between 2.2 x 10^{-3} and 22.7 x 10^{-3} A/m (mean 10.52 x 10^{-3} A/m) for the locality SI. Pilot specimens were demagnetized thermally, with the AF and combined method. Results obtained with all methods were very similar for almost all samples. AF method was used for demagnetization of the remaining part of the collection. NRM in the locality SU consists of a single component (labelled SU) which is nicely demagnetized between 5 and 120 mT (Fig. 25A). Unblocking temperature spectrum of this component is between 350 and 450°C (Fig. 25B).

In the samples from the locality SI the 20 mT and 300°C demagnetization step removed a low stability component (Fig. 26). High stability component SI was demagnetized between 350 and 450°C (Fig. 26A) or between 20 and 90 mT (Fig. 26B). Sample mean directions of the components SU and SI are presented in the Figures 27 and 28, and in the Tables 7 and 8.

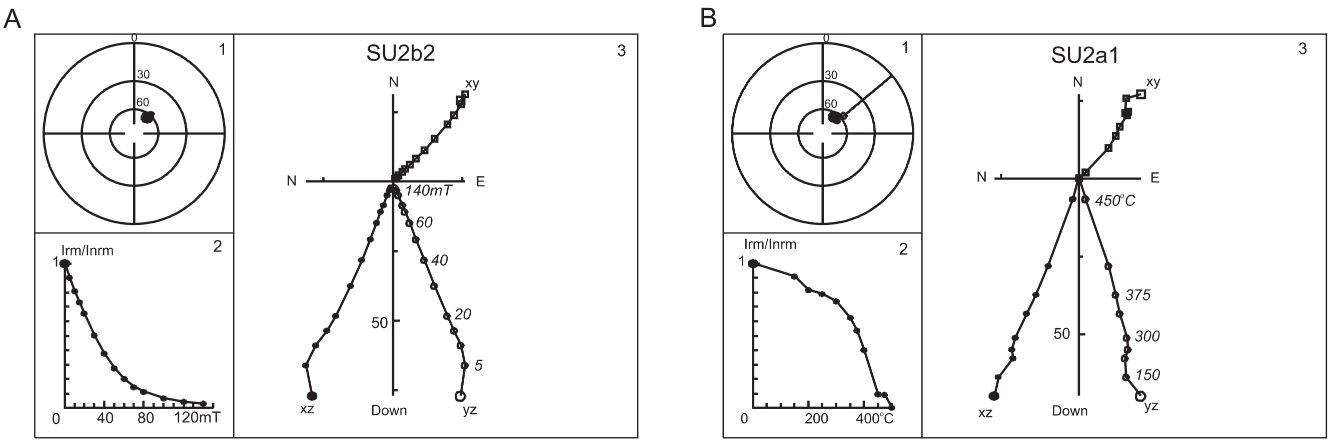


Fig. 25. Alternating field (A) and thermal (B) demagnetization of the Ladinian dolomites from the locality SU (after tectonic correction)

1 — stereographic projection of the demagnetization path; black (white) dots — lower (upper) hemisphere directions; *2* — intensity decay curve; *3* — orthogonal projection (Zijderveld diagram); x,y,z — the planes of projection, NRM intensity in 10 A/m

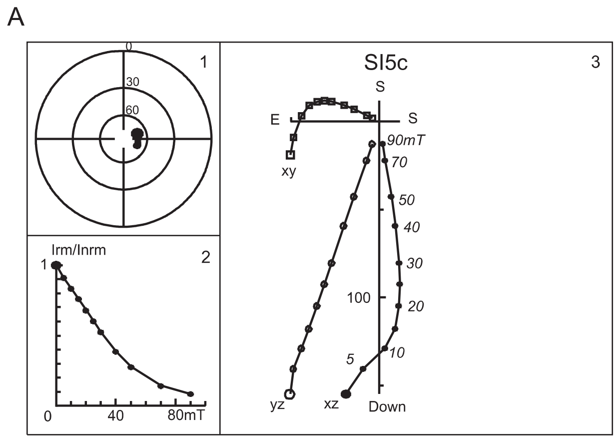


Fig. 26. AF demagnetizaton (A) and mixed AF + thermal demagnetization (B) of the Ladinian dolomites from locality SI (after tectonic correction)

Explanations as in the Figure 25

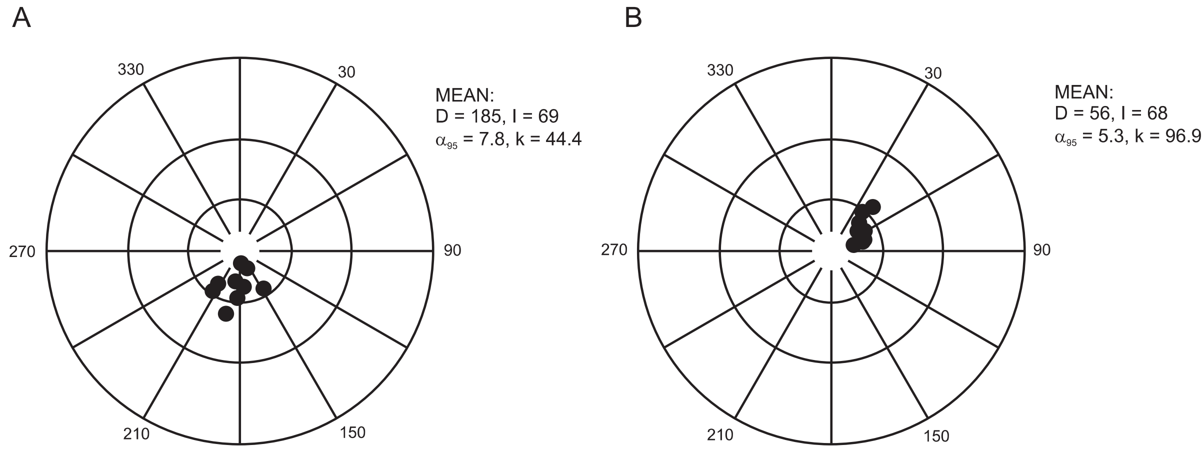


Fig. 27. Stereographic projection of the sample mean directions of the component SU before (A) and after tectonic correction (B)

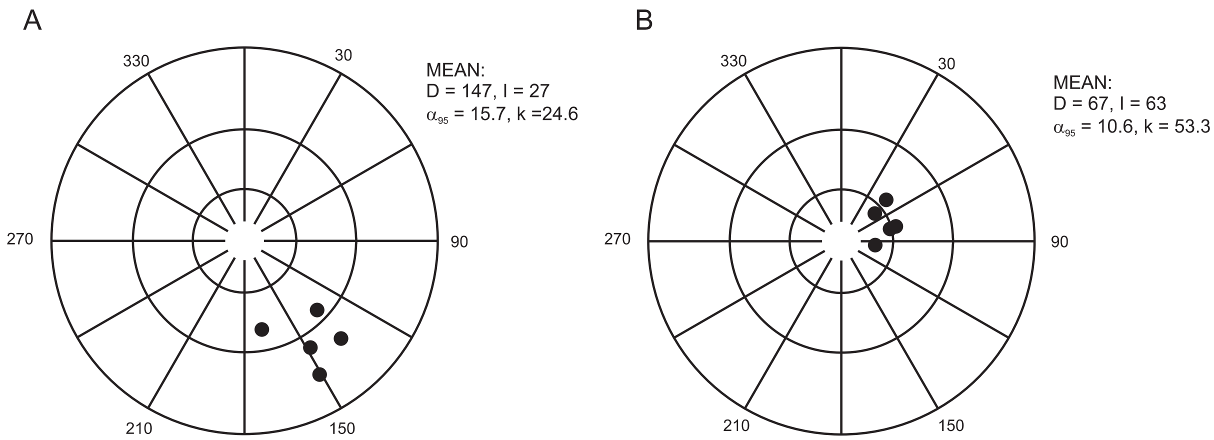


Fig. 28. Stereographic projection of the sample mean directions of the component SI before (A) and after tectonic correction (B)

Table 7

Sample mean directions of the component SI

Sample	D	I	Dc	Ic	α_{95}	k	n
SI1	136	19	77	58	2.8	392	8
SI2	134	35	47	55	6.4	376.7	3
SI3	169	40	50	65	–	–	1
SI4	149	24	76	61	7.3	160.9	4
SI5	151	13	98	71	4.1	885.4	3

	D	I	α_{95}	k	Dc	Ic	α_{95}	k	N
Mean	147	27	15.7	24.6	67	63	10.6	53.3	5

D (I) — declination (inclination) before tectonic correction, Dc (Ic) — declination (inclination) after tectonic correction, α_{95}, k — Fisher statistics parameters, n — number of specimens used for calculation of characteristic directions, N — number of hand samples

Table 8

Sample mean directions of the component SU

Sample	D	I	Dc	Ic	α_{95}	k	n
SU1	192	53	76	77	11.3	119	3
SU2	153	78	39	61	3.3	792	4
SU3	149	64	61	68	5.7	470	3
SU4	212	68	67	71	15.1	68	3
SU5	214	62	74	71	10.7	134	3
SU6	174	69	45	67	9.8	160	3
SU7	186	72	54	72	8.7	201	3
SU8	178	73	58	62	6.2	394	3
SU9	183	63	71	69	–	–	1

	D	I	α_{95}	k	Dc	Ic	α_{95}	k	N
Mean	185	69	7.8	44.4	56	68	5.3	96.9	9

Explanations as in the Table 7

4.2. WIERCH SPALENISKO

4.2.1. LOCALITY DESCRIPTION

Nine hand samples were taken from the gullies in the area of the Wierch Spalenisko (RT), in the western slopes of the Lejowa Valley. Entire Rhaetian profile was sampled from the Keuper dolomites in the bottom (sample RT9) up to the Hettangian marly mudstones (samples RT 1–2). Samples RT 3–8 were taken from the thick bedded dark limestones from five stratigraphical horizons. Detailed description of profile was published by Gaździcki (1974). The area is cut by numerous faults and the outcrops are quite poor. The strata dip 30–45° to the NE (Tab. 6).

4.2.2. PETROGRAPHY AND ROCK MAGNETISM

Thin sections were prepared from samples RT1 and RT6. Sample RT1 is a carbonate mudstone almost devoid of bioclasts, cut by several calcite veins. The veins do not contain opaque minerals which are distributed in the muddy matrix. SEM investigations revealed the presence of pyrite, zircon and rutile. The two latter minerals indicate influx of detrital terrigenous material to the basin. RT6 is a fossiliferous grainstone (with echinoderms, molluscs and foraminifers) that originated in a high energy shallow water environment with no influx of terrigenous grains. Unaltered pyrite occurring in framboids and regular grains is the most abundant non-transparent mineral. It resides in the rock matrix as well as the fillings in bioclasts and in stylolitic seams. Fine grained iron oxides or hydroxides of post-pyrite origin are occasionally observed.

IRM experiments (stepwise acquisition and 3-axes thermal demagnetization) were performed for the sample RT2. Low coercivity minerals prevail in this sample. The 95% saturation is achieved in the field 250 mT (Fig. 29A-1). Single magnetic phase with maximum unblocking temperature 500–550°C is observed in the IRM thermal demagnetization curves (Fig. 29B-1). These features are characteristic for fine grained magnetite. The minor amounts of high coercivity minerals are present but there is no evidence for presence of hematite, because unblocking temperatures are not higher than 550°C. Magnetic susceptibility does not change during thermal treatment between 20 and 400°C. Above 400°C its stepwise increase is observed (Fig. 29C-1). It might be indicative for transformation of pyrite to magnetite (Van Velzen, 1992).

Entirely different magnetic properties reveals the sample RT9 (Keuper dolomite). Only high coercivity minerals occur in this sample — saturation is not reached in the field 1.3 T (Fig. 29A-2). Maximum unblocking temperature is 650°C (Fig. 29B-2). These properties point to hematite as the magnetic carrier. There is no indication for presence of magnetite or iron sulphides. Sharp susceptibility increase is observed about 600°C (Fig. 29C-2). It might be interpreted as production of magnetite from clay minerals or transformation of hematite to magnetite (Van Velzen, 1992).

Thermomagnetic analysis was carried out for the sample RT7 (Fig. 30A). The magnetization decays smoothly up to

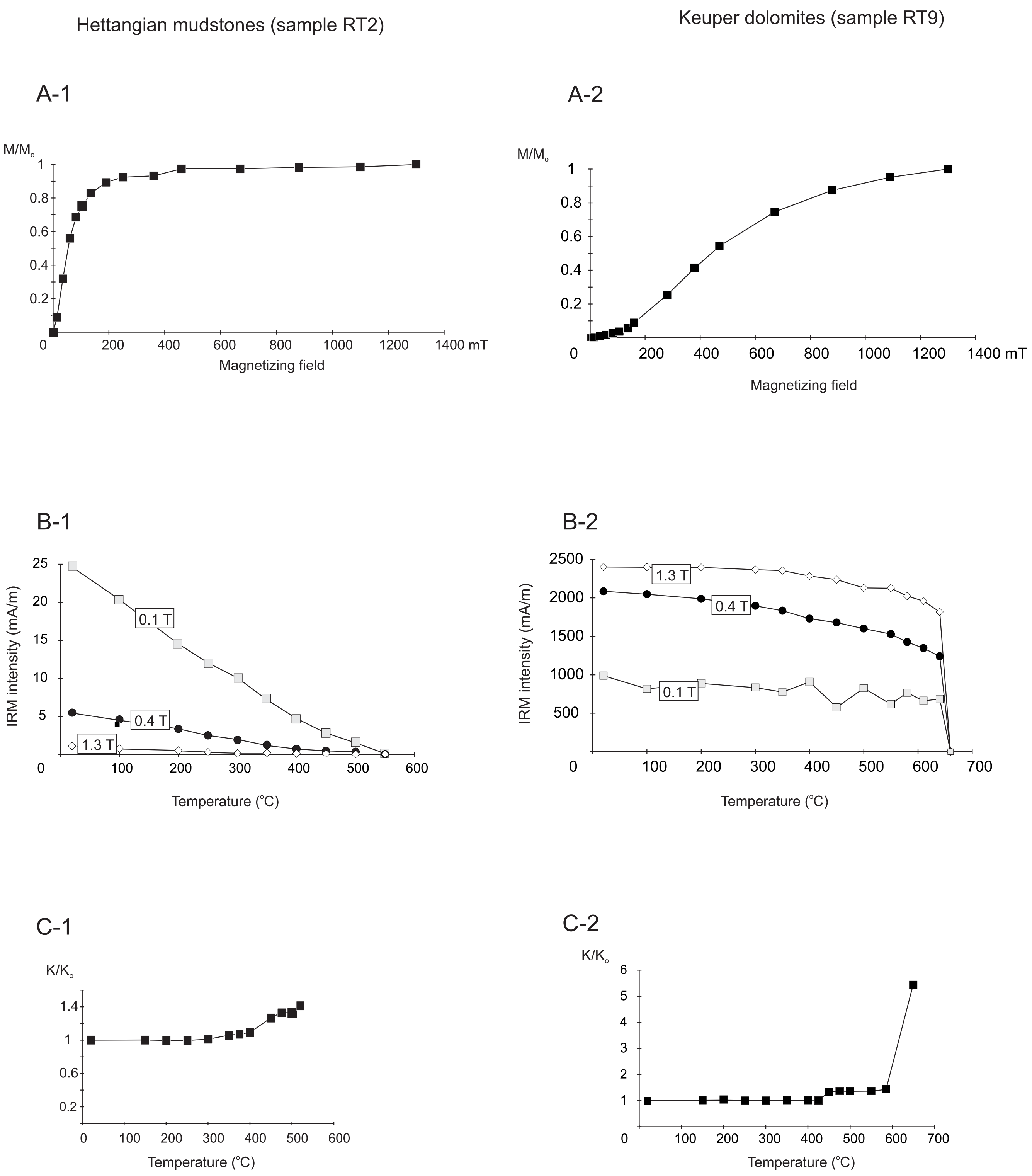


Fig. 29. Rock magnetic properties of the Hettangian mudstones and Keuper dolomites from the locality RT

A — IRM acquisition curve; **B** — thermal demagnetization of the 3-axes IRM acquired in the fields 0.1T, 0.4T and 1.3T; **C** — susceptibility changes during thermal treatment

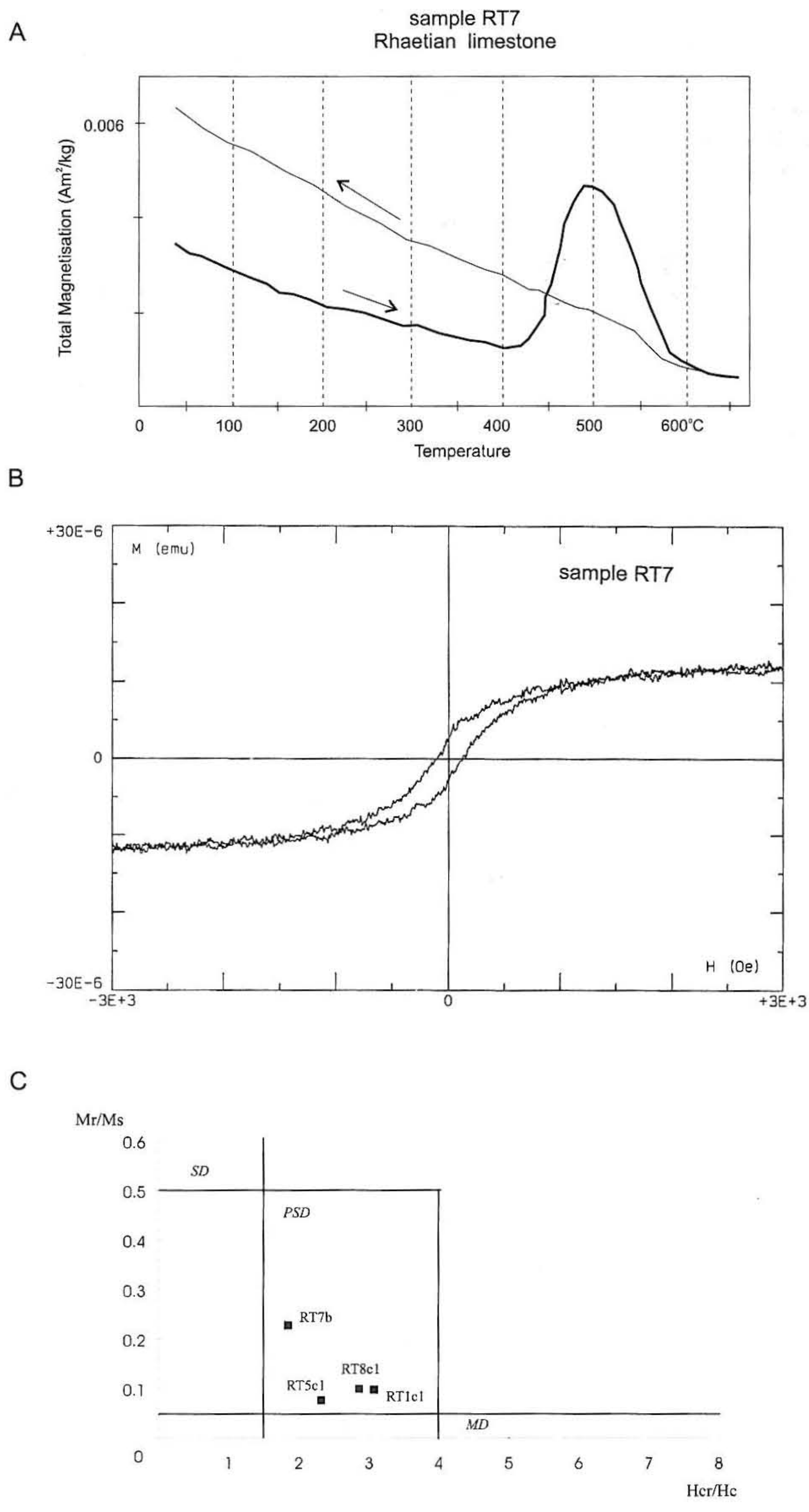


Fig. 30. Rock magnetic properties of the Rhaetian/Hettangian rocks at the locality RT

A — thermomagnetic analysis; bold line — heating curve, thin line — cooling curve; **B** — hysteresis loop; **C** — hysteresis ratios plotted on a Day *et al.* (1977) diagram

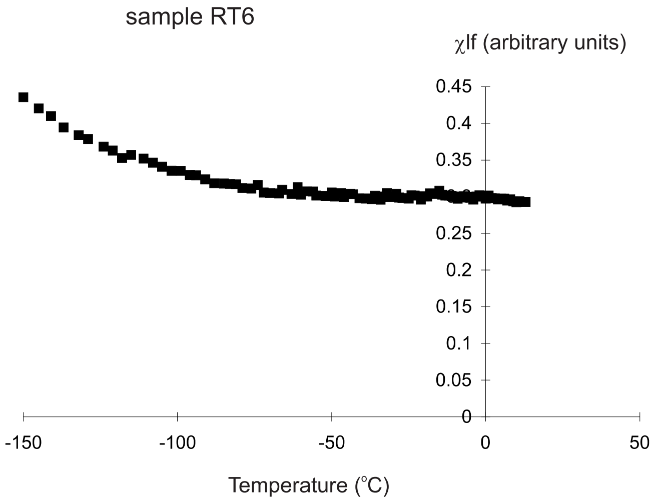


Fig. 31. Low temperature susceptibility measurements (sample RT6 — Rhaetian limestones)

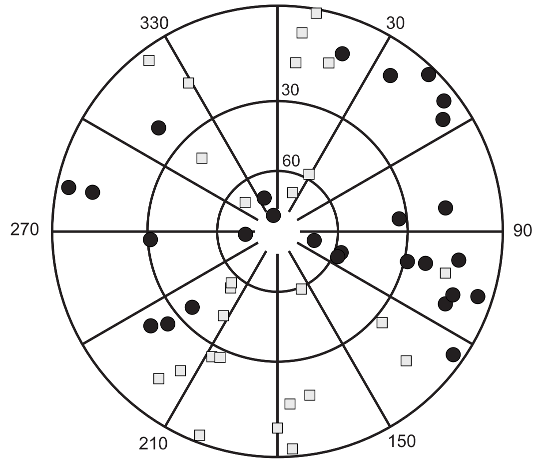


Fig. 32. Lower hemisphere projection of the maximum (squares) and minimum (dots) susceptibility axes at the locality RT (after tectonic correction)

410°C. Above this temperature the production of new magnetic mineral takes place which is manifested by increase of magnetization with temperature up to 495°C. Then the magnetization decays quickly up to 575°C. Slight decrease of magnetization is observed also between 575 and 660°C. The shape of the heating curve confirms the presence of magnetite. Curie points characteristic for pyrrhotite (300–350°C) are not observed. The peak on the heating curve is most likely related to decomposition of pyrite and origin of new magnetite (Mullender *et al.*, 1993). Magnetite is demagnetized up to its Curie point, but decay of magnetization between 575 and 660°C indicates to the presence of hematite. It is not certain whether this hematite is primary or originated due to thermal oxidation of magnetite. The latter case seems to occur because heating and cooling curves below 600°C are not identical. Obviously large part of magnetite that originated between 410 and 495°C was removed (oxidized to hematite).

The magnetic hysteresis measurements were performed for samples RT1, RT5, RT7 and RT8. Hysteresis pictured in the Fig. 30B is quite smart and very gently wasp-waisted. This could result from admixture of hematite or SP magnetite (Tauxe *et al.*, 1996). The hystereses loops for other three samples were poorly defined. Hysteresis parameters Mr/Ms and Hcr/Hc fall in the range of PSD magnetite with quite low Mr/Ms ratio (Fig. 30C).

Magnetic susceptibility in the liquid nitrogen temperature increases, as indicated in the Figure 31. It is interpreted as influence of paramagnetic minerals (Forster *et al.*, 1994) which dominate the low field susceptibility. Its values are between 40 and 300 x 10^{-6} SI units. Distribution of the principal susceptibility axes is quite complicated (Fig. 32). K1 axes tend to create N–S girdle, while K3 axes are distributed mostly along WNW–ESE striking girdle. Generally magnetic fabric in this locality must be regarded as poorly defined, but most probably of tectonic origin.

4.2.3. DEMAGNETIZATION RESULTS

NRM intensities were variegated ranging from 2.4 to 48.4 x 10^{-4} A/m (mean 13.38 x 10^{-4} A/m). Pilot specimens were demagnetized using thermal and AF method (Fig. 33A and B). Both methods seemed to be equally efficient for samples RT1–8 and bulk of collection was treated with the AF. Low stability, probably viscous component, was removed up to 250°C and 20 mT. Its intensity was low in comparison with the second characteristic component RT. This component was stable between 300 and 475°C and between 10 and 90 mT. Before tectonic correction it revealed southerly declina-

Table 9

Sample mean directions of the component RT

Sample	D	I	Dc	Ic	α_{95}	k	n
RT1–2	197	88	37	59	7	35.6	13
RT3	200	44	159	70	14	78	3
RT4	207	62	60	71	15.9	24	5
RT5	197	57	82	71	12.8	28.4	6
RT6	176	60	86	70	10.7	39.7	6
RT7–8	197	71	79	74	10.9	26.5	8
RT9	225	61	31	74	4.6	229	6

	D	I	α_{95}	k	Dc	Ic	α_{95}	k	N
Mean	200	64	11.4	29	69	74	11.2	29.7	7

Explanations as in the Table 7

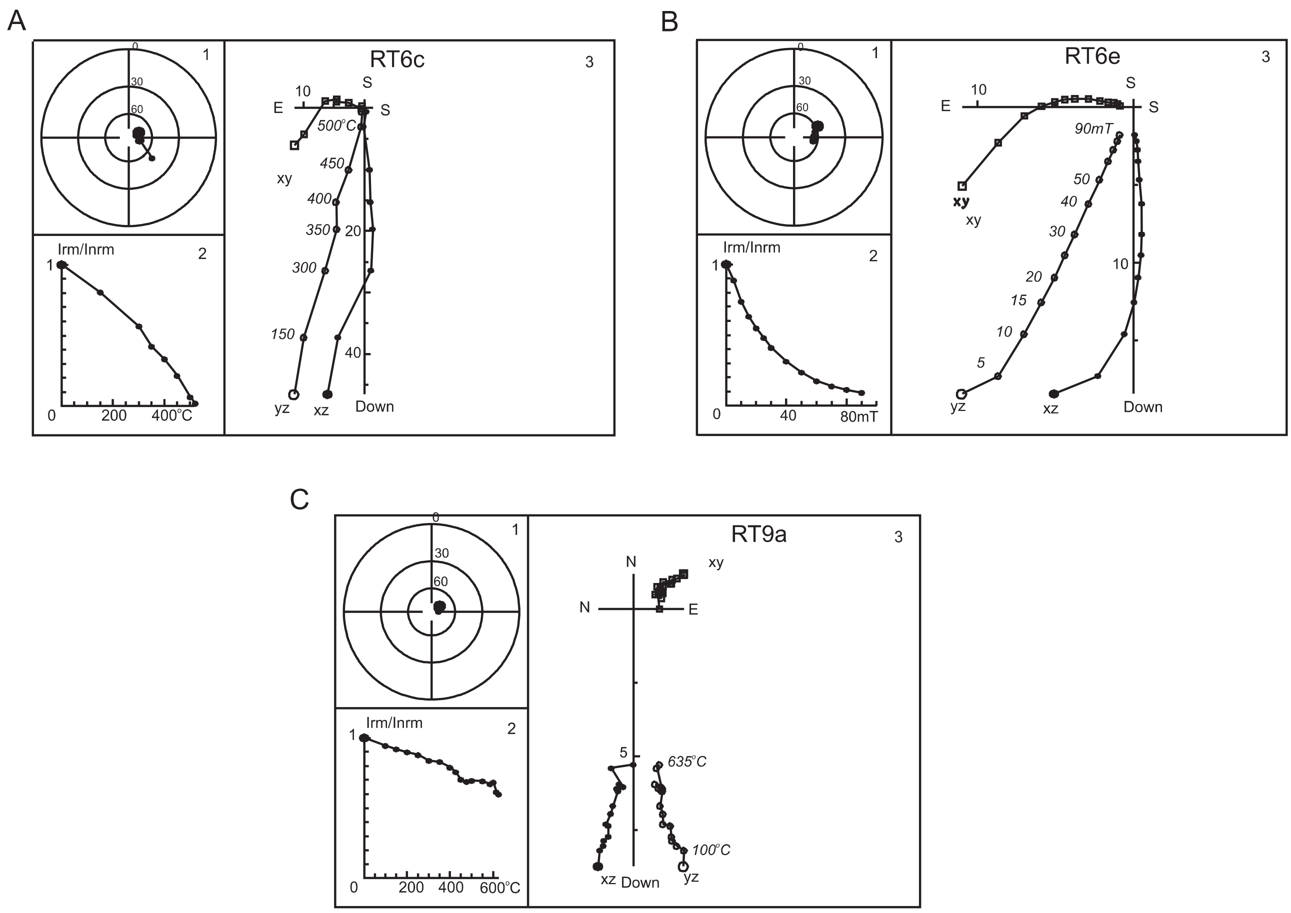


Fig. 33. Thermal (A) and AF demagnetization (B) of the Rhaetian limestone and thermal demagnetization of the Keuper dolomite (C — specimen desintegrated after 635°C) from the locality RT (after tectonic correction)

Explanations as in the Figure 25

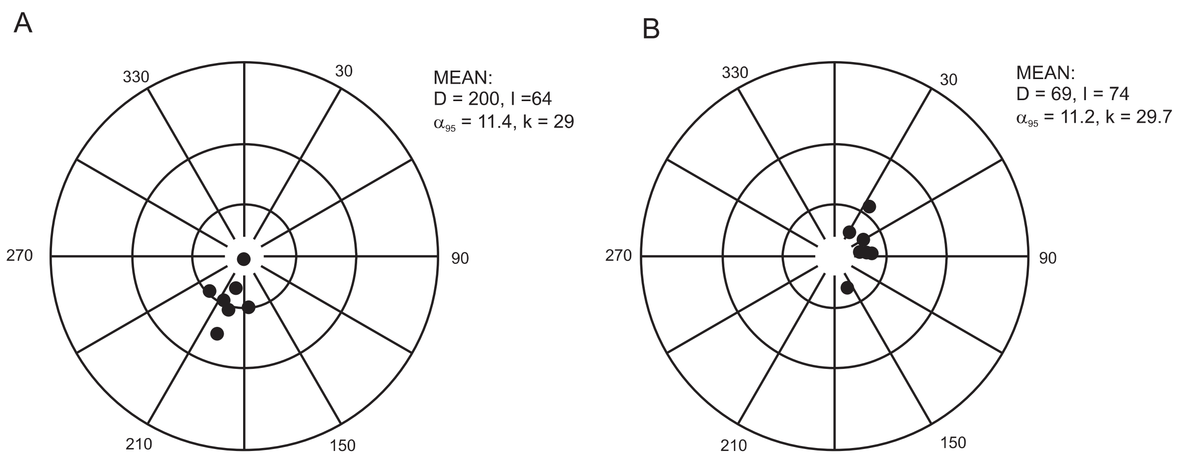


Fig. 34. Stereographic projection of the sample mean directions of the component RT before (A) and after tectonic correction (B)

tion and quite steep (ca. 70°) inclination. After tectonic correction the declination becomes ENE and inclination remains steep. The distribution of the component RT in both coordinates is presented in the Figure 34 and mean direction calculated in the Table 9.

As could be expected from the results of rock magnetic studies (section 4.2.2) dolomitic sample RT9 revealed different behaviour during thermal demagnetization. A single component was observed, similar to the RT, which was stable up to the temperatures 630–660°C (Fig. 33C). In this sample component RT is based on hematite.

4.3. DOLINA DŁUGA — LIASSIC

4.3.1. LOCALITY DESCRIPTION

Six hand samples (DDL10–15) of dark micritic (sometimes spotty) limestones were taken from the outcrops in the upper part of the Długa Valley (DDL) along the footpath. The rocks belong to the Sołtysia Marlstone Formation (Lefeld *et al.*, 1985) and their age is Sinemurian–Pliensbachian. The strata are tilted N to NNE with dip varying between 23 and 68°.

4.3.2. PETROGRAPHY AND ROCK MAGNETISM

Thin sections were prepared from samples DDL12 and DDL13. The rock is strongly recrystallized wackestone with sparse bioclasts. Authigenic pyrite (framboids and cubic grains) is abundant. Marcasite intergrowths in pyrite are observed. Opaque minerals are distributed in the rock matrix and do not occur in the calcite veins.

SEM investigations were carried out for the sample DDL13. Only pyrite grains were identified. Zircons or titanium oxides which could be of detrital origin were not observed.

One specimen from the sample DDL11 was subjected to the IRM experiments (stepwise acquisition of the IRM and thermal demagnetization of the 3 axes IRM). Low coercivity magnetic minerals prevail in the sample. 95% of saturation is achieved in the field of 250 mT (Fig. 35A). However, the shape of the IRM curve indicates also presence of high coercivity minerals. Thermal demagnetization of 3-axis IRM (Fig. 35B) reveals the presence of three magnetic phases with maximum unblocking temperatures 350°C, 500–550°C and above 600°C. The latter is undoubtedly hematite because it reveals also the highest coercivities. The 500–550°C phase is most likely magnetite. The 350°C phase could be either maghemite or pyrrhotite. The presence of titanomagnetite is rather unlikely as no detrital minerals were observed in the SEM. Magnetic susceptibility rises during thermal treatment after 400°C (Fig. 35C) what might indicate the alteration of iron sulphides to magnetite (Van Velzen, 1992).

One thermomagnetic analysis was performed for the rock chip from the sample DDL10 (Fig. 36A). Curie temperatures between 300 and 350°C are not observed, thus there is no evidence for presence of pyrrhotite in this sample. Above 440°C a new magnetic mineral originates which is demagnetized between 510 and 590°C. It is most likely magnetite that origi-

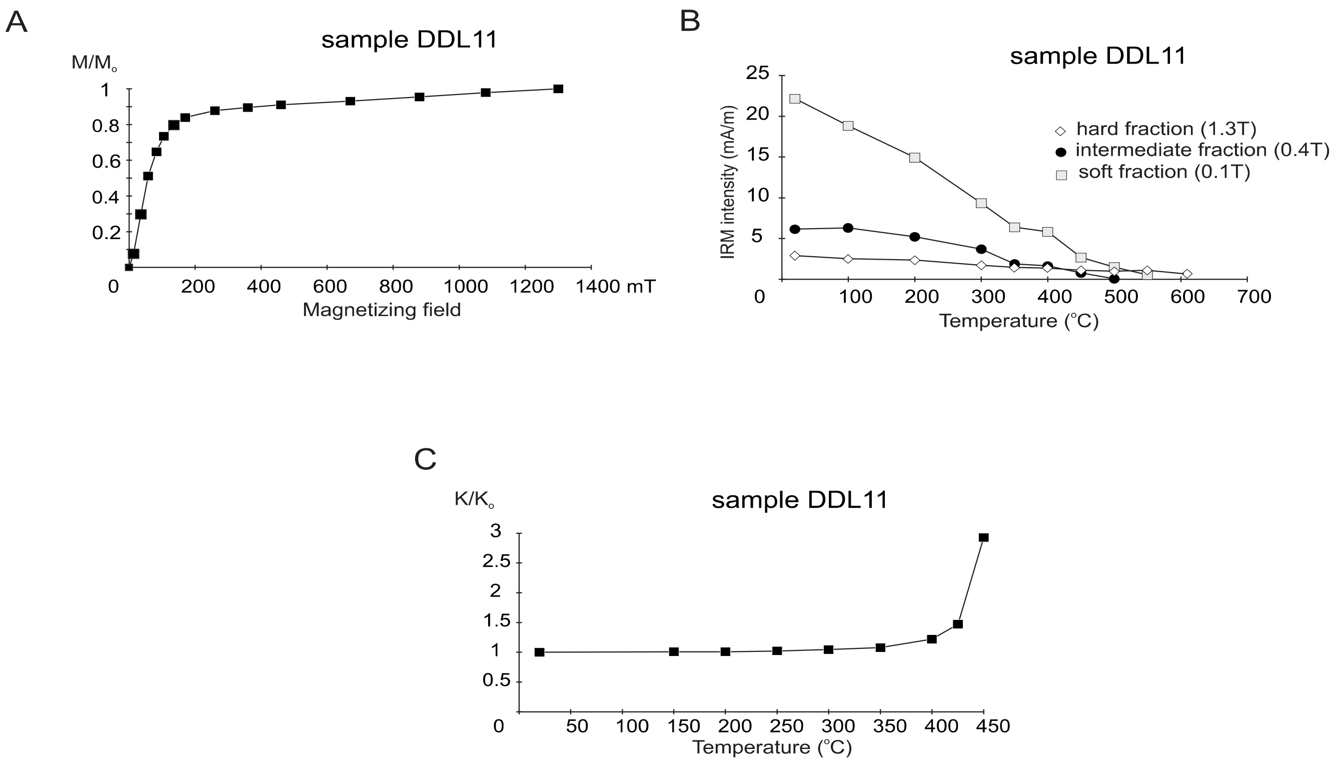


Fig. 35. Rock magnetic properties of the Liassic limestones from the locality DDL

A — IRM acquisition curve; **B** — thermal demagnetization of the 3-axes IRM acquired in the fields 0.1T, 0.4T and 1.3T; **C** — susceptibility changes during thermal treatment

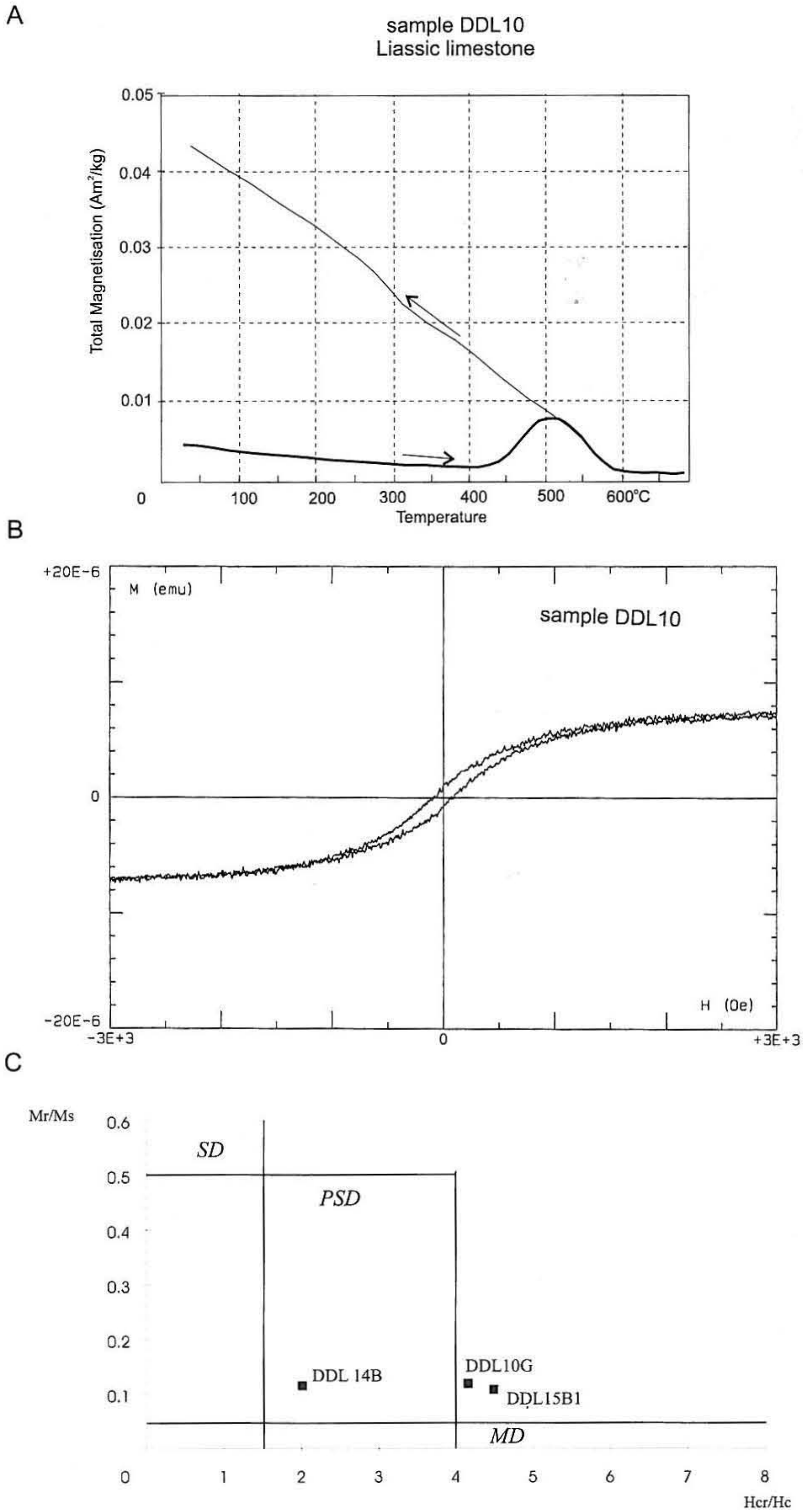


Fig. 36. Rock magnetic properties of the Liassic limestones at the locality DDL

A — thermomagnetic analysis; bold line — heating curve, thin line — cooling curve; **B** — hysteresis loop; **C** — hysteresis ratios plotted on a Day *et al.* (1977) diagram

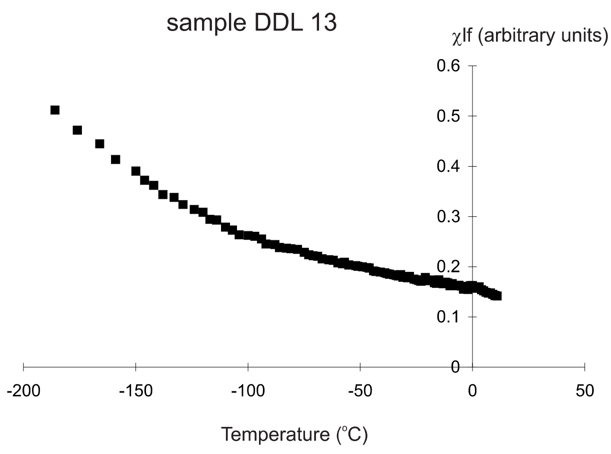


Fig. 37. Low temperature susceptibility measurements (sample DDL 13 — Liassic limestones)

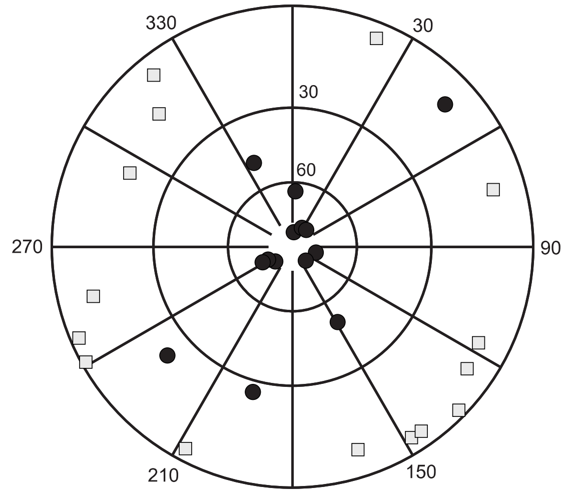


Fig. 38. Lower hemisphere projection of the maximum (squares) and minimum (dots) susceptibility axes at the locality DDL (after tectonic correction)

nates after breakdown of pyrite. The production of secondary magnetite is confirmed by the shape of the cooling curve.

Magnetic hysteresis measurements were performed for three rock chips from the samples DDL10, DDL14 and DDL15. An example is presented in the Figure 36B. Hysteresis ratios of sample DDL 14b fall in the PSD range with low Mr/Ms ratio. The parameters of other two samples are outside the PSD range revealing Hcr/Hc ratio slightly higher than expected for this range (Fig. 36C). This might be explained by presence of minor amounts of hematite. According to Channell and McCabe (1994) presence of hematite increases the Hcr/Hc ratio in magnetite bearing limestones.

As in the locality Wierch Spalenisko (RT) the low field susceptibility seems to be carried by paramagnetic minerals, what is inferred from the higher susceptibility values at low temperatures (Fig. 37). Room temperature susceptibility values are not higher than 80 x 10^{-6} SI units.

The AMS parameters are typical for weakly to moderately deformed sedimentary rocks. In the samples where anisotropy is well defined primary magnetic fabric with bedding parallel foliation is preserved.

K3 axes group roughly perpendicular to the bedding plane while directions of the K1 axes are more scattered (Fig. 38).

4.3.3. DEMAGNETIZATION RESULTS

NRM intensities ranged between 1.88 and 8.52 x 10^{-4} A/m (mean 4.97 x 10^{-4} A/m). Pilot specimens were demagnetized thermally and with AF. Both methods gave similar results (Fig. 39A and 39B) and AF was applied to entire collection. In most samples a low stability component was removed up to 300°C and 20 mT. Higher stability component DDL was almost completely demagnetized between 300 and 400°C and between 20 and 90 mT. Before tectonic correction it reveals southerly declination and steep inclinations. After tectonic correction inclination amounts to 64° and the declination becomes north-easterly. Considerable improvement of statistical parameters is observed after tectonic correction. (Fig. 40 and Tab. 10).

Table 10

Sample mean directions of the component DDL

Sample	D	I	Dc	Ic	α_{95}	k	n
DDL10	76	80	15	59	17.3	29.1	4
DDL11	139	70	54	74	5.7	476.4	3
DDL12	145	66	65	60	3.6	283.2	6
DDL13	201	55	30	57	13.4	47.7	4
DDL14	165	81	33	74	8.7	112.4	4
DDL15	174	71	18	53	5.3	297	4

	D	I	α_{95}	k	Dc	Ic	α_{95}	k	N
Mean	162	74	12.7	28.6	33	64	10.7	40.3	6

Explanations as in the Table 7

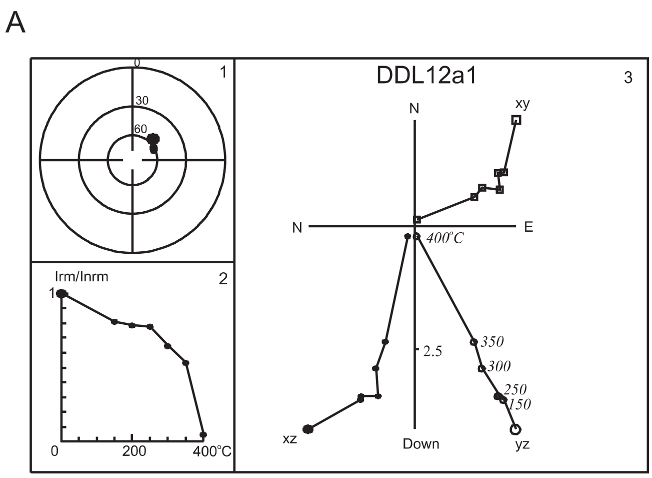


Fig. 39. Thermal (B) and AF (A) demagnetization of the Liassic limestone from the locality DDL (after tectonic correction)

Explanations as in the Figure 25

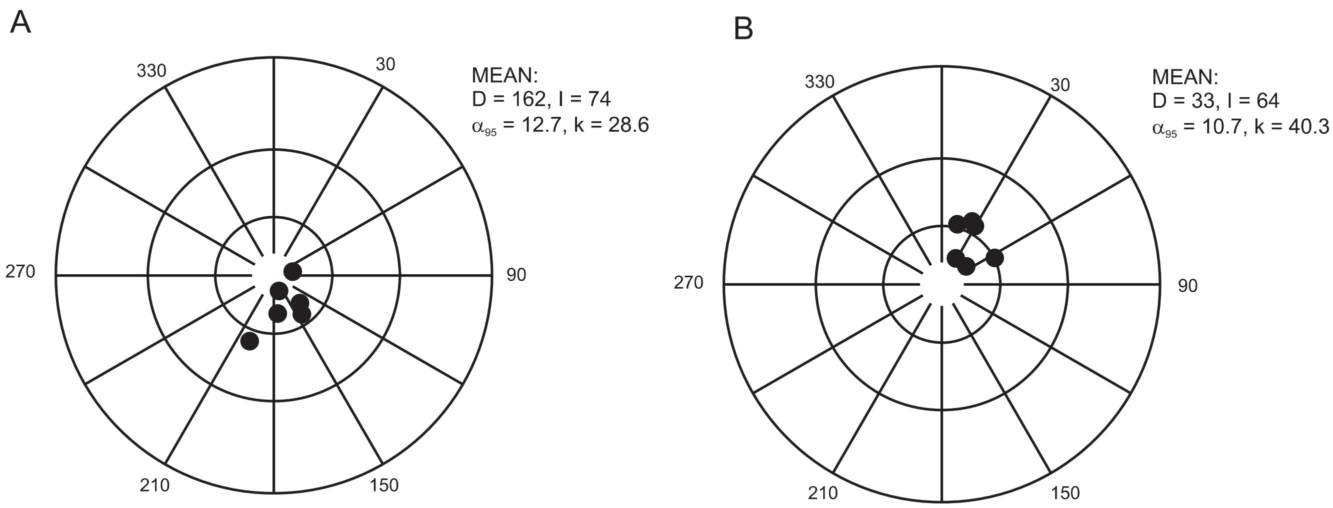


Fig. 40. Stereographic projection of the sample mean directions of the component DDL before (A) and after tectonic correction (B)

4.4. DOLINA KRYTA

4.4.1. LOCALITY DESCRIPTION

Six hand samples of greenish-grey marly limestones were taken from the middle part of the Kryta Valley (KR) from the series of outcrops in its bottom, along the footpath. The rocks belong to the lower part of the Pieniny Limestone Formation (Lefeld *et al.*, 1985) and their age could be estincted as Lower Tithonian, after occurrence of globochaete–saccocoma–radiolarian microfacies (Pszczółkowski, pers. commun.). They dip 38–47° to NNE. The rocks reveal very distinct foliation which is parallel to the bedding plane.

4.4.2. PETROGRAPHY AND ROCK MAGNETISM

Thin sections were prepared from samples KR2 and KR4. The rocks are pelagic wackestones with nannofossils. Quite numerous fine detrital grains were identified: quartz, feldspars, micas and litoclasts. Non-transparent minerals are scarce. They are distributed parallel to the bedding parallel foliation planes, which are of stylolitic character. Ferruginous substance is also distributed in the matrix. Pyrite or post-pyrite grains seem to be totally absent. Opaque minerals do not occur in the calcite veins.

The stepwise IRM acquisition and 3-axes thermal demagnetization were performed for one specimen from sample KR5. Low coercivity minerals prevail in the sample. The 95% of saturation was achieved in the field of 200 mT (Fig. 41A). The maximum unblocking temperatures of the magnetic minerals were between 500 and 585°C (Fig. 41B). Hard and intermediate coercivity fraction constituted only minor part of the IRM and they were also unblocked up to 585°C. Magnetic susceptibility rises during thermal demagnetization after 450°C (Fig. 41C). Maximum values of unblocking temperatures and quick saturation indicate that magnetite or titanomagnetite is the only magnetic mineral.

Thermomagnetic analysis were performed for two samples: KR4 and KR5 (Fig. 42A). The Curie temperature is slightly higher than 600°C, implying that magnetite may be partially oxidized during heating (Mullender *et al.*, 1993). Curie temperatures characteristic for titanomagnetites were not observed. Cooling curve reveals lower magnetizations than the heating curve what confirms the assumption that a part of magnetite might be replaced by a less magnetic mineral (possibly hematite) during thermal treatment. There is no indication of pyrite or pyrrhotite.

Hysteresis measurements were performed for three rock chips from samples KR3, KR4 and KR5. The hysteresis loop is very imperfect (Fig. 42B) but the Mr/Ms and Hcr/Hc ratios are quite consistent and fall in the PSD range (Fig. 42C). The Mr/Ms ratio is very low and close to the MD field. Low temperature susceptibility studies indicate paramagnetic origin of the susceptibility signal — susceptibility in the liquid nitrogen temperature is higher than in the room temperature (Fig. 43).

Susceptibility values are moderately low (between 80 and 150 x 10^{-6} SI units). Magnetic fabric in this locality is very typical for weakly deformed sedimentary rocks. It is perfectly developed with bedding parallel foliation and WNW–ESE directed lineation (Fig. 44).

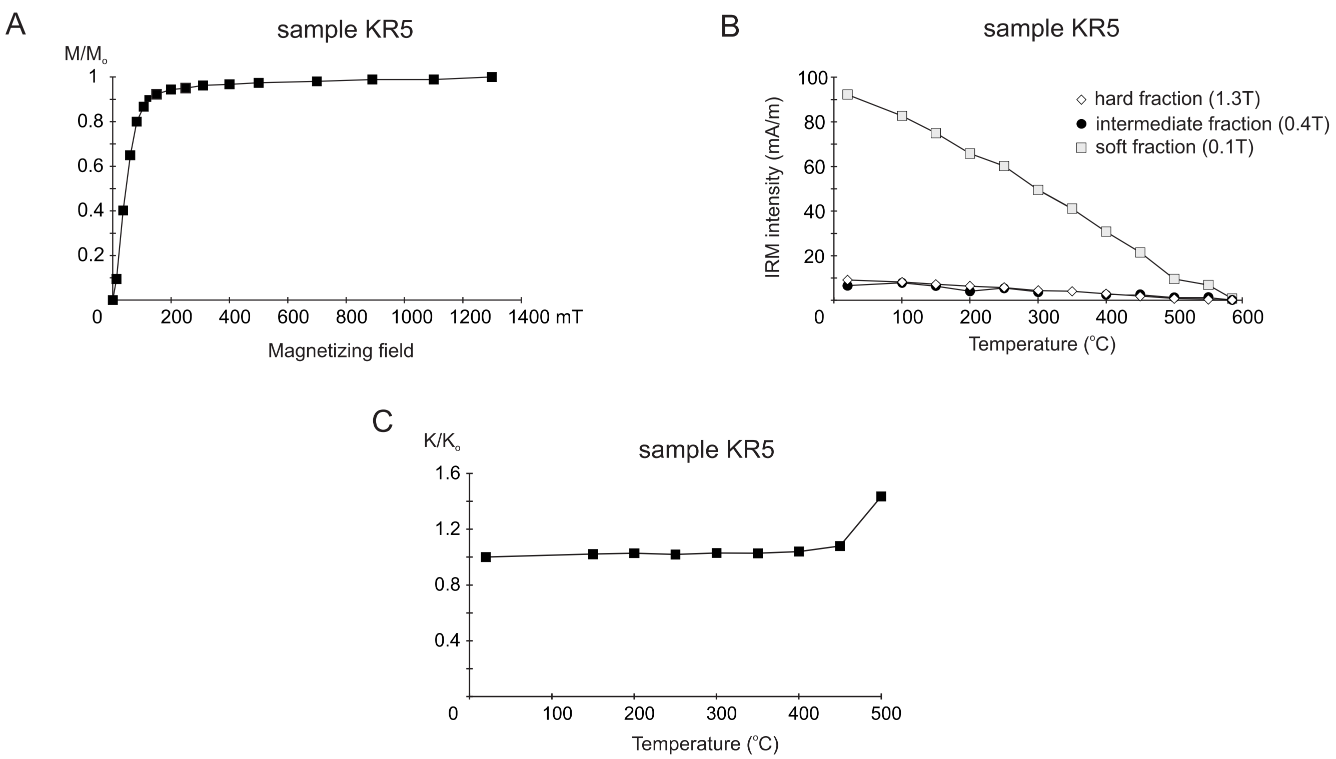


Fig. 41. Rock magnetic properties of the Tithonian marly limestones from the locality KR

A — IRM acquisition curve; **B** — thermal demagnetization of the 3-axes IRM acquired in the fields 0.1T, 0.4T and 1.3T; **C** — susceptibility changes during thermal treatment

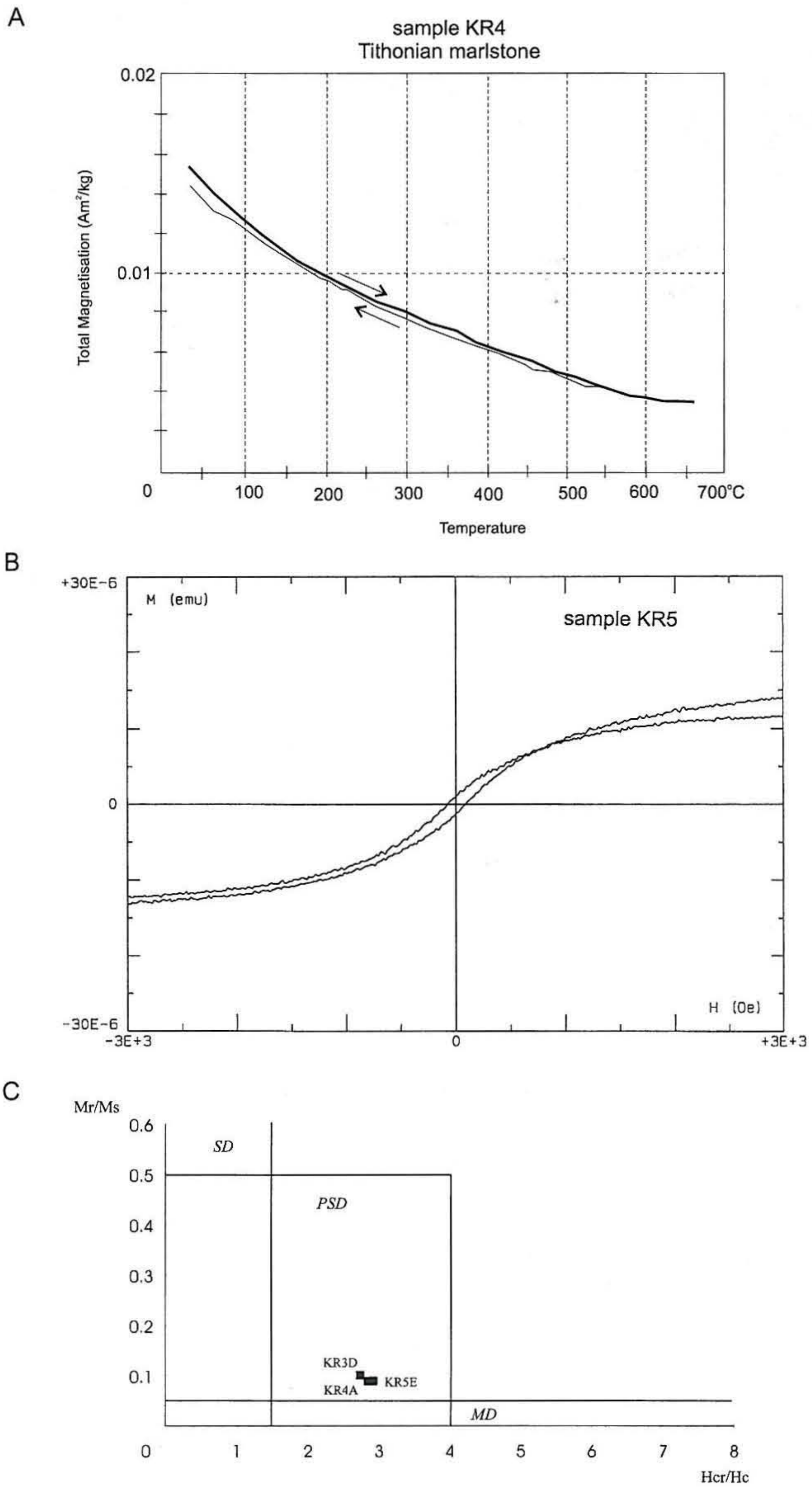


Fig. 42. Rock magnetic properties of the Tithonian marly limestones at the locality KR

A — thermomagnetic analysis; bold line — cooling curve, thin line — heating curve; **B** — hysteresis loop; **C** — hysteresis ratios plotted on a Day *et al.* (1977) diagram

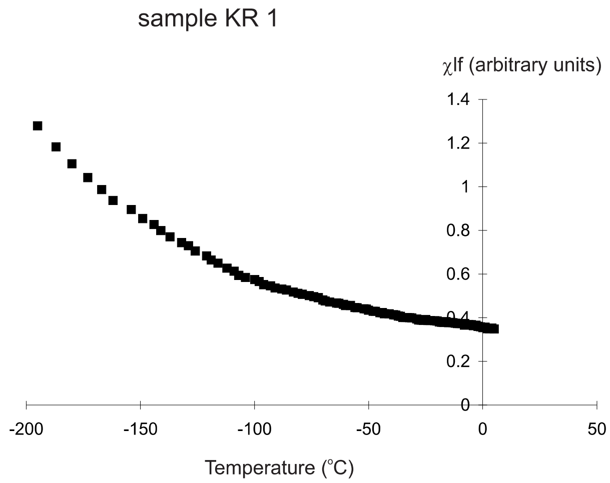


Fig. 43. Low temperature susceptibility measurements (sample KR1 — Tithonian marly limestones)

4.4.3. DEMAGNETIZATION RESULTS

NRM intensities were between 3.2 x 10^{-4} A/m and 6.97 x 10^{-4} A/m (mean 4.51 x 10^{-4} A/m). Pilot specimens were demagnetized thermally and by means of the AF (Fig. 45A and B). Thermal demagnetization revealed the presence of stable palaeomagnetic direction up to 475–500°C however the demagnetization path was not very smooth (Fig. 45B). Moreover the specimens easily disintegrated during heating. Better results were obtained using the AF demagnetization (Fig. 45A). Smooth decay of the NRM pointed to its univectorial nature. At the step 60 mT the initial NRM intensity fell down to 10%. The ChRM (component KR) was well defined till the end of demagnetization process. Before tectonic correction the component KR reveals southerly directed declinations and steep downwards inclinations. After tectonic correction the declinations become north-eastern and inclination values are close to 60° (Tab. 11; Fig. 46).

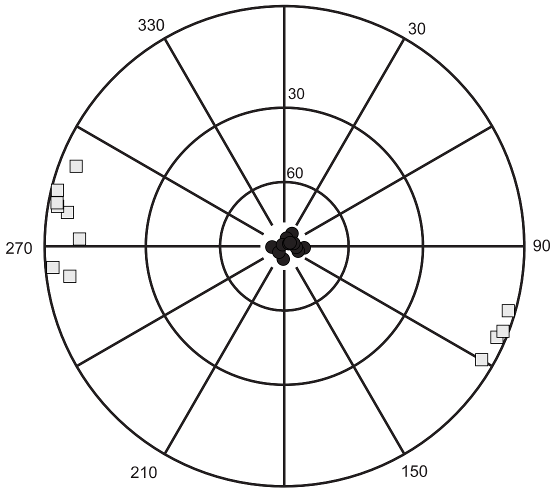


Fig. 44. Lower hemisphere projection of the maximum (squares) and minimum (dots) susceptibility axes at the locality KR (after tectonic correction)

Table 11

Sample mean directions from the locality KR

Sample	D	I	Dc	Ic	α_{95}	k	n
KR1	187	68	16	65	4.1	157.1	9
KR2	180	79	27	54	7.2	114.2	5
KR3	136	82	26	54	8.6	114.3	4
KR4	133	69	45	56	10	84.7	4
KR5	150	74	33	63	8.3	121.8	4
KR6	128	67	52	53	8.1	129.6	4

	D	I	α_{95}	k	Dc	Ic	α_{95}	k	N
Mean	151	75	8	70.4	34	58	7.3	85.7	6

Explanations as in the Table 7

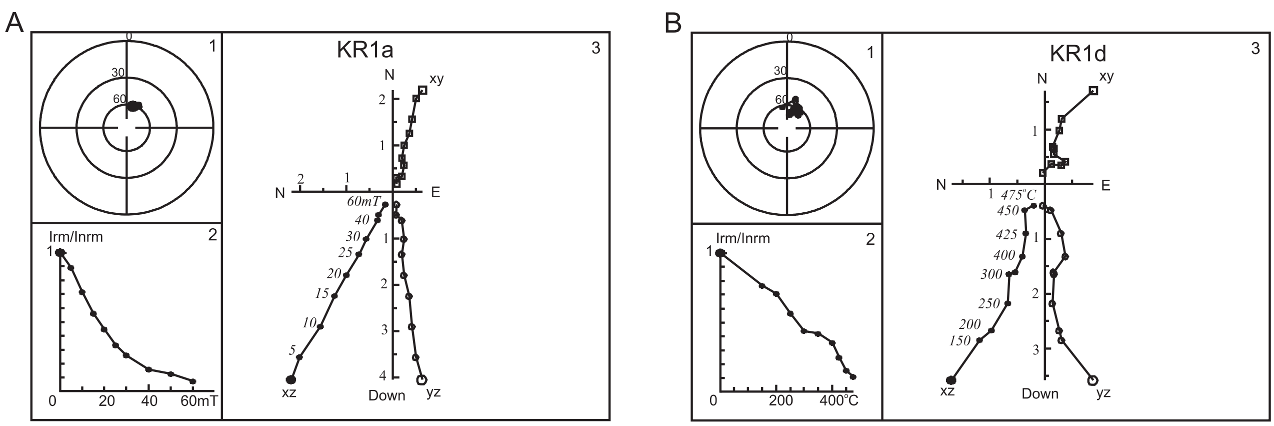


Fig. 45. AF (A) and thermal demagnetization (B) of the Tithonian marly limestone from the locality KR (after tectonic correction)

Explanations as in the Figure 25

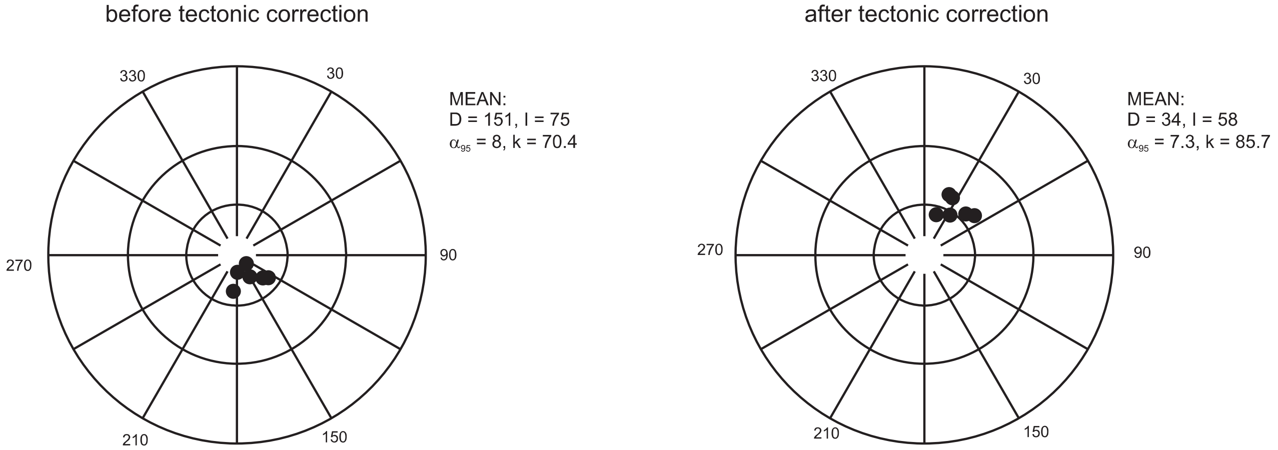


Fig. 46. Stereographic projection of the sample mean component KR

4.5. DOLINA DŁUGA — MALM/NEOCOMIAN

4.5.1. LOCALITY DESCRIPTION

Eight hand samples were taken from the ridge separating the Długa and Kryta valleys (DDM), between the Chochołowska Valley and Polish-Slovakian border. The rocks were grey-greenish pale limestones assigned to the upper part of the Pieniny Limestone Formation (Lefeld *et al.*, 1985) of Tithonian–Berriasian age. Detailed biostratigraphy of the profile, studied by Pszczółkowski (1996 and pers. commun.) revealed the presence of *Calpionella eliptica* which indicates the Berriasian age of investigated samples. Each hand sample was taken from separate limestone bed. The beds dip 30–65° to N–NE (Tab. 6).

4.5.2. PETROGRAPHY AND ROCK MAGNETISM

Thin sections were prepared from samples DDM4 and DDM7. The rocks are pelagic wackestones with calpio-

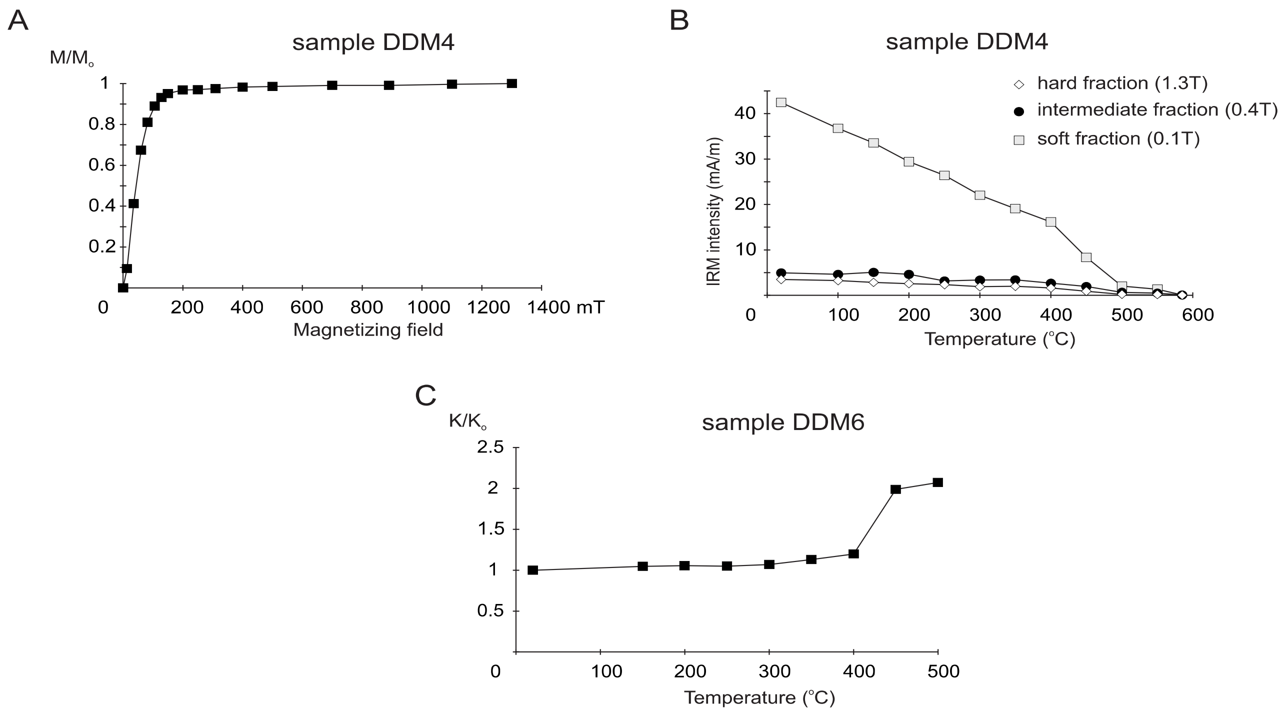


Fig. 47. Rock magnetic properties of the Berriasian limestones from the locality DDM

A — IRM acquisition curve; **B** — thermal demagnetization of the 3-axes IRM acquired in the fields 0.1T, 0.4T and 1.3T; **C** — susceptibility changes during thermal treatment

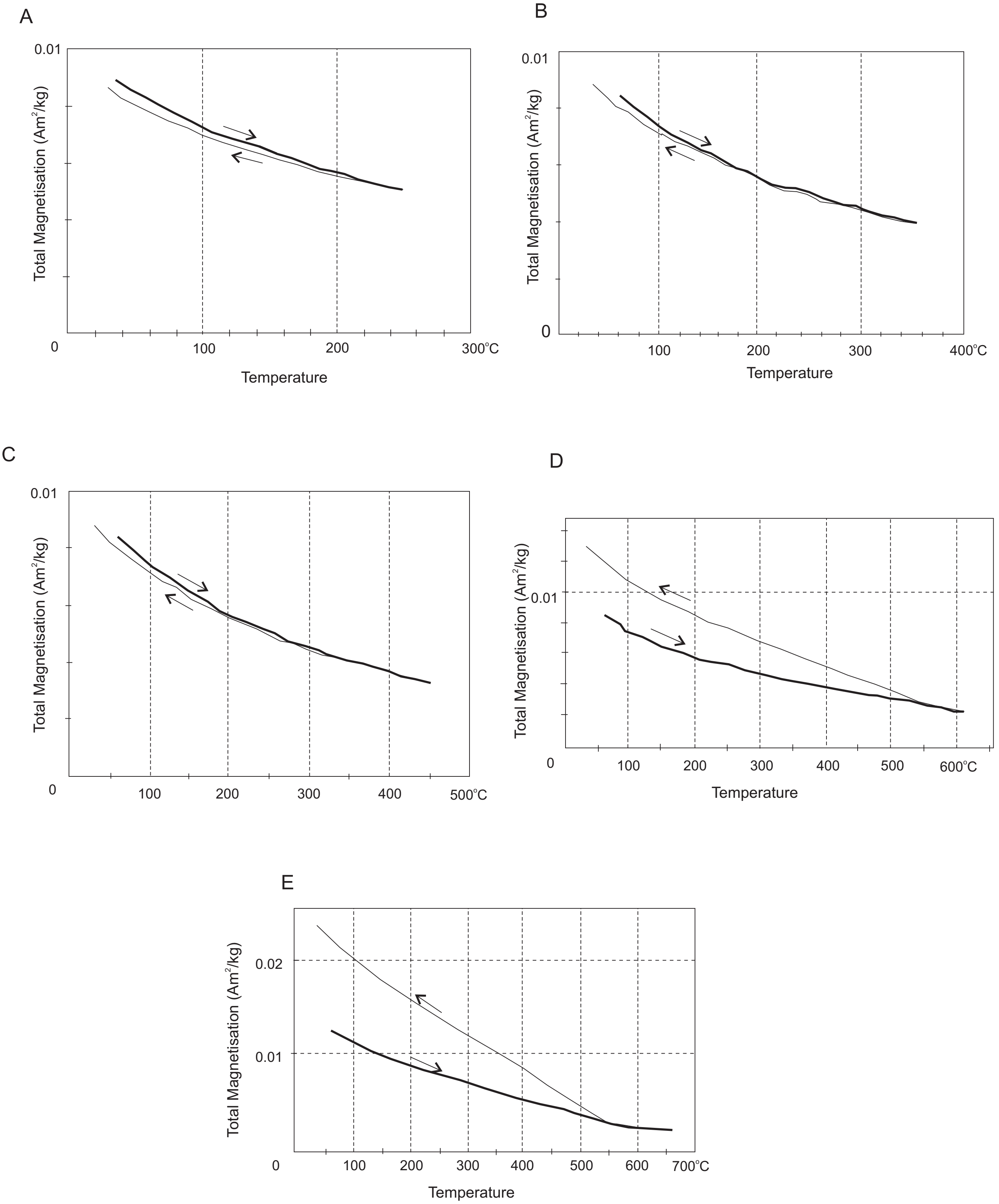


Fig. 48. Thermomagnetic analysis of the rock chip from the locality DDM (sample DDM5)

A — between 20 and 250°C; **B** — between 20 and 350°C ; **C** — between 20 and 450°C; **D** — between 20 and 620°C; **E** — between 20 and 660°C; bold line — heating curve, thin line — cooling curve

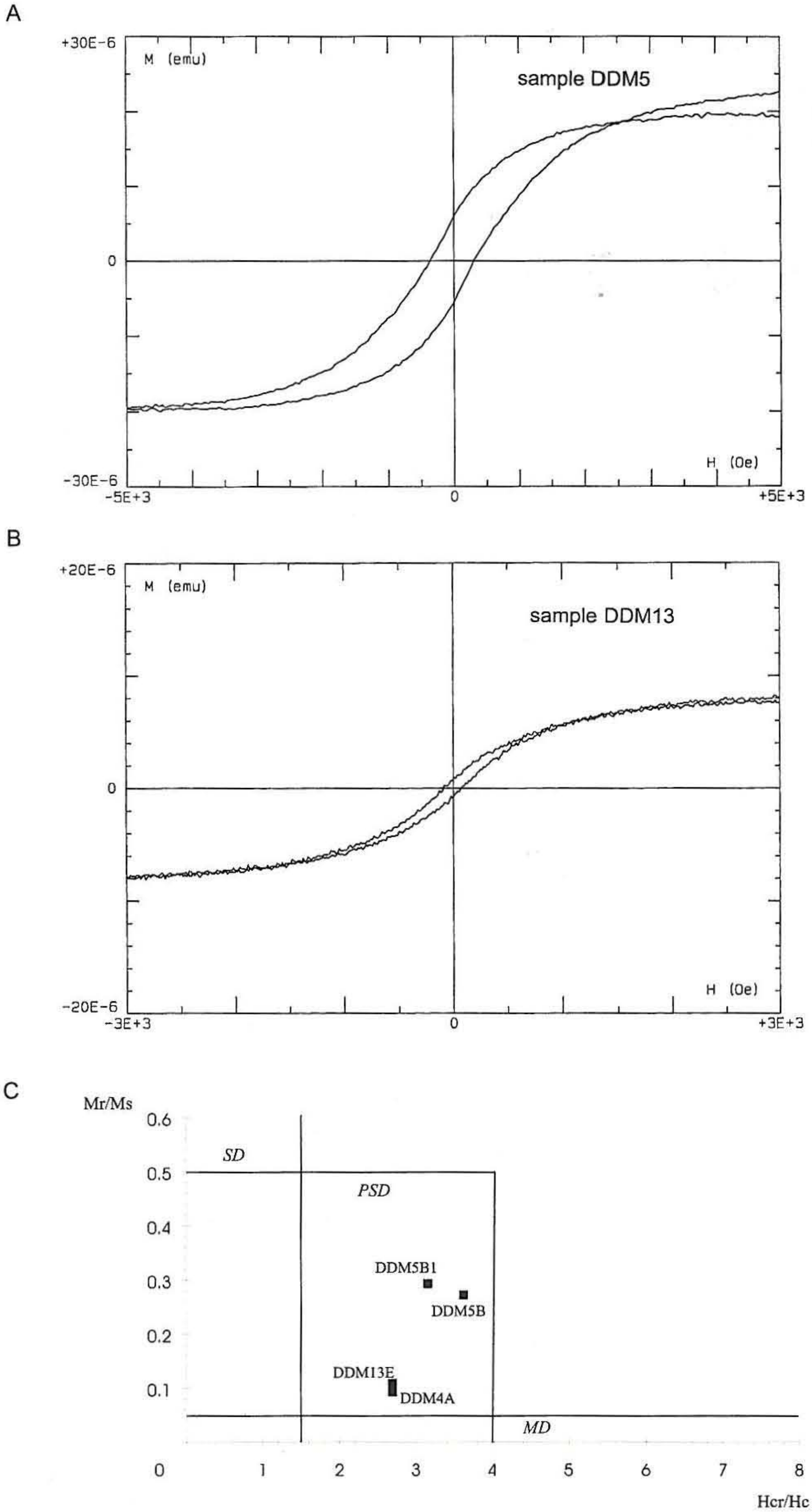


Fig. 49. Hysteresis loops from the locality DDM (A, B); C — hysteresis ratios plotted on a Day *et al.* (1977) diagram

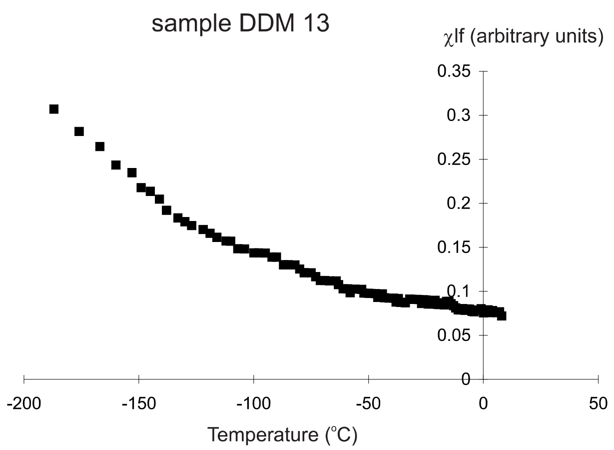


Fig. 50. Low temperature susceptibility measurements (sample DDM13 — Berriasian limestones)

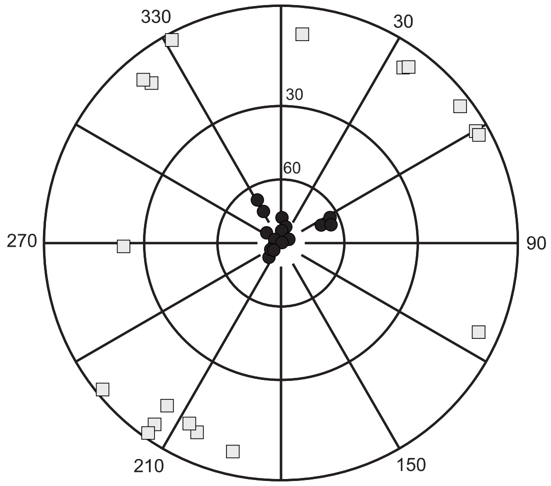


Fig. 51. Lower hemisphere projection of the maximum (squares) and minimum (dots) susceptibility axes at the locality DDM (after tectonic correction)

nellids. Detrital quartz is absent. Rare opaque minerals are distributed in the rock matrix. These are pyrite and products of its alteration: unidentified iron oxides or hydroxides. Calcite veins do not contain opaque grains. Ferruginous substance is present in the stylolitic seams. SEM analysis did not enable more detailed identification of opaque minerals due to their very fine grain size (<2 μm).

One specimen from the sample DDM4 was subjected to IRM experiments (stepwise IRM acquisition and thermal demagnetization of the 3-axes IRM). Only low corecivity magnetic minerals are present in the sample. IRM acquisition curve indicates 95% saturation in the field 200 mT (Fig. 47A). Thermal demagnetization of the soft fraction of the IRM reveals two magnetic phases: with maximum unblocking temperature between 400 and 500°C and between 550 and 585°C (Fig. 47B). This indicates the presence of magnetite. Unblocking temperatures lower than 500°C might point to maghemite or titanomagnetite. Hard and intermediate coercivity fractions contribute only little to the IRM. Their maximum unblocking temperatures are not higher than 600°C thus hematite is absent. Magnetic susceptibility increases during heating above 400°C (Fig. 47C) or above 450°C, probably due to varying content of iron sulphides in hand samples.

One rock chip from the sample DDM5 was subjected to thermomagnetic analysis. Five thermomagnetic runs were applied one after another, each time to a higher maximum temperature of 250, 350, 450, 620 and 660°C respectively. First three runs are not fully reversible: cooling curves reveal slightly lower intensities than heating curves (Figs. 48A–C). This indicates the alteration of a magnetic carrier between 20 and 450°C. This might be interpreted as transformation of maghemite to hematite. Between 450 and 620°C a new magnetic mineral originates (cooling curve runs "above" the heating curve — Fig. 48D). The shape of the cooling curve points to presence of magnetite. Curie temperatures characteristic for titanomagnetites are not observed. Between 620 and 660°C the creation of new magnetic fraction proceeds (Fig. 48E). Final products of alterations are magnetite and minor amounts of hematite as indicates the cooling curve in the Figure 48E. The detailed scenario of alteration processes during thermal treatment is not possible to reconstruct at the present stage of studies. There are considerable differences between the transformations in the sample DDM5 and sample STR3f (typical pyrite bearing limestone) described in the section 4.7.2. Pyrite occurs only sporadically in the DDM limestones. It can not be excluded that new magnetite originates due to breakdown of iron bearing clay minerals (e.g. illite and smectite, Van Velzen, 1992).

Magnetic hysteresis measurements were performed for four rock chips from the samples DDM4, DDM5, DDM13 and DDM15 (Fig. 49A and 49B). Hysteresis parameters fall into the PSD field of magnetite grains. However the Mr/Ms ratios between hand samples are variegated (Fig. 49C).

Susceptibility values are between 20 and 100 x 10^{-6} SI. It is dominated by paramagnetic minerals, what is evidenced from increase of susceptibility value in the liquid nitrogen temperature (Fig. 50).

Magnetic fabric is very well developed. K3 axes cluster close to the bedding pole indicating bedding parallel foliation. Weak SSW–NNE trending lineation is defined by the K1 axes (Fig. 51). Foliation is always stronger than lineation. The magnetic fabric indicates very weak degree of internal deformation of rocks. In all samples compactional magnetic fabric was preserved. Tectonic origin of lineation systems is very likely. It might point to dominant NW–SE directed compression.

4.5.3. DEMAGNETIZATION RESULTS

NRM intensities ranged between 0.376 and 11.1 x 10^{-4} A/m (mean 2.99 x 10^{-4} A/m). Pilot specimens were demagnetized with the AF and thermal method. AF was not effective in isolating of the characteristic components due to overlapping of the coercivity spectra. Thermal treatment was used for demagnetization of the collection. Two types of behav-

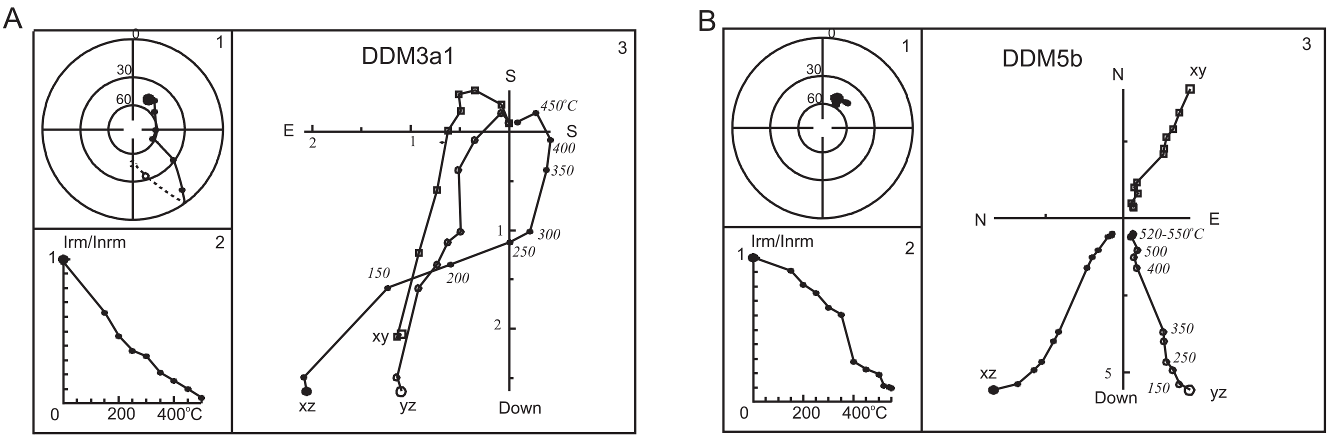


Fig. 52. Thermal demagnetization of the Berriasian limestone from the locality DDM (after tectonic correction)

A — specimen with reversed polarity component above 500°C; **B** — specimen with normal polarity component above 500°C; explanations as in the Figure 25

iour during thermal demagnetization were observed (Fig. 52A and B). In most samples three characteristic components occurred. A low stability component was demagnetized between 150 and 250°C (Fig. 52A). Its direction in the present coordinates is similar to the present day geomagnetic field direction. The second component DDM1 was demagnetized between 300 and 400°C (Tab. 12; Fig. 52A). The third component of reversed polarity DDM2r appeared between 400 and 500° C (Fig. 52A; Tab. 13 and 14). After 500°C the demagnetization paths became chaotic due to susceptibility increase (see Fig. 47C). Different behaviour of the NRM was observed in the samples DDM5 and DDM15. The NRM vector was stable up to 400°C (Fig. 52B) in these samples. The component DDM1 constituted the major part of the characteristic magnetization and the reversed polarity component was not observed. Above 400°C the NRM vector slightly changed its position. The presence of the component DDM2n of normal polarity is inferred (Tab. 13 and 14; Fig. 47). It was stable up to 550°C. It could be seen from the Fig. 47 and Tab. 13 that it is almost antiparallel to the component DDM2r (reversal test is performed in the section 5.3.2)

Table 12

Sample mean directions of the component DDM1

Sample	D	I	Dc	Ic	α_{95}	k	n
DDM3	193	59	48	61	16.1	33.5	4
DDM5	197	83	23	53	8	132.8	4
DDM6	265	68	21	64	13.8	44.9	4
DDM15	250	78	20	52	8.4	84.5	5

	D	I	α_{95}	k	Dc	Ic	α_{95}	k	N
Mean	225	75	17.7	27.5	27	58	10.2	82.2	4

Explanations as in the Table 7

The component DDM1 occurs in 4 samples and can be calculated from the fitted lines. It is projected in the Figure 53 (upper diagram) before and after tectonic correction.

DDM2 is presented in Figure 53 (lower diagram) as NRM vectors at specific demagnetization level (between 400 and 520°C) for each hand sample (see Tab. 13).

Table 13

Sample mean directions of the component DDM2

Sample	D	I	Dc	Ic	α_{95}	k	n	Remarks
DDM3	86	–64	180	–43	27.3	12.1	4	NRM at 500°C
DDM4	213	–69	193	–33	11.3	66.8	4	NRM at 500°C
DDM5	94	80	35	42	12.7	52.7	4	NRM at 500°C
DDM6	183	–67	211	–30	6.9	179.8	4	NRM at 400°C
DDM13	91	–74	182	–29	15.2	20.2	6	NRM at 520°C
DDM15	280	83	23	45	6.3	210.6	4	NRM at 520°C

Explanations as in the Table 7

Table 14

Mean directions of the normal and reversed component DDM2 (calculated from hand samples)

Comp.	D	I	α_{95}	k	Dc	Ic	α_{95}	k	N
DDM_{2n}	80	88	–	–	29	44	–	–	2
DDM_{2r}	143	–77	24	15.5	192	–35	15.5	38	4

Explanations as in the Table 7

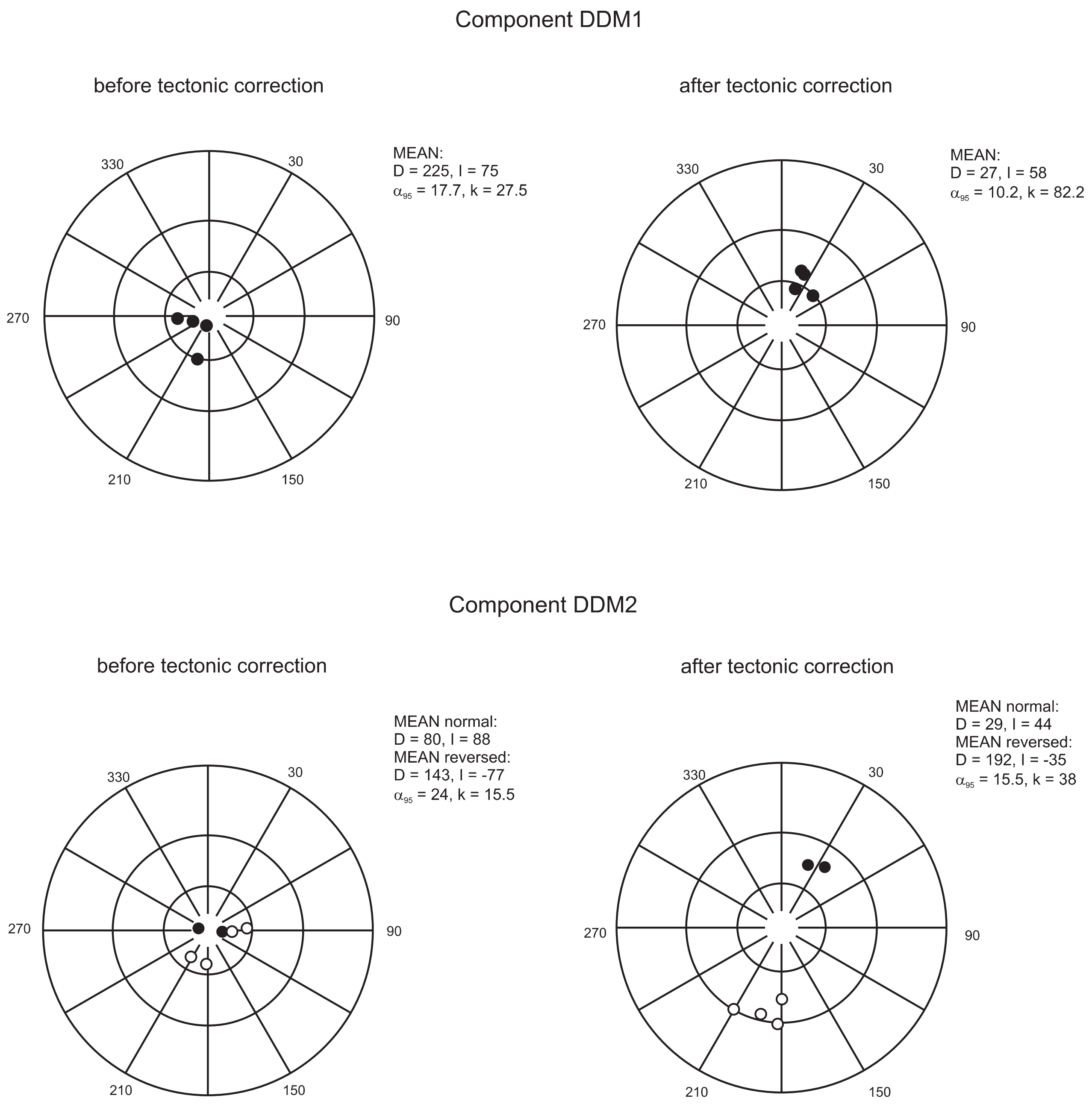


Fig. 53. Stereographic projection of the sample mean components at the locality DDM

4.6. SARNIA SKAŁA

4.6.1. LOCALITY DESCRIPTION

Three hand samples of dark Rhaetian limestones were taken along the tourist path from the Czerwona Przełęcz pass to the Sarnia Skała (SK) summit. The locality is situated within the minute tectonic slice which belongs to the Suchy Wierch unit (Iwanow, 1965). It is a part of highly complicated Czerwona Przełęcz tectonic zone where several tectonic slices occur between the main body of the Suchy Wierch unit and the higher Mała Świnica unit (Kibitlewski, 1972). Samples were taken from the section described by Gaździcki (1974). The strata dip 66° to the N (Tab. 6).

4.6.2. PETROGRAPHY AND ROCK MAGNETISM

Thin section was prepared from the sample SK4. The rocks are peloidal-organodetrital grainstones indicating shallow water high energy environment. Detrital minerals were not observed. Among non-transparent minerals only framboidal or cubic pyrite grains could be identified. They reside in between bioclasts or fill the pore space within. Enrichment in pyrite is observed within more intensely micritized bioclasts.

SEM analysis did not provide additional informations about the composition of opaque minerals. Titanium bearing minerals were not observed, what confirms the limestone are devoid of any detrital material.

Stepwise acquisition of the IRM and thermal demagnetization of the 3-axes IRM were performed for two specimens from samples SK4 and SK5. The results are similar for both samples. IRM acquisition curve indicates to predominance of low coercivity minerals (Fig. 54A), however high coercivity fraction is also present. Thermal demagnetization of the 3-axes IRM reveals that maximum unblocking temperatures of the soft and intermediate component do not exceed 550°C (Fig. 54B). This properties point to presence of magnetite. Hard component is persistent up to 650°C but its intensity is relatively low. This is the evidence for presence of hematite. Magnetic susceptibility increases sharply after heating above 350°C (Fig. 54C). This alteration is commonly attributed to transformation of pyrite to magnetite (Van Velzen, 1992).

Thermomagnetic analysis was carried out for a rock chip from the sample SK5 (Fig. 55A). The magnetization decays smoothly up to 440°. Above 440°C a new magnetic mineral originates which is demagnetized between 480 and 590°C. It is most likely magnetite that originates after breakdown of pyrite, similarly as in the locality RT (section 4.3.2). The production of secondary magnetite is confirmed by the shape of the cooling curve. It should be noted that magnetite is not oxidating to hematite during heating, unlike in the sample from the locality RT (section 4.3.2).

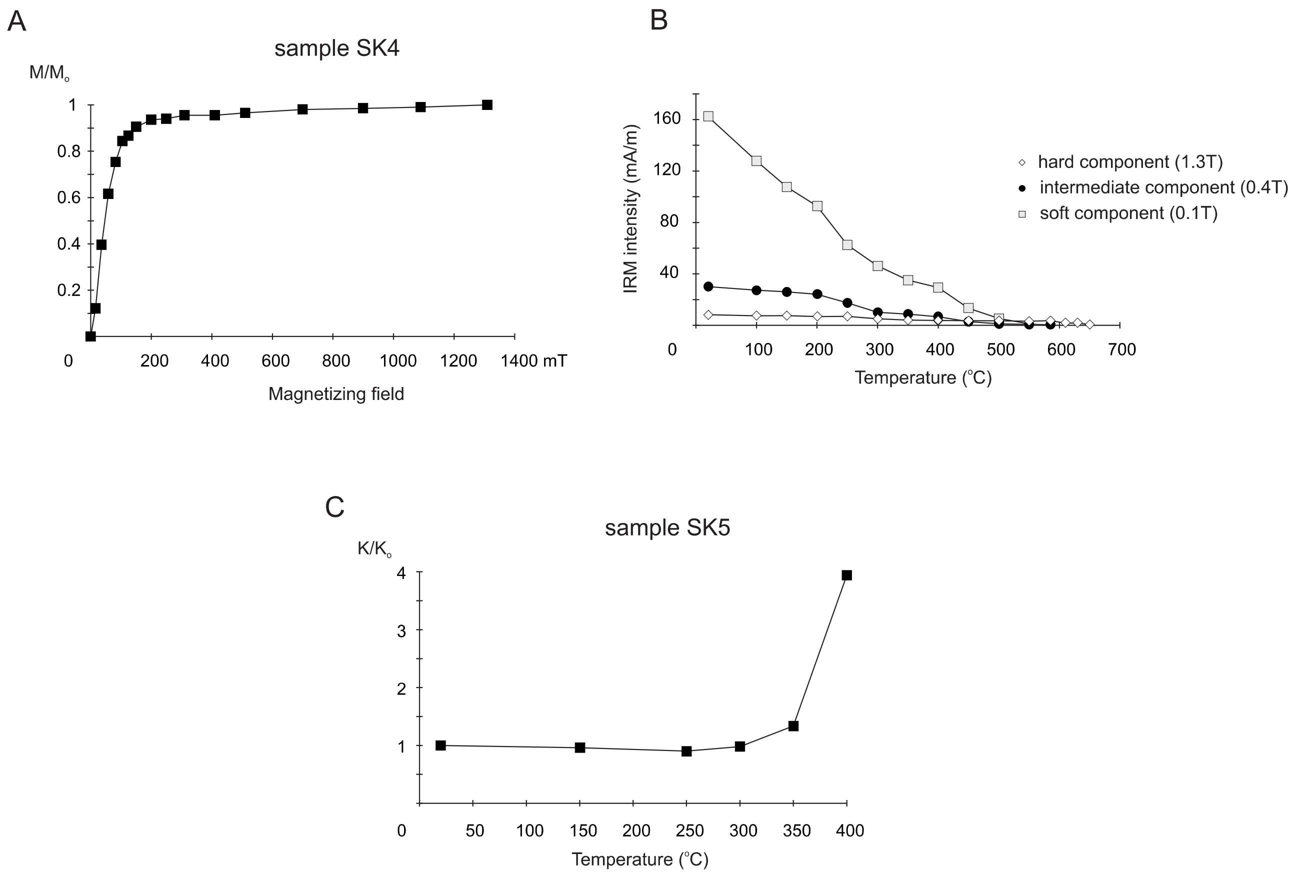


Fig. 54. Rock magnetic properties of the Rhaetian limestones from the locality SK

A — IRM acquisition curve; **B** — thermal demagnetization of the 3-axes IRM acquired in the fields 0.1T, 0.4T and 1.3T; **C** — susceptibility changes during thermal treatment

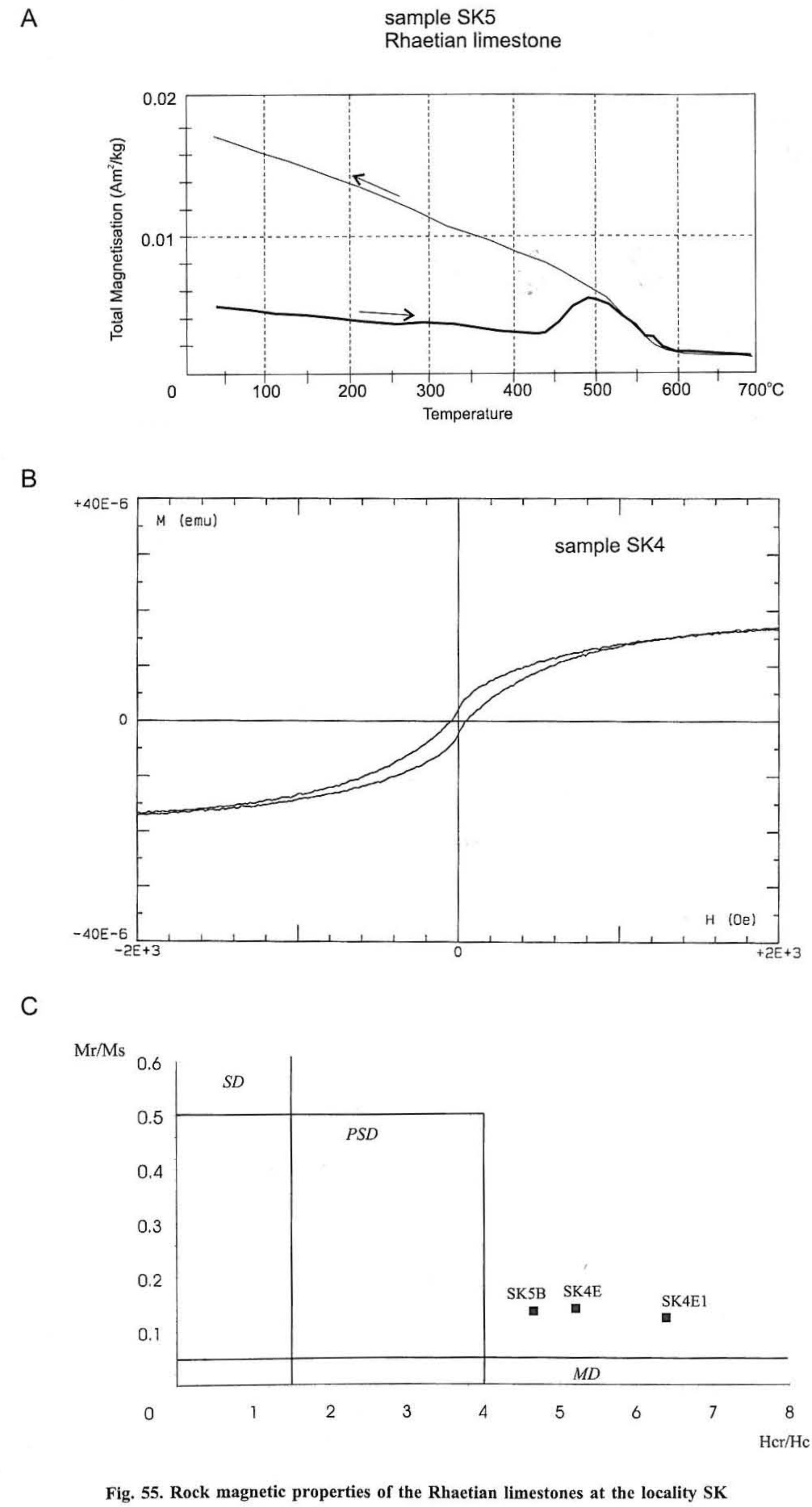


Fig. 55. Rock magnetic properties of the Rhaetian limestones at the locality SK

A — thermomagnetic analysis; bold line — heating curve, thin line — cooling curve; **B** — hysteresis loop; **C** — hysteresis ratios plotted on a Day *et al.* (1977) diagram

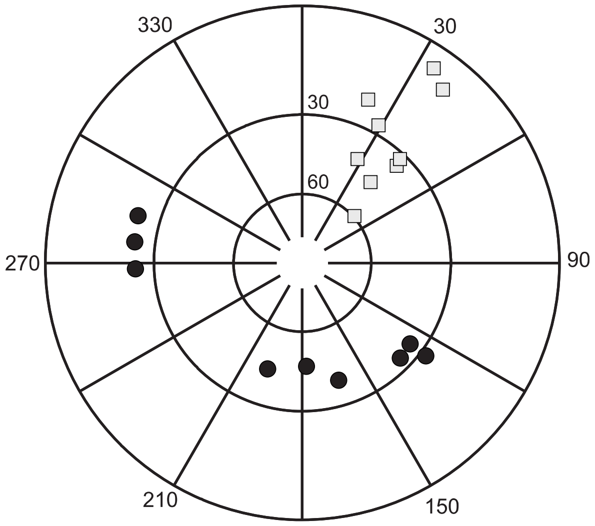


Fig. 56. Lower hemisphere projection of the maximum (squares) and minimum (dots) susceptibility axes at the locality SK (after tectonic correction)

Magnetic hysteresis measurements were performed for the samples SK4 (two rock chips) and SK5 (one rock chip). The hysteresis are characteristically wasp-waisted (Fig. 55B). The hysteresis ratios fall outside the PSD field with abnormally high Hcr/Hc values (Fig. 55C). This may be caused by presence of hematite. The wasp-waisted hysteresis and increased values of Hcr/Hc ratio were reported from magnetite bearing limestones with hematite admixture by Channell and McCabe (1994).

Mean susceptibility values were between 60 and 160 x 10^{-6} SI. Primary magnetic fabric has not been preserved in the locality. K1 axes reveal the NNE–SSW lineation. K3 axes are streaked along the plane roughly perpendicular to the direction of lineation (Fig. 56). The magnetic fabric might be tentatively interpreted as of tectonic origin.

4.6.3. DEMAGNETIZATION RESULTS

NRM intensities were between 6.7 and 33.5 x 10^{-4} A/m (mean 16.85 x 10^{-4} A/m). Pilot specimens were demagnetized using thermal demagnetization, AF demagnetization and the combination of both (Fig. 57A and B). The latter procedure brought the best results: thermal demagnetization up to 250°C and then AF up to 80–90 mT. In the temperatures

Table 15

Sample mean directions of component SK

Sample	D	I	Dc	Ic	α_{95}	k	n
SK3	118	64	28	33	6.9	95.9	6
SK4	76	71	19	18	7.9	94.3	5
SK5	121	71	19	33	7.9	43.7	9

	D	I	α_{95}	k	Dc	Ic	α_{95}	k	N
Mean	107	70	14.6	72.1	22	28	14.6	72.1	3

Explanations as in the Table 7

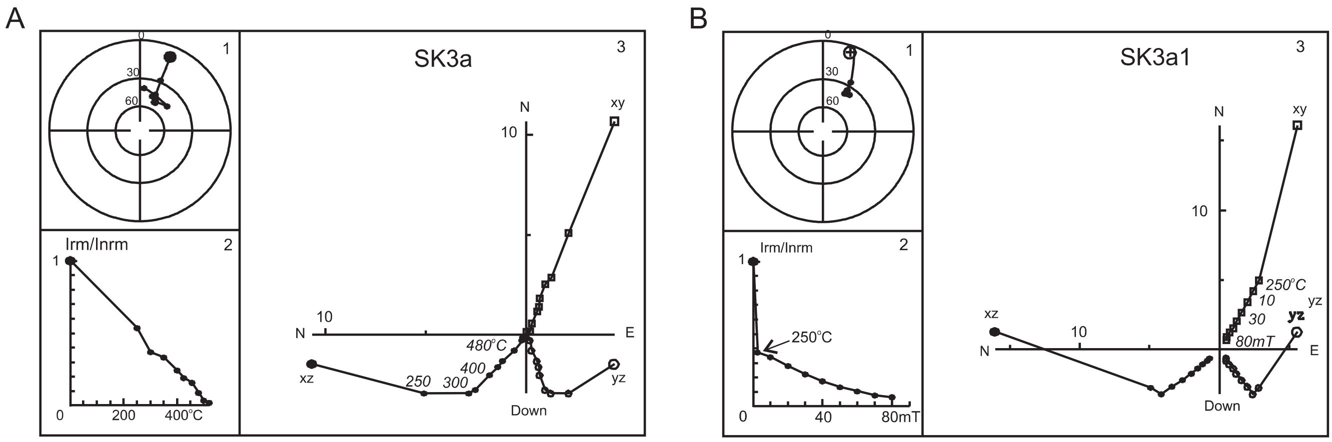


Fig. 57. Thermal (A) and mixed thermal + AF demagnetization (B) of the Rhaetian limestone from the locality SK (after tectonic correction)

Explanations as in the Figure 25

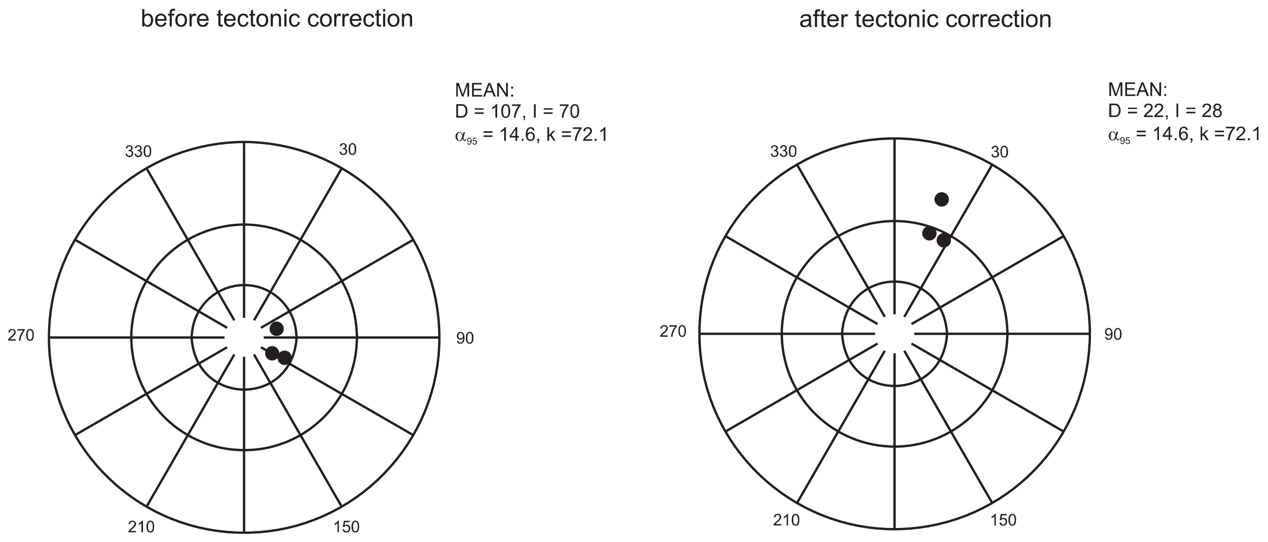


Fig. 58. Stereographic projection of the sample mean component SK

250–300°C the low stability component was removed. The second component SK was stable till the end of demagnetization process i.e. up to 80 mT and 480°C (Fig. 57A and B). This component is presented in Tab. 15 and Fig. 58. Before tectonic correction the declinations are directed towards ESE with inclination ca. 70°. After tectonic correction the declination changes to NNE and inclination is ca. 30°.

4.7. DOLINA STRĄŻYSKA

4.7.1. LOCALITY DESCRIPTION

Four hand samples of dark limestones were taken from the outcrop in the lowermost part of the Strążyska Valley (STR) on its eastern side, by the stream. The locality was described by Błaszyk and Gaździcki (1982). The limestones are of Hettangian age (Kopieniec Formation) and they occur as intercalations in mudstones. Each sample was taken from separate limestone bed. The extensive pyrite mineralisation occurs in this locality. The strata are overturned dipping steeply to the south (Tab. 6).

4.7.2. PETROGRAPHY AND ROCK MAGNETISM

Two thin sections were prepared from the samples STR2 and STR3. The rocks are organodetrital wackestones/mudstones (shallow water, high energy environment) with numerous detrital quartz grains. Very well developed pyrite crystals and framboids could be observed. Pyrite grains occur between bioclasts and quartz grains. Calcite veins do not contain opaque minerals. Brown ferruginous substance occurs in the matrix and stylolitic seams.

SEM analysis was carried out for the sample STR2. Abundance of pyrite grains was confirmed. Detrital rutiles and zircons were identified. Single grains containing Ti and Fe are most probably detrital ilmenites.

Stepwise IRM acquisition and thermal demagnetization of the 3-axes IRM were performed for two specimens from the sample STR1. The experiments gave contrasting results. Low coercivity magnetic minerals prevailed in the specimen STR1a. Almost complete saturation was achieved in 250 mT (Fig. 59A). Single magnetic phase with maximum unblocking temperature 500°C was present (Fig. 59B). All these features indicated presence of magnetite or titanomagnetite. The presence of titanomagnetite can not be excluded in this case because the rock contains variable assemblage of detrital grains. Specimen STR1c1 contained significant amount of high coercivity mineral (Fig. 60A) — the saturation was not achieved in 1400 mT. High coercivity phase had maximum unblocking temperatures between 600 and 650°C which is characteristic for hematite (Fig. 60B). Maximum unblocking temperatures of soft and intermediate fractions are about 575°C. This value confirms the presence of magnetite. Additionally mineral of unblocking temperature 350–400°C is present which might be maghemite or titanomagnetite.

Magnetic susceptibility increases sharply during thermal treatment after 400°C (Fig. 59C). This is very likely the alteration of pyrite to magnetite (Van Velzen, 1992).

One thermomagnetic analysis was performed for the rock chip from the sample STR3 (Fig. 61). Five thermomagnetic runs were applied one after another, each time to a higher maxi-

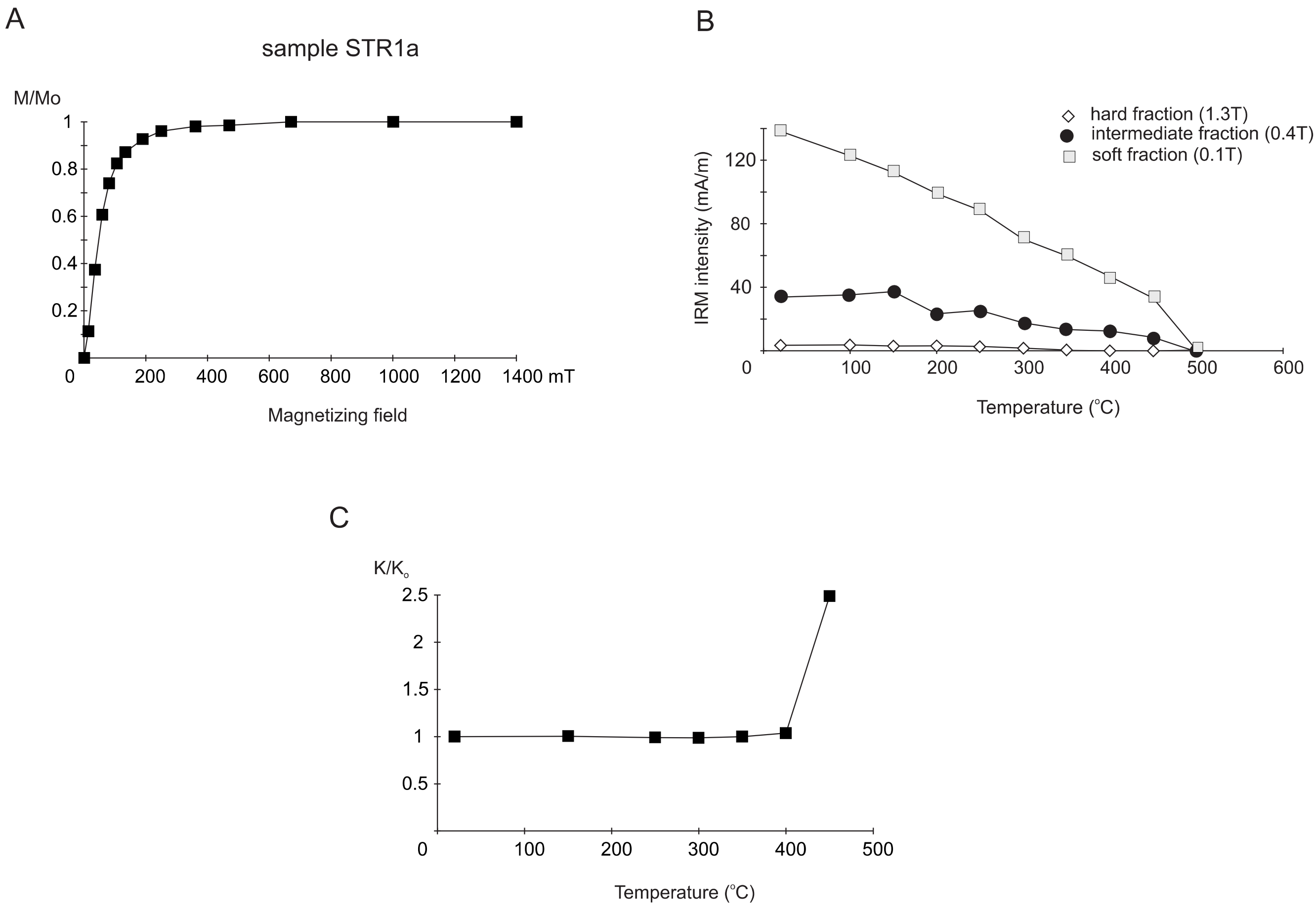


Fig. 59. Rock magnetic properties of the Hettangian sandy limestones from the locality STR: specimen containing magnetite

A — IRM acquisition curve; **B** — thermal demagnetization of the 3-axes IRM acquired in the fields 0.1T, 0.4T and 1.3T; **C** — susceptibility changes during thermal treatment

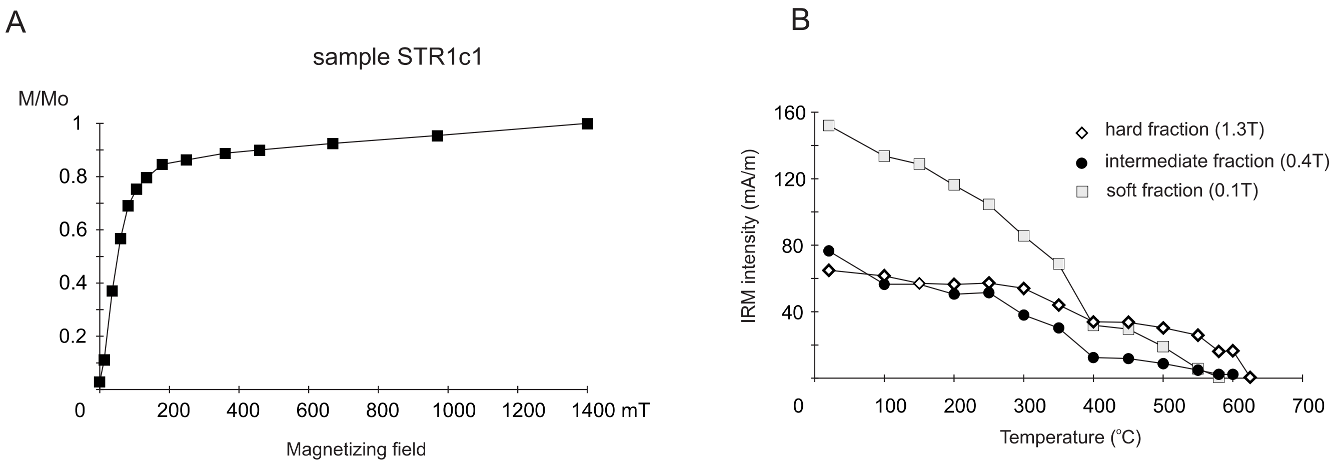


Fig. 60. Rock magnetic properties of the Hettangian sandy limestones from the locality STR: specimen containing magnetite and hematite

A — IRM acquisition curve; **B** — thermal demagnetization of the 3-axes IRM acquired in the fields 0.1T, 0.4T and 1.3T

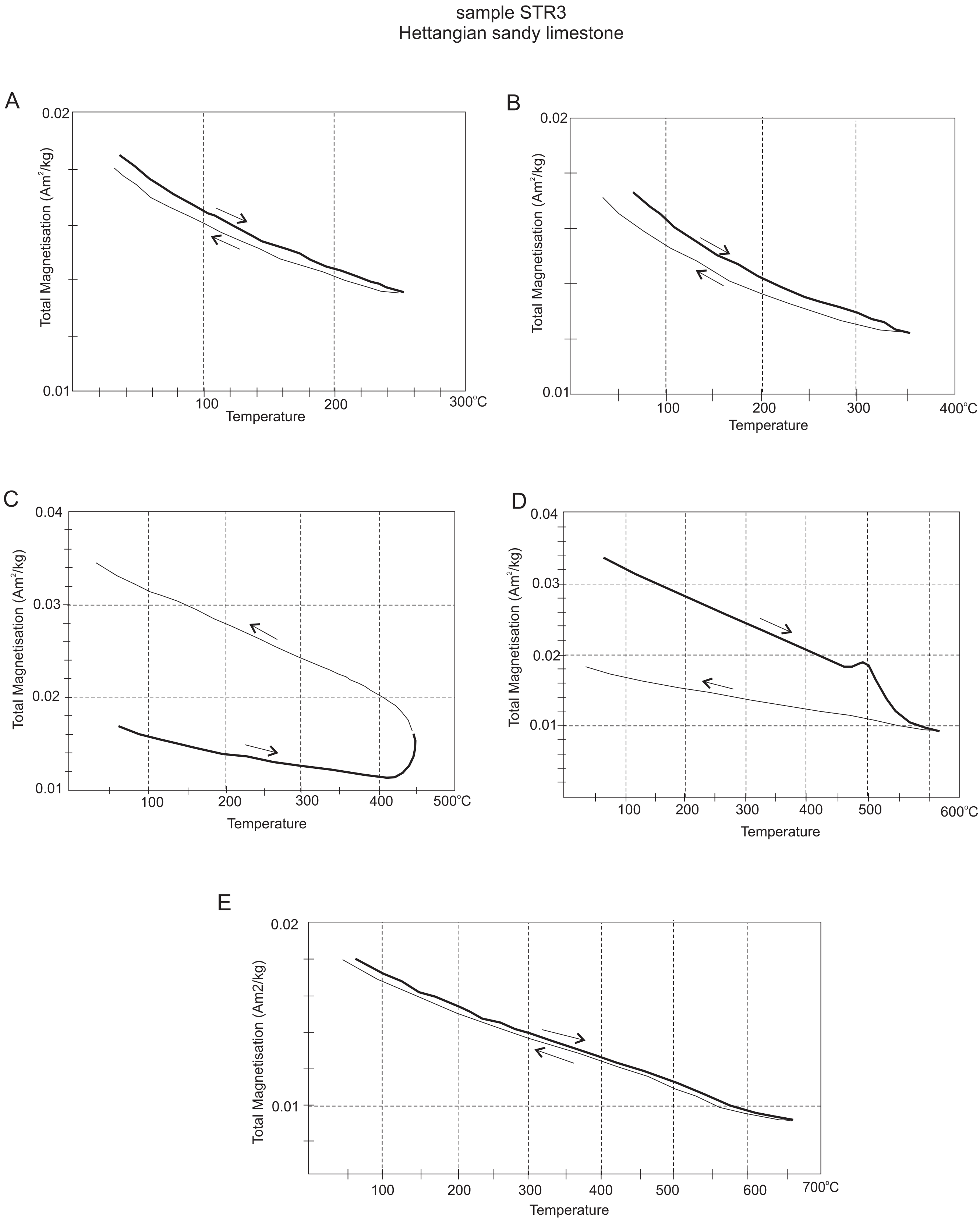


Fig. 61. Thermomagnetic analysis of the rock chip from the locality STR

A — between 20 and 250°C; **B** — between 20 and 350°C; **C** — between 20 and 450°C; **D** — between 20 and 620°C; **E** — between 20 and 660°C; bold line — heating curve, thin line — cooling curve

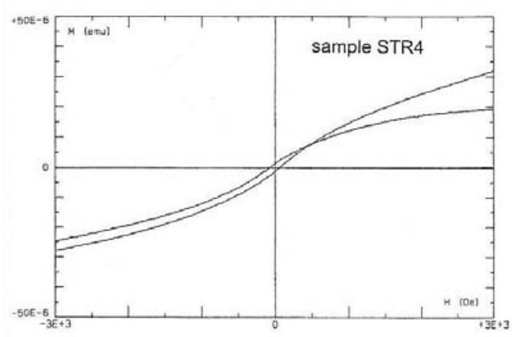


Fig. 62. Typical hysteresis curve of the rock chip
from the locality STR

Fig. 62. Typical hysteresis curve of the rock chip from the locality STR

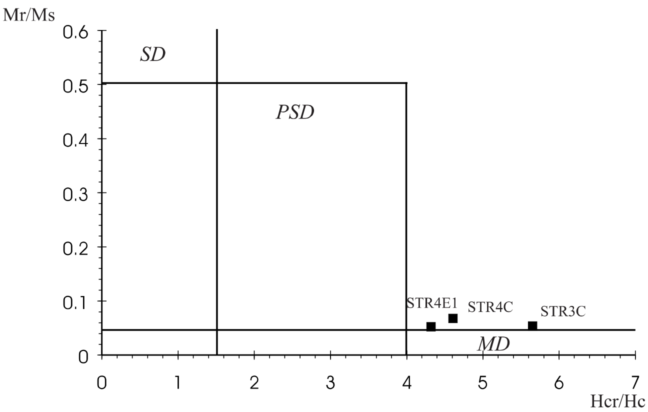


Fig. 63. Hysteresis ratios from the locality STR plotted on Day *et al.* (1977) diagram

mum temperature of 250, 350, 450, 620 and 660°C respectively. The first two runs indicate that between 20 and 350°C alterations occur which are manifested by decrease of total magnetization (Figs 61A and B). This might be caused by oxidation of maghemite to hematite. About 420°C new strongly magnetic mineral originates (Fig. 61C). The production of this mineral proceeds up to 495°C and afterwards the magnetization starts decaying (Fig. 61D). The transformations between 420 and 495°C might be interpreted as alteration of pyrite to magnetite. Above 495°C the total magnetization decreases due to approaching of magnetite Curie point but also due to physical removal (oxidation) of the magnetite, because the run in the Figure 61D is not reversible. The product of the oxidation is a mineral of Curie temperature above 650°C (hematite) what is visible in the last run (Fig. 61E). The entire cycle is almost identical to the model example of iron-sulphide bearing clay described by Mullender *et al*. (1993).

Magnetic hysteresis measurements were carried on for the samples STR3 (one chip) and STR4 (two chips). The hysteresis in all investigated samples is very weak (Fig. 62). The parameters Hcr/Hc and Mr/Ms fall very close to the MD range (Fig. 63).

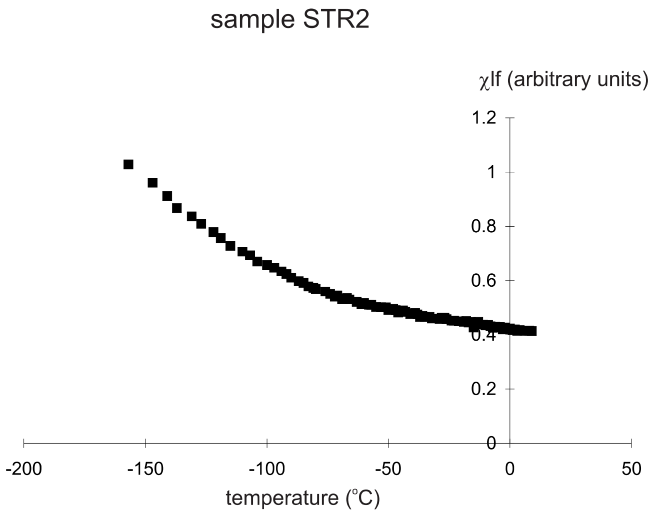


Fig. 64. Low temperature susceptibility measurements (sample STR2 — Hettangian sandy limestones)

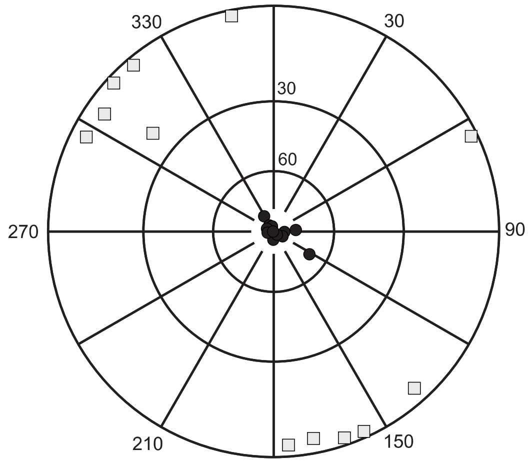


Fig. 65. Lower hemisphere projection of the maximum (squares) and minimum (dots) susceptibility axes at the locality STR (after tectonic correction)

Mean susceptibility values were quite high as for sedimentary rocks ranging from 100 to 500 x 10^{-6} SI. However low temperature investigations (Fig. 64) reveal that is based upon paramagnetic minerals, like in most other localities. Magnetic fabric is well developed and characteristic for weakly deformed sedimentary rocks. It is mostly planar with K3 axis perpendicular to the bedding plane (Fig. 65). NW–SE to NNW–SSE trending magnetic lineation is visible.

4.7.3. DEMAGNETIZATION RESULTS

NRM intensities were between 5.7 and 22.4 x 10^{-4} A/m (mean 14.1 x 10^{-4} A/m). Pilot specimens were demagnetized using AF, thermal and combined method (thermal demagnetization up to 250°C and then AF demagnetization). Demag-

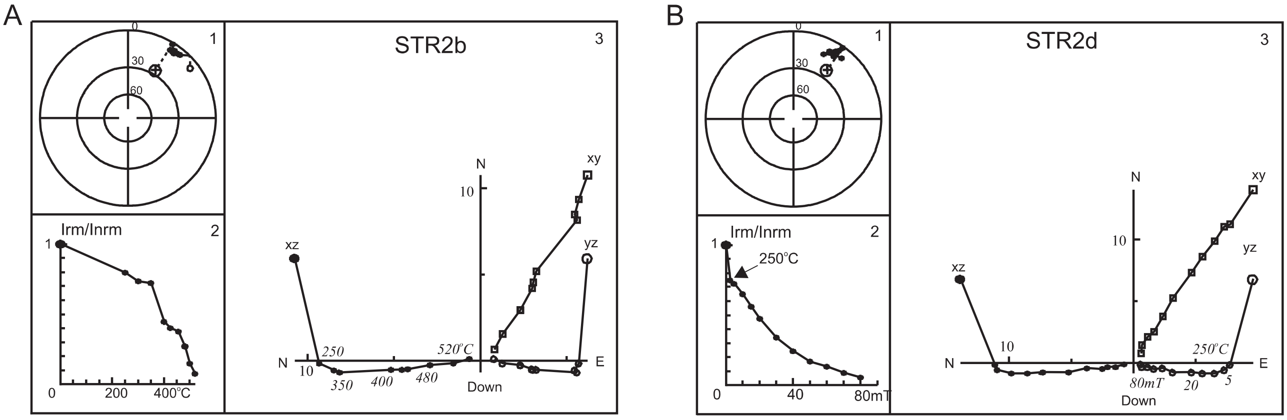


Fig. 66. Thermal (A) and mixed thermal + AF demagnetization (B) of the Hettangian sandy limestone from the locality STR (after tectonic correction)

Explanations as in the Figure 25

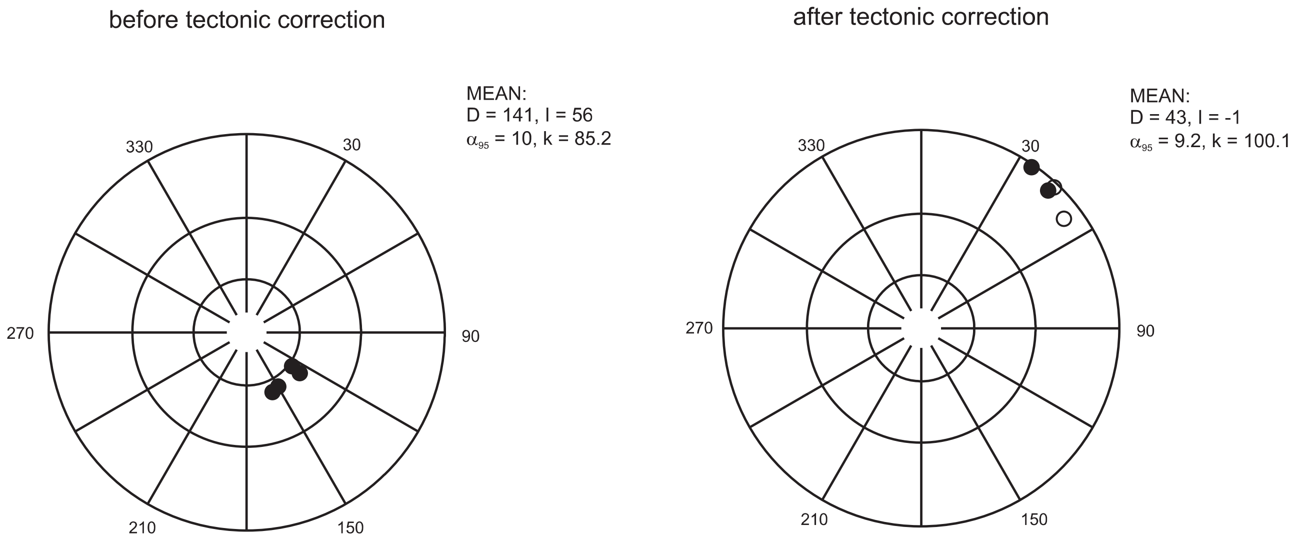


Fig. 67. Stereographic projection of the sample mean component STR

Table 16

Sample mean directions of the component STR

Sample	D	I	Dc	Ic	α_{95}	k	N
STR1	128	58	43	–2	7.6	148.7	4
STR2	157	54	35	0	6.5	199.4	4
STR3	151	55	43	3	11.8	32.9	6
STR4	129	53	52	–5	31.5	9.3	4

	D	I	α_{95}	k	Dc	Ic	α_{95}	k	N
Mean	141	56	10	85.2	43	–1	9.2	100.1	4

Explanations as in the Table 7

netization path was not very smooth when AF was applied. Thermal and combined methods gave better results (Fig. 66A, B) and the latter was applied to bulk part of collection. Low stability magnetization was removed up to 250°C and 20 mT. The second component STR was stabilised between 250 and 520°C and between 5 and 80 mT. It is well clustered with moderate downward inclination in the second quadrant of the stereonet. However quite poor statistics is observed within hand samples STR3 and STR4. After tectonic correction the declinations become north-easterly and inclinations change to equatorial (Fig. 67; Tab. 16). Both components are carried by magnetite.

4.8. WIELKIE KORYCISKA

4.8.1. LOCALITY DESCRIPTION

Five hand samples were taken from “Reifling” type limestone in the gully at the western termination of the Wielkie Koryciska Valley (KOR). Direct contact between “Reifling” limestone and “Partnach” marls crops out there. The locality is very famous due to rich macro- and microfauna which indicate Late Anisian/Early Ladinian age of rocks: ammonites, nautiloids and daonellas (Kotański, 1973), amphibians (Kotański, 1996), conodonts (Zawidzka, 1972) and foraminifers (Gaździcki, Zawidzka, 1973). The place was often described in geological guide-books (i.e. Kotański, 1971; Kotański, Iwanow, 1997). The strata dip 40° to N (Tab. 6).

4.8.2. PETROGRAPHY AND ROCK MAGNETISM

Thin sections were prepared from samples KOR1 and KOR3. The rock is biodetrital wackestones with slightly nodular appearance. Pyrite is the most abundant non-transparent mineral. It is confined to bioclasts fillings and occurs also in the rock matrix and stylolitic seams. Post-pyrite unidentified iron oxides or hydroxides are observed. Calcite veins do not contain opaque minerals.

SEM analysis was carried out for the sample KOR3. Only pyrite grains could be identified. Detrital minerals containing Ti are apparently absent.

Stepwise IRM acquisition and 3-axes thermal demagnetization was performed for one specimen of sample KOR4. Again low coercivity minerals prevail in this locality. The sample is magnetically saturated in the field of 300 mT (Fig. 68A). 3-axes demagnetization of the IRM (Fig. 68B) confirms the predomination of the soft magnetic fraction. The soft fraction curve reveals maximum unblocking temperature 500°C. Intermediate and hard components are completely demagnetized at 500°C and higher unblocking temperatures are not observed. Easy saturation and unblocking temperatures indicate magnetite or maghemite as the main magnetic mineral. The presence of titanomagnetite is not likely as the detrital material is apparently absent in the rock. Magnetic susceptibility during thermal treatment is stable up to 400°C and then rises sharply (Fig. 68C). This may account for presence of iron sulphides (Van

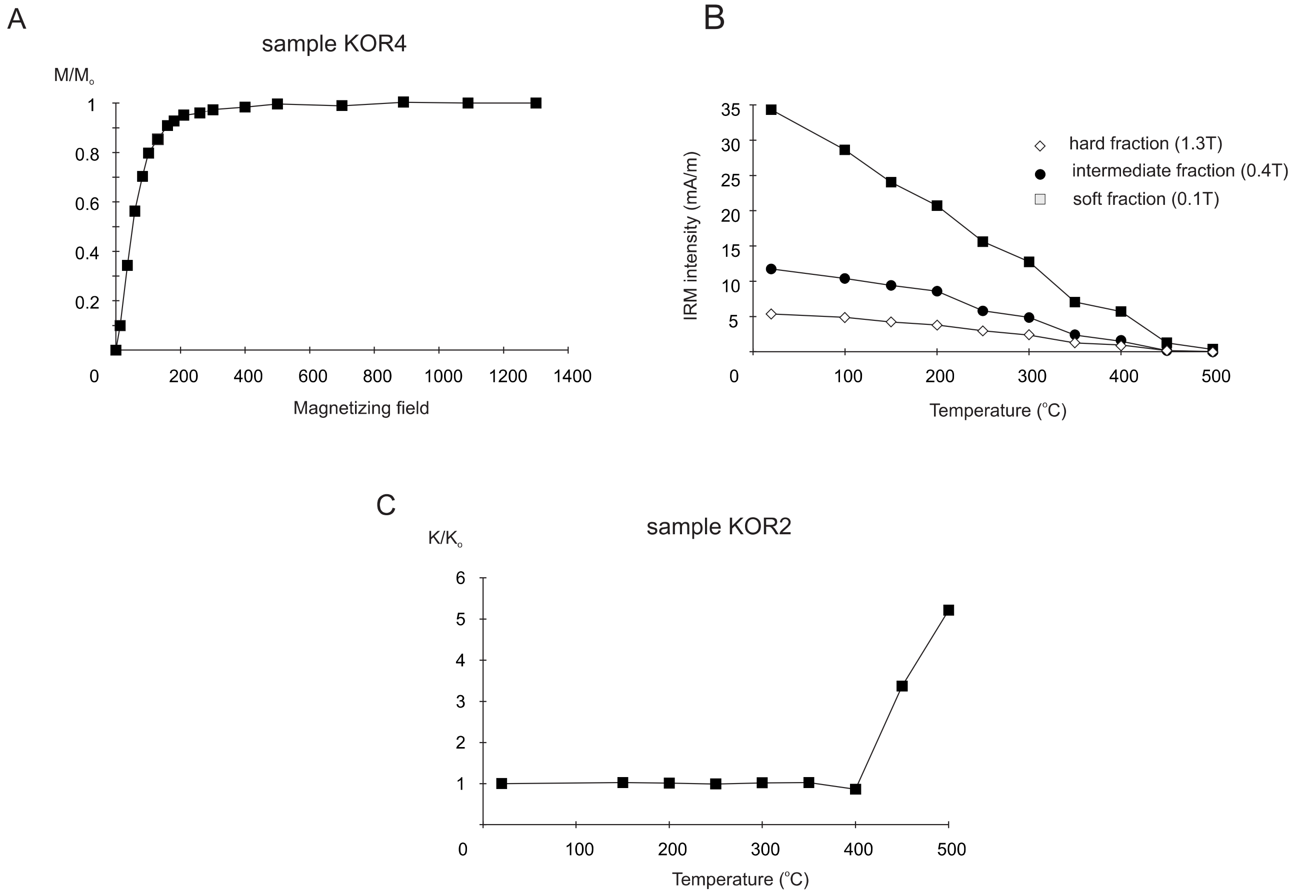


Fig. 68. Rock magnetic properties of the Anisian–Ladinian limestones from the locality KOR

A — IRM acquisition curve; **B** — thermal demagnetization of the 3-axes IRM acquired in the fields 0.1T, 0.4T and 1.3T; **C** — susceptibility changes during thermal treatment

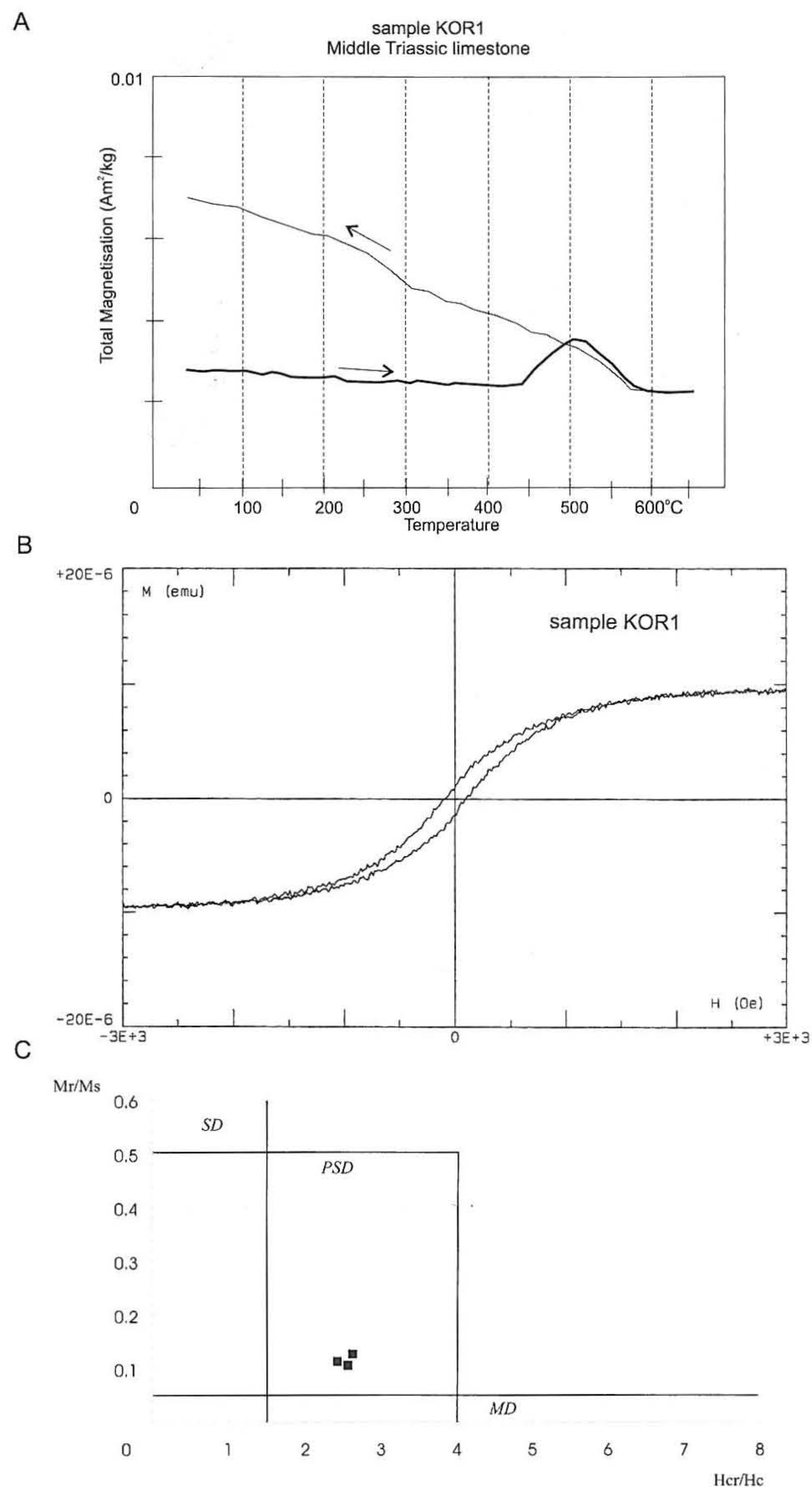


Fig. 69. Rock magnetic properties of the Anisian–Ladinian limestones at the locality KOR

A — thermomagnetic analysis; bold line — heating curve, thin line — cooling curve; **B** — hysteresis loop; **C** — hysteresis ratios plotted on a Day *et al.* (1977) diagram

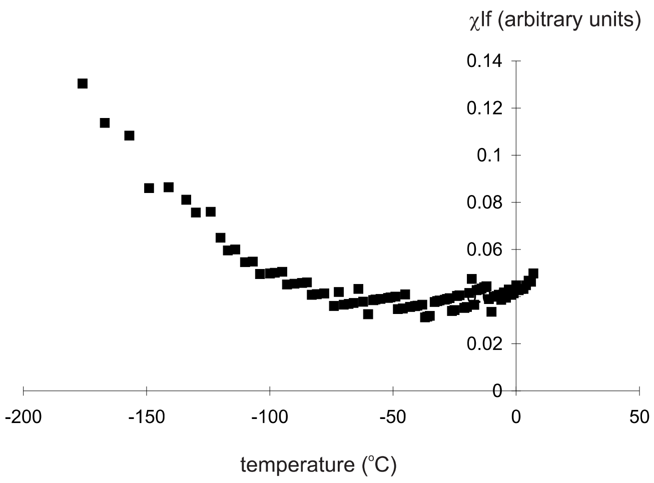


Fig. 70. Low temperature susceptibility measurements (sample KOR1 — Anisian–Ladinian limestones)

Velzen, 1992) and agrees with the presence of pyrite identified by SEM.

Thermomagnetic analysis was performed for the sample KOR1 (Fig. 69A). The magnetization decays smoothly up to 440°C. There is no indication of pyrrhotite with Curie temperature between 300 and 350°C. Above 440°C a new magnetic mineral originates in the sample with Curie temperature 585°C (magnetite). It is confirmed by the cooling curve which reveals much higher magnetizations than the heating curve. Chemical alterations above 440°C must be attributed to transformation of pyrite to magnetite (Mullender *et al*., 1993).

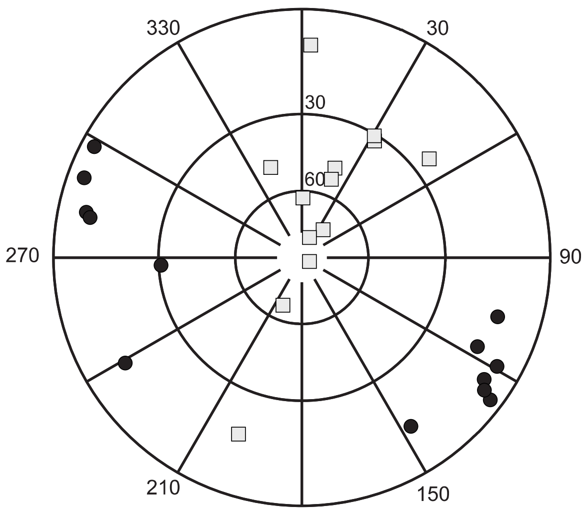


Fig. 71. Lower hemisphere projection of the maximum (squares) and minimum (dots) susceptibility axes at the locality KOR (after tectonic correction)

Hysteresis properties were measured for two chips from sample KOR1 (Fig. 69B) and one chip from sample KOR5. Hysteresis ratios Mr/Ms and Hcr/Hc point to PSD magnetite with low Mr/Ms ratio (Fig. 69C).

Susceptibility values were very low and they ranged between 5 and 30 x 10^{-6} SI units. Low temperature susceptibility studies indicate predominance of paramagnetic fraction (Fig. 70). Primary magnetic fabric with bedding parallel foliation was not preserved in this locality (Fig. 71). K3 axes are oriented WNW–ESE within the bedding plane, while K1 axes are streaked in the NNE–SSW direction.

4.8.3. DEMAGNETIZATION RESULTS

NRM intensities ranged between 1.17 x 10^{-4} A/m and 6.91 x 10^{-4} A/m (mean 4.71 x 10^{-4} A/m). Thermal demagnetization was applied to entire collection, because it was more

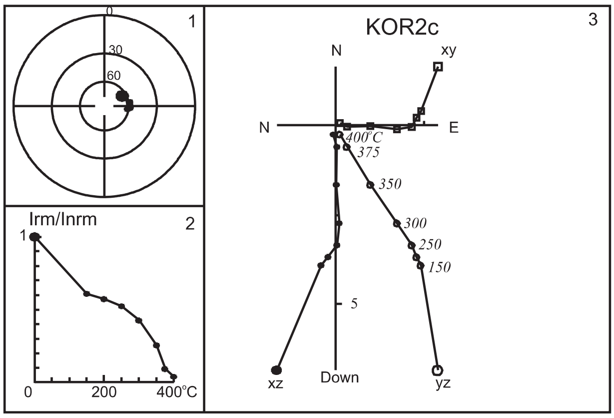


Fig. 72. Thermal demagnetization of the Anisian–Ladinian limestone from the locality KOR (after tectonic correction)

Explanations as in the Figure 25

Table 17

Sample mean directions of the component KOR

Sample	D	I	Dc	Ic	α_{95}	k	n
KOR1	145	39	89	68	14.1	42.9	4
KOR2	137	34	93	59	5.2	220.5	5
KOR3	140	45	93	51	11	70.8	4
KOR4	152	29	121	54	8.5	209.5	3
KOR5	146	43	95	63	12.4	55.6	4

	D	I	α_{95}	k	Dc	Ic	α_{95}	k	N
Mean	144	38	7.4	108.7	100	60	8.1	89.1	5

Explanations as in the Table 7

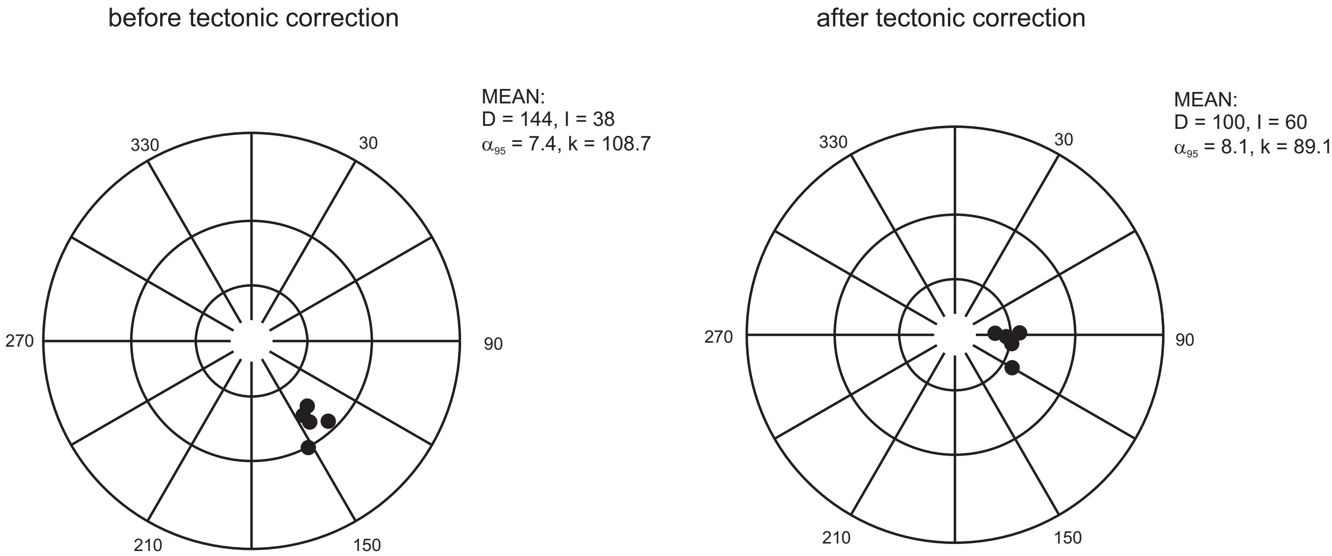


Fig. 73. Stereographic projection of the sample mean component KOR

effective than the AF demagnetization (Fig. 72) and the vector of magnetic remanence was more stable. Low stability component was removed up to 250°C. High stability component KOR was demagnetized between 250 and 400°C.

Before tectonic correction its declinations are south-eastern and inclinations amount to ca. 40°. After tectonic correction the declination become easterly and inclination values are about 60° (Tab. 17; Fig. 73).

5. INTERPRETATION AND DISCUSSION OF PALAEOMAGNETIC DATA

5.1. SUMMARY OF THE ROCK MAGNETIC PROPERTIES

Low coercivity minerals predominate in all studied localities. They are also carriers of the ChRM. Magnetite and maghemite are the most plausible candidates for the low coercivity magnetic fraction. Applied methods have not confirmed unambiguously the presence of pyrrhotite. Magnetite is identified by the Curie temperatures 585–600°C. Slightly elevated Curie temperatures are attributed to partial oxidation of magnetite grains during thermomagnetic experiment (Mullender *et al*., 1993) as for example in the localities SU (section 4.1.2) and KR (section 4.5.2). Unblocking temperatures lower than Curie point of magnetite: 500–585°C were frequently revealed during thermal demagnetization of the IRM. They might be attributed to the size variations of magnetite grains. Titanomagnetites might occur in the rocks with increased amount of detrital grains. This refers to the Hettangian sandy limestones from the locality STR where presence of Ti–Fe oxides was confirmed by SEM investigations. In all other localities titanomagnetites were not found. In the localities RT and KR influx of detrital material is manifested by presence of zircon and rutile which are paramagnetic minerals.

The presence of maghemite in some localities (especially SU and SI) is inferred from the behaviour of low field susceptibility during thermal treatment. Decrease of magnetic susceptibility above 400°C is usually attributed to the presence of maghemite which converts to hematite during heating (i.e. Opdyke, Channell, 1996; Marton *et al*., 1988; Tarling, Hrouda, 1993). Loose of magnetic remanence between 350 and 450°C is in agreement with the maghemite decomposition in this temperature range. It can not be excluded that very fine grained magnetite also contributes to this fraction. The occurrence of maghemite might be inferred in localities where unblocking temperatures of NRM or IRM were below 400°C (localities DDL, DDM and KOR). However as IRM experiments were not very numerous the presence of maghemite in these localities is accompanied by the question mark in the Table18.

Goethite is absent in all studied localities. Hematite occurs only locally. It is unambiguously identified by high saturation fields and unblocking temperatures 650°C and higher. It is an important magnetic carrier only in the Keuper dolomites of the RT locality (sample RT9). Subordinate amounts of hematite occur in some dark limestones (for example in the SK, STR and DDL localities, see Table 18).

Pyrite is very abundant in the dark limestones from localities RT, SK, STR, DDL and KOR (Table 18). It is identified by increase of susceptibility after 400°C and by characteristic shape of thermomagnetic curves (i.e. section 4.7.2) indicating

Table 18

Rock magnetic properties in the studied localities

Locality	Age and lithology	Magnetic minerals	Pyrite	Grain size from hysteresis ratio
SU	Ladinian dolomites	magnetite, maghemite	–	PSD
SI	Ladinian dolomites	magnetite, maghemite	–	PSD
RT	Keuper dolomites Rhaetian–Hettangian dark limestones	hematite magnetite	– +	PSD
SK	Rhaetian dark limestones	magnetite, minor hematite	+	outside expected fields (high Hcr/Hc)
STR	Hettangian dark sandy limestones	magnetite, titanomagnetite(?), hematite	+	outside expected fields (high Hcr/Hc) close to MD
DDL	Sinemurian–Pliensbachian dark limestones	magnetite, maghemite(?), minor hematite	+	PSD or outside expected fields (high Hcr/Hc)
KR	Tithonian marly limestones (greenish)	magnetite	– (?)	PSD
DDM	Berriasian pale limestones	magnetite, maghemite(?)	– (?)	PSD
KOR	Anisian–Ladinian dark limestones	magnetite, maghemite(?)	+	PSD

production of secondary magnetite during thermal treatment. It is also quite easy to identification in the reflected light microscopic investigations. Pyrite seems to be absent in the Middle Triassic dolomites (localities SU, SI), where it was almost completely oxidized to ferric oxides and hydroxides (section 4.1.2). Also pale Berriasian limestones of DDM locality and Tithonian marly limestones of KR locality contain just trace amounts of pyrite.

Origin of magnetic carriers can not be unambiguously evaluated at the present state of investigations. Occurrence of secondary (post-pyrite) magnetite and hematite is quite likely because pyrite is sometimes oxidized, as observed in ore microscope. More detailed discussion concerning the origin of magnetization is enclosed in the section 5.4.

5.2. SUMMARY OF THE SUSCEPTIBILITY AND AMS DATA

AMS studies revealed variegated magnetic fabric in the studied localities. The susceptibilities generally did not exceed 200 x 10^{-6} SI units thus contribution of paramagnetic minerals to the AMS is very likely (Rochette, 1987). This was confirmed by the susceptibility measurements in the liquid nitrogen temperature: in all but two localities the susceptibility is dominated by paramagnetic matrix. Susceptibility signal of the Middle Triassic dolomites in localities SU and SI is carried by SP magnetite.

The AMS pattern should be considered for each tectonic unit separately. The most extensive data set was obtained in the Bobrowiec unit. The new data from the western slopes of Chochołowska Valley confirm previous suggestions that upper part of the Bobrowiec unit is weakly deformed (Grabowski, 1996). The relict sedimentary ("primary") magnetic fabric with very uniform bedding parallel foliation and single lineation system was encountered in the Tithonian marly limestones at KR locality (section 4.5.2). Similar pattern of magnetic fabric was encountered also in the DDM locality (section 4.4.2) in the Berriasian limestones that directly overlie the sediments from KR. Here again mostly oblate shape of susceptibility ellipsoid and bedding parallel foliation indicate low degree of tectonic deformation. Also in the radiolarian limestones (Oxfordian) relict sedimentary fabric with weak tectonic overprint was described (Grabowski, *op. cit.*). In the Late Liassic spotty limestones at locality DDL the relicts of sedimentary fabric were preserved as well (section 4.3.2). The Keuper–Hettangian succession at RT locality (section 4.2.2) reveals quite anomalous, probably tectonically changed magnetic fabric.

Only single localities were studied from other tectonic units of the KriŽna and higher nappes and their AMS interpreatation is summarized in the Table 19 and in the chapter 4 for each locality separately.

A question must be posed if ChRMs could be disturbed by internal deformations of host rock manifested by the AMS ellipsoid? Examples of reorientation of the magnetic remanence due to local strain effect have been thoroughly described in palaeomagnetic studies of orogens (i.e. Kligfield *et al.*, 1983; Hirt *et al.*, 1986; Stamatakos, Kodama, 1991; Kirker, McClelland, 1997). In these investigations strain modified palaeomagnetic directions were accompanied by development of tectonic magnetic fabric reoriented within the cleavage plane (Kligfield *et al.*, 1983; Hirt *et al.*, 1986), kink bands (Kirker, McClelland, 1997) or showed deflection with stepwise increase of tectonic deformations (Vetter *et*

Table 19

Summary of the AMS data in the studied localities

Tectonic unit	Locality	Age and lithology	K_m (x 10^{-6} SI)	Magnetic fabric	Susceptibillity carrier
Suchy Wierch	SU	Upper Ladinian dolomites	20–160	deformational	SP magnetite
	SI	Ladinian dolomites	20–110	relict sedimentary	SP magnetite
Bobrowiec	RT	Keuper dolomites Rhaetian–Hettangian limestones	40–300	deformational	paramagnetic
	DDL	Sinemurian–Pliensbachian limestones	10–80	relict sedimentary	paramagnetic
	*	Oxfordian radiolarian limestones	47–271	relict sedimentary	not studied
	KR	Tithonian marly limestones	80–150	relict sedimentary	paramagnetic
	DDM	Berriasian limestones	20–100	relict sedimentary	paramagnetic
Sarnia Skała**	SK	Rhaetian limestones	60–160	deformational	not studied
Samkowa Czuba	STR	Hettangian limestones	100–500	relict sedimentary	paramagnetic
Furkaska–Koryciska	KOR	Anisian–Ladinian limestones	5–30	deformational	paramagnetic

* data after Grabowski (1996)

** denotes the name of locality. The rocks belong to unnamed tectonic slice between Suchy Wierch and Mała Świnica units

al., 1989). The geometry of these magnetizations was often synfolding what could lead to some palaeomagnetic and geological misinterpretations.

Relict sedimentary magnetic fabric developed within the bedding plane occurs in several localities in the Tatra Mts. This indicates that internal deformation of the rock structure was very small and therefore reorientation of palaeomagnetic directions due to physical rotation of magnetic grains is rather unlikely. Lack of "primary" magnetic fabric does not imply that ferromagnetic minerals were subjected to strain. Magnetic fabric in the SU and KOR localities is of tectonic origin differing between particular hand samples. On the other hand, ChRMs in all samples are very consistent (Figs. 27 and 73). Therefore it seems that the magnetization in these localities was not influenced by the factors that contributed to the origin of magnetic fabric. Some uncertainties could be associated with the locality SK. Directions of ChRM and K1 coincide in this locality (sections 4.6.2 and 4.6.3) and it is not unlikely that the magnetization was reoriented towards tectonic stretching direction. This locality is situated within a tectonized zone where several minor tectonic slices occur between two major units (Suchy Wierch and Mała Świnica units, see section 4.6.1). Thus ChRM from this locality might be "suspected" and should be treated with some caution while used for palaeotectonic reconstructions.

5.3. TILT TESTS AND THE AGE OF MAGNETIZATION

Generally the age of palaeomagnetic component is estimated using two methods:

— fold or conglomerate test (Graham, 1949; McFadden, 1990; Morris, 1996) which enables to establish the age of magnetization in relation to folding processes or conglomerate formation,

— comparison of palaeomagnetic components with time sequence of reference directions from the studied area (see section 2.5).

Application of both methods in the tectonically complicated orogen as is the case of the Tatra Mts. meets some problems. Reference data for the Central West Carpathians do not exist and the obtained palaeomagnetic components should be compared with reference data from European and African plates, as well as smaller areas incorporated in the Alpine-Carpathian orogenic belt (i.e. Apulia, Northern Calcareous Alps). Complete fold structures are extremely rare in the KriŽna and higher tectonic units of the Western Tatra Mts. The area is composed of numerous independent tectonic slices or partial nappes which dip monoclinally northwards (see section 2.3). The slices could have been rotated independently not only around vertical or horizontal axes but also around the inclined axes. Fold or rather tilt tests were performed, when possible, for tectonic units according to the correlation test of McFadden (1990). However deformations in the Tatra Mts. were polyphased (Late Cretaceous thrusting and Neogene uplift). Positive result of fold test means only that the sedimentary rocks within single locality or tectonic unit were magnetized when the beds were parallel, but not necessarily horizontal. Although the fold test is positive, the magnetization can be synfolding (or rather syntectonic) as well. Therefore the special attention was paid to inclinations of the characteristic components and comparison with expected palaeoinclinations of major plates (section 2.5). Mean directions will be calculated for each tectonic sub-unit separately in order to evaluate possible amount of rotation between them.

In all localities a normal polarity component was revealed. Additionally in the locality DDM of Bobrowiec unit a mixed polarity component occurs. Both magnetizations will be discussed separately.

5.3.1. NORMAL POLARITY COMPONENT A

5.3.1.1. Suchy Wierch unit

The SU and SI localities differ in tectonic position of beds (Table 6). It is evident from Fig. 74 that the clustering of directions improves after 100% tectonic correction.

Correlation test according to definition 2 of McFadden (1990) was applied for the entire data set from Suchy Wierch unit. Each hand sample was treated as a site. The results are presented in the Table 20. The fold test is positive at 99% confidence level.

The magnetization A can not be primary of Middle Triassic age. The expected inclinations for Middle Triassic should be about 40° (section 2.5). It is considerably steeper than any expected Mesozoic inclination (either European or African) and is close to the Late Tertiary/Quaternary inclinations. There is no geological evidence that the sedimentation took place on the topographical slope (Kotański, 1963) that could result in inclination bias after tectonic correction to the horizontal. Therefore the magnetization of Ladinian dolomites in this tectonic unit must be interpreted as secondary. Remagnetization took place when Suchy Wierch unit had not been internally deformed yet (revealed layer-parallel structure). The secondary remanence resides in PSD magnetite and/or maghemite, but SP magnetite is present as well. The age of remagnetization is discussed below.

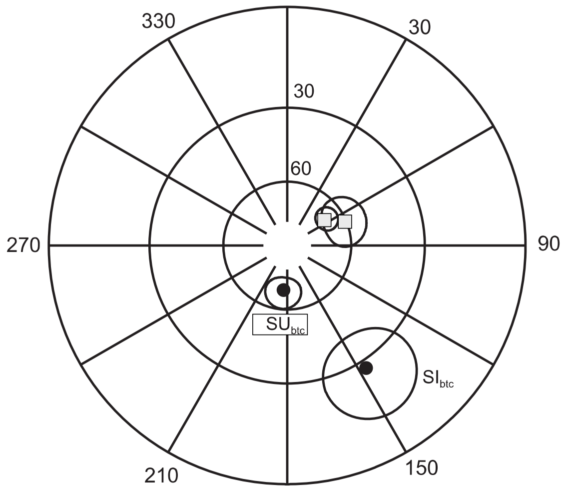


Fig. 74. Stereographic projection of the mean directions from the Suchy Wierch unit

dots — before tectonic correction; squares — after tectonic correction; results of the tilt test are presented in the Table 20

Table 20

Tilt test for normal polarity components SU and SI from the Suchy Wierch unit

N	D	I	α_{95}	k	ξ	Dc	Ic	α_{95}	k	ξ
14	163	55	13.5	9.6	13.431	62	67	4.4	83.1	2.272

Critical values of ξ for 14 sites are: 4.358 for 95% and 6.087 for 99% confidence level

5.3.1.2. Bobrowiec unit

Mean direction for entire tectonic unit was calculated from the localities investigated in this study and Upper Jurassic localities of the Bobrowiec unit studied previously (Grabowski, 1995a) which are included to Table 21. The stereogram is presented in the Figure 75. It is unlikely that the magnetization is post-folding. *In situ* inclinations are too steep even for inferred Late Tertiary/Quaternary age of the component. After tectonic correction ChRMs of six (from seven) localities cluster very well with the declination 30° and inclination about 60° (Fig. 75). Only component RT deviates from the main cluster. The locality RT is situated in the eastern part of the Bobrowiec unit in the zone affected by the faults (Bac, 1971). It can not be ruled out that it underwent more complicated tectonic history than the western part of the Bobrowiec unit. It should be noted that clustering of directions improves after tectonic correction. Correlation test according to definition 2 of McFadden (1990) was applied for the entire data set from Bobrowiec unit. Each hand sample was treated as a site. The results are presented in the Table 22. The fold test is positive at 99% confidence level. Therefore it is assumed that magnetization was acquired, alike in the Suchy Wierch unit, before the internal deformations within Bobrowiec unit took place.

Mean direction for the Middle-Upper Jurassic of KriŽna nappe obtained also from the localities in Bobrowiec unit by

Table 21

Locality mean directions of the normal polarity component from the Bobrowiec unit calculated from samples

Comp.	D	I	α_{95}	k	Dc	Ic	α_{95}	k	N
RT	200	64	11.4	29	69	74	11.2	29.7	9
DDL	162	74	12.7	28.6	33	64	10.7	40.3	6
B1*	71	83	9	28	29	52	9	28	7
B3*	99	83	8	80	31	52	8	80	6
B4*	207	85	9	50	34	59	9	50	7
KR	151	75	8	70.4	34	58	7.3	85.7	6
DDM1	225	75	17.7	27.5	27	58	10.2	82.2	4

* localities in the Bobrowiec unit studied by Grabowski (1995a)

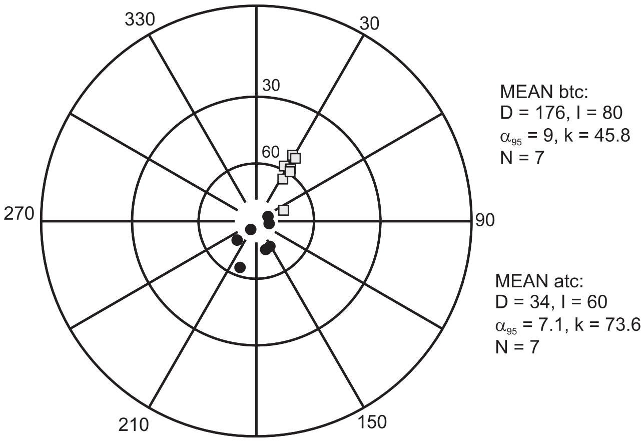


Fig. 75. Stereographic projection of the mean directions from the Bobrowiec unit

dots — before tectonic correction; squares — after tectonic correction; results of the tilt test are presented in the Table 22

Table 22

Tilt test for the normal polarity component from the Bobrowiec unit

N	D	I	α_{95}	k	ξ	Dc	Ic	α_{95}	k	ξ
45	172	81	4.7	21.7	25.889	34	60	4	28.7	1.082

Critical values of ξ for 45 sites are: 7.803 for 95% and 11.033 for 99% confidence level

Kądziałko-Hofmokl and Kruczyk (1987) (D = 25, I = 60, α_{95} = 6.3, k = 94) fits very well the mean direction in Figure 75. The magnetization must be interpreted as secondary. It is rather unlikely that rocks of variegated age sampled in continuous profile (from Rhaetian to Lower Cretaceous) reveal the same normal polarity. The characteristic inclination after 100% tectonic correction (60°) is slightly steeper than the European reference inclination for the Late Mesozoic. It corresponds to Eocene-Miocene inclinations of Africa and Europe (section 2.5).

5.3.1.3. Samkowa Czuba unit and Sarnia Skała

Only one locality — Dolina Strążyska (STR) was studied within the Samkowa Czuba unit. Post-folding age of the component STR is unlikely because this would require large tectonic rotation of the unit in the Neogene or later. Pre-folding

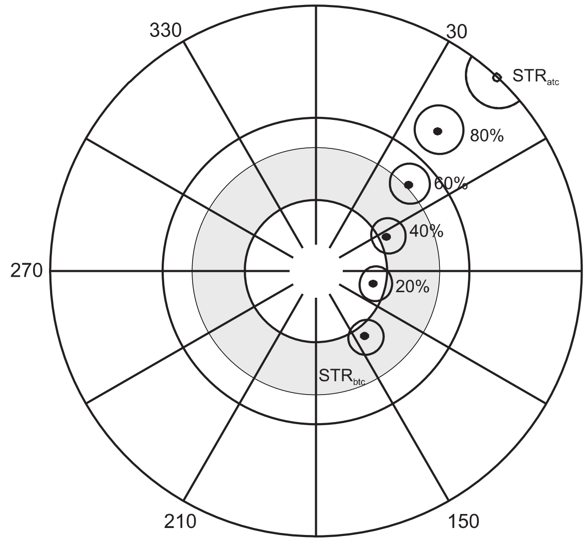


Fig. 76. Stepwise unfolding of the component STR (in 20% intervals). Shaded area corresponds to palaeoinclinations between 40 and 60° which are expected for synfolding magnetization of the Late Cretaceous to Late Tertiary age

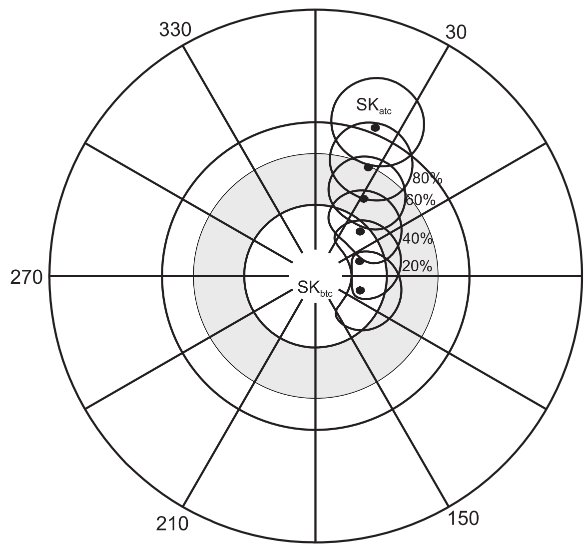


Fig. 77. Stepwise unfolding of the component SK

Explanations as in the Figure 76

Table 23

Incremental unfolding of the component STR

% unfold.	0	10	20	30	40	50	60	70	80	90	100
Dec/Inc	143/56	127/62	106/64	83/62	67/56	57/49	50/40	46/31	43/20	42/10	43/0

Table 24

Incremental unfolding of the component SK

% unfold.	0	10	20	30	40	50	60	70	80	90	100
Dec/Inc	107/70	87/71	71/70	55/67	45/63	37/57	32/52	28/46	26/41	23/35	22/28

age of this component must also be rejected because of its equatorial inclinations. The only plausible solution is synfolding age of magnetization. Results of stepwise unfolding of the component STR are presented in the Figure 76 and Table 23. As synfolding magnetization must be of Late Cretaceous to Late Tertiary age (90–15 Ma), its expected palaeoinclinations should be between 40 and 60° (see Fig. 18 for reference data). These values are reached between 40 and 60% of unfolding with declination values between 67 and 50° (Fig. 76). This implies that the strata in the locality STR were magnetized dipping 45–67° to NE.

Similar situation is encountered in the locality SK. The component SK can not be post-folding because this would require a large local rotation in the Neogene or Quaternary which is improbable. In the pre-folding coordinates the inclination is too shallow for any Late Triassic or later time (see Figure 18 for reference palaeoinclinations). Therefore also in this locality a synfolding age of magnetization must be considered. Expected 40–60° palaeoinclinations are obtained after 40–80% tectonic correction with north-easterly declination values between 45 and 26° (Fig. 77; Table 24). In this case the strata would be magnetized dipping 13–40° to N. However, as was mentioned in the section 5.2, the coincidence of directions between K1 axes and the ChRM brings a question, if component SK is not a "false synfolding" magnetization.

5.3.1.4. Furkaska–Koryciska unit

Only one locality (KOR) was investigated in this tectonic unit. The component can not be post-folding because the inclination (38°) does not match the expected Late Tertiary/Quaternary inclinations (63–70°). After full tectonic correction inclination (60°) corresponds to the expected Tertiary inclinations of Africa and Europe. Large clockwise rotation of declination is observed. The magnetization can not be primary because, as in the case of the Suchy Wierch unit, its inclination is steeper than expected Middle Triassic palaeoinclinations and there is no evidence that sedimentation took place on the topographic slope.

5.3.1.5. Age of the normal polarity component

Component STR must be interpreted as true synfolding (5.3.1.3 and Fig. 76), reasoning from its comparison with reference data from major plates (2.5). The component could be acquired either in the Late Cretaceous (after Turonian) during the main thrusting event in the Tatra Mts. or between Late Cretaceous thrusting and Miocene uplift. Normal polarity compo-

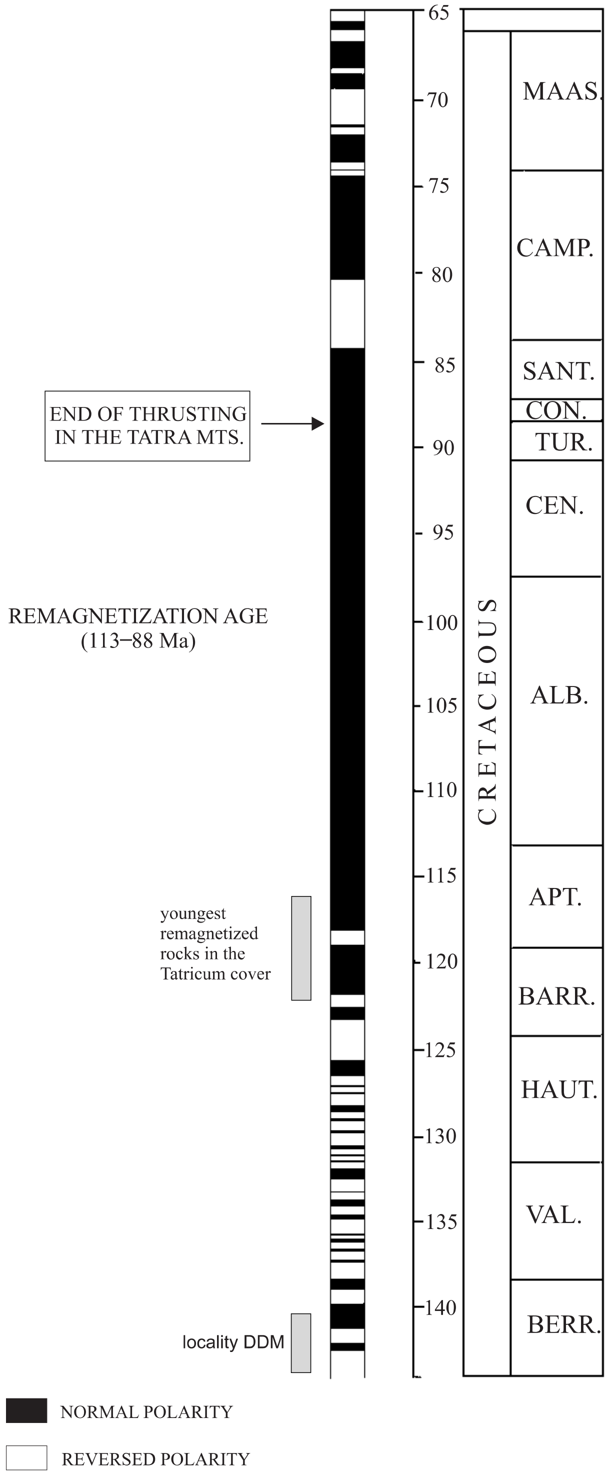


Fig. 78. Cretaceous magnetostratigraphic time scale (after Hailwood, 1989) with possible remagnetization period indicated

nents from all other tectonic units presented in this section could also be interpreted as synfolding. The tectonically corrected inclinations from the Suchy Wierch, Bobrowiec and Furkaska units correspond to expected Tertiary inclinations. However, assuming synfolding age of the component they could be easily matched also with the Late Cretaceous reference inclinations (see below and section 6.1 for discussion). Thus the age of magnetization can not be established unambiguously. Presence of synfolding remagnetizations poses particular problems for interpretation because rocks could be magnetized at various stages of their tectonic history. Therefore all interpretations must be to some degree speculative and simplified. The following assumptions have been made while interpreting the age of the normal polarity component:

1. The component is synchronous in all localities and tectonic units.
   There is no proof that the statement is false. The one exception could be the magnetization residing in hematite in the sample RT9. This problem is discussed in the section 5.4. It must be kept in mind that remagnetization phenomena could be significantly extended in time. For example some Devonian carbonates from the Holy Cross Mts. (Central Poland) reveal post-folding remagnetizations of Late Palaeozoic age which were acquired over the time span about 35 Ma (Grabowski, Nawrocki, 1996).
2. The deformation trajectories were combination of rigid body rotations around vertical and horizontal axes. Tectonic correction to the horizontal is performed rotating the beds around their present strike.
   This assumption is surely a simplification because there is no independent evidence that rotations were about horizontal and vertical axes only. Intermediate stages of folding and thrusting are not known. We do not know what were the deformation trajectories of particular tectonic units during the Late Cretaceous thrusting and what was the geometry of tectonic structures between Late Cretaceous and Neogene (see section 2.3).

Results of previous palaeomagnetic studies might be helpful in estimation of the magnetization age. Secondary normal polarity component was already reported from several localities in the High Tatric units (component A — Grabowski, 1997a) and Gładkie Upłaziańskie tectonic slice (Grabowski, 1995a) which is the southernmost KriŽna sub-unit in the Tatra Mts. (locality GU in the Fig. 11). Both directions are listed in the Table 25. In the Gładkie Upłaziańskie the normal polarity component was undoubtedly pre-thrusting. It is the most likely interpretation that normal polarity component from this study is synchronous with the component from Gładkie Upłaziańskie and High Tatric units — thus it is also of the Late Cretaceous age. Moreover such component have not been encountered in the Eocene and Oligocene rocks of the Podhale basin (Grabowski, 1997a; Marton *et al.*, 1999) what might imply that it was acquired before Eocene. It should be noted that the component is "pre-thrusting" at the Gładkie Upłaziańskie and synfolding (thus perhaps "syn-thrusting") in the Samkowa Czuba, Bobrowiec and Suchy Wierch units. According to this model the tectonic units were affected by remagnetization at various stages of the thrusting. The onset of thrusting was diachronous in the Central West Carpathians as could be deduced from the time when sedimentation ceased in the main tectonic units. In the StraŽov and Choč units the youngest members are Tithonian/Lower Neocomian (Mahel, 1968), in the KriŽna units — Aptian, in the High Tatric units — Turonian (Lefeld *et al.*, 1985). The remagnetization must be post-Aptian (age of the youngest rocks where remagnetization was proved, see Grabowski, 1997a) and pre-Coniacian (the thrusting in the Tatra was completed by this time) thus between 113 and 88 Ma. When comparing these time interval with Cretaceous polarity

Table 25

Mean palaeomagnetic directions of the syntectonic component from the main tectonic units of the Tatra Mts.

No.*	Nappe system	Tectonic unit	Dc	Ic	α_{95}	k	N	References
1	High Tatric	Parautochthonous	23	50	6	256	4	Grabowski, 1997a
2		Czerwone Wierchy	312	48	–	–	1	Grabowski, 1997a
3	KriŽna	Bobrowiec	34	60	7.1	73.6	7	this study; Grabowski, 1995a
4		Gładkie Upłaziańskie	54	49	–	–	1	Grabowski, 1995a
5		Hawrań	40	59	–	–	1	Kruczyk *et al.*, 1992
5a		Kopy Sołtysie**	216	–55	–	–	1	Kądziałko-Hofmokl and Kruczyk, 1987
6		Suchy Wierch	62	66	–	–	2	this study
7		Sarnia Skała***	22	28	–	–	1	this study
8		Samkowa Czuba	43	–1	–	–	1	this study
9	Choč	Furkaska–Koryciska	100	60	–	–	1	this study

* numbering of palaeomagnetic vectors (see Fig. 80); ** secondary nature of magnetization in this locality should be verified (see section 5.3.1.5 for discussion); *** denotes the name of locality. The rocks belong to unnamed tectonic slice between Suchy Wierch and Mała Świnica units. It is possible that magnetization is tectonically modified here (see 5.2. and 5.3.1.3); N — number of localities

Table 26

Mean parameters of the component DDM2 calculated from reversed and normal samples

Component	D	I	α_{95}	k	Dc	Ic	α_{95}	k	Pole	N
DDM2	327	82	15.3	19.9	17	38	11.2	36.7	167°lg. E, 59°lat. N	6

time scale it appears that remagnetization most probably took place in the long Cretaceous Quiet Zone of normal polarity (Fig. 78) which lasted from Aptian till Santonian (120–84 Ma). This might explain why the ubiquitous secondary component reveals only normal polarity.

It is necessary to discuss the concept of Cretaceous remagnetization outlined in this section with previous interpretations of palaeomagnetic results from the Tatra Mts. (Kądziałko-Hofmokl, Kruczyk, 1987; Kruczyk *et al.*, 1992). As was already indicated in section 5.3.1.2. secondary magnetization revealed in the Rhaetian–Lower Cretaceous rocks from Bobrowiec unit corresponds exactly to direction from the Middle-Late Jurassic rocks of this unit described by Kądziałko-Hofmokl and Kruczyk (1987). Their direction revealed exclusively normal polarity, similarly as components RT, DDL, DDM and KR. Some of rocks sampled by Kądziałko-Hofmokl and Kruczyk (*op. cit.*) were restudied by Grabowski (1995a). The unblocking temperature range 300–500°C are the same as in other remagnetized rocks of Bobrowiec unit. Therefore it must be assumed that the results of Kądziałko-Hofmokl and Kruczyk (*op. cit.*) from Bobrowiec unit, which had been previously interpreted as primary Middle–Late Jurassic magnetizations, represent in fact Cretaceous remagnetizations. The problem is more complex in the eastern part of the Sub-Tatric units which were palaeomagnetically studied: the Kopy Sołtysie and the Havrań unit. There are two results available from these units (Table 25). They yield both normal and reversed polarity directions which are almost anti-parallel. In original papers (Kądziałko-Hofmokl, Kruczyk, op. cit.; Kruczyk et al., 1992) they were also interpreted as primary. However it can not be excluded that they are secondary directions as well. Reversed polarity component from the Kopy Sołtysie unit (Dolina Filipka) is quite confusing because it does not fit the assumed model that all Sub-Tatric rocks were remagnetized in the normal Cretaceous Quiet Zone. Thus it is still possible that in the Oxfordian radiolarites from the Kopy Sołtysie magnetization was preserved. However it might be postulated as well that these rocks were remagnetized within one of reversed zones in the Late Barremian/Early Aptian. The Kopy Sołtysie unit yields a very complicated structure with overturned sequence and numerous secondary folds (Sokołowski, 1978). Convincing documentation of primary magnetization must include a tilt test, or at least, discussion of ChRMs directions before and after tectonic correction. As these conditions are not met in the original papers (Kądziałko-Hofmokl *et al.*, 1985; Kądziałko-Hofmokl, Kruczyk, 1987), a question mark must be posed if reversed polarity direction from Dolina Filipka represents a truly primary Late Jurassic magnetization. A detailed palaeomagnetic study comprising entire profile of the Kopy Sołtysie and Havran units in Poland and Slovakia would be very desired.

5.3.2. MIXED POLARITY COMPONENT

This component occurs only in the locality Dolina Długa (DDM, component DDM2). The maximum unblocking temperatures are higher than 500°C. Because of susceptibility rise between 400 and 500°C it was impossible to calculate this component from the fitted lines.

The component DDM2 can not be post-folding. It reveals almost vertical inclinations in the geographical coordinates (Tabs. 13 and 26; Fig. 47 — lower diagram). After full tectonic correction the component reveals north-easterly declinations and moderate inclinations (about 38°). The mean inclination values of the reversed and normal component differ by ca. 10° (Table 13). The reversal test (McFadden, McElhinny, 1990) has been performed for the bedding corrected component DDM2.

Reversal test is applied in order to check whether two sets of observation (one with normal polarity and the other with reversed polarity) could have been drawn from distribution which are 180° apart. This may not be the case if the magnetizations are contaminated with another component. The parameter which is decisive in the reversal test is the angle γc between the mean directions of the two sets of observations. The positive reversal test could be classified to category “A” if $\gamma c \leq 5°$, “B” if $5° < \gamma c \leq 10°$, “C” if $10° < \gamma c \leq 20°$. The test is indeterminate if $\gamma c > 20°$.

The reversal test performed on the sample level is positive with value $\gamma c = 16.3°$ which is however close to the critical value of 20° above which the test is “indeterminate”.

The mixed polarity, high unblocking temperatures and relatively shallow inclinations distinguish that component from the ubiquitous normal polarity component which was interpreted as synfolding (section 5.3.1). It is possible that the DDM2 represents a primary Early Cretaceous magnetization.

The component could be compared with the primary directions obtained by Houša *et al.* (1996) in the same Tithonian/Berriasian limestones at Brodno in the Pieniny Klippen Belt of Western Slovakia (D = 236, I = 45, $\alpha_{95} = 5.6$, k = 9.8). The remagnetization was not reported from the Brodno limestones. The limestones belong to the “Biancone” facies which is widespread in the Central Carpathians (Wieczorek, 1988). The preservation of possibly primary magnetization in these rocks points that they are especially suitable for palaeomagnetic investigations. The mean inclination values at Brodno is slightly higher than at Dolina Długa.

Some doubts about true dual polarity nature of the component might arise from hysteresis measurements (section 4.4.2). Sample with normal polarity (DDM5) revealed different hysteresis ratios (typical PSD) than two samples of reversed polarity which fall closer to the MD field (Fig. 43C). Moreover samples with normal polarity were quite stable during thermal demagnetization up to 540–560°C while reversed samples behaved chaotically already in temperature 520°C and sometimes even lower. These observations indicate that rock magnetic properties of normal and reversed samples are different. It can not be excluded that normal polarity magnetization is not primary but is still contaminated with a typical Late Cretaceous remagnetization described in section 5.3.1. In order to establish if polarity changes in this locality reflect truly primary magnetisation more detailed investigations with more samples should be carried out. Additionally systematic hysteresis studies should be performed for all normally and reversely magnetized samples.

5.4. LINKS BETWEEN MAGNETIC MINERALOGY AND REMAGNETIZATION

Preliminary palaeotemperature estimations using conodont alteration index (CAI) indicate that rocks in the locality KOR reveal index CAI = 2 (Grabowski *et al*., 1999). This value corresponds to palaeotemperatures 50°C through 100 Ma, 70–80°C through 10 Ma or 110°C through 1 Ma (Epstein *et al*., 1977). Palaeotemperature estimations based on unblocking temperature — relaxation time relationship are usually performed using nomograms of Pullaiah *et al*. (1975) and Middleton, and Schmidt (1982). Both are elaborated for SD magnetite but using different single domain theories. The nomograms of Middleton and Schmidt (1982) appear to fit better experimental data for rocks containing PSD and MD magnetite (Kent, 1985). Therefore this nomogram is applied here for evaluation if remagnetization in the locality KOR is of thermoviscous or chemical origin (magnetite in this locality is in the PSD state — Fig. 69C). Maximum unblocking temperature of the Late Cretaceous component KOR is 400°C (section 4.8.3). If this component was a thermoviscous remagnetization the observed unblocking temperatures would correspond to ca. 150°C heating through 10 Ma or 125°C through 100 Ma. (Fig. 79). It is almost twice a temperature revealed from CAI index. Thus it might be postulated that secondary direction of Late Cretaceous age in the locality KOR is a chemical remagnetization. The locality KOR belongs to the highest (Choč or StraŽov) Sub-Tatric nappe and its tectonic burial was shallower than that of the KriŽna nappe or High Tatric units. It can not be excluded that palaeotemperatures in these units reached interval 150–200°C, therefore thermoviscous remagnetization in remaining localities can not be excluded.

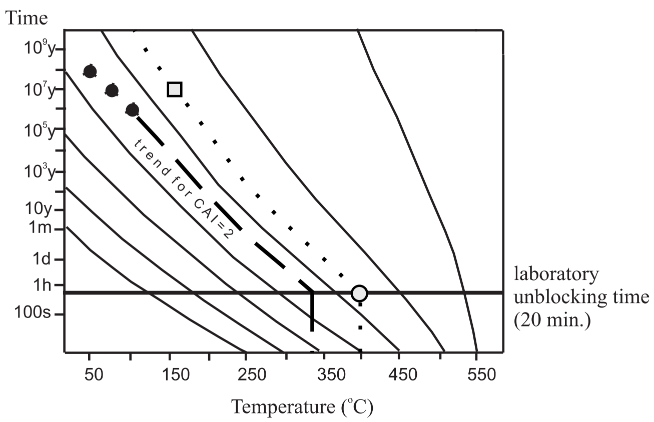


Fig. 79. Theoretical relaxation time — unblocking temperature relations for magnetite, after Middleton and Schmidt (1982)

Black dots and dashed curve represent trend for CAI = 2 (laboratory unblocking temperature ca. 350°C).Grey circle represents laboratory unblocking conditions for the component KOR (400°C) while grey square indicates time–temperature conditions of acquisition of relevant thermoviscous magnetization.

Syntectonic magnetization acquired during the Late Cretaceous thrusting is carried by variegated mineral assemblage: essentially magnetite, but also maghemite and locally hematite. Occurrence of the same magnetization in three different magnetic minerals deserves explanation. In the Middle Triassic dolomites (localities SU and SI) where presence of maghemite and magnetite was suspected in the same samples (section 5.1) there is only one high stability component which apparently resides in both fractions. This observation is in good agreement with experimental observations of Heider and Dunlop (1987). The maghemite is a product of the low temperature (100–250°C) oxidation of magnetite. According to Heider and Dunlop (*op. cit.*) when acicular SD magnetite converts to maghemite during a single-phase oxidation, maghemite preserves the magnetization direction of the parent magnetite. In the locality RT the same characteristic direction resides in magnetite (samples RT1–8 — Rhaetian–Hettangian limestones and mudstones) and in hematite (sample RT9 — Keuper dolomite). Occurrence of the same secondary magnetization in magnetite and hematite is difficult to explain because these minerals originate in different geochemical environments (Machel, 1995). The following explanation might be suggested. Remagnetization took place most probably during the long period of the normal polarity of the geomagnetic field ("Cretaceous Quiet Zone" — Fig. 78) thus hematite and magnetite components, although they have the same declination and inclinations, could be slightly heterochronous. Magnetization residing in hematite could be acquired later than the magnetite component when oxidizing conditions started (possibly during some uplift).

It could hardly be speculated about the mechanisms of remagnetization. Carriers of magnetization have not been unambiguously identified either in the ore microscope or in the SEM because of their very fine grain size. Ubiquitous calcite veins seem to be devoid of any other mineralization therefore it is rather unlikely that remagnetization was related to their formation. However the migration of fluids released during orogeny still might be the likely cause of remagnetization as is the case of Paleozoic sedimentary rocks of North America (McCabe, Elmore, 1989). Activity of orogenic fluids during thrusting in the Tatra Mts. and other Central West Carpathian massifs was postulated by Jaroszewski (1982), Jacko and Sasvari (1990), Plašienka and Sotak (1996). In the locality SU the presence of authigenic potassium feldspars was documented (Skiba, Michalik, 1999). The feldspars are evidence for the late diagenetic brine migration which might be coeval with nappe overthrusting, as is in the case of Northern Calcareous Alps (Spötl *et al*., 1999).

6. GEOLOGICAL IMPLICATIONS

6.1. SYNTECTONIC REMAGNETIZATIONS OF LATE CRETACEOUS AGE

Having established the syntectonic nature of the normal polarity component and its Late Cretaceous age (section 5.3.1.5) its geological interpretation could be performed. Characteristic declinations and inclinations from each palaeomagnetically investigated tectonic unit in the Tatra Mts. (Tab. 25; Fig. 80) should be compared with the reference Cretaceous data for Africa and Europe (Table 27).

Palaeoinclinations of the High Tatric parautochthonous and Czerwone Wierchy units (no. 1 and 2 in the Table 25) are intermediate between the "European" and "African" values. These tectonic units as well as Gładkie Upłaziańskie tectonic slice (no. 4 in the Table 25) could be magnetized roughly in the horizontal position. Most of the palaeoinclinations from the Sub-Tatric units (excluding no. 7 and 8 in the Table 25) are steeper than 50°.

If the Late Cretaceous age of the component is accepted it must be concluded that the magnetization was acquired while the strata in the Bobrowiec, Hawrań, Suchy Wierch and Furkaska–Koryciska units were inclined southwards. Thus an additional tectonic correction (called here "overcorrection") is

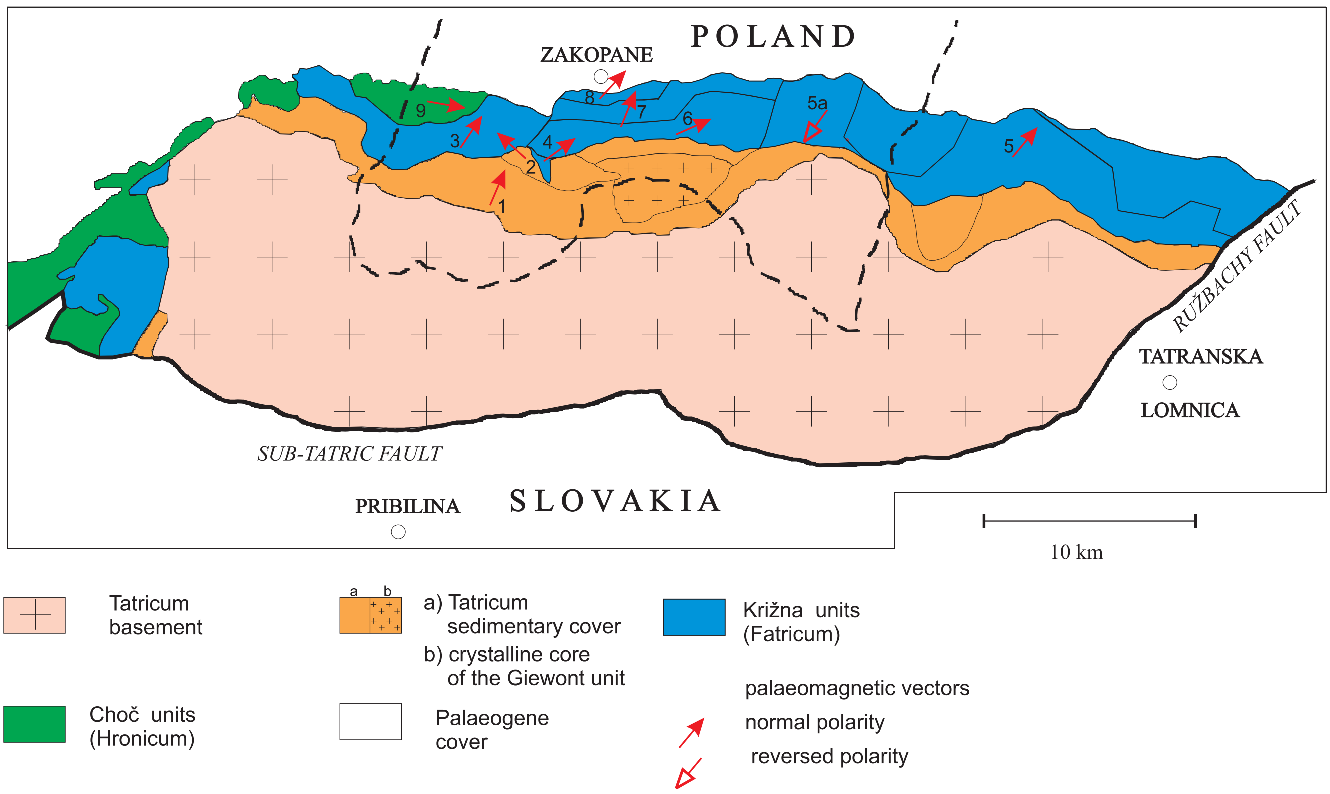


Fig. 80. Mean paleomagnetic declinations of the syntectonic component in the Tatra Mts.
Numbering of palaeomagnetic vectors as in the Table 25

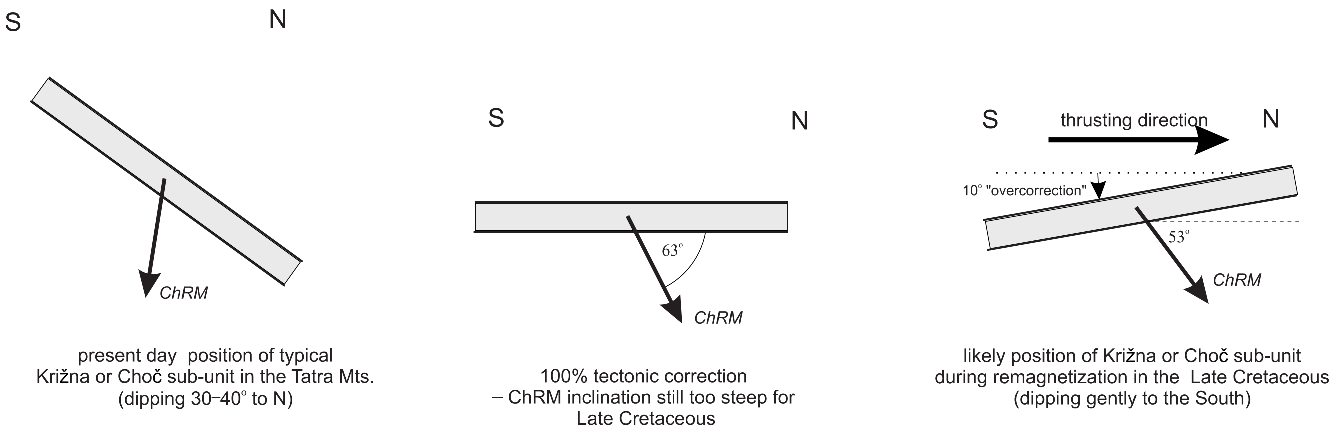


Fig. 81. Model explaining abnormally steep inclinations in the Križna and Choč units. The thrust sheets could have been remagnetized dipping about 10–20° to the south during thrusting episode

ChRM — Late Cretaceous syntectonic magnetization

required to match the inclinations observed in the pre-folding coordinates with expected Late Cretaceous paleoinclinations. The idea is presented in the Figure 81. The exact amplitude of "overcorrection" can not be established because it depends on the position of the strata during thrusting and magnetization which is not known. However, the minimum values of "overcorrection" could be estimated assuming SW dip of the overthrusted units during magnetization: 10–20° for Suchy Wierch unit, 5–15° for Bobrowiec and Hawrań units. On the other hand investigated beds of the Samkowa Czuba unit must have dipped north about 67–45° during magnetization process (section 5.3.1.3).

The separate problem are differences in palaeodeclinations between the tectonic units in the Table 25. They could indicate tectonic rotations that took place after remagnetizations. However also the amount of rotations can not be estimated directly because, as it was already pointed out, intermediate stages of folding (thrusting) are not exactly known. As an example Furkaska–Koryciska tectonic unit could be considered. Palaeodeclination of this unit after 100% unfolding apparently indicates a large clockwise rotation (65°) in relation to Bobrowiec unit. After applying stepwise tectonic "overcorrection" to match the palaeoinclination of the Furkaska–Koryciska unit with reference Late Cretaceous inclinations it appears that hypothetical tectonic rotation depends on subjectively chosen azimuth of tectonic "overcorrection" (Tab. 28; Fig. 82). If it is presumed that the tectonic element dipped 30–40° southwards at the intermediate stage of folding (azimuth of tectonic "overcorrection" 0°) the characteristic declination decreases to about 50° and the inferred amount of rotation in relation to Bobrowiec unit would be only 20°. On the other hand, if a gentle (10–20°) westward dip of the Furkaska–Koryciska unit is accepted (there is no geological evidences against this assumption) the 65° tectonic rotation must be introduced. The first option that requires less tectonic rotation is considered as more likely. Bearing in mind all constraints given above the syntectonic magnetization gives limited basis for geological interpretations. It is not suitable for palaeolatitude reconstructions. The width of hypothetical oceanic domains in the Late Cretaceous north of the Tatra Mts. (Vahicum and Magura oceans) can not be estimated directly. Therefore assumptions sug-

Table 27

Reference Cretaceous (124–76 Ma) palaeomagnetic directions calculated from data in the Tabs. 1, 2, 4 and 5

Plate	D	I	α_{95}	N	Palaeopole (lat. N/lg. E)
Europe	5	52	7	4	73/185
Africa + Adria	335	45	6	8	60/250

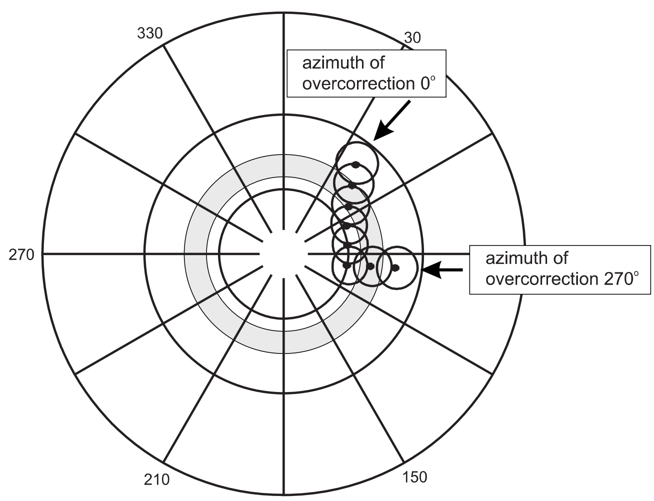


Fig. 82. Stepwise "overcorrection" of the component KOR in two alternative azimuths

Data after Table 28. Note the dependence of palaeodeclination on azimuth of the overcorrection. Shaded area indicates expected Late Cretaceous palaeoinclinations gested in the earlier papers (Grabowski, 1995b, 1997a, see also chapter 1) that Cretaceous palaeoinclinations situate the Tatra Mts. area closer to the European than African or Adriatic plate must be treated as most straightforward but not necessarily true. Even very moderate dip of the parautochthonous cover of the Tatra crystalline core during the Late Cretaceous remagnetization (5–10° to S) would change this interpretation: the palaeoinclination become closer to African/Adriatic. However the palaeodeclinations are definitely closer to the "European" values. They might be used for rough estimation of tectonic rotation amplitude. The parautochthonous unit reveals ca 18° (±5.6°) post-Late Cretaceous clockwise rotation in relation to the European Plate. The rotation seems still larger (30–50°) for the KriŽna and Furkaska–Koryciska units but this conclusion is hampered by indeterminate tectonic correction. The opposite sense of rotation (Tab. 25) is observed in the Czerwone Wierchy unit however it requires confirmation from more localities.

Table 28

Dependence of palaeodeclination on subjectively chosen azimuth of tectonic "overcorrection" for component KOR (see also Figure 82)

Overcorrection	0°	10°	20°	30°	40°
D/I (azimuth of over-correction 0°)	100/60	82/60	66/58	54/53	45/46
D/I (azimuth of over-correction 270°)	100/60	98/50	97/40	–	–

The fact that most Sub-Tatric units were magnetized dipping south (while moving northwards) might be an evidence that horizontal compression and not gravitational gliding was the most important emplacement mechanism.

6.2. PRIMARY(?) EARLY CRETACEOUS MAGNETIZATION

This component was revealed in only one locality DDM (Fig. 47; Tabs. 12 and 13). Primary nature of this component has not been definitively proved yet (section 5.3.2.) Therefore the geological implication must be treated as very preliminary. The azimuth of declination is almost the same as that of the secondary Late Cretaceous magnetization in the Bobrowiec unit (see Tabs. 25 and 26). This would indicate that no major tectonic rotation occurred in that unit between Berriasian and Late Cretaceous. This would also confirm the ca. 20° clockwise rotation after Late Cretaceous (section 6.1).

Mean palaeoinclination calculated from normal and reversed directions (Table 26) is 38°. The palaeoinclination from the "Biancone" limestones at Brodno near Žilina, western part of the Pieniny Klippen Belt (Houaa *et al*., 1996) amounts to 45°. Therefore the intermediate value 41–42° might be tentatively accepted as the reference Tithonian/Berriasian palaeoinclination for the Central West Carpathians. The palaeoinclination of the southern margin of European Plate should be 53° (± 3°) according to the reference APWP of Besse and Courtillot (1991). This inclination is based mainly on the good quality Berriasian palaeopole of Galbrun (1985). The coeval reference inclinations from the Adriatic plate would be about 40° (Channell, 1996). Thus palaeoinclinations from Biancone limestones seem to be concordant rather with the African/Adriatic than European domain. The existence of narrow (several hundred km) oceanic domain in the Late Jurassic/Early Cretaceous time between the Central West Carpathians and European Platform is quite likely. However it must be kept in mind that these quantitative estimations are close to the resolution limit of the palaeomagnetic method and should be treated with extreme caution until more data is available.

6.3. COMPARISONS OF PALAEOPOLES FROM THE TATRA MTS. TO MESOZOIC PALAEOPOLES FROM ADJACENT REGIONS

The syntectonic Late Cretaceous component from the KriŽna unit can not be used for broad palaeogeographic considerations because the attitude of strata during magnetization is not exactly known (section 6.1). Therefore only Late Cretaceous palaeopole from the parautochthonous unit (palaeopole A of Grabowski, 1997a) might be more safely used as reference for the Tatra Mts. since its error is probably lower. Additionally the component DDM2 from Bobrowiec unit will be used as Berriasian snap-shot although it is quite poorly constrained. It was already compared to the primary Tithonian–Berriasian magnetization from Pieniny Klippen Belt (section 6.2). In this chapter Mesozoic palaeomagnetic data from Western Carpathians, Northern Pannonia and Eastern Alps are briefly reviewed and discussed jointly with the data from the Tatra Mts.

6.3.1. CENTRAL WEST CARPATHIANS OF SLOVAKIA

The ubiquitous syntectonic Late Cretaceous component could be compared to data from the Lower to Upper Jurassic rocks of the KriŽna units in Slovakia obtained by Kruczyk *et al*. (1992) (Tab. 29). It was suggested earlier (Grabowski, 1997a) that these data represent Cretaceous remagnetizations acquired

Table 29

Palaeomagnetic data from the Križna nappe in Slovakia, after Kruczyk *et al.* (1992)

No.	Locality	Age of rocks	D	I	Dc	Ic	α_{95}	k	n/N	Pole lat. N/ lg. E
1	Mala Fatra	Bajocian–Tithonian	328	85	321	44	3	96	173/21	58/253
2	Nizke Tatry	Middle Jurassic	35	76	2	56	14	73	29/3	78/192
3	Magura Spišska	Middle Jurassic	312	70	75	46	12	114	12/3	30/102
4	Choč Hills	Lower Jurassic	116	80	39	63	12	104	15/3	63/105

Dc (Ic) — declination (inclination) after tectonic correction, n — number of samples, N — number of outcrops (sites?)

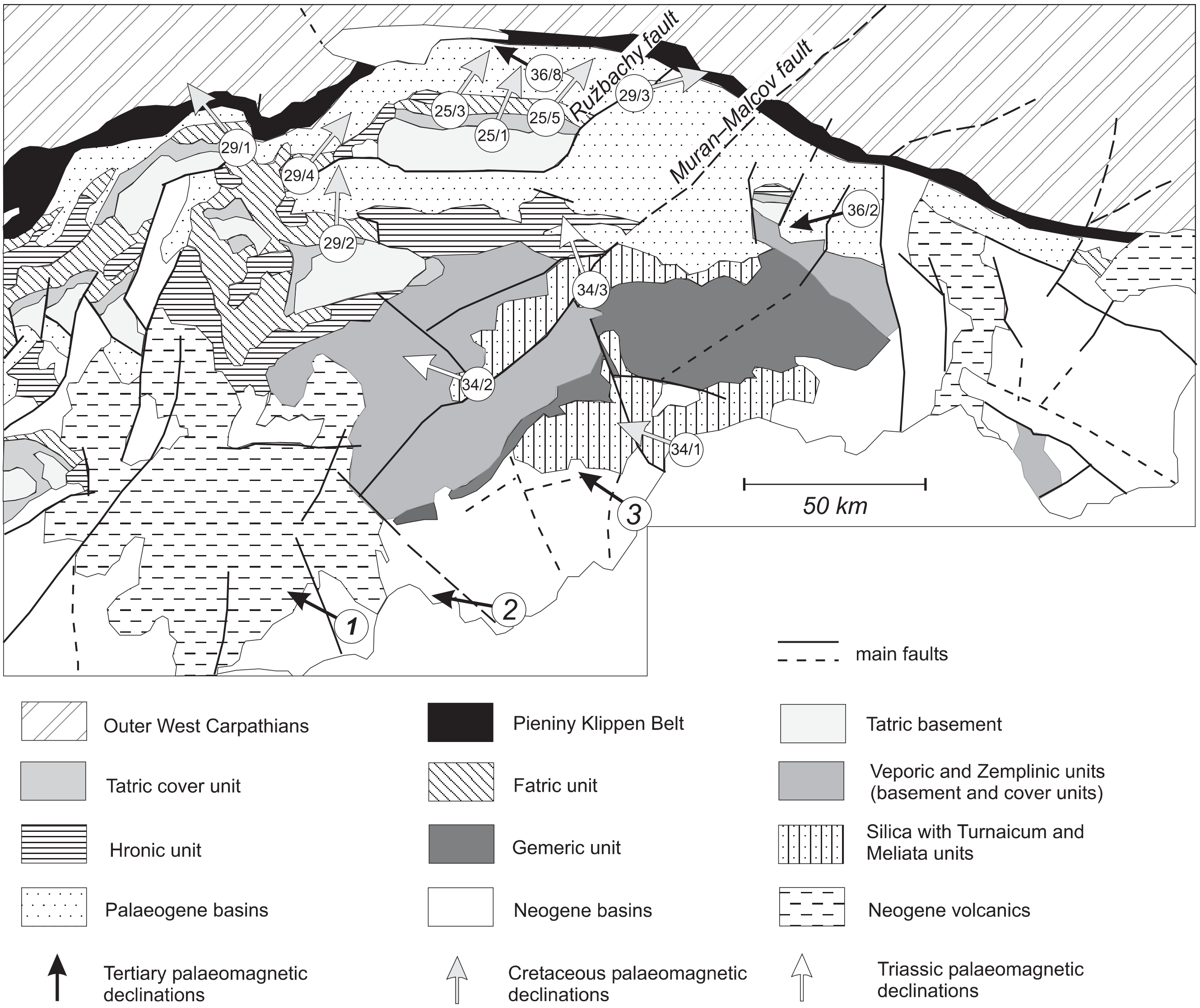


Fig. 83. Tectonic sketch map of the Central and Inner West Carpathians of Poland and Slovakia with palaeomagnetic declinations

Palaeomagnetic results quoted in the Text–Tables are indicated by no. of Table vs. no. of entry (e.g. 25/3). Palaeomagnetic results not quoted in the Text–Tables: 1, 2 and 3 (all from South Slovak Tertiary Basin, after Marton *et al*., 1996) in the "Cretaceous Quiet Zone". The evidences were: (1) exclusively normal polarity of magnetization from rocks belonging to different parts of the Jurassic and (2) unblocking temperatures 350–450°C which are typical for undoubtedly remagnetized carbonates in the Tatra Mts. Additional argument for syntectonic nature of the components in Table 29 is quite a large scatter of pre-folding inclinations: from 44 to 63°. This problem was pointed out also by Kądziałko-Hofmokl *et al.* (1990) and by Kruczyk *et al.* (1992). It could be easily explained assuming syntectonic (syn-thrusting) age of magnetization. The KriŽna nappe fragments were magnetized in variegated position either dipping up to 10° north (in Mala Fatra) or south (in Choč Hills). These components rather should not be used for palaeolatitudinal reconstructions as was attempted in the earlier papers (Grabowski, 1995b, 1997a). It is noteworthy that most of them reveal clockwise rotations similar to the typical secondary components from the Tatra Mts. Differences in declinations between localities indicate tectonic rotations of Late Cretaceous or Tertiary age. Distribution of localities and declinations is presented in the Figure 83.

6.3.2 OUTER WEST CARPATHIANS

Entirely different pattern of rotation is revealed by Cretaceous palaeomagnetic directions from the Outer West Carpathians (Tab. 30). The data were obtained in the western part of the Flysch Belt.

The inclination values are quite dispersed from 44 to 73° but declinations reveal very consistent counter-clockwise rotation by 40–60°. The same amount of rotation is expressed by the results from Tertiary (Tab. 35, section 6.4). Palaeopoles are located close to the African APWP as was already remarked by Krs *et al.* (1991, 1993). This indicates almost 90° difference between the Cretaceous declinations from the Outer West Carpathians and Tatra Mts. As yet there is no good theory explaining this disagreement. The reliability of data from the Table 30 is variegated. The result from the Biele Karpaty (no. 1) is of good quality because rocks were carefully demagnetized and both normal and reversed polarities were observed. Data no. 2–5 were obtained mostly in 1970s and laboratory treatment was not sufficient (Krs *et al.*, 1982). The statistics is also weak ($k \approx 10$). Nevertheless consistency of results is remarkable. Some ideas how to explain the difference of palaeomagnetic record between Central and Outer West Carpathians are put forward in the section 6.4.3.

6.3.3. EASTERN ALPS

Senonian/Danian palaeopoles from the Northern Calcareous Alps are quite similar to the Late Cretaceous pole from the parautochthon of the Tatra Mts. (Fig. 84A). Tithonian/Neocomian palaeopoles are rotated 55–70° clockwise (Table 31) but the palaeoinclinations are in good agreement with coeval palaeoinclinations from Brodno (Houša *et al.*, 1996) and only slightly steeper than in DDM (this study). Surprisingly high palaeoinclination for upper Neocomian (no. 3 in Table 31) might be linked to the rapid northward drift of the area but certainly not to the opening of the South Penninic ocean as was suggested by Mauritsch and Marton (1995). It is well known that palaeopoles from the Northern Calcareous Alps are generally closer to the European than to African/Adrian palaeopoles (Vandenberg, Zijderveld, 1982; Heller *et al.*, 1989; Marton, Mauritsch, 1990; Marton, 1993; Mauritsch, Marton, 1995) but the geological interpretations are ambiguous. Large clockwise rotation of the Northern Calcareous Alps took place between Late Neocomian and Senonian. Most geological reconstructions suggest that the Central West Carpathians and Austroalpine realm had constituted a coherent palaeogeographic unit in the Mesozoic (Häusler *et al.*, 1993; Kozur, Mock, 1996) although their models differ in details (section 2.2.2). The difference between Mesozoic palaeopoles from the Northern Calcareous Alps and Central West Carpathians must arise from the different Tertiary kinematics of the areas. It is remarkable that palepoles from the Tatra Mts. are closer to the palaeopoles from the Northern Calcareous Alps than to the palaeopoles from the Outer Flysch Carpathians. This problem is discussed in the section 6.4.3.

Table 30

Palaeomagnetic data from the Cretaceous rocks of the Outer West Carpathians, after Krs *et al.* (1982, 1996)

No.	Locality	Age of rocks	Lithology	Dc	Ic	a_{95}	k	n/N	Pole: lat. N/ lg. E
1	Biele Karpaty	Upper Senonian	grey sandstones	321	44	5.6	86	75/9	52/265
2	NE Moravia, Silesia	Coniacian	grey sandstone	300	46	10	3.1	94/?	40/285
3	NE Moravia, Silesia, Ondrejnik Mt.	Cenomanian/Turonian	red claystones	318	73	4.5	11	101/20	64/323
4	NE Moravia, Silesia,	Cenomanian/Turonian	red claystones	312	53	3	9	228/10	52/281
5	NE Moravia, Silesia	Hauterivian/Barremian	teschenites	295	56	5.6	6.4	116/12	42/297

Explanations as in the Table 29

Table 31

Post Middle Jurassic palaeomagnetic data from the eastern Northern Calcareous Alps, after Mauritsch and Marton (1995). Palaeopoles calculated for geographic coordinates 48° N, 13° E

No.	Age of rocks	Dc	Ic	α_{95}	k	n/N	Pole: lat. N/ lg. E
1	Danian	8	42	–	–	?/2	–
2	Senonian	22	50	15	17	49/7	66/141
3	Upper Neocomian	74	63	17	22	5/1	41/78
4	Lower Neocomian	72	44	26	9	5/1	31/97
5	Tithonian	66	47	12	63	4/?	36/99

n — number of sites, N — number of localities; other explanations as in the Table 29

Palaeomagnetic data from the Upper Austroalpine nappes to the south of the Northern Calcareous Alps show counter-clockwise rotations (Table 32). The Late Cretaceous sediments in two Gosau basins reveal 50–70° counter-clockwise rotation relative to Europe but almost no rotation relative to the Southern Alps or Africa. These observations indicate that no observable rotation occurs across Periadriatic line (Heller *et al.*, 1989; Marton, 1990).

6.3.4. TRANSDANUBIAN CENTRAL MTS.

Considerable differences are observed when comparing Mesozoic palaeopoles from the Tatra Mts. and Transdanubian Central Mts. in the Northern Hungary (Tab. 33). The Tithonian/Neocomian palaeodeclinations differ by 110° and the Middle–Late Cretaceous by 90°. On the other hand, palaeoinclinations from both regions are quite consistent. Late Senonian palaeoinclination (no. 1 in the Table 33) matches the Late Cretaceous palaeoinclination from para-autochthonous unit (Tab. 25) while Tithonian/Hauterivian palaeoinclinations are identical to the Berriasian palaeoinclinations from Bobrowiec unit (DDM2).

6.3.5. INNER WEST CARPATHIANS OF SLOVAKIA

Mesozoic palaeomagnetic directions from the Inner West Carpathians also reveal north-westerly declinations (Tab. 34). Primary Triassic magnetization was identified in several localities (e.g. in the Muranska Planina and Stratenska Hornatina Mts. — no. 2 and 3 in the Tab. 34, see also Fig. 83). Synfolding components of Early/Late Cretaceous age were established in the Slovak Karst (no. 1 in Tab. 34). They must be almost coeval with the remagnetizations in the Tatra Mts. 60–70° difference in declination is observed between synfolding directions from the Slovak Karst and Tatra Mts. The palaeoinclinations from the Slovak Karst are slightly lower than in the Tatra Mts. However as tilt correction is not precisely known in both areas the inclination values can not be used for quantitative interpretations.

Table 32

Senonian palaeomagnetic data from the Eastern Alps, south of the Northern Calcareous Alps (after Mauritsch and Becke, 1987)

No.	Locality	Dc	Ic	α_{95}	k	N	Pole: lat. N/ lg. E
1	Krappfeld (Carinthia)	316	43	–	–	2	55/259
2	Kainach (Styria)	317	59	26	24	3	59/301

Explanations as in the Table 29

Table 33

Post Middle Jurassic/pre-Tertiary palaeomagnetic data from the Transdanubian Central Mts., after Mauritsch and Marton (1995)

No.	Age of rocks	Dc	Ic	α_{95}	k	n/N	Pole: lat. N/ lg. E
1	Upper Senonian	318	53	7	195	34/4	56/280
2	Albian	300	38	15	71	39/3	36/281
3	Aptian	284	46	11	132	24/3	29/298
4	Berriasian–Hauterivian	277	38	18	46	72/3	20/298
5	Tithonian	265	38	6	121	275/6	12/306

Explanation as in the Table 29

Table 34

Post Middle Triassic/pre-Tertiary palaeomagnetic data from the Inner West Carpathians of Slovakia

No.	Age of rocks	Dc	Ic	α_{95}	k	n/N	References
1*	Ladinian–Norian	320	42	5.9	168	52/5	Marton *et al.*, 1991
2	Carnian	290	51	12.7	14	11/1	Marton *et al.*, 1991
3	Carnian	160	–55	9.3	53	6/1	Marton *et al.*, 1991
4	Upper Triassic	289	59	11	38	45/6	Marton *et al.*, 1988

* synfolding Early/Late Cretaceous remagnetization; explanations as in the Table 29

6.4. MESOZOIC PALAEOMAGNETIC DATA FROM THE TATRA MTS. IN THE LIGHT OF TERTIARY KINEMATICS OF THE CARPATHIAN-PANNONIAN DOMAIN

Palaeomagnetic data obtained only from Mesozoic rocks could be used for local palaeotectonic reconstructions as was attempted in the section 6.1. However for reconstruction of palaeogeographic position of the Tatra Mts. in the Mesozoic and Tertiary Tethys the effects of Tertiary kinematics of the area should be considered. Geologists postulate large mobility of the Carpathian–Pannonian domain between Palaeogene and Early Miocene (see section 2.2.1 and Figs. 4–7). Therefore Mesozoic palaeomagnetic directions from the Tatra Mts. should be rotated according to trends of Tertiary palaeomagnetic rotations. Their final interpretation should be performed in the pre-Tertiary coordinates.

6.4.1. REVIEW OF TERTIARY PALAEOMAGNETIC DATA FROM THE WESTERN CARPATHIANS AND NORTHERN PANNONIA

The data are summarized and critically evaluated in the Tables 35–38 for following regions:

1. Outer West Carpathians.
2. Central Carpathian and Peri-Klippen Palaeogene Basins.
3. Northern Pannonian Palaeogene Basin (NPPB).
4. Inner West Carpathians.

There is a dramatic difference between the palaeomagnetic directions from the Mesozoic overthrusted units of the Central West Carpathians and Cainozoic directions of the entire Western Carpathians. All Tertiary directions reveal uniform counter-clockwise rotations.

Not numerous data from the Outer West Carpathians (Tab. 35) point to about 40° counter-clockwise rotation after Eocene. The magnitude of rotation is in agreement with the Cretaceous data from the area (Tab. 30). The rotation must have been completed before the time of intrusions of Pieniny andesites. The declination from the Mount Wżar (no. 4 in Table 35) is close to the present day geomagnetic declination of the area.

Also data from the Central Carpathian and Peri-Klippen Palaeogene (*sensu* Mahel, 1968) indicates large magnitude of post-Eocene counter-clockwise rotations: 90–110°. These data should be treated with some caution because flysch rocks are extremely difficult material for palaeomagnetic study. Post-folding remagnetizations (no. 3 and 4 in the Tab. 36) are present and rock magnetism of these rocks is not fully understood. Great variations of the inclination parameter (from 21° to 66°) could point to not complete removal of secondary component or even a synfolding magnetization. Data presented by Kovač and Tunyi (1995) are strongly dispersed after tectonic correction (k = 4–13). Before tectonic correction the ChRM are close to the present day geomagnetic field direction what indicates possibility of post-folding remagnetizations as well. However the amount of data is still increasing — Marton *et al*. (1999) reported ca. 60° counter-clockwise rotated palaeomagnetic declination from the Podhale and Levoča basins (no. 8 in the Tab. 36). Therefore this rotation must have also affected the Mesozoic sedimentary rocks in the Tatra Mts., what must be kept in mind while interpreting the palaeomagnetic directions of Mesozoic age (section 6.4.3).

The most extensive results from the Tertiary sediments and volcanites were obtained in the Northern Pannonian Palaeogene Basin. They were presented by Marton and Marton (1996). The amount of data is so great that only mean directions from selected stratigraphical intervals are presented in the Table 37. According to the authors no major rotation occurred between the Eocene and Ottnangian. Mean palaeomagnetic directions from the Eocene and pre-Ottnangian Miocene are alike (Tab. 37). Mean Oligocene palaeodirection is slightly deviated but the data are scarce and dispersed. In the Ottnangian 40–60° counter-clockwise rotation took place. Second counter-clockwise rotation occurred between Karpatian and Ba-

Table 35

Tertiary palaeomagnetic data from the Outer West Carpathians

No.	Region	Age and lithology	Dc	Ic	α_{95}	k	n/N	References
1	Dukla unit E. Slovakia	Eocene red claystones	159	–40	3.5	10.4	165/5	Korab *et al.*, 1981 (*fide* Krs *et al.*, 1982)
2	Magura unit, Biele Karpaty W. Slovakia	Upper Palaeocene flysch	121 126 135	–42 –40 –35	11.4 7.5 6.4	46.2 13 10.3	31/5 31/5 18/2	Krs *et al.*, 1993
3	Magura unit, Oravska Magura W. Slovakia	Up. Palaeocene–Mid. Eocene flysch	123	–59	9.8	5.3	49/4	Krs *et al.*, 1991
4	Pieniny, Mount Wżar	Miocene andesites	191	–73	9.4	17.5	52/15	Birkenmajer, Nairn, 1968; Kruczyk, 1970

n — number of samples, N — number of sites; other explanations as in the Table 29

Table 36

Tertiary palaeomagnetic data from the Central Carpathian and Peri-Klippen Palaeogene

No.	Region	Age and lithology	Dc	Ic	α_{95}	k	N/N$_o$	References
1	Omastina, Banovska Kotlina W. Slovakia	Eocene flysch	83	–28	10	24	10/25	Tunyi, Marton, 1996
2	Demjata, E. Slovakia	Eocene flysch	75	–54	19	9	9/13	Tunyi, Marton, 1996
3*	Horne Srnie, W. Slovakia	Eocene flysch	85** 119	55** 8	18	15	5/8	Tunyi, Marton, 1996
4*	Sambron, E. Slovakia	Eocene flysch	93** 142	55** 9	14	21	7/8	Tunyi, Marton 1996
5	Solosnica, Male Karpaty	Eocene limestone	106	–66	13	13	11/15	Marton *et al.*, 1992
6	Roh Motel	Eocene mudstone	250	21	13	29	6/6	Marton *et al.*, 1992
7	Male Karpaty	Miocene clastics	315	64	10	35	7 localities	Kovač, Tunyi, 1995
8	Podhale Basin	Oligocene flysch	298	53	6	121	33/47 6 localities	Marton *et al.*, 1999

N/N$_o$ — number of used/collected samples; other explanations as in the Table 29

*not fully removed post-folding remagnetization, **declination and inclination before tectonic correction

Table 37

Tertiary palaeomagnetic data from the Northern Pannonian Palaeogene Basin

Age	Dc	Ic	α_{95}	k	N	PLAT	References
Eocene	108	–53	5.9	61.4	11	34	Marton, Mauritsch, 1990 Marton *et al.*, 1992 Marton, Marton, 1996
Oligocene	121	–51	19.9	15.6	5	32	Marton, Mauritsch, 1990 Marton *et al.*, 1992 Marton, Marton, 1996
Miocene pre-Ottnangian	101	–51	6.2	27.4	21	32	Marton, Marton, 1996
Ottnangian–Karpatian	153	–42	4.8	28.3	33	25	Marton, Marton, 1996
Badenian–Sarmatian	158	–76	14	77.6	3		Marton, Marton, 1996

N — number of localities; PLAT — palaeolatitude; other explanations as in the Table 29

Table 38

Post-folding Tertiary remagnetizations of the Triassic rocks in the Inner West Carpathians after Kruczyk *et al.* (1998)

Magnetization age	D	I	α_{95}	k	N
Oligocene	98	–69	5	12	1
Middle Miocene	324	54	7	113	5

N — number of localities; other explanations as in the Table 29

denian. Striking feature is quite low inclination of Eocene–Miocene directions from the Northern Pannonian Palaeogene Basin. Mean value is about 50°. Even lower inclination (42°) for the Ottnangian–Karpatian might be the effect of magnetic foliation (see Fig. 13 in Marton, Marton, *op. cit.*). The Badenian–Sarmatian palaeoinclinations are on the other hand slightly steeper than expected. They are in sound with the inclinations from the Pieniny andesites.

Recently obtained palaeomagnetic data from the Slovak Karst (Inner West Carpathians, Kruczyk *et al.*, 1998) were interpreted as post-folding Tertiary remagnetizations (Tab. 38). Timing of remagnetization was referred to the kinematics of the Northern Pannonian Unit outlined above (after Marton, Marton, 1996). Reversed component was acquired before the Ottnangian rotation (probably in the Oligocene). Age of the normal component was established as Middle Miocene, between two rotational phase (between Ottnangian and Badenian).

6.4.2. GEOLOGICAL INTERPRETATION OF TERTIARY PALAEOMAGNETIC DATA

Attempts of geological interpretations of Tertiary palaeomagnetic data were made by Balla (1987), Marton (1993), Nemčok, Nemčok (1994), Kovač, Tunyi (1995), Mauritsch, Marton (1995), Marton, Fodor (1995) and Marton *et al.* (1996). Generally all models link the counter-clockwise rotated declinations in the Carpathian-Pannonian area (Tabs.

35–38) with its eastward tectonic escape from the domain of Alpine collision into the unconstrained realm of the Carpathian subduction zone. The tectonic escape started after Late Cretaceous and was completed by the Middle Miocene but different time constraints are suggested by various authors (for review see Marton, 1993). Different mechanisms of rotation were proposed as well. According to Balla (1987) the entire North Pannonian unit rotated as a rigid body and the rotation was forced by the clockwise rotation of the Southern Pannonian (Tisza) unit. The alternative model is presented in the Figs. 4–7 (redrawn from Csontos *et al.*, 1992), where orientation of the main boundaries of the North Pannonian unit remains constant in respect to the stable European framework. This would imply that the rotations were accomodated by internal deformation i.e. by the rotation of small blocks limited by strike slip faults (Marton, Fodor, 1995). Intermediate model was proposed by Marton and Fodor (*op. cit.*) in which rigid body as well as internal deformations are responsible for counter-clockwise rotated declinations in the Northern Pannonian unit. Tectonic escape model becomes a dominant paradigma in the interpretations of Tertiary kinematics of the Eastern Alpine-Carpathian-Pannonian area (Ratschbacher *et al.*, 1991). It is beyond the scope of this study to evaluate its reliability and possible alternatives (e.g. Hejl, 1996). Nevertheless the mobility of the Central West Carpathians in Tertiary is well established by palaeomagnetic data. All palaeomagnetic directions of Mesozoic age obtained in the Central West Carpathians, should be corrected for the Tertiary rotations (see section 6.4.3).

Mauritsch and Marton (1995) and Marton *et al.* (1996) discuss the problem of anomalously shallow palaeomagnetic inclinations in the Tertiary of the Northern Pannonian Palaeogene Basin and Transdanubian Range which indicate more southerly position of the area in the Eocene–Early Miocene. Northward shift occurred after Ottnangian and before Middle Badenian (Marton *et al.*, 1996).

The problem of southerly position of the Northern Pannonia in the Tertiary deserves special attention. The palaeolatitude 32–34° (see Tab. 37) places the area in the middle of the Eocene–Miocene Mediterranean Basin much closer to the African Plate than to the southern margin of Europe. It is at variance with Tertiary plate tectonic reconstructions (i.e. Scotese *et al.*, 1988) where entire Apulian-Dinarid-Hellenic region is already amalgamated to the European plate. Similar problem was described by Beck and Schermer (1994) in respect to Aegean region, where anomalously shallow palaeoinclinations were also encountered. The authors argue that Aegean blocks drifted significantly northwards in the middle Tertiary and/or later, therefore anomalous palaeoinclinations indicate indeed southerly position of the area. Thus southerly position of the Northern Pannonia in the Tertiary is not unlikely. Geological as well as geophysical evidences (for example possibility of the non-dipolar components in the geomagnetic field, see Westphal, 1993) should be critically evaluated.

To summarize, palaeomagnetic data point to a very high mobilism of the Carpathian-Pannonian region, especially in the Early Miocene. Latitudinal position of the area between Eocene and Early Miocene is not clear.

6.4.3. MESOZOIC POLES FROM THE TATRA MTS. CORRECTED FOR TERTIARY MOVEMENTS

As was already mentioned the amplitude of Tertiary counter-clockwise tectonic rotation in the Podhale area is estimated as ca. 60° (Marton *et al.*, 1999). Its application brings following implications:

1. Mesozoic palaeopoles from the Tatra Mts. are shifted to the cluster of Late Jurassic/Neocomian palaeopoles from the Northern Calcareous Alps (which probably did not rotate much in the Tertiary, see Mauritsch, Marton, 1995) (Fig. 84B).

2. About 90° difference persists between the palaeopoles from the Tatra Mts. and coeval poles from the Outer West Carpathians, Inner West Carpathians and Northern Pannonia (Fig. 85).

Mesozoic palaeopoles from the Tatra can not be treated as reference poles for the entire Central West Carpathians because post-Late Cretaceous tectonic rotations took place between particular Central Carpathians massifs (Kruczyk *et al.*, 1992; Grabowski, Nemčok, 1999). Nevertheless the palaeopoles from the Central West Carpathians are closer to the palaeopoles from the Northern Calcareous Alps than from the Inner West Carpathians and Outer West Carpathians. There is a general agreement between geologists that the Austroalpine units and Central West Carpathians constituted a single tectonic domain in the Mesozoic (Tollmann, 1990; Schmidt *et al.*, 1991; Häusler *et al.*, 1993; Haas *et al.*, 1995; Kozur, Mock, 1996). The agreement of pre-Gosau palaeopoles from the Northern Calcareous Alps and Tatra Mts. after subtraction of Tertiary rotation of Podhale Basin would be a first palaeomagnetic evidence supporting this concept (Fig. 84B). Upper Austroalpine units from the Northern Calcareous Alps were subjected to about 70° clockwise rotation between Neocomian and Danian (Mauritsch, Marton, 1995). This movement did not affect the area of the Northern Pannonia, Inner West Carpathians and southern Alps (Adriatic plate). On the other hand the Central West Carpathians very likely participated in this rotation as indicated by generally clockwise trend of rotated palaeomagnetic declinations. The model seems to be promising because it incorporates geologic as well as palaeomagnetic observations. Separation of the Central West Carpathians from the Northern Calcareous Alps must have taken place between the Late Cretaceous and Eocene. In the Northern Calcareous Alps important northward thrusting took place in the Late Eocene (Unrug, 1982; Windley, 1997). This event did not affect the Central West Carpathians where the Late Cretaceous tectonic structures are still preserved and they have not been subjected to intensive Tertiary deformations (Tomek, 1993). This is well explained by the “tectonic escape” model: the Central West Carpathians in the Eocene must have been already outside the main Alpine collisional zone. Existence of a Mesozoic microplate consisting of the Central West Carpathians and Austroalpine units might be accepted as working hypothesis which could be palaeomagnetically tested. It requires an extensive palaeomagnetic study in Slovakia in order to establish the magnitude and age of local tectonic rotations in the Central West Carpathians block which certainly took place during the Late Cretaceous thrusting, Gosau basin formation

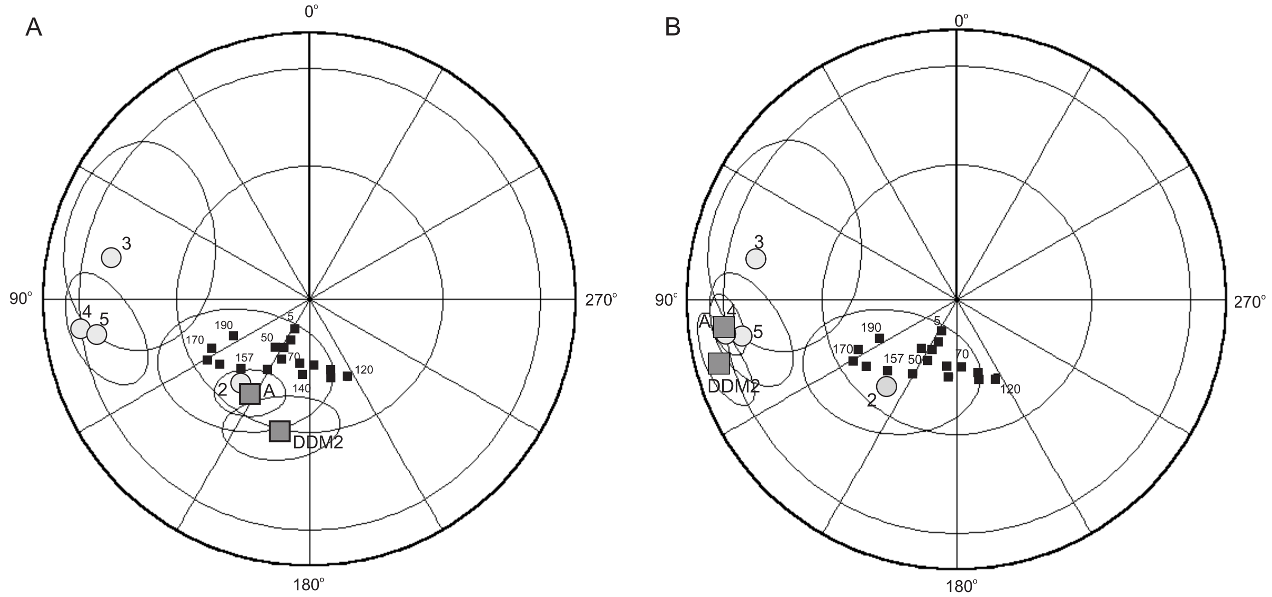


Fig. 84. Comparison of Mesozoic palaeopoles from the Tatra Mts. (big squares) and Northern Calcacerous Alps (NCA) (grey dots) with the European APWP of Besse and Courtillot (1991)

A — all palaeopoles in their present day coordinates; **B** — palaeopoles from the NCA in the present day coordinates, palaeopoles from the Tatra Mts. rotated to their pre-Oligocene position

A — Late Cretaceous syntectonic component from the parautochthonous unit (Grabowski, 1997a); DDM2 — primary(?) Berriasian component from Bobrowiec unit (section 5.3.2). Numbering of NCA palaeopoles the same as in the Table 31

and Tertiary tectonic escape. Existence of these rotations were preliminarily reported by Kruczyk *et al*. (1992). Counter--clockwise rotations are expected in the western part of the Central West Carpathians (e.g. Mala Fatra) while clockwise rotations prevail in their central and eastern part. The interpretation of rotation patterns in the Central West Carpathians must also take into account activity of large strike-slip faults of Tertiary age (Doktór *et al*., 1985; Bac-Moszaszwili, 1993). It is quite likely that palaeomagnetic direction from Magura Spišska (No. 3 in Tab. 29, see also Fig. 83) is affected by local rotation

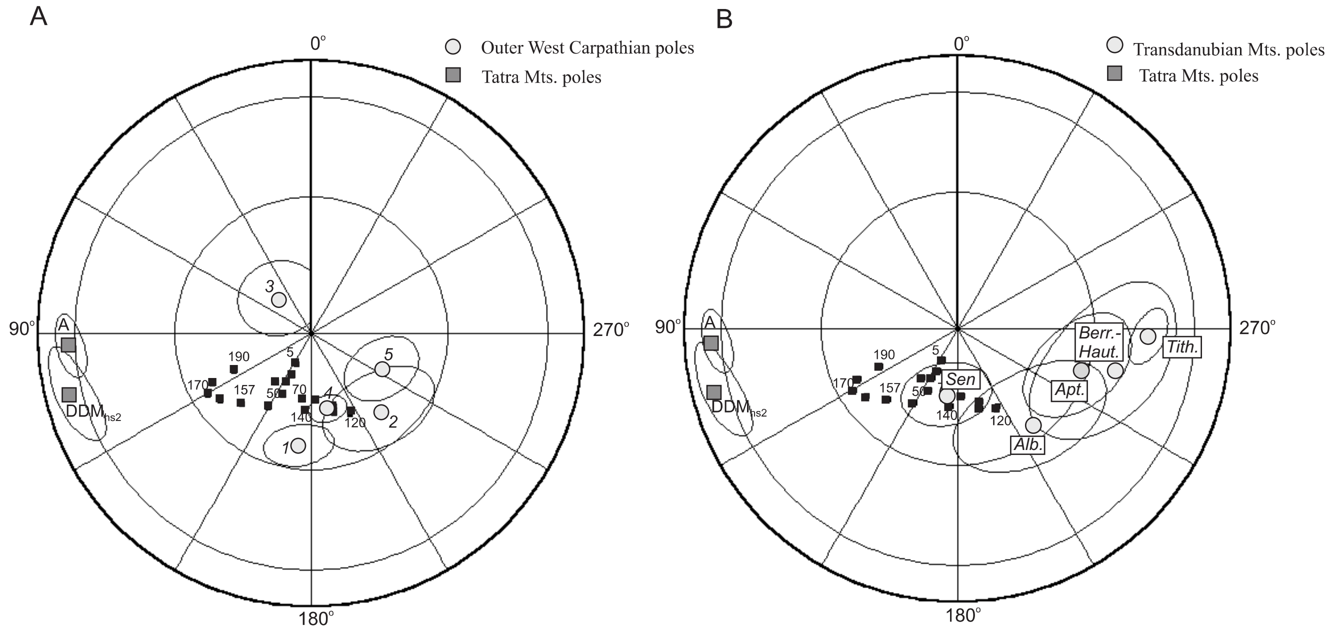


Fig. 85. Comparison of Mesozoic palaeopoles from the Tatra Mts. (big squares) with palaeopoles from the Outer West Carpathians (OWC) (Fig. A — grey dots) and Transdanubian Mts. (Fig. B — grey dots)

All palaeopoles are rotated to their pre-Tertiary position. European APWP of Besse and Courtillot (1991) is given as a reference. Palaeopoles for OWC and Transdanubian Mts. are listed in the Tables 30 and 33

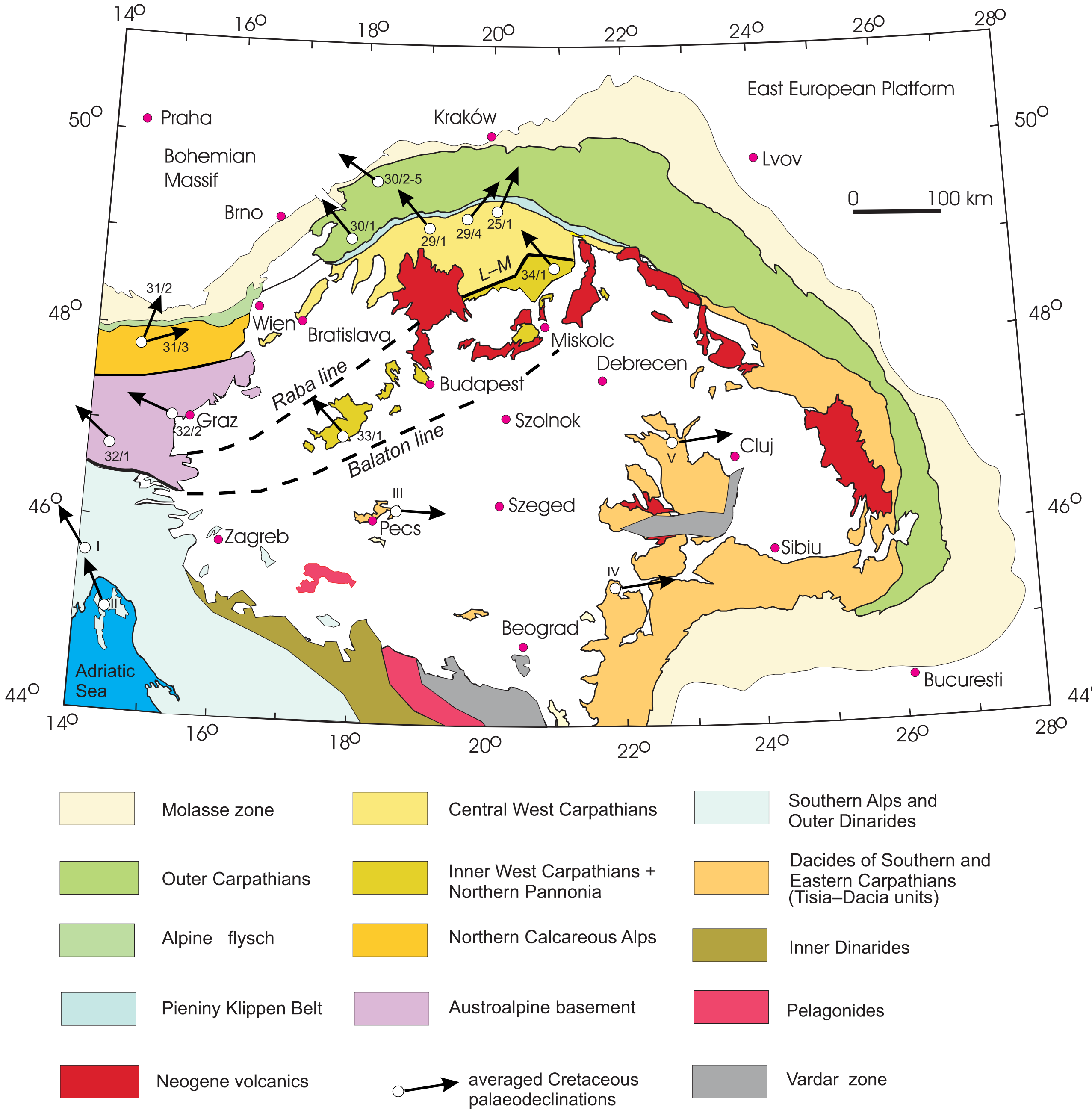


Fig. 86. Cretaceous palaeomagnetic declinations in the Alpine-Carpathian-Pannonian domain

L-M — Lubenik–Margecany line

Palaeomagnetic results quoted in the Text–Tables are indicated by no. of Table vs. no. of entry (e.g. 30/1). Palaeomagnetic results not quoted in the Text–Tables: I — Valanginian–Maastrichtian limestones north of "autochthonous" Istria (Marton, Veljovic, 1987); II — Lower to Upper Cretaceous limestones of Cres and Krk Islands (Marton *et al*., 1990); III — Cretaceous volcanites of Mecsek Mts. (Marton, Mauritsch, 1990); IV and V — Upper Cretaceous magmatic rocks of Apuseni and Banat areas (Patrascu *et al*., 1992)

in the vicinity of the RuŽbachy fault (Grabowski, Nemčok, 1999). A better control on palaeomagnetic data from Eastern Alps is also desired. It seems that palaeomagnetic data from the Northern Calcareous Alps can not be treated as reference for entire Austroalpine domain. Results obtained from Permian and Upper Cretaceous sediments between Periadriatic line and Northern Calcareous Alps (in Carinthia and Styria, see Table 32 and Fig. 86) indicate counter-clockwise rotation, similarly as Southern Alpine realm situated south of the Periadriatic fault (Mauritsch, Becke, 1987) and alike as in the Inner West Carpathians and Transdanubian Mts.

Cretaceous palaeomagnetic declinations in the Alpine-Carpathian-Pannonian domain (in the present day coordinates) are presented in the Figure 86. Disagreement of the Cretaceous palaeomagnetic poles from the Outer West Carpathians and Tatra Mts. remains a major puzzle. It is still too early for its geological interpretation because there is a lack of palaeomagnetic data from Outer West Carpathians along the Kraków–Zakopane traverse. However, it is not excluded that the Outer West Carpathians would reveal systematically higher magnitudes of the counter-clockwise rotations of Cretaceous palaeodeclinations than the Central West Carpathians where clockwise rotations prevail. This would be an evidence that pre-Tertiary clockwise rotation of the Central West Carpathians took place along the Pieniny Klippen Belt. The sense of rotation would be in agreement with that postulated by Birkenmajer (1985), however there is a dramatic difference in the age interpretations: according to Birkenmajer (*op. cit.*) the clockwise rotation of the Central West Carpathians took place in the Early Miocene. Clearly the Pieniny Klippen Belt and the neighbouring Outer Carpathian flysch nappes require a separate palaeomagnetic study in order to verify the Tertiary kinematics of the area. At present state of palaeomagnetic investigations the sinistral strike-slip movements documented by Birkenmajer (*op. cit.*) on the Pieniny Klippen Belt boundary faults can not be interpreted as result of the Early Miocene clockwise rotation or the Central Carpathian block around local vertical axis.

Counter-clockwise rotations of Mesozoic palaeomagnetic declinations in the Inner West Carpathians and Northern Pannonia (Tabs. 33 and 34) indicate rotational movements between these areas and the Central West Carpathians. The age of rotation would be probably Cretaceous/pre-Eocene since Eocene and later palaeodeclinations are quite concordant in the entire Western Carpathian–Pannonian realm. The nature of this rotation is still poorly understood. However, the rotation might be tentatively linked with the rotation which took place between the Southern Alps and Northern Calcareous Alps in the Cretaceous (Channell *et al.*, 1992a). Dextral transpression operated at the southern margin of the Northern Calcareous Alps after closure of the Hallstatt trough. These transpressional phenomena might take place also at the Central West Carpathians/Inner West Carpathians interface after closure of the Meliata ocean, which is regarded as prolongation of the Hallstatt through to the east (Kozur, 1991). Here again the problem of correlation between the Alpine and the Central/Inner West Carpathian tectonic units is approached (Häusler *et al.*, 1993; Kozur, Mock, 1996) which should be redefined including the still increasing palaeomagnetic data base from the Alpine-Carpathian-Pannonian realm.

7. CONCLUSIONS

7.1. PALAEOMAGNETISM OF MESOZOIC ROCKS IN THE TATRA MTS.

All palaeomagnetically investigated sedimentary rocks in the Tatra Mts. were remagnetized. In most localities from the High-Tatric and Sub-Tatric units two components of magnetization of different stabilities were observed. They are carried by magnetite and maghemite(?) (subordinately by hematite). The low stability component was demagnetized up to 20 mT and 250°C. It consists mostly of recent viscous overprint and it was not used for palaeotectonic interpretations. High stability component was demagnetized between 20 and 60–100 mT and between 350 and 500°C. It occurred in the rocks of variegated age — from Middle Triassic up to Lower Cretaceous, revealed always normal polarity and predominantly moderately steep (50–70°) inclinations. It was interpreted as syntectonic magnetization acquired during the Late Cretaceous thrusting. Fold test in some KriŽna (Bobrowiec and Suchy Wierch) units revealed that the remagnetization had occurred before the internal deformations of these units took place. In one locality ("Biancone" limestones of Berriasian age) the mixed polarity component DDM2 was noted which revealed maximum unblocking temperatures higher than 500°C. The component passed the reversal test and was preliminarily interpreted as primary. However, different hysteresis parameters and maximum unblocking temperatures in normally and reversely magnetized samples indicate complex rock magnetic properties and further investigations should be performed to prove the primary nature of the component DDM2.

7.2. GEOLOGICAL IMPLICATIONS OF PALAEOMAGNETIC RESULTS

Identification of syntectonic Late Cretaceous remagnetization let to determine the dip of strata in the investigated tectonic units during thrusting. The most numerous and reliable data were obtained in this and earlier studies from the High Tatric parautochthon and Bobrowiec, Hawrań, and Suchy Wierch units belonging to the KriŽna nappe. Parautochthon was remagnetized in roughly horizontal position (±5°) while the mentioned KriŽna units were dipping at least 10–20° to the S to SW during magnetization. This implies that horizontal compression might be an important factor of their emplacement. However, in the Samkowa Czuba unit palaeomagnetic directions indicate that rocks were magnetized dipping 20–60° to the north thus the attitude of strata during overthrusting was complex. Palaeolatitude of the Tatra Mts. in the Late Cretaceous amounts to 30–36° N. Possibly primary component isolated in the Berriasian limestones indicate palaeolatitude 21–23° N which is closer to the African/Adriatic than European plate. Palaeodeclinations of Mesozoic components reveal

20–50° clockwise rotation of the parautochthonous unit and KriŽna nappe in relation to the European platform. These are most likely resultant values of ca. 60° counterclockwise rotation after Oligocene and 80–110° clockwise rotation between Cenomanian–Turonian and Eocene. After subtracting the effect of Tertiary rotation Mesozoic palaeopoles from the Tatra Mts. are matched with pre-Gosau palaeopoles from the Northern Calcareous Alps. These two rotational events are most probably characteristic also for the Central West Carpathians in Slovakia, however, their magnitude is variable due to local tectonic effects. Existing palaeomagnetic data point to palaeotectonic affinity of the Central West Carpathians and Northern Calcareous Alps in the Mesozoic. On the other hand, the Central West Carpathians reveal different rotation pattern than the areas belonging to the Adriatic plate (Southern Alps, Inner West Carpathians and Northern Pannonia). It seems that different azimuth of Cretaceous palaeodeclinations between the Central West Carpathians (predominantly clockwise rotations) and the Inner and Outer West Carpathians (exclusively counter-clockwise rotations) point to Cretaceous rotational movements along the Pieniny Klippen Belt and Meliata suture zone.

7.3. CONSIDERATIONS FOR FUTURE INVESTIGATIONS

Palaeomagnetic method might be used for solving local problems concerning palaeotectonic reconstructions in the Tatra Mts. More detailed palaeomagnetic study comprising possibly complete profile of rocks (similar to that performed for the parautochthonous and Bobrowiec units) should be done for other major tectonic units: at least Czerwone Wierchy, Giewont and Hawrań units as well as some minor tectonic units in the Zakopane part of the KriŽna nappe (more detailed study of the Suchy Wierch, Mała Świnica and Samkowa Czuba units). Such investigations would help to resolve the problem of geological structure of the Tatra Mts. area in the Late Cretaceous: what was the attitude of strata in the overthrusted units, what was the likely relief of the High Tatric substratum during overthrusting of the Sub-Tatric units and what was the part of gravitational component in the nappe emplacement.

Mesozoic rocks from the Tatra Mts. are generally not suitable for magnetostratigraphic investigations. The exception are “Biancone” limestones of KriŽna nappe where it seems possible to establish the polarity pattern at the Jurassic/Cretaceous boundary.

Further rock magnetic studies (especially SEM/STEM and XRD investigations of magnetic extracts and isotopic analyses) would shed some light on mechanisms of chemical remagnetization and diagenetic/burial history of sedimentary rocks in the Tatra Mts. Some kind of rocks revealed quite well defined AMS ellipsoids which might be used (in combination with structural study) for evaluation of the Mesozoic and Tertiary stress reconstructions of the area.

As was already indicated palaeomagnetic results from the Tatra Mts. are not representative for the Central West Carpathians as a whole. Therefore, systematic palaeomagnetic study should be performed in the “core mountains” of Slovakia comprising the investigations of Mesozoic rocks of the parautochthonous and overthrusted units together with their Tertiary cover in Palaeogene and Neogene basins. The question of local tectonic rotations within the Central West Carpathians block and problem of its Tertiary palaeolatitude should be revealed in these studies. It is not until then that the synthetic Mesozoic-Tertiary apparent polar wander path for the Central West Carpathians might be reconstructed. The separate question is the reconstruction of the Cretaceous/Tertiary kinematics of the Pieniny Klippen Belt area and possible rotations between the Central and Outer West Carpathians. A systematic palaeomagnetic traverse of Carpathians at the meridian of Cracow would contribute to the solution of the problem.

Acknowledgements. I am deeply grateful to supervisor of my Ph.D. thesis, doc. dr hab. Magdalena Kądziałko-Hofmokl for her guidance and support throughout this work, especially at its final stages.

I would like to express my thanks to the palaeomagnetic staff at the Institute of Geophysics, Polish Academy of Sciences, for fruitful scientific discussions and enabling to perform a part of palaeomagnetic work at the Laboratory. Furthermore, I would like to thank Dr Mark Dekkers (University of Utrecht), Prof. Dr Friedrich Heller (ETH Zürich), Dr Mark Hounslow and Dr Barbara Maher (University of East Anglia — Norwich) for their help and permission to use the instrumentation and facilities in palaeomagnetic laboratories in Utrecht, Zürich and Norwich. Doc. dr hab. M. Jeleńska (Inst. of Geophysics, Pol. Acad. of Sci.) and Prof. dr hab. T. Peryt (Polish Geological Institute) revised the manuscript, which is gratefully acknowledged. Thanks are due to Jerzy Nawrocki (PGI) for stimulating discussions and his support during preparation of this thesis. Special thanks should be given to Prof. dr hab. Michał Szulczewski (Warsaw University) who encouraged me to undertake the palaeomagnetic study in the Tatra Mts. during my employment at the University.

I am indebted to colleagues of EUROPROBE — PANCARDI group for important and valuable discussions concerning the palaeomagnetism and tectonics of Carpathian area. Especially constructive comments on geological interpretations of palaeomagnetic results were given by Emö Marton, Michal Nemčok and Čestmir Tomek.

I am grateful to the Polish Committee for Scientific Research for the financial support of palaeomagnetic investigations in the Tatra Mts. (research grant No. 6P04D 020 11 Rock magnetic properties of the sedimentary rocks in the Tatra Mts.). It is a part of the PANCARDI sub-project: Palaeomagnetic constraints of the Tertiary Carpathian evolutionary model.

I wish to thank the authorities of the Tatra National Park for the permission to carry out the field work.

I acknowledge the help of all my friends during the field work. I am very thankful to Andrzej Jagielski, Asia and Jarek Przygoda, Maciek Tyniec, Kinga and Krzyś Walasek, Monika Wiechowska, Andrzej Wilamowski, Basia Wiśniewska, Rafał Zawadzki and finally to Wojtek Dudziak for his hospitality in the hut at Siwa Polana.

REFERENCES

ARGAND E., 1924 — La tectonique de l'Asie. Proceedings of the International Geological Congress, XIII: 171–372.

BAC M., 1971 — Tectonics of the Bobrowiec unit in the Western Tatra Mts. [Eng. Sum.]. *Acta Geol. Pol.*, **21**: 279–317.

BAC-MOSZASZWILI M., 1993 — Structure of the western termination of the Tatra Massif. *Ann. Soc. Geol. Pol.*, **63**: 167–193.

BAC-MOSZASZWILI M., 1995 — Diversity of Neogene and Quaternary tectonic movements in the Tatra Mountains [Eng. Sum.]. *Folia Quaternaria*, **66**: 131–144.

BAC-MOSZASZWILI M., 1998 — Geology of the Subtatric units, Western Tatra Mts., Poland [Eng. Sum.]. *Stud. Geol. Pol.*, **111**: 113–136.

BAC-MOSZASZWILI M., BURCHART J., GŁAZEK J., IWANOW A., JAROSZEWSKI W., KOTAŃSKI Z., LEFELD J., MASTELLA L., OZIMKOWSKI W., RONIEWICZ P., SKUPIŃSKI A., WESTWALEWICZ-MOGILSKA E., 1979 — Mapa geologiczna Tatr Polskich w skali 1:30 000. Wyd. Geol. Warszawa.

BAC-MOSZASZWILI M., JAROSZEWSKI W., PASSENDORFER E., 1984 — On the tectonics of Czerwone Wierchy and Giewont area in the Tatra Mts., Poland [Eng. Sum.]. *Ann. Soc. Geol. Pol.*, **52**: 67–88.

BALLA Z., 1987 — Tertiary palaeomagnetic data for the Carpatho--Pannonian region in the light of Miocene rotation kinematics. *Tectonophysics*, **139**: 67–98.

BECK M. E., SCHERMER E. R., 1994 — Aegan palaeomagnetic inclination anomalies. Is there a tectonic explanation? *Tectonophysics*, **231**, 281–292.

BESSE J., COURTILLOT V., 1991 — Revised and synthetic apparenat polar wander paths of the African, Eurasian, North American and Indian Plates, and true polar wander since 200 Ma. *Jour. Geoph. Res.*, **96**, B3: 4029–4050.

BIELIK M., 1994 — Continental convergence in the area of the Western Carpathians on the basis of density modelling. *Geol. Carpath.*, **46**: 3–12.

BIELY A., 1990 — The geological structure of the West Carpathians. *In*: Mem. Soc. Geol. France, Paris, Nouvelle Serie, **154** (II): 51–57.

BIRKENMAJER K., 1976 — The Carpathian orogen and plate tectonics. *Publ. Inst. Geophys. Pol. Acad. Sc. A-2*, **101**: 43–53.

BIRKENMAJER K., 1984 — Interrelation of Neogene tectonics and volcanism in the Pieniny Klippen Belt of Poland. *Acta Geodaet., Geophys., et Montanist. Hung.*: 37–48.

BIRKENMAJER K., 1985 — Major strike-slip faults of the Pieniny Klippen Belt and the Tertiary rotation of the Carpathians. *Publ. Inst. Geophys. Pol. Acad. Sc.*, **A-16** (175): 101–115.

BIRKENMAJER K., 1986 — Stages of structural evolution of the Pieniny Klippen Belt, Carpathians. *Stud. Geol. Pol.*, **88**: 7–32.

BIRKENMAJER K., 1988 — Exotic Andrusov Ridge: its role in plate tectonic evolution of the West Carpathian fold belt. *Stud. Geol. Pol.*, **91**: 7–37.

BIRKENMAJER K., DELITALA M. C., NICOLETTI M., PETRUCCIONI C., 1987 — K-Ar dating of andesite intrusions (Miocene). Pieniny Klippen Belt, Carpathians. *Bull. Acad. Pol. Sc.. Ser. Sc. Terr.*, **35**: 11–19.

BIRKENMAJER K., NAIRN A. E. M., 1968 — Palaeomagnetic studies of Polish rocks. III. Neogene igneous rocks of the Pieniny Mountains, Carpathians. *Ann. Soc. Geol. Pol.*, **38**: 475–489.

BŁASZYK J., GAŻDZICKI A., 1982 — Lower Lias ostracodes of the Tatra Mts (West Carpathians). *Acta Palaeont. Pol.*, **27**: 129–136.

BULLARD E. C., EVERETT J. E., SMITH A. G., 1965 — A symposium on continental drift — IV. The fit of the continents around the Atlantic. *Phil. Trans. Roy. Soc.*, **258**, 41–51.

BURCHART J., 1968 — Rubidium-Strontium isochron ages of the crystalline core of the Tatra Mountains, Poland. *Amer. J. Sci.*, **266**: 895–907.

BURCHART J., 1972 — Fission-track age determinations of accessory apatite from the Tatra Mountains, Poland. *Earth. Planet. Sc. Lett.*, **15**: 418–422.

BURCHFIEL B. C., 1980 — Eastern European Alpine system and the Carpathian orocline as an example of collision tectonics. *Tectonophysics*, **63**: 31–61.

CHANNELL J. E. T., 1996 — Palaeomagnetism and palaeogeography of Adria. *In*: A. Morris, D. H. Tarling (Eds.), Palaeomagnetism and Tectonics of the Mediterranean region. *Geol. Soc. Sp. Publ.*, **105**: 119–132.

CHANNELL J. E. T., BRANDNER R., SPIELER A., STONER J. S., 1992 — Palaeomagnetism and palaeogeography of the Northern Calcareous Alps (Austria). *Tectonics*, **11**: 792–810.

CHANNELL J. E. T., D'ARGENIO B., HORVATH F., 1979 — Adria, the African promontory, in Mesozoic Mediterranean palaeogeography. *Earth Sc. Rev.*, **15**: 213–292.

CHANNELL J. E. T., DOGLIONI C., 1994 — Early Triassic palaeomagnetic data from Dolomites (Italy). *Tectonics*, **13**: 157–166.

CHANNELL J. E. T., DOGLIONI C., STONER J. S., 1992 — Jurassic and Cretaceous palaeomagnetic data from the Southern Alps (Italy). *Tectonics*, **11**: 811–822.

CHANNELL J. E. T., KOZUR H., 1997 — How many oceans? Meliata, Vardar and Pindos oceans in Mesozoic Alpine palaeogeography. *Geology*, **25** (2): 183–186.

CHANNELL J. E. T., McCABE C., 1994 — Comparison of magnetic hysteresis parameters of remagnetized and unremagnetized limestones. *J. Geoph. Res.*, **99**: 4613–4623.

CSASZAR G., GALACZ A., HAAS J., KAZMER M., KOVACS S., NAGYMAROSY A., SZENTGYÖRGYI K., VÖRÖS A., 1990 — Palaeogeography of the Pannonian Basin. *In*: *Mem. Soc. Geol. France*, Paris, *Nouvelle Serie*, **154** (II): 63–89.

CSONTOS L., NAGYMAROSY A., HORVATH F., KOVAČ M., 1992 — Tertiary evolution of the Intra-Carpathian area: a model. *Tectonophysics*, **208**: 221–241.

DAY R., FULLER M., SCHMIDT V. A., 1977 — Hysteresis properties of titanomagnetites: grain-size and compositional dependence. *Phys. Earth Planet. Int.*, **13**: 260–267.

DERCOURT J., ZONENSHAIN L. P., RICOU L.-E., KAZMIN V. G., LE PICHON X. *et al.*, 1986 — Geological evolution of the Tethys belt from the Atlantic to the Pamirs since the Lias. *Tectonophysics*, **123**: 241–315.

DOKTÓR S., DORNIČ J., GRANICZNY M., REICHWALDER P., 1985 — Structural elements of Western Carpathians and their Foredeep on the basis of satellite interpretation. *Geol. Quarter.*, **29**: 129–138.

DOWNES H., 1996 — Neogene magmatism and tectonics in the Carpatho--Pannonian region. Mitteil. der Geol. and Bergbaustud. in Osterreich, 41: 104–105. PANCARDI Workshop 1996.

DUMONT T., WIECZOREK J., BOUILLIN J-P., 1996 — Inverted Mesozoic rift structures in the Polish Western Carpathians (High-Tatric units). Comparison with similar features in the Western Alps. *Eclogae geol. Helv.*, **89**/1: 181–202.

EDEL J. B., DURINGER P., 1997 — The apparent polar wander path of the European Plate in Upper Triassic-Lower Jurassic times and the Liassic intraplate fracturing of Pangea: new palaeomagnetic constraints from NW France and SW Germany. *Geophys. J. Int.*, **128**: 331–344.

EPSTEIN A. G., EPSTEIN J. B., HARRIS L. D., 1977 — Conodont Color Alteration — an Index to Organic Metamorphism. *U. S.Geol. Surv. Prof. Paper*, **995**: 1–27.

FARYAD S. W., 1997 — Petrological model for blueschist facies metamorphism in the Pieniny Klippen Belt. *In*: P. Grecula, D. Hovorka, M. Putiš (Eds.), Geological evolution of the Western Carpathians. Mineralia Slovaca — Monograph, Bratislava: 155–162.

FORSTER TH., EVANS M. E., HELLER F., 1994 — The frequency dependence of low field susceptibility in loess sediments. *Geophys. J. Int*., **118**: 636–642.

FRISCH W., 1979 — Tectonic progradation and plate tectonic evolution of the Alps. *Tectonophysics*, **60**: 121–139.

GALBRUN B., 1985 — Magnetostratigraphy of the Berriassian stratotype section (Berrias, France). *Earth Planet. Sc. Lett*., **74**: 130–136.

GAWĘDA A., 1995 — Geochemistry and Rb/Sr isochron age of pegmatites from the Western Tatra Mts, (S-Poland). *Geol. Carpath*., **46**: 95–99.

GEALEY W. K., 1988 — Plate tectonic evolution of the Mediterranean--Middle East region. *Tectonophysics*, **155**: 285–306.

GARCES M., PARES J. M., CABRERA L., 1996 — Inclination error linked to sedimentary facies in Miocene detrital sequences from the Valles — Penedes Basin (NE Spain). *In*: A. Morris, D. H. Tarling (Eds.), 1996, Palaeomagnetism and tectonics of the Mediterranean region. *Geol. Soc. Sp. Pub*., **105**: 91–99.

GAŽDZICKI A., 1974 — Rhaetian microfacies, stratigraphy and facies development in the Tatra Mts. *Acta Geol. Pol*., **24**: 17–96.

GAŽDZICKI A., LEFELD J., 1997 — Seria reglowa dolna. Jura dolna. *In*: J. Lefeld, A. Gaździcki (Eds.) Przewodnik LXVIII Zjazdu Polskiego Towarzystwa Geologicznego. Zakopane 2–4 października 1997: 59–61.

GAŽDZICKI A., ZAWIDZKA K., 1973 — Triassic foraminifer assemblages in the Choč nappe of the Tatra Mts. *Acta Geol. Pol*., **23**: 483–490.

GIALANELLA P. R., HELLER F., INCORONATO A., 1994 — Rock magnetism of deformed Upper Triassic limestones from the Lagonegro Basin (Southern Appenines, Italy). *Geophys. Res. Lett*., **21**(24): 2665–2668.

GRABOWSKI J., 1995a — New palaeomagnetic data from the Lower Sub-Tatric radiolarites, Upper Jurassic (Western Tatra Mts.). *Geol. Quarter*., **39**: 61–74.

GRABOWSKI J., 1995b — Palaeomagnetic investigations of the High Tatric series in the Tatra Mts. [Eng. Sum.]. *Prz. Geol*., **43**(2): 113–116.

GRABOWSKI J., 1996 — Magnetic fabric of the Upper Jurassic sediments, KriŽna unit, Tatra Mts., Poland. *Geol. Carpath*., **47**: 331–337.

GRABOWSKI J., 1997a — Palaeomagnetic results from the cover (High Tatric) unit and Nummulitic Eocene in the Tatra Mts. (Central West Carpathians, Poland) and their tectonic implications. *Ann. Soc. Geol. Pol*., **67**: 13–23.

GRABOWSKI J., 1997b — New palaeomagnetic data from Fatricum and Hronicum in the Tatra Mts. (Poland) — further evidences for Cretaceous remagnetization in the Central West Carpathians. *Prz. Geol*., **45**(10): 1074. PANCARDI workshop 1997.

GRABOWSKI J., 1999 — Pierwotna(?) pozostałość magnetyczna w wapieniach tytano-beriasu płaszczowiny kriźniańskiej w Tatrach Zachodnich — komunikat wstępny. *Prz. Geol*., **47**, 6: 536–537.

GRABOWSKI J., NAWROCKI J., 1996 — Multiple remagnetizations in the Devonian carbonates in the north-western part of the Kielce region (Holy Cross Mts, southern part). *Geol. Quarter*., **40**: 47–64.

GRABOWSKI J., NEMČOK M., 1999 — Summary of palaeomagnetic data from the Central West Carpathians of Poland and Slovakia: evidence for the Late Cretaceous–Early Tertiary transpression. *Phys. Chem. Earth* (*A*), **24**(8): 681–685.

GRABOWSKI J., NARKIEWICZ K., POPRAWA P., 1999 — First results of palaeomagnetic and palaeothermal (CAI) investigations of the highest Sub-Tatric units in the Polish Tatra Mts [Eng. Sum.]. *Prz. Geol*., **47**(2): 153–158.

GRAHAM J. W., 1949 — The stability and significance of magnetism in sedimentary rocks. *J. Geoph. Res*., **54**: 131–167.

GUZIK K., KOTAŃSKI Z., 1963 — La tectonique de la zone subtatrique de Zakopane [French Sum.]. *Acta Geol. Pol*., **13**: 387–424.

HAAS J., KOVACS S., KRYSTYN L., LEIN R., 1995 — Significance of Late Permian–Triassic facies zones in terrane reconstructions in the Alpine-North Pannonian domain. *Tectonophysics*., **242**: 19–40.

HAILWOOD E. A., 1989 — Magnetostratigraphy. *The Geological Society. Special Report*, **19**.

HÄUSLER H., PLAŠIENKA D., POLAK M., 1993 — Comparison of Mesozoic Successions of the Central Eastern Alps and the Central Western Carpathians. *Jb. Geol. B-A*., **138**: 715–739.

HEIDER F., DUNLOP D. J., 1987 — Two types of chemical remanent magnetization during oxidation of magnetite. *Phys. Earth Planet. Int*., **46**: 24–45.

HEJL E., 1996 — The lateral extrusion of the Eastern Alps: fact or fiction? Mitteil. der Geol. and Bergbaustud. in Osterreich, 41: 111. PANCARDI Workshop 1996.

HELLER F., LOWRIE W., HIRT A. M., 1989 — A review of palaeomagnetic and magnetic anisotropy results from the Alps. *In*: M.P. Coward, D. Dietrich, R. G. Park (Eds.). Alpine Tectonics, *Geol. Soc. Sp. Publ*., **45**: 399–420.

HIRT A. M., LOWRIE W., PFIFFNER O. A., 1986 — A palaeomagnetic study of tectonically deformed red beds of the Lower Glarus nappe complex, eastern Switzerland. *Tectonics*, **5**: 723–731.

HOUŠA V., KRS M., KRSOVA M., PRUNER P., 1996 — Magnetostratigraphy of Jurassic-Cretaceous limestones in the Western Carpathians. *In*: A. Morris, D. H. Tarling (Eds.), Palaeomagnetism and tectonics of the Mediterranean region. *Geol. Soc. Sp. Publ*., **105**: 185–194.

HOVORKA D., 1996 — Mesozoic non-ophiolitic volcanics of the Carpathian arc and Pannonian Basin. *Geol. Carpath*., **47**: 63–72.

IWANOW A., 1965 — La structure geologique de l'unite de Suchy Wierch entre la vallee de Bystra et la vallee de Mała Łąka dans les Tatras [French Sum.]. *Acta Geol. Pol*., **15**: 393–428.

JACKO S., SASVARI T., 1990 — Some remarks to an emplacement mechanism of the West Carpathian palaeo-alpine nappes. *Geol. Zb.-Geol. Carpath.*, **41**: 179–197.

JANAK M., 1994 — Variscan uplift of thr crystalline basement, Tatra Mts., Central West Carpathians: evidence from Ar/ Ar laser probe dating of biotite and P-T-t paths. *Geol. Carpath*., **45**: 293–300.

JAROSZEWSKI W., 1982 — Hydrotectonic phenomena at the base of the KriŽna nappe, Tatra Mts. *In*: M. Mahel (Ed.), Alpine Structural Elements: Carpathians–Balkan–Caucasus–Pamir Orogene Zone. Veda, Bratislava: 137–148.

JELINEK V., 1977 — The statistical theory of measuring anisotropy of magnetic susceptibility of rocks and its application. Geofyz. Brno: 1–88.

JUAREZ M. T., OSETE M. L., MELENDEZ G., LOWRIE W., 1995 — Oxfordian magnetostratigraphy in the Iberian Range. *Geoph. Res. Lett*., **22**: 2889–2892.

KAZMER M., KOVACS S., 1985 — Permian–Palaeogene palaeogeography along the eastern part of the Insubric-Periadriatic Lineament system: evidence for continental escape of the Bakony–Drauzug unit. *Acta Geol. Hung*., **28**: 69–82.

KĄDZIAŁKO-HOFMOKL M., KRUCZYK J., 1987 — Palaeomagnetism of Middle-Late Jurassic sediments from Poland and implications for the polarity of the geomagnetic field. *Tectonophysics*, **139**: 53–66.

KĄDZIAŁKO-HOFMOKL M., KRUCZYK J., LEFELD J., 1985 — Results of palaeomagnetic investigations and the tectonic position of Jurassic radiolarites and limestones of the lower Sub-Tatric nappe in the Tatra Mts. *Publs. Inst. Geophys. Pol. Acad. Sc.*, **A-16** (175): 117–130.

KĄDZIAŁKO-HOFMOKL M., KRUCZYK J., LEFELD J., PAGAČ P., TUNYI I., 1990 — Palaeomagnetism of the KriŽna nappe Jurassic sediments from the Mala Fatra Mts. and tectonic implications. *Acta Geophys. Pol*., **38**(4) 457–470.

KENT D. V., 1985. — Thermoviscous remagnetization in some Appalachian limestones. *Geophys. Res. Lett.*, **12**: 805–808.

KIBITLEWSKI S., 1972 — Jointing of post nappic age in the Sub-Tatric series of the Tatra Mts [Eng. Sum.]. *Acta Geol. Pol*., **22**: 83–91.

KIRKER A., McCLELLAND E., 1996 — Application of net tectonic rotations and inclination analysis to a high resolution palaeomagnetic study in the betic Cordillera. *In*: A. Morris, D. H. Tarling (Eds.), Palaeomagnetism and tectonics of the Mediterranean region. *Geol. Soc. Sp. Publ*., **105**: 19–32.

KIRKER A. I., McCLELLAND E., 1997 — Deflection of magnetic remanence during progressive cleavage development in the Pembrokeshire Old Red Sandstone. *Geophys. J. Int*., **130**: 240–250.

KIRSCHVINK J. L., 1980 — The least square line and plane and the analysis of palaeomagnetic data. *Jour. Roy. Astr. Soc*., **62**: 699–718.

KLIGFIELD R., LOWRIE W., HIRT A., SIDDANS W. B., 1983 — Effect of progressive deformation on remanent magnetization of Permian red beds from the Alpes Maritimes (France). *Tectonophysics*, **98**: 59–85.

KORAB T., KRS M., KRSOVA M., PAGAČ P., 1981 — Palaeomagnetic investigations of Albian(?) — Palaeocene to Lower Oligocene sediments from the Dukla unit, East Slovakian Flysch, Czechoslovakia. *Zapadne Karpaty, seria geologia*, Geological Institute of Dionyz Stur, Bratislava, **7**: 127–149.

KOTAŃSKI Z., 1961 — Tectogenese et reconstitution de la palaeogeographie de la zone haut-tatrique dans les Tatras [French Sum.]. *Acta Geol. Pol*., **11**: 187–476.

KOTAŃSKI Z., 1963 — Stratigraphie et lithologie du Trias subtatrique de la region de Zakopane [French Sum.]. *Acta Geol. Pol*., **13**: 317–385.

KOTAŃSKI Z., 1971 — Przewodnik geologiczny po Tatrach. 280 pp. Wyd. Geol. Warszawa.

KOTAŃSKI Z., 1973 — Upper and Middle Sub-Tatric nappe in the Tatra Mts. *Bull. Acad. Pol. Sc*., **21**: 75–83.

KOTAŃSKI Z., 1976 — Triassic. The Carpathians. *In*: S. Sokołowski (Ed.), Geology of Poland. vol. I. Stratigraphy, part 2 Mesozoic: 105–124.

KOTAŃSKI Z., 1996 — History of discovery and age of labyrinthodont remains in the Tatra Mts, Poland. *Pr. Muzeum Ziemi* **43**: 47–52.

KOTAŃSKI Z., IWANOW A., 1997 — Trasy terenowe. Trasa A-7. *In*: J. Lefeld, A. Gaździcki (Eds.), Przewodnik LXVIII Zjazdu Polskiego Towarzystwa Geologicznego. Zakopane 2–4 października 1997: 133–140.

KOVACH A., SVINGOR E., GRECULA P., 1986 — Rb–Sr isotopic ages of granitoid rocks from the Spissko-Gemerske Rudohorie Mts., Western Carpathians, Eastern Slovakia. *Miner. Slovaca*, **18**: 1–14.

KOVAČ M., BIELIK M., LEXA J., PERESZLENYI M., ŠEFARA J., TUNYI I., VASS D., 1997 — The Western Carpathian intramountane basins. *In*: P. Grecula, D. Hovorka, M. Putiš (Eds.), Geological evolution of the Western Carpathians. Mineralia Slovaca — Monograph, Bratislava: 43–64.

KOVAČ M., KRAL J., MARTON E., PLAŠIENKA D., UHER P., 1994 — Alpine uplift history of the Central Western Carpathians: geochronological, palaeomagnetic, sedimentary and structural data. *Geol. Carpath*., **45**: 83–96.

KOVAČ M., NAGYMAROSY A., SOTAK J., ŠUTOVSKA K., 1993 — Late Tertiary palaeogeographic evolution of the Western Carpathians. *Tectonophysics*, **226**: 401–415.

KOVAČ M., TUNYI I., 1995 — Interpretation of palaeomagnetic data from Western part of the Central West Carpathians [Eng. Sum.]. *Mineralia Slovaca*, **27**/3: 213–220.

KOVACS S., 1992 — Tethys "western ends" during the Late Palaeozoic and Triassic and their possible genetic relationships. *Acta Geol. Hung*., **35**: 329–369.

KOZUR H., 1991 — The evolution of the Meliata–Hallstatt ocean and its significance for the early evolution of the Eastern Alps and Western Carpathians. *Palaeogeogr., Palaeoclimatol., Palaeoecol*., **87**: 109–135.

KOZUR H., MOCK R., 1996 — New palaeogeographic and tectonic interpretations in the Slovakian Carpathians and their implications for correlations with the Eastern Alps. Part I: Central West Carpathians. *Miner. Slov*., **28**: 151–174.

KRS M., CHVOJKA R., POTFAJ M., 1993 — Palaeomagnetic investigations in the Biele Karpaty Mts. unit, flysch belt of the Western Carpathians. *Geol. Carpath*., **45**: 35–43.

KRS M., KRSOVA M., CHVOJKA R., POTFAJ M., 1991 — Palaeomagnetic investigations of the flysch belt in the Orava region, Magura unit, Czechoslovak Western Carpathians. *Geologicke prace*, Geological Institute of Dionyz Stur, Bratislava, **92**: 135–151.

KRS M., KRSOVA M., PRUNER P., 1996 — Palaeomagnetism and palaeogeography of the Western Carpathians from the Permian to the Neogene. *In*: A. Morris, D. H. Tarling (Eds.), Palaeomagnetism and tectonics of the Mediterranean region. *Geol. Soc. Sp. Publ*., **105**: 175–184.

KRS M., MUŠKA P., PAGAČ P., 1982 — Review of palaeomagnetic investigations in the West Carpathians of Czechoslovakia. *Geologicke prace, Spravy*, Geological Institute of Dionyz Stur, Bratislava, **78**: 39–58.

KRUCZYK J., 1970 — Palaeomagnetic investigations of andesitic intrusions from Mt. Wżar [Rus. abstract]. *Publ. Inst. Geophys. Pol. Acad. Sci*., **35**: 35–83.

KRUCZYK J., JELEŃSKA M., KĄDZIAŁKO-HOFMOKL M., 1980 — Palaeomagnetism of Jurassic sediments from the Western Tatra Mts. *Acta Geophys. Pol*., **28**: 297–308.

KRUCZYK J., KĄDZIAŁKO-HOFMOKL M., 1988a — Polarity of the geomagnetic field during the Callovian. 2nd Int. Symp. on Jurassic Stratigraphy: 1139–1150, Lisbon.

KRUCZYK J. , KĄDZIAŁKO-HOFMOKL M., 1988b — Palaeomagnetism of Oxfordian sediments from the halokinetic structure in Kujawy, Central Poland. 2nd Int. Symp. on Jurassic Stratigraphy: 1151–1178, Lisbon.

KRUCZYK J., KĄDZIAŁKO-HOFMOKL M., LEFELD J., PAGAČ P., TUNYI I., 1992 — Palaeomagnetism of Jurassic sediments as evidence for oroclinal bending of the Inner West Carpathians. *Tectonophysics*, **206**: 315–324.

KRUCZYK J., KĄDZIAŁKO-HOFMOKL M., TUNYI I., PAGAČ P., MELLO J., 1998 — Palaeomagnetic study of Triassic sediments from the Silica nappe in the Slovak Karst, a new approach. *Geol. Carpath*., **49**: 33–43.

KRZYWIEC P., JOCHYM P., 1997 — Characteristics of the Miocene subduction zone of the Polish Carpathians: results of flexural modelling [Eng. Sum.]. *Prz. Geol*., **45**(8): 785–792.

KSIĄŻKIEWICZ M., 1977 — The tectonics of the Carpathians. *In*: W. Pożaryski (Ed.), Geology of Poland, vol. IV, Tectonics. Państw. Inst. Geol., Warsaw: 476–608.

LEFELD J., 1974 — Middle–Upper Jurassic and Lower Cretaceous biostratigraphy and sedimentology of the Sub-Tatric succession in the Tatra Mts. (Western Carpathians). *Acta Geol. Pol*., **24**: 277–364.

LEFELD J., 1997 — Tektogeneza Tatr. Cykl alpejski. *In*: J. Lefeld, A. Gaździcki (Ed.), Przewodnik LXVIII Zjazdu Polskiego Towarzystwa Geologicznego. Zakopane 2–4 października 1997: 16–21.

LEFELD J., 1999 — Tectonics of the Subtatric Units, Eastern Tatra Mts. *Studia Geol. Pol*., **115**: 139–166.

LEFELD J. (Ed.), GAŻDZICKI A., IWANOW A., KRAJEWSKI K., WÓJCIK K., 1985 — Jurassic and Cretaceous lithostratigraphic units of the Tatra Mountains. *Stud. Geol. Pol*., **84**: 3–93.

LEFELD J., JANKOWSKI J., 1985 — Model of deep structure of the Polish Inner Carpathians. *Publs. Inst. Geophys. Pol. Acad. Sc*., **A-16** (175): 71–100.

LEWANDOWSKI M., NOWOŻYŃSKI K., WERNER T., 1997 — PDA — a package of FORTRAN programs for palaeomagnetic data analysis (manuscript).

LOWRIE W., 1986 — Palaeomagnetism and the Adriatic promontory: a reappraisal. *Tectonics*, **5**: 797–807.

LOWRIE W., 1990 — Identification of ferromagnetic minerals in a rock by coercivity and unblocking temperature properties. *Geophys. Res. Lett*., **17**(2), 159–162.

MACHEL H. G., 1995 — Magnetic mineral assemblages and magnetic contrasts in diagenetic environments — with implications for studies of palaeomagnetism, hydrocarbon migration and exploration. *In*: P. Turner, A. Turner (Eds.), Palaeomagnetic applications in Hydrocarbon exploration and production. *Geol. Soc. Spec. Publ*., **98**: 9–29.

MAHEL M., 1968 — The Tatry Mountains. *In*: A. Matejka (Ed.), Regional Geology of Czechoslovakia, part II, The West Carpathians. Academia, Praha: 152–166.

MAHEL M., 1981 — Island character of Klippen Belt; Vahicum — continuation of Southern Penninicum in West Carpathians. *Geol. Zbor. Geol. Carpath*., **32**: 293–305.

MALUSKI H., RAJLICH P., MATTE P., 1993 — Ar–Ar dating of the Inner Carpathians Variscan basement and Alpine mylonitic overprinting. *Tectonophysics*, **223**: 313–337.

MARTON E., 1990 — Palaeomagnetism — an overview of the Central Mediterranean. *Mem. Soc. Geol. France, Paris, Nouvelle Serie*, **159** (II), 223–244.

MARTON E., 1993 — The itinerary of the Transdanubian Central Range: an assesment of relevant palaeomagnetic observations. *Acta Geol. Hung*., **37**: 135–151.

MARTON E., FODOR L., 1995 — Combination of palaeomagnetic and stress data — a case study from North Hungary. *Tectonophysics*, **242**: 99–114.

MARTON E., MARTON P., 1983 — A refined apparent polar wander curve for the Transdanubian Central Mountains and its bearing on the Mediterranean tectonic history. *Tectonophysics*, **98**: 43–57.

MARTON E., MARTON P., 1996 — Large scale rotations in North Hungary during the Neogene as indicated by palaeomagnetic data. *In*: A. Morris, D. H. Tarling (Eds.), Palaeomagnetism and tectonics of the Mediterranean region. *Geol. Soc. Spec. Public*., **105**: 153–173.

MARTON E., MARTON P., LESS G., 1988 — Palaeomagnetic evidence of tectonic rotations in the southern margin of the Inner West Carpathians. *Phys. Earth Planet. Int.*, **52**: 256–266.

MARTON E., MAURITSCH H. J., 1990 — Structural applications and discussion of a post-Palaeozoic data base for the Central Mediterranean. *Phys. Earth. Planet. Int*., **62**: 46–59.

MARTON E., MILICEVIC V., VELJOVIC D., 1990 — Palaeomagnetism of the Kvarner islands, Yugoslavia. *Phys. Earth Planet. Int*., **62**: 70–81.

MARTON E., PAGAČ P., TUNYI I., 1992 — Palaeomagnetic investigations on Late Cretaceous–Cenozoic sediments from the NW part of the Pannonian basin. *Geol. Carpath*., **43**, 6, 363–368.

MARTON E., TOKARSKI A. K., MASTELLA L., 1999 — Palaeomagnetic results from Tertiary Podhale flysch, Polish West Carpathians. *Phys. Chem. Earth* (A), **24**(8): 645–649.

MARTON E., VASS D., TUNYI I., 1996 — Rotation of the South Slovak Palaeogene and Lower Miocene rocks indicated by palaeomagnetic data. *Geol. Carpath*., **47**/1, 31–42.

MARTON E., VELJOVIC D., 1987 — Palaeomagnetism of Cretaceous carbonates from the northwestern part of the Dinaric fold belt. *Tectonophysics*, **134**: 331–338.

MARTON P., ROZLOŽNIK L., SASVARI T., 1991 — Implications of palaeomagnetic study of the Silica nappe, Slovakia. *Geophys. J. Int*., **107**: 67–75.

MAURITSCH H. J., BECKE M., 1987 — Palaeomagnetic investigations in the Eastern Alps and the Southern Border Zone. *In*: H.W. Flügel, P. Faupl (Eds.), Geodynamics of the Eastern Alps. Deuticke, Vienna: 282–305.

MAURITSCH H. J., MARTON E., 1995 — Escape models of the Alpine-Carpathian-Pannonian region in the light of palaeomagnetic observations. *Terra Nova*, **7**: 44–50.

McCABE C., ELMORE R. D., 1989 — The occurrence and origin of Late Palaeozoic remagnetization in the sedimentary rocks of North America. *Rev. Geophys*., **27**: 471–494.

McFADDEN P. L., 1990 — A new fold test for palaeomagnetic studies. *Geophys. J. Int.*, **103**: 163–169.

McFADDEN P. L., McELHINNY M. W., 1990 — Classification of the reversal test in palaeomagnetism. *Geophys. J. Int.*, **103**: 725–729.

MICHALIK J., GAŽDZICKI A., 1980 — Is the StraŽov nappe present in the Tatra Mts.? [Eng. Sum.]. *Prz. Geol*., **28**(11): 616–619.

MIDDLETON M. F., SCHMIDT P. W., 1982. — Palaeothermometry of the Sydney Basin. *J. Geophys. Res*., **87**: 5351–5359.

MOCK R., 1978 — Knowledge recently gained about the southern parts of the West Carpathians. *In*: J. Vozar (Ed.), Palaeogeographical evolution of the West Carpathians. 321–341. Geol. Ustav D. Stura, Bratislava.

MORRIS A., 1996 — Glossary of basic palaeomagnetic and rock magnetic terms. *In*: A. Morris, D.H. Tarling (Eds.), Palaeomagnetism and tectonics of the Mediterranean region. *Geol. Soc. Sp. Publ*., **105**: 401–415.

MORRIS A., TARLING D. H., 1996 — Palaeomagnetism and tectonics of the Mediterranean region: an introduction. *In*: A. Morris, D. H. Tarling (Eds.), 1996, Palaeomagnetism and tectonics of the Mediterranean region. *Geol. Soc. Sp. Publ.*, **105**: 1–18.

MULLENDER T. A. T., VAN VELZEN A. J., DEKKERS M. J., 1993 — Continuous drift correction and separate identification of ferrimagnetic and paramagnetic contributions in thermomagnetic runs. *Geophys. J. Int*., **114**: 663–672.

NARĘBSKI W., 1990 — Early rift stage in the evolution of the western part of the Carpathians: geochemical evidence from limburgite and teschenite rock series. *Geol. Zbor. Geol. Carpath*., **41**: 521–528.

NEMČOK J. (Ed.) *et al*., 1995 — Geologicka Mapa Tatier (Geological Map of the Tatra Mts.), 1:50 000. Geologicky Ustav Dionyza Stura, Bratislava.

NEMČOK M., NEMČOK J., 1994 — Late Cretaceous deformation of the Pieniny Klippen Belt, Western Carpathians. *Tectonophysics*, **233**: 81–109.

NEMČOK M., KEITH jr J. F., NEESE D. G., 1996 — Development and hydrocarbon potential of the Central Carpathian Palaeogene Basin, West Carpathians, Slovak Republic. *In*: P. A. Ziegler, F. Horvath (Eds.), Peri-Tethys Memoir 2: Structure and Prospects of Alpine Basins and Forelands. *Mem. Mus. natn. Hist. nat*., **170**: 321–342, Paris.

NEMČOK M., POSPIŠIL L., LEXA J., DONELICK R. A., 1998 — Tertiary subduction and slab break-off model of the Carpathian-Pannonian region. *Tectonophysics*, **295**: 307–340.

NEUBAUER F., 1996 — Middle Jurassic subduction related volcanism and Cretaceous kinematics in Meliata units of the eastern Northern Calcareous Alps. *Mitteil. der Geol. and Bergbaustud. in Österreich*, **41**: 123. PANCARDI Workshop 1996.

NEY R., 1976 — The Carpathians and plate tectonics. *Prz. Geol*., **6**: 309–316.

OGG J.G., STEINER M.B., WIECZOREK J., HOFFMAN M., 1991 — Jurassic magnetostratigraphy, 4, Early Callovian through Middle Oxfordian of the Kraków Uplands (Poland). *Earth. Planet. Sc. Lett*., **104**: 488–504.

OPDYKE N. D., CHANNELL J. E. T., 1996 — Magnetic stratigraphy. Academic Press, San Diego, 346 pp.

OSZCZYPKO N., 1992 — Late Cretaceous through Palaeogene evolution of the Magura Basin. *Geol. Carpath*., **43**: 333–338.

OSZCZYPKO N., 1997 — The Early–Middle Miocene Carpathian peripheral foreland basin (Western Carpathians, Poland). *Prz. Geol*., **45**(10): 1054–1063.

OSZCZYPKO N., ŚLĄCZKA A., 1985 — An attempt to palinspastic reconstruction of Neogene basins in the Carpathian Foredeep. *Ann. Soc. Geol. Pol*., **55**: 55–76.

PASSENDORFER E., 1983 — Jak powstały Tatry. Wyd. Geol., Warszawa, 286 pp.

PATRASCU S., BLEAHU M., PANAIOTU C., PANAIOTU C. E., 1992 — The palaeomagnetism of the Upper Cretaceous magmatic rocks in the Banat area of South Carpathians: tectonic implications. *Tectonophysics*, **213**: 341–352.

PIOTROWSKI J., 1978 — Mesostructural analysis of the main tectonic units of the Tatra Mountains along the Kościeliska Valley [Eng. Sum.]. *Stud. Geol. Pol*., **55**: 3–90.

PLAŠIENKA D., 1995 — Mesozoic evolution of Tatric units in the Male Karpaty and PovaŽsky Inovec Mts.: Implications for the position of the Klape and related units in Western Slovakia. *Geol. Carpath*., **46**: 101–112.

PLAŠIENKA D., 1996 — Mid-Cretaceous (120–80 Ma) orogenic processes in the Central West Carpathians: brief review and interpretation of data. *Slovak Geol. Magaz*., 3–4/96: 319–324.

PLAŠIENKA D., PROKEŠOVA D., 1996 — Towards an evolutionary model of the KriŽna cover nappe (Western Carpathians, Slovakia). *Slovak Geol. Magaz*., 3–4/96: 279–286.

PLAŠIENKA D., SOTAK J., 1996 — Rauhwackized carbonate tectonic breccias in the West Carpathian nappe edifice: introductory remarks and preliminary results. *Slovak Geol. Magaz*., 3–4/96: 287–291.

PLAŠIENKA D., SOTAK J., SPIŠIAK J., 1995 — Penninic units of the Central Western Carpathians: lithological, structural and metamorphic signatures. *Geol. Soc. Greece, Sp. Publ*., **4**, 1995. Proceedings of the XV Congress of the Carpatho-Balcan Geological Association, September 1995, Athens, Greece.

PLAŠIENKA D., GRECULA P., PUTIŠ M., KOVAČ M., HOVORKA D., 1997a — Evolution and structure of the Western Carpathians: an overview. *In*: P. Grecula, D. Hovorka, M. Putiš (Eds.), Geological evolution of the Western Carpathians. Mineralia Slovaca — Monograph, Bratislava: 1–24.

PLAŠIENKA D., HAVRILA M., MICHALIK J., PUTIŠ M., REHAKOVA D., 1997b — Nappe structure of the western part of the Central Carpathians. *In*: Alpine evolution of the Western Carpathians and related areas. Introductory articles to the excursion: 139–161. Bratislava, Slovakia, September 11–14th 1997.

PLAŠIENKA D., PUTIŠ M., KOVAČ M, ŠEFARA J., HRUSECKY I., 1997c — Zones of Alpidic subduction and crustal underthrusting in the Western Carpathians. *In*: P. Grecula, D. Hovorka, M. Putiš (Eds.), Geological evolution of the Western Carpathians. Mineralia Slovaca — Monograph, Bratislava: 35–42.

PSZCZÓŁKOWSKI A., 1996 — Calpionellid stratigraphy of the Tithonian–Berriasian pelagic limestones in the Tatra Mts. (Western Carpathians). *Stud. Geol. Pol*., **109**: 103–130.

PULLAIAH G., IRVING E., BUCHAN K. L., DUNLOP D. J., 1975 — Magnetisation changes caused by burial and uplift. *Earth Planet. Sc. Lett*., **28**: 133–143.

PUTIŠ M., 1991 — Geology and petrotectonics of some shear zones in the West Carpathian crystalline complexes. *Min. Slov*., **23**: 459–473.

PUTIŠ M., 1992 — Variscan and Alpidic nappe structures of the Western Carpathian crystalline basement. *Geol. Carpath*., **43**: 369–380.

RABOWSKI F., 1959 — High-Tatric series in Western Tatra [Eng. Sum.]. *Pr. Inst. Geol*., **27**: 3–178.

RATSCHBACHER L., FRISCH W., LINZER H-G., MERLE O., 1991 — Lateral extrusion in the Eastern Alps. Part 2: Structural analysis. *Tectonics*, **10**: 257–271.

RICOU L.E., DERCOURT J., GEYSSANT J., GRANDJACQUET C., LEPVRIER C., BIJU-DUVAL B., 1986 — Geological constraints on the Alpine evolution of the Mediterranean Tethys. *Tectonophysics*, **123**: 83–122.

ROCHETTE P., 1987 — Magnetic susceptibility of the rock matrix related to magnetic fabric studies. *J. Struct. Geol*., **9**: 1015–1020.

ROWLEY D. B., LOTTES A. L., 1988 — Plate-kinematic reconstructions of the North Atlantic and Arctic: Late Jurassic to Present. *Tectonophysics*, **155**: 73–120.

ROYDEN L. H., BURCHFIEL B. C., 1989 — Are systematic variations in thrust belt style related to plate boundary processes? (The Western Alps versus the Carpathians). *Tectonics*, **8**(1): 51–61.

SAVOSTIN L. A., SIBUET J-C., ZONENSHEIN L. P., LE PICHON X., ROULET M.-J., 1986 — Kinematic evolution of the Tethys Belt from the Atlantic Ocean to the Pamirs since the Triassic. *Tectonophysics*, **123**: 1–35.

SCHMIDT T., BLAU J., KAZMER M., 1991 — Large scale strike-slip displacement of the Drauzug and the Transdanubian Mountains in early Alpine history: evidence from Permo-Mesozoic facies belts. *Tectonophysics*, **200**: 213–232.

SCOTESE C. R., GAHAGAN L. M., LARSON R. L., 1988 — Plate tectonic reconstructions of the Cretaceous and Cenozoic ocean basins. *Tectonophysics*, **155**: 27–48.

SIKORA W. J., 1976 — Cordileres of the Western Carpathians in the light of plate tectonics [Eng. Sum.]. *Prz. Geol*., **24**(6): 336–349.

SKIBA M., MICHALIK M., 1999 — Origin of non-carbonate components in Triassic carbonate rocks from the KriŽna unit in the Tatra Mts. *Miner. Soc. of Poland — Sp. Pap*., **14**: 122–123.

SOKOŁOWSKI S., 1948 — Les Tatry Bielskie. La Geologie de leurs versants meridionaux. *Pr. Państw. Inst. Geol*., **4**.

SOKOŁOWSKI S., 1978 — Geology of the Sub-Tatric zone of the Polish Tatras between Kopy Sołtysie and Białka stream [Eng. Sum.]. *Pr. Muzeum Ziemi*, **28**: 35–61.

SOTAK J., 1992 — Evolution of the Western Carpathian suture zone — principal geotectonic events. *Geol. Carpath*., **43**: 355–362.

SPERNER B., 1996 — Computer programs for the kinematic analysis of brittle deformation structures and the Tertiary evolution of the Western Carpathians (Slovakia). *Tübinger Geowissenschaftliche Arbeiten, Reihe A*, Band 27.

SPÖTL C., KUNK M. J., RAMSEYER K., LONGSTAFFE F. J., 1998 — Authigenic potassium feldaspar: a tracer for the timing of palaeofluid flow in carbonate rocks, Northern Calcareous Alps, Austria. *In*: J. Parnell (Ed.), Dating and Duration of Fluid Flow and Fluid-Rock Interaction. *Geol. Soc. London, Sp. Publ.*, **144**: 107–128.

STAMATAKOS J., KODAMA K. P., 1991 — Flexural flow folding and the palaeomagnetic fold test: an example of strain reorientation of remanence in the Mauch Chunk formation. *Tectonics*, **10**, 807–819.

STAMPFLI G. M., MARCHANT R. H., 1995 — Geodynamic evolution of the Tethyan margins of the Western Alps. *In*: Lehner *et al*. (Eds.), Deep structure of Switzerland — results from NFP 20. Birkhäuser AG, Basel.

TAUXE L., MULLENDER T. A. T., PICK T., 1996 — Potbellies, wasp-waists and superparamagnetism in magnetite hysteresis. *J. Geoph. Res*., **101**: 571–583.

TARLING D. H., HROUDA F., 1993 — The magnetic anisotropy of rocks. Chapman & Hall, London, 217 pp.

TOKARSKI A. K., 1980 — Dynamics of Outer Carpathian Tertiary orogenesis. *Publ. Inst. Geophys. Pol. Acad. Sc. A-8*, **130**: 129–142.

TOLLMANN A., 1990 — Palaeogeographic maps and profiles in the Eastern Alps and the relationship of the Eastern Alps to neighboring terrain. *Mem. Soc. Geol. France*, Paris, *Nouvelle Serie*, **154** (II): 23–49.

TOMEK Č. , 1993 — Deep crustal structure beneath the Central and Inner West Carpathians. *Tectonophysics*, **226**: 417–431.

TOMEK Č. , HALL J., 1993 — Subducted continental margin imaged in the Carpathians of Czechoslovakia. *Geology*, **21**: 535–538.

TORSVIK T. H., SMETHURST M. A., 1994 — GMAP for Windows (version 1.0). Geographic mapping and palaeoreconstruction package (manuscript).

TUNYI I., MARTON E., 1996 — Indications for large Tertiary rotations in the Carpathian-Northern Pannonian region outside the North Hungarian Palaeogene Basin. *Geol. Carpath*., **47**: 43–50.

UCHMAN A., 1993 — Lower Jurassic carbonate sedimentation controlled by tilted blocks in the Choč unit in the Tatra Mts., Poland. *Zentralbl. Geol. Palaont*., Teil I, 7/8: 875–883.

UNRUG R., 1979 — Palinspastic reconstruction of the Carpathian arc before Neogene tectogenesis. *Ann. Soc. Geol. Pol.*, **49**: 3–21.

UNRUG R., 1982 — Geodynamic evolution of the Carpathians. *Ann. Soc. Geol. Pol*., **52**: 39–66.

VANDENBERG I., ZIJDERVELD J. D. A., 1982 — Palaeomagnetism in the Mediterranean area. *In*: B. Berckhemer, K. Hsu (Eds.), Alpine Mediterranean geodynamics. *Am. Geophys. Un./Geodyn*. Ser., **7**: 83–112.

VAN DER VOO R., 1993 — Palaeomagnetism of the Atlantic, Tethys and Iapetus Oceans. Cambridge University Press, 411 pp.

VAN VELZEN A., 1992 — Magnetic minerals in Pliocene and Pleistocene marine marls from Southern Italy. Rock magnetic properties and alteration during thermal demagnetisation. *Geologica Ultraiectina*, **122**: 154 pp.

VETTER J. R., KODAMA K.P., GOLDSTEIN A., 1989 — Reorientation of remanent magnetization during tectonic fabric development: an example from the Waynesboro Formation, Pennsylvania, USA. *Tectonophysics*, **165**: 29–39.

VILLALAIN J. J., OSETE M. L., VEGAS R., GARCIA-DUENAS V., HELLER F., 1996 — The Neogene remagnetization in the western Betics: a brief comment on the reliability of palaeomagnetic directions. *In*: A. Morris, D. H. Tarling (Eds.), Palaeomagnetism and tectonics of the Mediterranean region. *Geol. Soc. Sp. Publ*., **105**: 33–41.

WAGREICH M., FAUPL P., SCHLAGINTWEIT F., 1995 — Heavy minerals from Urgonian limestone pebbles of the Northern Calcareous Alps (Austria, Bavaria): further evidence for an Intra-Austroalpine suture zone. *Geol. Carpath*., **46**: 197–204.

WESTPHAL M., 1993 — Did a large departure from the geocentric axial dipole hypothesis occur during the Eocene? Evidence from the magnetic polar wander path of Eurasia. *Earth Planet. Sc. Lett*., **117**: 15–28.

WIECZOREK J., 1988 — Maiolica — a unique facies of the Western Tethys. *Ann. Soc. Geol. Pol.*, **58**: 255–276.

WIECZOREK J., 1995 — Mesozoic evolution of the Tatra Mts. (Western Carpathians). *Geol. Soc. Greece, Sp. Publ*., **4**, Proceedings of the XV Congress of the Carpatho-Balkan Geological Association, September 1995, Athens, Greece.

WIECZOREK J., BARBACKI A., 1997 — Przekrój geologiczny przez nieckę podhalańską i jej podłoże na podstawie danych sejsmicznych i wiertniczych; implikacje dla geotermii. *Prz. Geol*., **45**(7): 715–720.

WINDLEY B. F., 1997 — The evolving continents. John Wiley, Chichester.

WINKLER W., ŚLĄCZKA A., 1992 — Sediment dispersal and provenance in the Silesian, Dukla and Magura flysch nappes (Outer Carpathians, Poland). *Geol. Rdsch*., **81**: 371–382.

WINKLER W., ŚLĄCZKA A., 1994 — A Late Cretaceous to Palaeogene geodynamic model for the Western Carpathians in Poland. *Geol. Carpath.*, **45**: 71–82.

YILMAZ P. O., NORTON I. O., LEARY D., CHUCHLA R. J., 1996 — Tectonic evolution and palaeogeography of Europe. *In*: P. A. Ziegler, F. Horvath, (Eds.), Peri-Tethys Memoir 2: Structure and prospects of Alpine basins and forelands. *Mem. Mus. natn. Hist. nat.*, **170**: 47–60 + Enclosures 1–13. Paris.

ZAWIDZKA K., 1972 — Stratigraphic position of the Furkaska limestones (Choč nappe, the Tatra Mts.). *Acta Geol. Pol.*, **22**: 459–466.

ZIEGLER P. A. SCHMID S. M., PFIFFNER A., SCHÖNBORN G., 1996 — Structure and evolution of the Central Alps and their northern and southern foreland basins. *In*: P. A. Ziegler, F. Horvath (Eds.), Peri-Tethys Memoir 2: Structure and prospects of Alpine basins and forelands. *Mem. Mus. natn. Hist. nat.*, **170**: 221–233.