MAGNETO- AND BIOSTRATIGRAPHY OF THE JURASSIC/CRETACEOUS BOUNDARY IN THE LÓKÚT SECTION (TRANSDANUBIAN RANGE, HUNGARY)

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ABSTRACT

Jurassic-Cretaceous sediments of Transdanubian Range in Northern Hungary mostly retain their primary magnetizations and are suitable for detailed bio- and magnetostratigraphic studies. The Lókút section, 13 m in thickness, is localized in the central part of the Transdanubian Range. It contains the Jurassic/Cretaceous boundary in pelagic carbonate facies. Although the colour of the rocks changes from reddish-pinkish in the bottom to almost white at the top of the section, magnetite was identified as a magnetic carrier without evidence of hematite. Integrated bio- and magnetostratigraphical investigations resulted in construction of chronostratigraphical scheme. The section, embraces magnetozones from M21r to M18r, of upper Lower Tithonian (Parastomiosphaera malmica Zone) to Lower Berriasian age (Calpionella alpina Subzone). Sedimentation rate of pelagic limestones increased from 1–3 m/My during Tithonian to 5–7 m/My during Berriasian. The sedimentation rate and its changes up the section are comparable to those from the Jurassic/Cretaceous boundary sections of Trento plateau (Southern Alps, Italy) - sedimentary environments of Trento plateau and central Transdanubian Range in that time might be similar. Sedimentation rate within Umbrian Apennine basins and Križna unit in the Western Tatra Mts. seems significantly higher. Analysis of rock magnetic parameters reveals that detrital input was much lower into the Lókút section than into Križna basin in the Tatra Mts. (Zliechov trough). Increase of sedimentation rate occurs in both sections in the Upper Tithonian - Lower Berriasian. It coincides with the onset of calpionellid limestone facies and is related to increased productivity of calcareous micro- and nannoplankton. Detailed correlation of both sections basing on rock magnetic parameters and susceptibility changes is, however, not possible. They are dependent mostly on the local sedimentary conditions (Bakony Mts. - deep water plateau; Križna unit - deep water trough) and correlation with any "global" paleoenvironmental (climatic, eustatic) trends is not straightforward.

Keywords: Jurassic/Cretaceous boundary, magnetostratigraphy, calpionellids, Transdanubian Range, Hungary

1. INTRODUCTION

Magnetostratigraphic investigations of the Jurassic/Cretaceous boundary started in the early 1980s (e.g. *Channell et al., 1982*; *Lowrie and Channell, 1984*; *Cirilli et al., 1984*). The main goal of these studies was the correlation of land sections to the M sequence of oceanic anomalies but also correlation of magnetozones to biostratigraphic subdivisions. The most thoroughly studied was the Maiolica Formation in the Southern Alps and Apennines. It comprises pelagic carbonate sediments rich in micro- and nannofossils, providing the fundament of a very detailed biostratigraphy. Maiolica Formation appeared to be very suitable for magnetostratigraphic studies. It revealed weak but stable magnetization based upon magnetite and lack of regional remagnetizations which are very common in fold-and thrust belts (*Lowrie and Heller, 1982*). The first successful attempt of correlation of magnetozones to calpionellid based biostratigraphy was performed by *Ogg and Lowrie (1986)*. Boundary between calpionellid zones A and B, now widely accepted as J/C boundary (*Gradstein et al., 2004*), was correlated with middle part of M19n. Boundary of zones B and C falls within magnetozone M17r, while boundary C/D in magnetozone M16r (Fig. 1). The pattern was subsequently tested and verified by *Channell and Grandesso (1987)* and *Channell et al. (1987)* by studies of sections within Trento plateau, Belluno and Lombardian Basins (Southern Alps, Italy). All results were summarized by *Ogg et al. (1991)* who presented a database of all the sections studied with facies changes and biotic events at the J/C boundary correlated with magnetic stratigraphy. Recently, *Houša et al. (2004)* and *Speranza et al. (2005)* supplied new data on some sections in Apennines, generally confirming the magnetostratigraphic scheme elaborated by previous authors. Magnetostratigraphic investigations of Upper Jurassic and Lower Cretaceous in Hungary started almost as early as in Italy. These embraced sections in the Transdanubian Range; Sümeg (*Márton, 1982*) and Borzavár and Hárskút (*Márton, 1985, 1986*). The presence of primary magnetization and polarity changes was convincingly proved. Unfortunately the studies did not contain full biostratigraphic and lithological documentation which hampers the correlation between magneto- and biozones. Moreover, the correlation of the J/C boundary with lower part of magnetozone M16n, accepted by *Márton (1982)*, was abandoned only few years later (*Ogg and Lowrie, 1986*). In the last few years several papers dealing with magnetostratigraphy of the J/C boundary out of the classic areas of Southern Alps and Apennines were published. New results were reported from sections in Carpathians: Brodno (Pieniny Klippen Belt, Slovakia, *Houša et al., 1999*) and Lower Sub-Tatric (Križna) unit (Tatra Mts., Poland, *Grabowski and Pszczółkowski, 2006*). *Houša et al. (2007)* performed an attempt to correlate the J/C boundary in Boreal and Tethyan realms. Studies on Spanish and Austrian sections are in progress (*Pruner et al., 2009; Pruner et al., 2010*).

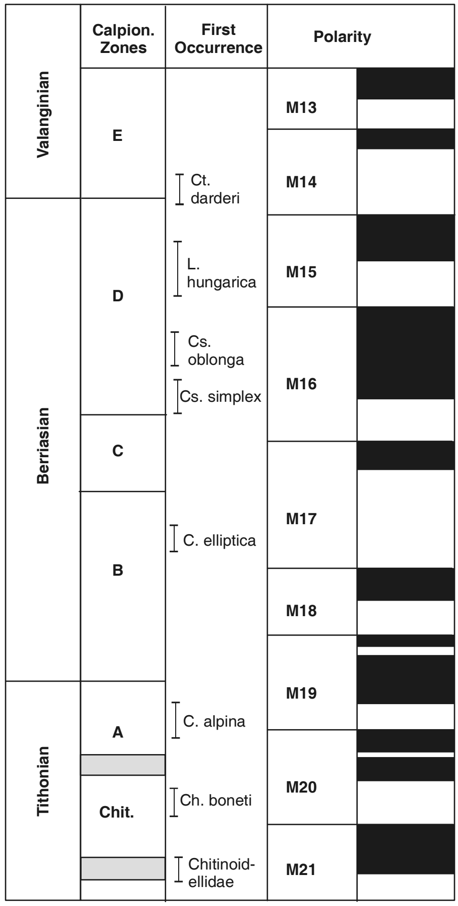

Fig. 1. Correlation of magnetozones around the Jurassic/Cretaceous boundary with calpionellid zonation (compiled after *Ogg et al., 1991*; *Opdyke and Channell, 1996*; *Houša et al., 1999*; *Gradstein et al., 2004*).

Bio- and magnetostratigraphic framework for the J/C boundary interval seems to be well established. It enables application of magnetic methods for dating and tracing paleoenvironmental (sedimentary, climatic, geochemical) changes on a global or regional scale, akin to those in Quaternary loess sequences (e.g. *Heller and Evans, 1995*). This was an approach of *Channell et al. (1993)* who correlated the Valanginian $\delta^{13}C$ event with the global polarity time scale. *Mayer (1999)* and *Mayer and Appel (1999)* integrated magneto- and cyclostratigraphic investigations which resulted in astrochronologic calibration of some Lower Cretaceous magnetozones. *Grabowski and Pszczółkowski (2006)* attempted to use petromagnetic properties of sediments for unraveling the nature of sedimentary changes in the Križna Basin (Tatra Mts.). It appeared that three formations of the Križna Basin, spread between Upper Tithonian and Upper Berriasian, reveal distinct rock magnetic contrasts which might be interpreted in terms of environmental changes on a basinal scale. In this paper we applied similar methodology for magnetostratigraphic calibration and dating of sedimentary changes in the Lókút section, located in the central part of the Transdanubian Range, Northern Hungary.

1.1. Geological Setting and Sampling

Northern Hungary belongs to the Alcapa megaterrane that is a large composite structural unit including the Eastern Alps, the Western Carpathians and the northern part of the basement of the Pannonian Basin (*Csontos and Vörös, 2004*). Displaced terranes of South Alpine and Dinaridic origin that are located at present at the southeastern part of the Alcapa were named as the Pelso Composite Terrane. The Transdanubian Range Unit is one of the components of the Pelso Terrane (*Kovács et al., 2000*) (Fig. 2). In the Mesozoic, the Transdanubian Range unit was situated between the South Alpine and Austroalpine realms (*Császár et al., 1990*; *Schmid et al., 1991*; *Haas et al., 1995*; *Vörös and Galácz, 1998*; *Haas, 2001*; *Csontos and Vörös, 2004*). The Transdanubian Range unit approached its present-day setting as a result of large scale tectonic motions which took place, according to some authors (e.g. *Csontos and Nagymarosy*, *1998*) before the Early Miocene, or based on paleomagnetic observations, during two tectonic phases in the 18.5−17.7 Ma and in the 16.0−14.5 Ma time intervals (*Márton and Fodor, 2003*). Transdanubian Range reveals a synclinal structure (Fig. 2) formed by tectonic deformations that occurred mainly in Early Albian but started already in the Late Aptian (*Fodor et al., 2005*). Middle Cretaceous structures are unconformably covered by Upper Cretaceous and Eocene formations.

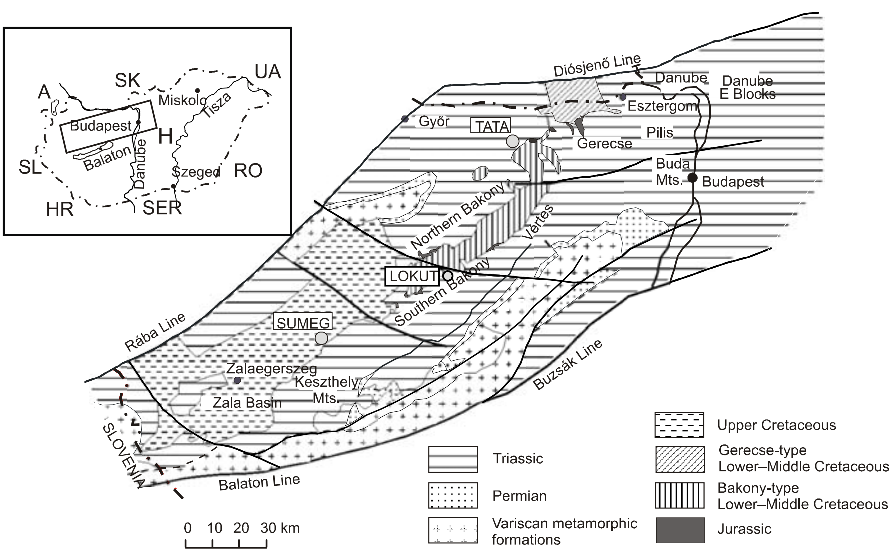

Fig. 2. Tectonic sketch of Transdanubian Range (without Tertiary cover) with Lókút and Sümeg sections indicated. Insert map indicates the position of Transdanubian Range (rectangle) in the contour map of Hungary.

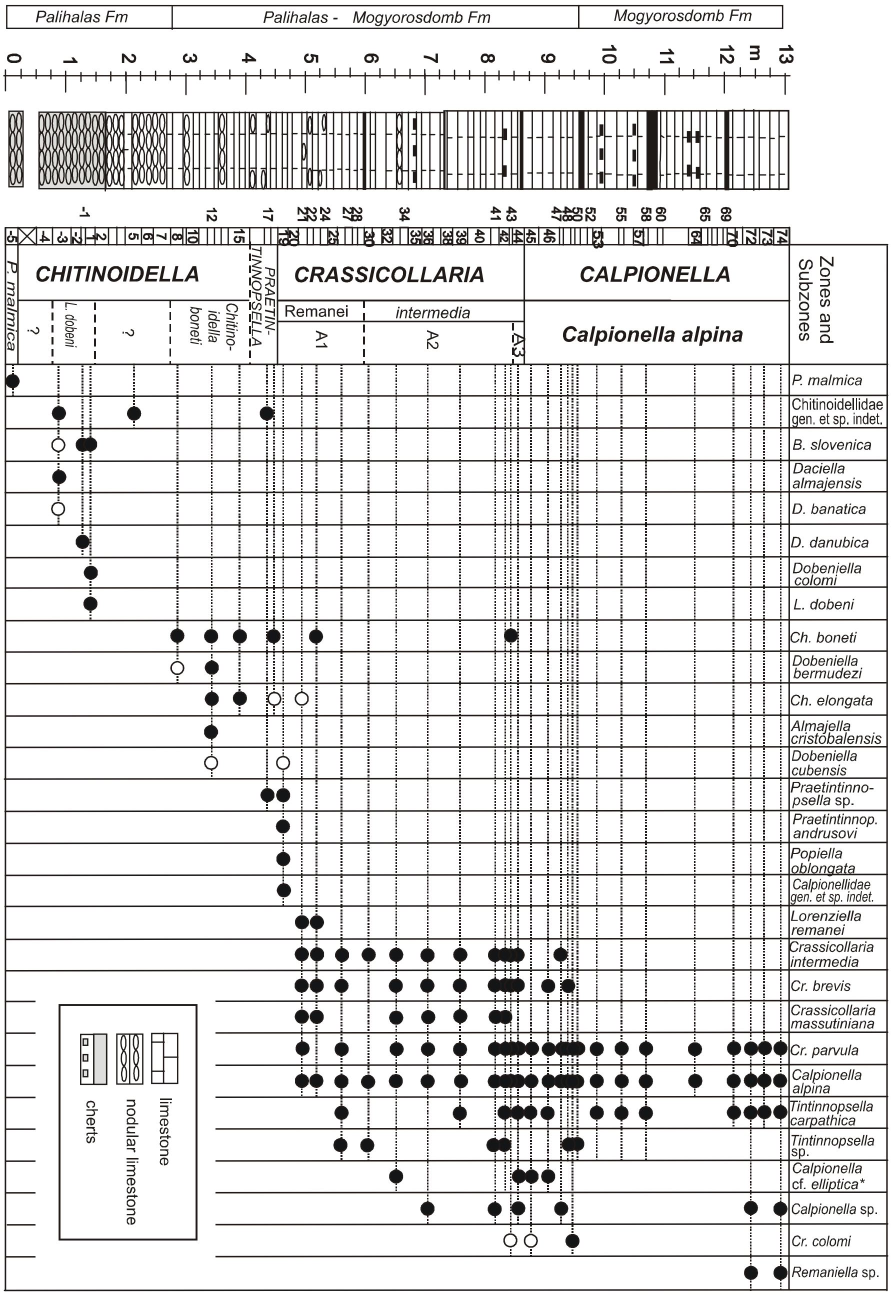

Fig. 3. Lithology, litho- and biostratigraphy of the Lókút section. Red colour of beds (grey shading in this figure) terminates at the top of bed 2.

The Lókút section is located in the Bakony Mts. that is a part of the Transdanubian Range. The area belongs to the southern limb of the main syncline, near to its axis. The exposure is an artificial trench on the southern slope of Lókút Hill, east of village Lókút (geographical coordinates: 47°12′17″N; 17°52′56″E). Strata dip gently to the north with mean orientation 357°/20°, calculated as a mean of tens of measurements.

The basal part of the measured section comprises red and yellowish to white nodular limestones of the Pálihálás Formation (Fig. 3), rich in *Saccocoma* ossicles. Based on ammonites the Pálihálás Formation was assigned to the Kimmeridgian to Lower Tithonian (*Vígh, 1984*). Upper Tithonian to Lower Berriasian interval is represented by white cherty limestones of Maiolica-type, containing calpionellids, radiolarians and aptychi (Mogyorósdomb Fm). Facies distribution and thickness data of biostratigraphically well correlable sections suggest that in the latest Jurassic to earliest Cretaceous the SW part of the Bakony area (environs of town Sümeg) may have been located in the inner part of a deep pelagic basin, whereas its central and eastern part belonged to a much shallower submarine plateau (*Fülöp, 1964*; *Haas et al*., *1985*). The Lókút area belonged to the transitional zone between the deep basin and the shallower plateau. This reflects in the microbiofacies of the succession. The biomicrite and biomicrosparite packstones contain large amount of calpionellids and globochaetes and usually only very few radiolarians in Lókút, whereas large amount of radiolarians and related chert nodules characterizes the typical deep basin successions in Sümeg (*Haas et al., 1994*).

First paleomagnetic investigations of the Lókút section (“white Tithonian limestone”) were performed by *Márton and Márton (1981)*. Rock magnetic investigations indicated that magnetite was the magnetic carrier. The authors documented presence of pre-folding component of normal polarity, which was interpreted as primary (see Table 1 of *Márton and Márton, 1981)*. They mentioned that limestones are partially remagnetized, however they did not discuss the secondary component in detail.

For the purpose of present study, all the beds in the section were numbered. Uncovered part of the section consists of 79 beds: the lowest −5, the highest 74 (Fig. 3). Samples were taken either with a gasoline powered drill or as block (hand samples). Sampling was performed in two steps. In the first step, the Hungarian team collected samples from beds −3, −1, 15, 20, 30, 40 and 53. The pilot collection of cylindrical samples was divided in two parts, which were demagnetized in the paleomagnetic laboratories in Warsaw and Budapest. The results obtained in two laboratories were fully concordant. In the second step, the section was sampled bed by bed. 74 beds were sampled, drill cores were taken from 50 beds while hand samples were taken from 24 beds. The main part of the samples was demagnetized and analyzed in the Paleomagnetic Laboratory of the Polish Geological Institute in Warsaw.

1.2. Litho- and Biostratigraphy

Precise lithostratigraphic assignment and subdivision of the Lókút section is not easy due to paleogeographic setting of the study area i.e. it was located in a transitional zone between a deep basin in the west and a shallower plateau in the east. This setting reflects in the transitional litho- and biofacies characteristics of the formations. The lower part of the section, comprising reddish nodular limestones, should be assigned to Pálihálás Formation (Fig. 3). Above bed no. 2, the limestones become yellowish or white, nodularity is however visible up to bed no. 7. This feature decreases up the section, but single nodular beds are observed as high as bed no. 33. Therefore top of the Pálihálás Formation was arbitrarily accepted at the top of bed no. 7. The interval between beds no. 7 and 50 was assigned to transitional "Pálihálás - Mogyorósdomb Fm.", because it reveals features characteristic for both formations (nodular horizons and chert layers). The highest part of the section, above the bed no. 50, becomes more cherty, which is characteristic for the Mogyorósdomb Fm.

Study of thin sections from samples collected at Lókút (Bakony Mts.) has shown that the lowermost exposed bed (−5) belongs to the Early Tithonian *Parastomiosphaera malmica* Zone (Fig. 3). The *Chitinoidella* Zone begins with the bed −3; this is probably the *Longicollaria dobeni* Subzone. The interval between beds −1 and 15 also belongs to the *Chitinoidella* Zone. Samples 8 to 15 represent the (earliest) Late Tithonian *Chitinoidella boneti* Subzone. Sample taken from bed 17 contains Chitinoidellidae together with a few specimens of *Praetintinnopsella* sp., thus it is placed in the *Praetintinnopsella* Zone (sensu *Borza, 1984*). Sample from bed 19 yielded *Praetintinnopsella andrusovi* Borza, 1969, but poorly preserved calpionellids indicate for the base of Late Tithonian *Crassicollaria* Standard Zone. Microfossil data from beds 19 and 21 indicate lowermost interval of the Remanei Subzone (Fig. 3). The subsequent beds (22−29) also belong to the Remanei Subzone of the *Crassicollaria* Zone (=A1 in the subdivision of *Remane, 1963, 1964, 1985*).

The beds 30 and 33 have been assigned to the *Crassicollaria intermedia* Subzone of the *Crassicollaria* Zone. The sample from the latter limestone bed is rich in calpionellids and *Saccocoma* ossicles. There are some specimens in this calpionellid assemblage (assigned to *Calpionella* cf. *elliptica*, Fig. 3), which are more slender than those described by *Nagy (1986)* as *C. grandalpina* and *C. elliptalpina* from the Upper Tithonian strata of Mecsek Mountains. Interestingly, the discussed *Calpionella* specimens from the Late Tithonian bed 33 of the Lókút section are similar to *C. alpelliptica* reported from the Early Berriasian *C. alpina* Subzone (*Nagy, 1986*). The Intermedia Subzone terminates with the bed 44, which is characterized by a very high share of *C. alpina* (84.6%) and presence of infrequent specimens of *Cr. intermedia*, *Cr. brevis* and *Cr. parvula* (Fig. 4).

The calpionellid assemblage of the next bed (45) is mainly composed of *C. alpina* (84.9%) and *Cr. parvula* (10.2%), therefore was assigned to the Early Berriasian *C. alpina* Subzone of the *Calpionella* Standard Zone (Figs. 3 and 4), despite a presence of one specimen identified as *Crassicollaria* cf. *colomi* Doben. Therefore, the Tithonian/Berriasian boundary (and J/K boundary) is placed between beds 44 and 45 following the criteria of *Remane et al. (1986).* Nevertheless, rare specimens belonging to *Cr. intermedia, Cr. brevis* and *Cr.* cf. *colomi* still occur in beds 45−53. Relatively high share of *Cr. parvula* is characteristic for the calpionellid assemblage of some Early Berriasian limestones, with the maximum value (about 35%) recorded in sample 49 (Fig. 4).

A tentative (lithology-based) correlation of the microfossil zonation (Fig. 3) with the ammonite zones reported for the Lókút section (*Vigh, 1984*) suggests that the bulk of the Early Berriasian strata (our beds 45−74) was not included in this earlier macrofaunal study. The basal Tithonian ammonite zone (*Hybonoticeras hybonotum* Zone) reported by *Vigh (1984)* is not represented in our section because this part of the Pálihálás Formation is currently not exposed.

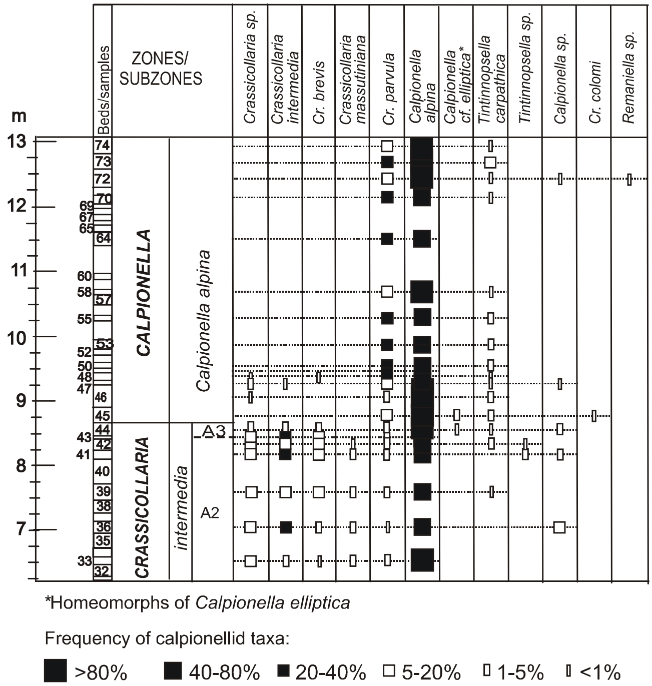

Fig. 4. Calpionellid taxa frequencies for the latest Tithonian-Early Berriasian limestones of the Lókút section determined from thin sections analysis.

1.3. Paleomagnetic Methods

Standard cylindrical specimens of 2.2 cm height and 2.5 cm of diameter were prepared from drill cores and hand samples. Most paleomagnetic experiments were performed in the paleomagnetic laboratory of Polish Geological Institute. Natural Remanent Magnetization (*NRM*) was measured with a JR6A spinner magnetometer and kappabridge KLY2 was used for magnetic susceptibility measurements. Specimens were demagnetized almost exclusively by thermal method using MMTD1 oven. Occasionally alternating field demagnetization was also applied, by means of Molspin tumbler. Results of measurements were further processed using the Remasoft software (*Chadima and Hrouda, 2006*). The VGP latitudes of primary components from each bed, were calculated as angular distances between VGPs from individual beds and mean paleopole for the section (see Table 1). Rock magnetic investigations comprised measurements of Isothermal Remanent Magnetization (IRM) applied along the Z axis in the field of 1 T, and then antiparallel in the field of 100 mT (using MMPM pulse magnetizer). The S parameter calculated as ratio of IRM intensities applied in both fields was indicative for proportions of low and high coercivity minerals. S ratio was calculated for all beds sampled. In samples from selected beds, stepwise acquisition of the IRM (in the

Table 1. Characteristic magnetizations from the Lókút section.

Component	*D*/*I*	α_{95}	*k*	D_c/I_c	N/N_o	Paleopole	*dp*/*dm*
A	18/60	4.4	23.8	11/41	48/74	75.7°N/126.8°E	5.0/6.7
B	263/49	3.5	32.4	285/46	53/74	29.5°N/297.6°E	2.9/4.5
Cn	255/38	3.0	45.8	277/39	51/74		
Cr	71/–30	15.3	9.0	84/–34	12/74		
C	255/36	3.6	25.7	270/38	63/74	15°N/303°E	2.5/4.3

D/*I* - declination/inclination before tectonic correction, D_c/I_c - declination/inclination after tectonic correction; α_{95}, *k* – Fisher statistics parameters, N/N_o - number of beds investigated/used for calculation of mean direction; *dp*/*dm* - confidence oval of paleopole estimation.

maximum field of 1 T) was performed, followed by thermal demagnetization of three axes IRM acquired in the fields of 1 T, 0.36 T and 0.12 T, for identification of magnetic minerals by unblocking temperatures (*Lowrie, 1990*). These experiments were carried out in the ELGI paleomagnetic laboratory, using JR5A spinner magnetometer for remanent magnetization measurements, non-magnetic oven Schonsted TSD1 for thermal demagnetization and MOLSPIN pulse magnetizer for IRM acquisition.

2. DEMAGNETIZATION RESULTS

NRM intensities of most samples were between 5 and 50×10^{-4} A/m. Some samples revealed anomalously high *NRM* intensities of magnitude 10^{-2} A/m (Fig. 5a). Demagnetization of pilot collection revealed presence of normal and reversed polarity directions which indicated that primary magnetization was preserved (*Márton and Márton, 1981*). In most samples three components A, B and C are present (Figs. 6 and 7). Component A has low unblocking temperatures (maximum 200°C). Its direction before tectonic correction is close to the present day geomagnetic field direction in northern Hungary (Fig. 7a). Component B of westerly declination and moderate positive inclination is stable between 200°C and 525°C (Fig. 6a,b,d). Both components A and B might be calculated using a fitted line methods (*Kirschvink, 1980*). The third component C with unblocking temperatures higher than 525°C reveals dual polarity (Cn - normal and Cr - reversed). Its presence is more evident in the case of Cr (Fig. 6c,d). Its calculation using the fitted line method is often impossible because of magnetic susceptibility increase above 550°C and overlapping unblocking temperature spectra with component B. However *NRM* direction before magnetic susceptibility increase is different than component B. It tends towards SW (Cn) and NE (Cr) quadrant of the stereonet and its inclinations are lower than that of B component. Therefore component C was calculated as last *NRM* directions on demagnetization paths, just before final susceptibility increase (Fig. 7c). Samples with anomalously high *NRM* intensity (see Fig. 5a) behaved in a different way. Thermal treatment alone did not lead to isolation of the three components A, B and C. Demagnetization paths created great circles directed towards component B

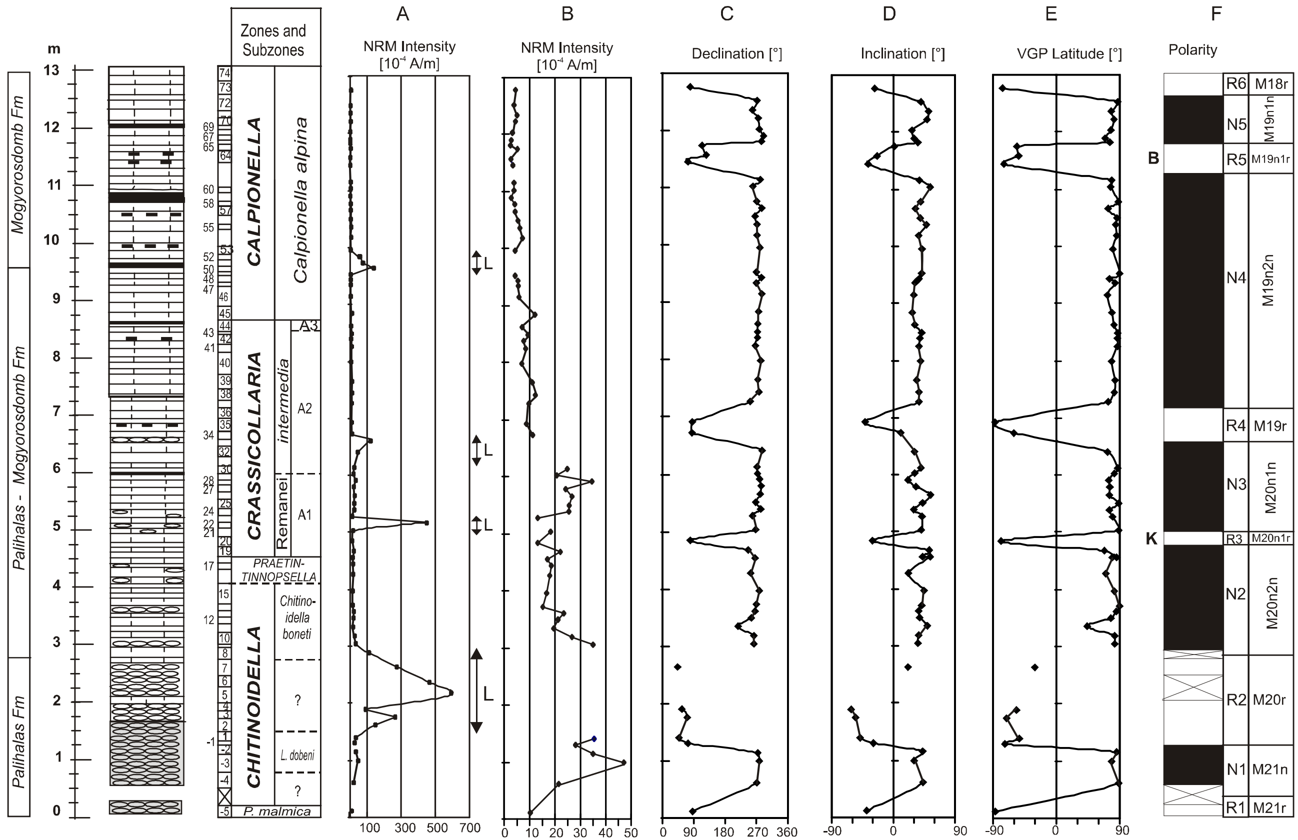

Fig. 5. Lókút section. (A) NRM intensities of all samples within the section; intervals most probably affected by lightning effect are indicated by L. (B) NRM intensities after removal of anomalous peaks caused by lightning effect; (C) declination variations of component C; (D) inclination variations of component C; (E) computed VGP latitude; (F) magnetic polarity. K – Kysuca magnetosubzone; B – Brodno magnetosubzone.

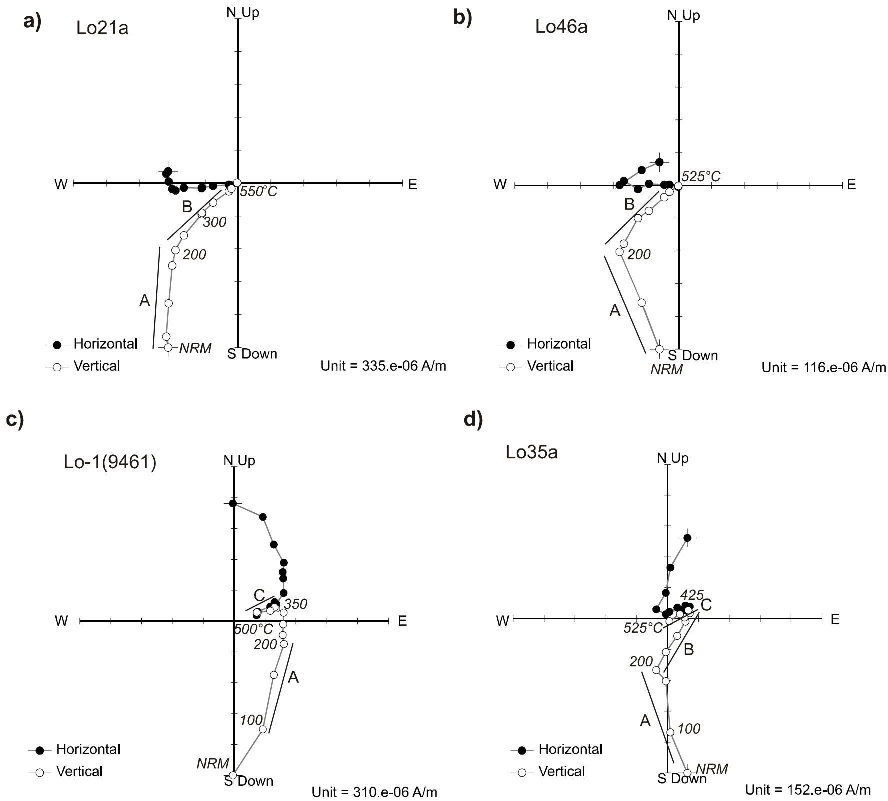

Fig. 6. Thermal demagnetization of typical samples. Orthogonal projection (Zijderveld diagram). **(a)** Lo-21a, Upper Tithonian, magnetozone M20n1n; **(b)** Lo-46a, lower Berriasian, magnetozone M19n2n; **(c)** Lo-1(9461), Lower Tithonian, magnetozone M20r; **(d)** Lo-35a, Upper Tithonian, magnetozone M19r. All projections before tectonic correction. Magnetization components indicated on Zijderveld diagrams.

and C, but stable end point was not reached. *NRM* in natural state revealed an equatorial inclination and this component was hard to remove using thermal demagnetization alone (Fig. 8a). However it was very easily removed with alternating field. In most cases after AF treatment up to 20 mT, components B and C emerged. In this case further demagnetization was carried out with thermal method (Fig. 8b,c).

3. AGE OF MAGNETIZATION COMPONENTS

Component A, in the geographical coordinates (Fig. 7a), situates very close to the present day geomagnetic field direction of the study area ($D = 6°$, $I = 61°$). This observation, as well as low unblocking temperatures (100–200°C) account for a viscous remanent magnetization acquired during last 700 thousand years (Brunhes epoch). Also the component B, of exclusively normal polarity (Fig. 7b), is most probably secondary. Presence of this component was reported in earlier studies (*Márton and Márton, 1981*). They suggested that it was carried by pigmentary hematite and mentioned that even after demagnetization to temperatures as high as 450°C primary antipodal direction could not be isolated. Present results confirm this observation - double polarity of component C (Figs. 6c,d and 7c) is observed after demagnetization to 500°C. *Márton and Márton (1981)* claimed that component B is of pre-Tertiary age. Post-folding Tertiary age of component B is considered as unlikely. Reference Tertiary direction for Transdanubian Range reveals declination 300–330° (*Márton and Mauritsch, 1990*; *Márton and Fodor, 2003*), thus almost 30–60° different than that of component B in post-folding coordinates (Fig. 7b). More precise dating of component B might be performed basing on reference

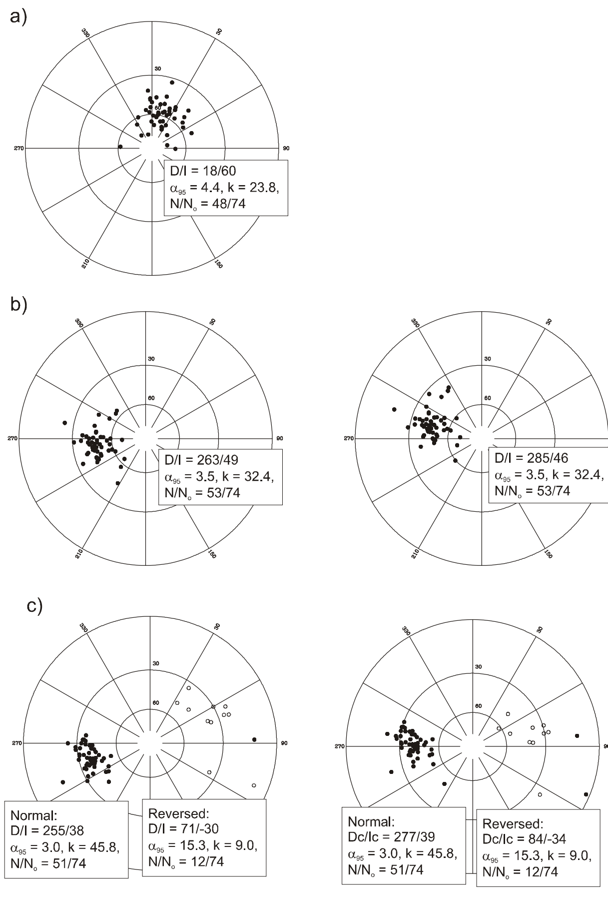

Fig. 7. Stereographic projection of characteristic components of magnetization. **a)** component A, before tectonic correction; **b)** component B before (left) and after tectonic correction (right); **c)** component C before (left) and after tectonic correction (right).

Mesozoic paleodirections for Transdanubian Range, obtained by *Márton and Márton (1983,* Fig. 9). In geographical coordinates, component B is situated close to expected Tithonian magnetizations, while after tectonic corrections it is shifted towards Aptian–Albian directions (see Figs. 7b and 9a). Post-folding age of component B must be excluded because folding in the Transdanubian Range took place between Aptian and Albian (*Fodor et al., 2005*). Pre-folding origin of component B is accepted because it might be acquired during maximum burial of the studied formations in the Aptian, just before the onset of the folding. Component C is considered as primary. It reveals double polarity (Cn - normal, Cr - reversed) and its direction is close to the reference Tithonian direction of the area (Fig. 9b), calculated by *Márton and Márton (1983)* from several localities in the Transdanubian Mts. with positive result of fold test. Therefore this component was used for construction of magnetic polarity time scale (Fig. 5e,f).

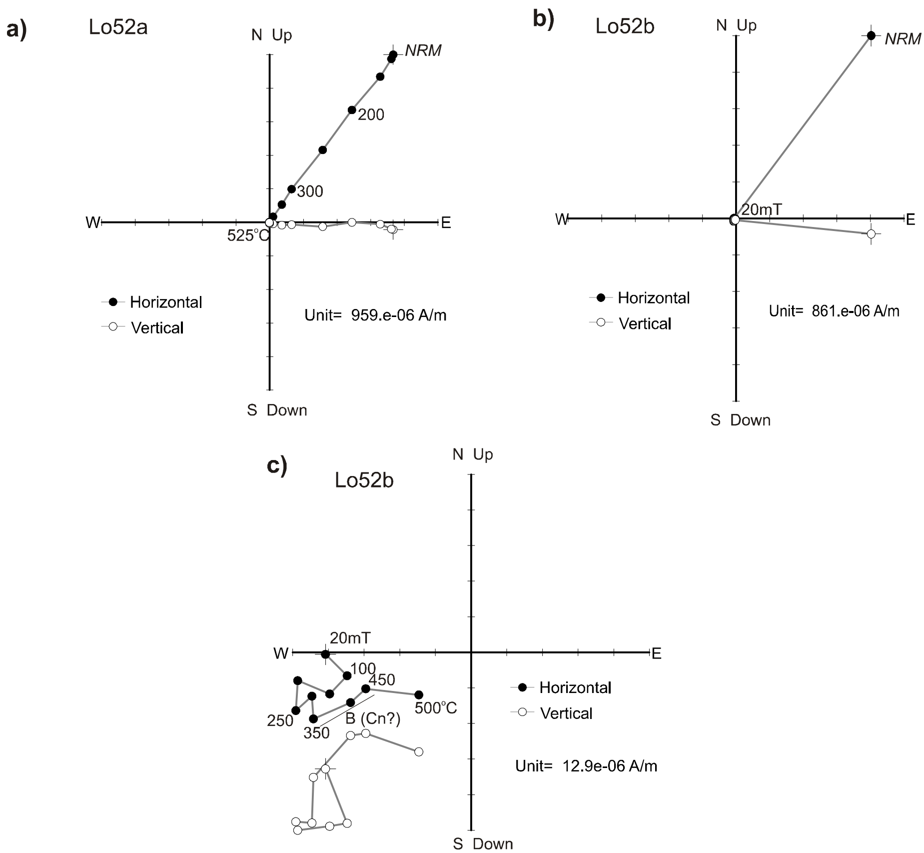

Fig. 8. Demagnetization of anomalous specimen of high *NRM* intensity. Sample Lo-52, Lower Berriasian, magnetozone M19n2n. **a)** thermal demagnetization. Components B and C not isolated; **b)** and **c)** - mixed AF (20 mT) and thermal demagnetization. After AF and very large drop in *NRM* intensity a component B or Cn is revealed between 350 and 450°C.

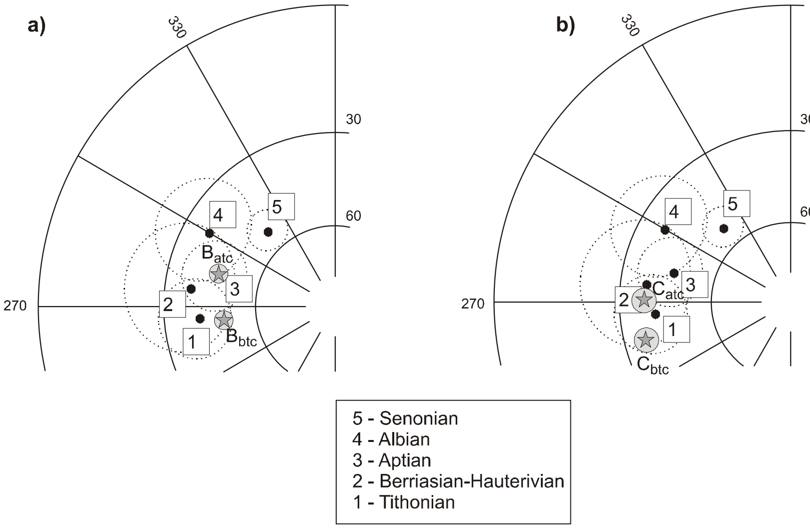

Fig. 9. Primary magnetizations of Tithonian–Senonian age from Transdanubian Range (numbers 1–5, after *Márton and Márton, 1983*) and its comparison with components B **(a)** and C **(b)** from Lókút section, before (btc) and after tectonic correction (atc).

Directions Cn and Cr are almost antipodal, however dispersal of Cr is greater than Cn (Fig. 7c). This is a very common situation observed in paleomagnetism of Mesozoic carbonate rocks from Carpathians (e.g. *Márton and Márton, 1981*; *Grabowski, 2005*; *Lewandowski et al., 2005*) and Apennines (*Houša et al., 2004*). It results from contamination of component C by component B even at the highest demagnetization steps. Components B and Cn are situated close to each other, therefore contamination of B component does not affect the clustering of Cn - Fisherian parameters of B and Cn are comparable, see Fig. 7b,c). Cr is situated on the opposite quadrant of stereonet and admixture of component B, of variegated intensity in particular samples, is the main reason of its rather poor clustering. That is also the reason why component B has apparently lower stability against thermal demagnetization in the reversed polarity than in normal polarity samples (see Fig. 6b,d). However the *McFadden and McElhinny (1990)* reversal test for component C is positive (with category C, the lowest), so it represents a reliable approximation of ancient geomagnetic field direction at the J/C boundary in the study area.

Nature of anomalous directions with very high *NRM* intensity is not certain. In our opinion they represent the lightning effects causing acquisition of an IRM. They may partially or completely obscure all other components within the rocks. Magnetizations with similar properties (high intensities, very low coercivity, anomalous directions) were reported by *Grabowski (2000)* from the Mesozoic of Tatra Mts., especially from higher areas not covered by vegetation, which makes the lightning model even more plausible.

4. MAGNETOSTRATIGRAPHY AND CORRELATION WITH GLOBAL POLARITY TIME SCALE (GPTS)

Normal (N) and reversed (R) polarity intervals within the Lókút section (according to polarity of component C) were numbered (see Fig. 5c–f). Bed −5, the lowest in the section, reveals reversed polarity of component C (interval R1). The three beds above (−4 to −2) are of normal polarity (N1). Beds −1 and 1 are again of reversed polarity (R2). Beds 2 to 8 reveal very high *NRM* intensities related to "lightning effect" (Fig. 5a). Beds 5, 6 and 8 are completely remagnetized by lightning, however in beds 2, 3 and 7 a reversed polarity of component C might be established, although characteristic component from bed 7 is somehow "transitional" (see Fig. 5). Therefore these beds were also included to reversed polarity interval R2. Beds numbered 11 to 19 are of predominantly normal polarity. Bed no. 20 is of reversed polarity (R3). Up to the bed no. 32 next normal polarity interval occur (N3). Bed no. 33 reveals a 'lightning effect" and then beds no. 34 and 35 are of reversed polarity - R4. Bed no. 36 was not sampled, while between beds no. 37 and 61 there is a long normal polarity interval (N4). Beds no. 50-52 are affected by lightning, however in bed 52 it was possible to restore the primary direction of normal polarity (Fig. 8). Bed no. 62 was not sampled. Beds no. 63−65 are of reversed polarity (R5) and 66−72 again of normal polarity (N5). The highest bed no. 73 is reversely magnetized (R6).

Microbiostratigraphically indicated J/C boundary falls between beds no. 44 and 45 - in the middle part of normal polarity interval N4 which must be correlated with magnetosubzone M19n2n (Fig. 5f). Basing on reference sections from Italy (*Ogg et al., 1991*; *Houša et al., 2004*; *Speranza et al., 2005*) and Carpathians (*Houša et al., 1999*; *Grabowski and Pszczółkowski, 2006*) a boundary of standard calpionellid zones A/B (boundary of subzones *intermedia*/*alpina*) is situated within this magnetosubzone (Fig. 1). Consequently, the magnetic intervals might be correlated to the GPTS down and up the section, verifying position of magneto(sub)zones with help of microfossil biostratigraphy. Reversed polarity interval R4 should be correlated with magnetozone M19r. It is situated entirely within subzone *intermedia* (A2) in the Upper Tithonian, which is conformable with standard Italian sections (*Ogg et al., 1991*) as well as Carpathian results (*Houša et al., 1999; Grabowski and Pszczółkowski, 2006*). Normal polarity intervals N3 and N2 might be correlated to magnetosubzones M20n1n and M20n2n respectively. They are separated by a short reversed polarity event (R3) which is most probably magnetosubzone M20n1r (Kysuca). It is situated within the *remanei* subzone (A1), similarly as in the Brodno section (*Houša et al., 1999*). Magnetosubzone M20n2n contains *Praetintinopsella* Zone and large part of *Ch. boneti* Subzone, which again agrees well with the Carpathian sections. Further down the section, reversed polarity interval R2 might correspond to magnetozone M20r. It is situated entirely within Chitinoidella Zone which is also a case in the Italian sections (*Ogg et al., 1991*). The chitinoidellids in the Brodno and Tatra Mts. sections do not occur below this magnetozone (*Houša et al., 1999*; *Grabowski and Pszczółkowski, 2006*). The M20n2n/M20r boundary correlates with boundary between Lower and Upper Tithonian. The lowermost normal polarity interval N1 should be assigned to a magnetozone M21n. In Lókút, as well as in Italian sections (*Ogg et al., 1991*), the lower boundary of the *Chitinoidella* Zone occurs within M21n. Reversed polarity interval in the bottom of the section, R1, should be correlated with magnetozone M21r.

Several magnetozones in the upper part of the section, above the J/C boundary, are situated entirely witin the *C. alpina* subzone of Lower Berriasian. Thus reversed polarity interval R5 should be correlated with a short magnetosubzone M19n1r (Brodno) and normal polarity interval N5 with magnetosubzone CM19n1n. The highest reversed polarity interval, R6, is most probably the lowermost part of M18r (Fig. 5f).

5. ROCK MAGNETISM

Magnetic susceptibility values range from 30×10^{-6} SI in the lower part of the section to 0 and negative values in the upper part (Fig. 10a). Also IRM values decrease up the section (Fig. 10b). There is a good correlation between both values for samples with $IRM_{1T} < 300$ mA/m (Fig. 11a), thus for beds younger than no. 8. This might indicate that magnetic susceptibility might be related to ferromagnetic minerals. However a relative contribution of para- and ferromagnetic minerals to magnetic susceptibility was not evaluated. Low coercivity minerals dominate within the section which might be deduced from negative values of *S* ratio. However variations of *S* ratio indicate admixture of high coercivity minerals. Samples with contrasting values of IRM_{1T} and *S* ratio (Fig. 11b) were subjected to IRM experiments using *Lowrie (1990)* method. It appeared that magnetic mineral assemblage is similar throughout the section. Low coercivity fraction consists of a mineral with maximum unblocking temperature below 610°C (Fig. 11c,d). It is interpreted as magnetite, perhaps slightly oxidized because IRM still persists in the 590°C, i.e. above Curie temperature of pure magnetite. Presence of oxidized phases is confirmed by distinct increase followed by decrease of magnetic susceptibility between 350 and 500°C: this may be interpreted as evidence for goethite which, on heating, converts to maghemite. High gradient of IRM decrease on intermediate and high coercivity curves between 20 and 150°C accounts for the presence of goethite. It is worth mentioning that samples with anomalously high *NRM* intensities (see Fig. 5a) do not reveal high amount of magnetic minerals. This supports our model that their unusually strong magnetization might be an IRM produced naturally by lightning strikes.

6. SEDIMENTATION RATE AND COMPARISON WITH OTHER TETHYAN SECTIONS

Magnetostratigraphic calibration enables a rough estimation of sedimentation rate which might be of importance in paleoenvironmental and paleogeographical reconstructions. The calculations are presented in the Table 2 and Fig. 10d. They should be treated as approximations only and not absolute values. Age assignment to bulk of magnetozones is performed by interpolation assuming constant spreading rate in the Hawaiian anomalies block. This is because only few magnetozones are reliably radiometrically dated (*Opdyke and Channell, 1996; Gradstein et al., 2004*).

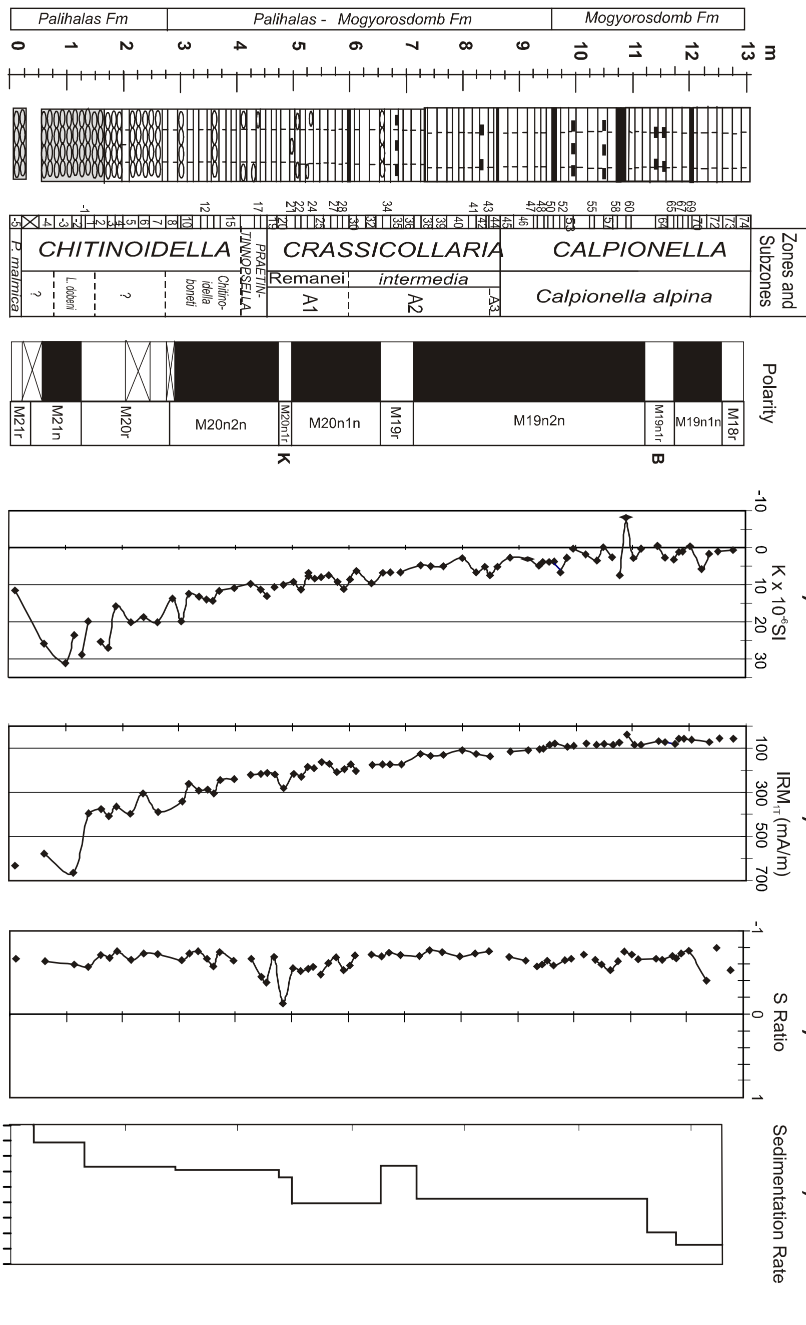

Fig. 10. Rock magnetic parameters and sedimentation rate in the Lókút section: **(a)** K - magnetic susceptibility; **(b)** IRM_{1T} - isothermal remanent magnetization acquired in the field of 1 T; **(c)** S - ratio of low to high coercivity minerals; **(d)** mean sedimentation rate for specific magneto(sub)zones from Table 2; K - Kysuca magnetosubzone; B - Brodno magnetosubzone.

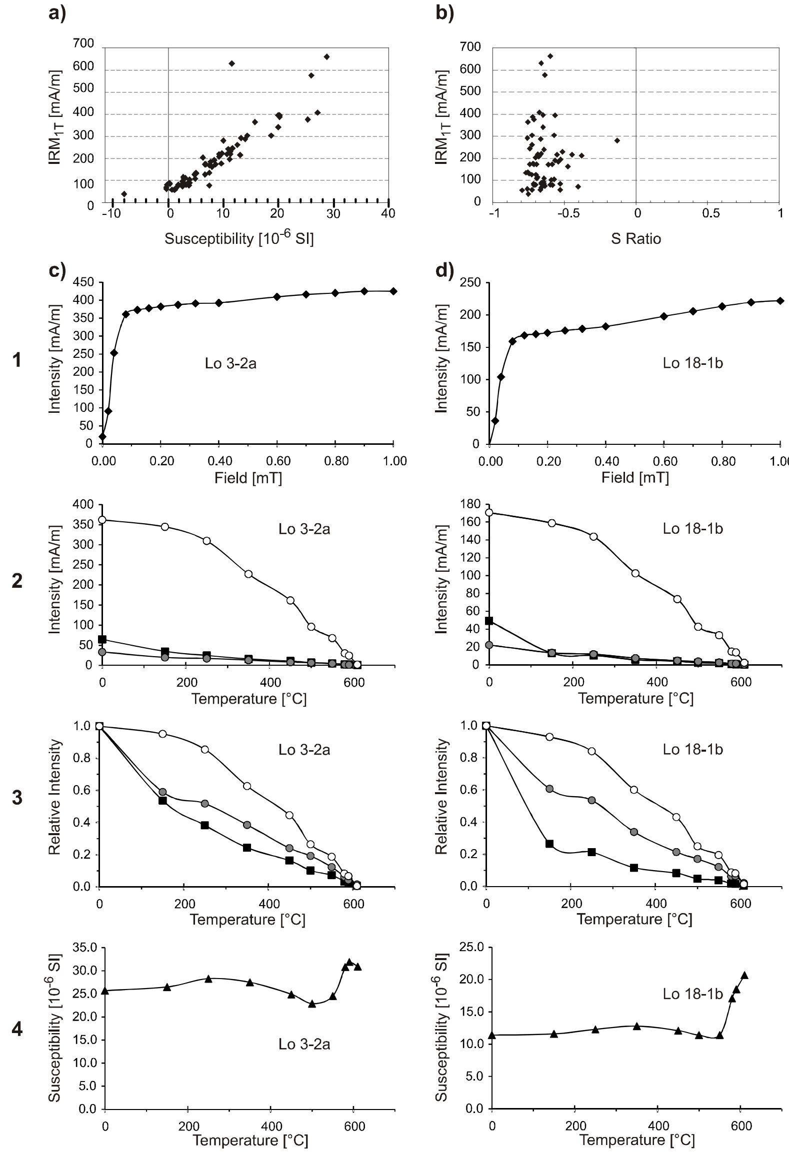

Fig. 11. Correlation of IRM_{1T} with k **(a)** and S ratio **(b)**, and petromagnetic properties of selected samples: **(c)** - sample Lo3-2 (bed no. 3, $k = 27.06 \times 10^{-6}$ SI, $IRM_{1T} = 407$ mA/m, $S = -0.67$). **(d)** - sample Lo18-1 (bed no. 18, $k = 13.1 \times 10^{-6}$ SI, $IRM_{1T} = 213.6$ mA/m, $S = -0.38$). 1 - stepwise acquisition of the IRM, 2 - thermal demagnetization of the IRM acquired in the fields of 0.12 T, 0.36 T and 1 T in three perpendicular directions, 3 - relative intensity decay of IRM acquired along three axes, 4 - magnetic susceptibility changes during thermal treatment.

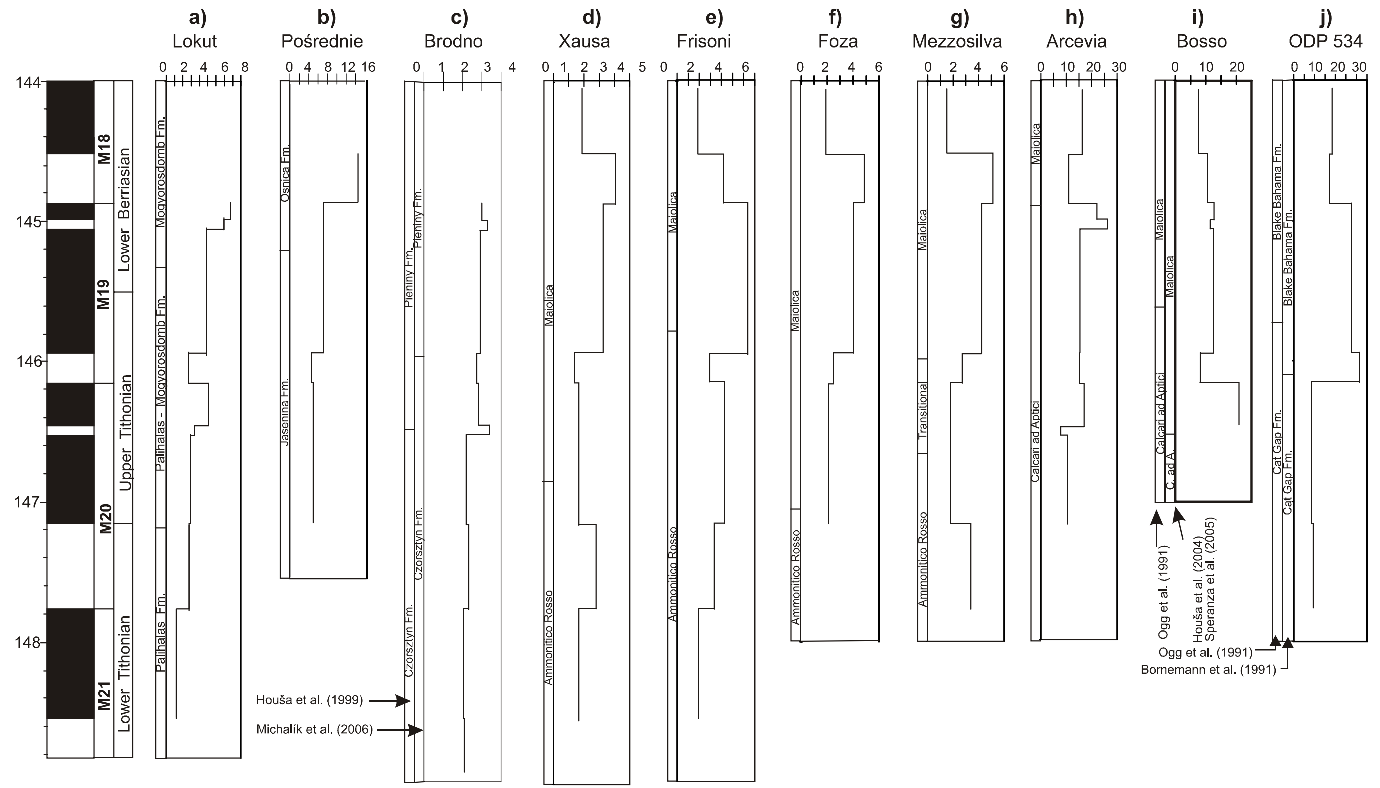

Fig. 12. Sedimentation rate around the Jurassic/Cretaceous boundary of representative sections from Carpathians **(a–c)**, Southern Alps - Trento Plateau **(d–g)** and Apennines **(h, i)**. **(a)** Lókút (this study); **(b)** Pośrednie (*Grabowski and Pszczółkowski, 2006*); **(c)** Brodno (*Houša et al., 1999*); **(d)** Xausa (*Channell et al., 1987*); **(e)** Frisoni (*Channell and Grandesso, 1987*); **(f)** Foza (*Ogg et al., 1991*); **(g)** Mezzosilva (*Ogg et al., 1991*); **(h)** Arcevia (*Speranza et al., 2005*); **(i)** Bosso (*Houša et al., 2004*; *Speranza et al., 2005*); **(j)** ODP 534 (*Ogg et al., 1991*).

Table 2. Attempt of sedimentation rate estimation in the Lókút section (timescale after *Gradstein et al., 2004*).

Magnetozone	Interval [m]	Thickness [m]	Duration [My]	Sedimentation Rate [m/My]
CM19n1n	11.8–12.65	0.85	0.11 (144.99–144.88)	7.73
CM19n1r[1]	11.21–11.8 11.42–11.8	0.59 0.38	0.07 (145.06–144.99)	8.43 5.43
CM19n2n[2]	7.03–11.42 7.19–11.21	4.39 4.57	0.88 (145.95–145.06)	4.99 4.57
CM19r	6.53–7.03 6.53–7.19	0.50 0.66	0.22 (146.16–145.95)	2.27 3.00
CM20n1n	4.94–6.53	1.59	0.31 (146.47–146.16)	5.13
CM20n1r	4.77–4.94	0.17	0.05 (146.52–146.47)	3.40
CM20n2n[3]	2.83–4.77 3.03–4.77	1.94 1.74	0.64 (147.16–146.52)	3.03 2.72
CM20r	1.25–2.83 1.25–3.03	1.58 1.78	0.61 (147.77–147.16)	2.59 2.92
CM21n	0.2–1.25 0.5–1.25	1.05 0.75	0.77 (148.54–147.77)	1.36 0.97

[1] No results from bed 62.
[2] No results from bed 36.
[3] No results from bed 8.

Sedimentation rate generally increases upsection. It is relatively low in the bottom of the section in the Lower Tithonian (magnetozone M21n - 1–1.4 m/My), while it amounts to 7–8 m/My in the Lower Berriasian (magnetosubzones M19n1r and 19n1n). Increase of sedimentation rate is not stepwise. A decrease is observed in magnetozone M19r, comparing to neighboring magnetosubzones M20n1n and M19n2n (almost 5 m/My). The values might be compared to other Carpathian and Alpine sections which are magnetostratigraphically calibrated (Fig. 12). The resolution of sampling in the Pośrednie section from the Križna unit of the Tatra Mts. (*Grabowski and Pszczółkowski, 2006*) was not as high as in Lókút. Nevertheless, the increase of sedimentation rate is observed from Upper Tithonian to Lower Berriasian, with slightly higher values than in Lókút (Fig. 12b). The section Brodno (PKB, Slovakia) is much more condensed (Fig. 12c). Czorsztyn Limestone Formation, between M21r and M20n2n, reveals sedimentation rate slightly more than 2 m/My, thus comparable with typical Pálihálás Formation in the Lókút section. However Pieniny Limestone Formation around the J/C boundary sedimented only slightly faster (ca. 3 m/My) - almost two times slower than Osnica (Tatra Mts.) and Mogyorósdomb Fms (Lókút).

The overall sedimentation rate within the Lókút section is comparable with Southern Alpine sections (*Channell and Grandesso, 1987; Channel et al., 1987; Ogg et al., 1991*;

Speranza et al., 2005). Four out of five Southern Alpine sections embracing J/C boundary, which were magnetostratigraphically calibrated (Xausa, Frisoni, Foza and Mezzosilva), are situated in the Trento Plateau. The fifth section Valle de Mis occurs in the Belluno Basin, east of Trento Plateau. All the Trento Plateau sections reveal relatively low sedimentation rate, no more than 7 m/My and in all of them an increase in sedimentation rate is observed - at least between magnetozones M20n and M19n (Fig. 12d–g). In three of them the sedimentation rate increases also in magnetozone M18r. The increase amounts from 1–2 m/My in magnetozone M20n to 4–6 m/My in magnetozones M19n–18r. The results are in good agreement with our data from Lókút section. Sedimentary conditions in both areas (Bakony Mts. and Trento Plateau) must have been similar. It supports the paleogeographic model that Transdanubian Range was in the Mesozoic a part of Adriatic plate (*Márton and Márton, 1983*) and Bakony Mts. constituted a northern prolongation of Trento plateau (*Schmid et al., 1991*; *Vörös and Galácz, 1998*). On the other hand, sedimentation rates in the sections from Appenines, i.e. Arcevia and Bosso (*Houša et al., 2004*; *Speranza et al., 2005*) and Central Atlantic ODP534 (*Ogg et al., 1991*) are much higher. In Bosso a decrease is observed already from the Upper Tithonian (Fig. 12i). In ODP534 the sedimentation rate dramatically increases close to the boundary of Cat Gap (red marlstones) and Blake Bahama formations (greyish limestone) in the lowermost part of magnetozone CM19r (Fig. 12j). However in younger magnetozones in the Lower Berriasian a decrease of sedimentation rate is observed which concerns also all Trento Plateau sections, in magnetozone M18n (Fig. 12d–g).

The boundary between the red limestone formations (Ammonitico Rosso or Calcari ad Aptici) and Maiolica is not always reflected in a significant change in the sedimentation rate (Fig. 12). The boundary is claimed to be diachronous (*Ogg et al., 1991*; *Speranza et al., 2005*), however this diachronism might be, at least partially, related to subjectively determined formation boundaries; different teams commonly proposed different position of the boundaries. For example *Houša et al. (2004)* and *Speranza et al. (2005)* drew the boundary between Calcari ad Aptici and Maiolica in the Bosso section (Apennines) just above short M20n1r (Kysuca) magnetosubzone. However in the earlier studies (*Ogg et al., 1991*) the boundary was situated in the lower part of M19n. The same authors indicated the boundary between Cat Gap Formation and Blake Bahama Formation in ODP534 in the lower part of M19n. In more recent papers the boundary is drawn in the lowermost part of M19r (*Bornemann et al., 2003*; *Tremolada et al., 2006*). Also in the Brodno section, the boundary between Czorsztyn and Pieniny Formations, originally defined at the top of M20n1r (Kysuca) magnetosubzone (*Houša et al., 1999*), was shifted closer to J/C boundary (upper part of M19r) by *Michalík et al. (2006).* These discrepancies might arise from stepwise sedimentary changes within the sections, just like as we found in the Lókút section.

7. MAGNETIC MINERALOGY AND SEDIMENTARY CHANGES

It might be of interest, if inferred changes of sedimentation rate are somehow reflected by rock magnetic parameters within the sections. For the Lókút section, there is a general inverse correlation between magnetic susceptibility κ and IRM_{1T} values, and sedimentation rate (Fig. 10). The correlation is not perfect, the magnetozones M20n1n and

19r do not fully conform to the pattern. However it is quite clear that para- and ferromagnetic matrix, responsible for κ and IRM_{1T}, is somehow diluted upsection. *Mayer (1999)* and *Mayer and Appel (1999)* proved that there is a close negative correlation between magnetic susceptibility and carbonate content for Maiolica Formation in the Southern Alps. Therefore we believe that the upward decreasing trend of magnetic susceptibility in our section is also related to enhanced carbonate productivity. Calcium carbonate content increase in pelagic sediments around the J/C boundary seems to be a regional event in low latitude seas. It was documented in several DSDP sites in Central Atlantic (*Bornemann et al., 2003*; *Tremolada et al., 2006*). It was interpreted as result of a calcareous nannoplankton bloom (Nannofosil Calcification Event), which started in the Middle and Upper Tithonian and is correlated with magnetozone M20n. In the Lókút section it coincides with termination of the typical Pálihálás Formation (see Figs. 10 and 12). The sedimentation rate in DSDP 105 (*Bornemann et al., 2003*) correlates well with absolute abundance of nannofossils. The magnetic susceptibility decrease in the Upper Tithonian, of similar magnitude as in Lókút, was also reported from other sections: Brodno (*Houša et al., 1999*), Bosso (*Houša et al., 2004*) and Nutzhof (*Pruner et al., 2009*). Also in the Pośrednie section from the Tatra Mts. there is a marked contrast of susceptibility between Upper Tithonian (Jasenina Formation) and Lower Berriasian (Osnica Formation) (*Grabowski and Pszczółkowski, 2006*). It was interpreted as result of carbonate productivity increase too. Then it might be concluded that magnetic parameters of pelagic carbonates at the J/C boundary might reflect some global (or at least more than regional) paleoenvironmental changes. More detailed comparison of the magnetic record between Lókút and Pośrednie sections reveals some differences that might result from local sedimentary conditions. Sedimentary changes in the Pośrednie and other sections from the Tatra Mts. were clearly reflected in changes of rock magnetic parameters which coincide with litho- and biofacies boundaries and accordingly boundaries of formations (*Grabowski and Pszczółkowski, 2006*). On the other hand lithological and rock magnetic changes within the Lókút section are much more stepwise. The most important magnetic contrast in the Pośrednie section - magnetic mineralogy change at the boundary of Jasenina and Osnica Formations and sharp decrease of magnetic susceptibility within magnetozone M19n, close to Brodno magnetosubzone, is not recognized in the Lókút section at all (Fig. 10). Magnetic susceptibility in the sections from the Križna unit in the Tatra Mts. is higher than in Lókút. It concerns calpionellid limestones ($40–60 \times 10^{-6}$ SI in the of Osnica Formation and only $0–5 \times 10^{-6}$ SI in Mogyorósdomb Formation) as well as underlying Jasenina ($60–150 \times 10^{-6}$ SI) and Pálihálás or Mogyorósdomb-Pálihálás Formation ($5–30 \times 10^{-6}$ SI). This might result from different sedimentary settings. Sections studied in the Križna unit were situated in the deep water Zliechov trough (*Michalík, 2007*) typified by a significant input of terrigenous detrital material and higher sedimentation rate. Area of deposition of Lókút section was a moderately deep pelagic plateau typified by a very limited terrigenuos input. The difference in the paleogeographic setting is also reflected in different abundances of radiolarians. In Lókút radiolarians are scarce, while they are abundant in the Križna unit sections which were situated in a deeper basin.

8. CONCLUSIONS

Chronostratigraphic framework for the Lókút section within its Tithonian–Lower Berriasian part was constructed using integrated bio- and magnetostratigraphic analysis. Magnetozones M21r (Lower Tithonian, *P. malmica* Zone) to M18r (Lower Berriasian, *C. alpina* Subzone) were identified. Sedimentation rate increases from 1−2 m/My in Tithonian to 5−7 m/My in Berriasian, which is a trend corresponding to other J/C boundary sections in the Trento Plateau (Southern Alps). A regional trend of increasing carbonate content in pelagic sediments across J/C boundary is most probably reflected by magnetic susceptibility decrease in Lókút, as well as throughout the other Tethyan sections, so far investigated. This general trend might be obscured by local sedimentary conditions (detrital influx, redeposition etc.), therefore significant differences of magnetic properties and sedimentation rate patterns might be observed between several Carpathian sections or between Southern Alpine and Apennine sections. Combination of standard bio- and magnetostratigraphy with rock magnetic stratigraphy provides a useful tool in unraveling sedimentary changes of local and regional significance in ancient pelagic environments.

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