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Preliminary palaeomagnetic study of the High Tatra granites, Central Western Carpathians, Poland

Jacek GRABOWSKI, Aleksandra GAWĘDA

Grabowski J., Gawęda A. (1999) — Preliminary palaeomagnetic study of the High Tatra granites, Central Western Carpathians, Poland. Geol. Quart., 43 (3): 263–276. Warszawa.

Variscan granitoids of the High Tatra Mts. in Poland were the subject of palaeomagnetic, petrographical and rock magnetic investigations. The sampled rocks were granodiorites, rarely tonalites showing weak hydrothermal alterations (chloritisation, epidotisation). 31 hand samples from 7 localities were palaeomagnetically investigated. Stable palaeomagnetic directions of Late Palaeozoic age were isolated in four localities (mean direction: $D = 193°$, $I = 17°$, $\alpha_{95} = 12$, $k = 59$, palaeopole: 4°E, 31°S). The stable magnetisation resides in hematite. This mineral occurs in hematite-ilmenite intergrowths that exsolved in high temperatures (670–720°C) and as secondary hematite of hydrothermal origin. Because of heterogeneity of magnetic carriers it is possible that the characteristic magnetisation is shifted in time between localities. Question of tectonic tilt of the High Tatra granite is discussed. The age of characteristic magnetisation based on palaeoinclination estimations apparently fits the isotopic cooling age of the intrusion (330–300 Ma) if tectonic correction is not applied. The palaeopole is situated between the European and African Apparent Polar Wander Paths (APWP) and could be matched with both reference curves. After tectonic correction the palaeopole could be matched only with the African APWP at the point *ca.* 360 Ma. In this case the magnetisation related to the high temperature hematite would preceed the cooling ages recorded by Ar-Ar method.

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Key words: Central Western Carpathians, Tatra Mts., granite, palaeomagnetism, petrography, rock magnetism.

INTRODUCTION

Palaeomagnetic investigations have not been carried out yet in the crystalline massifs of the Central Western Carpathians (CWC) in Poland and Slovakia. In the CWC the Variscan granitoid intrusions are common, forming a crystalline basement of the Alpidic orogenic belt (I. Petrik *et al.*, 1994). The High Tatra massif is the northernmost crystalline core, ranked among the Tatric unit, uplifted in the Neogene (M. Kováč *et al.*, 1994). Palaeomagnetism might be a useful tool for unravelling tectonic history of crystalline massifs (e.g. J. L. Pereira *et al.*, 1996; C. L. Rosenberg, F. Heller, 1997). In this paper we report the results of pilot palaeomagnetic investigations of the Polish part of the High Tatra granitoid supported by petrographical and rock magnetic studies[1].

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GEOLOGICAL SETTING OF THE STUDY

The crystalline basement of the Tatra Mts. is composed of pre-Mesozoic metamorphic rocks and granitoids, overlain by sedimentary Mesozoic cover sequence and nappes and Paleogene (Fig. 1a). Two structural units of the basement, separated by a low angle thrust fault were distinguished (M. Janak, 1994). The lower unit is composed of the mica schists complex in the Western Tatra Mts. (Fig. 1a). The upper unit is formed by variegated gneisses and amphibolites intruded by granitoid pluton. Both units form a Variscan nappe pile, exhibiting an inverted metamorphic zonation.

The recently accepted tectonic model of granitic magma emplacement assumed that the granite intruded synkinematically in relation to the Variscan uplift of the upper tectonic unit (M. Kohut, M. Janak, 1994). The depth of magma intrusion, stated on the basis of thermobarometric calculations for envelope rocks, was delimited as 18–22 km what is adequate to the mid-crustal level ($P = 5–6$ kbar; $T = 450–550°C$). The thermal influence of the intrusion shifted isobarically the

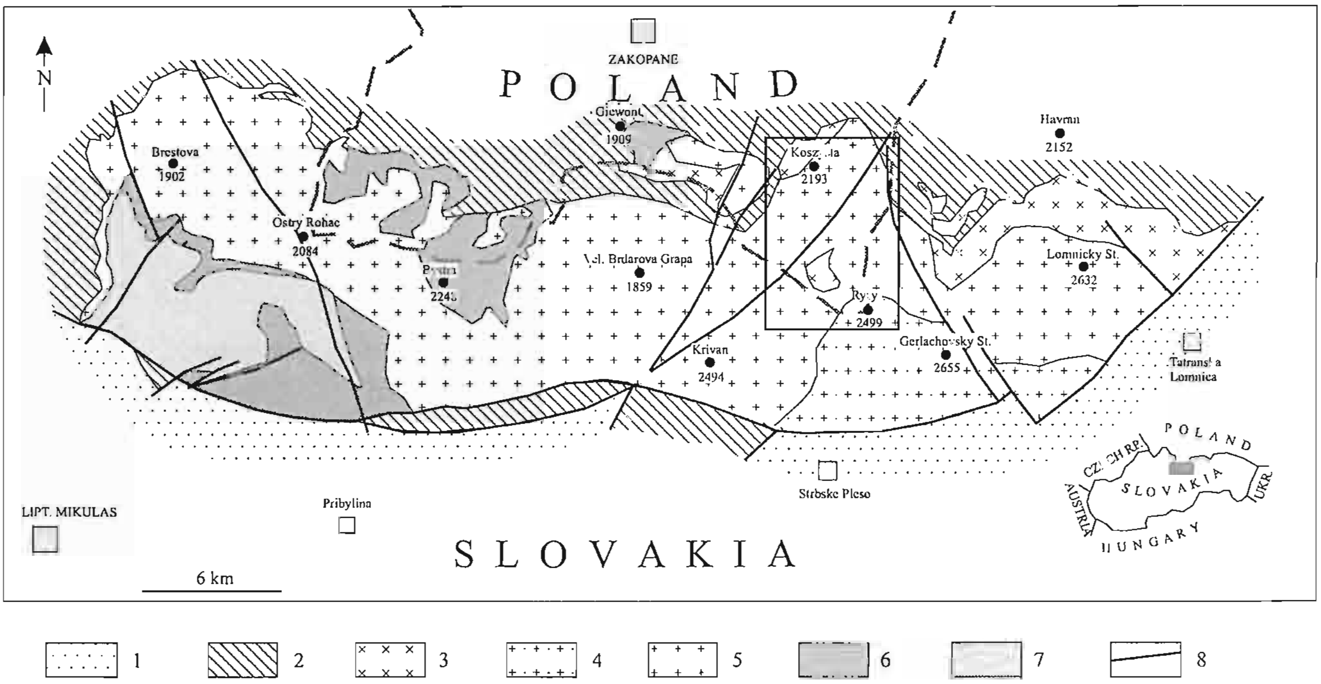

Fig. 1a. Tectonic sketch of the Tatra Mts. (after M. Kohut, M. Janak, 1994, modified)

1 — Paleogene; 2 — Mesozoic; 3–5 — granitoids: 3 — Goryczkowa type, 4 — High Tatra type, 5 — common Tatra type; 6 — migmatites and amphibolites; 7 — mica schists; 8 — faults; box indicates the area pictured in the Figure 1b

stability field to the T = 700–730°C (L. Ludhova, M. Janak, 1996; R. Piwkowski, A. Gawęda, 1996). The intrusion was emplaced into an extensional shear zone of the regional extent, with the general dextral (E–W) sense of shearing (M. Kohut, M. Janak, 1994). Local thrust zones within granitic body of the Polish part of the High Tatra Mts., with SW to NE directed tectonic transport were reported by K. Piotrowska (1997).

The isotopic cooling ages of the granitoid intrusion of the Tatra Mts. (Fig.1a) range between 300–330 Ma, according Ar-Ar and Rb-Sr methods (J. Burchart, 1968; H. Maluski *et al.*, 1993; M. Janak, 1994)[2]. The intrusion itself is petrographically heterogeneous. Four types of granitoid rocks were distinguished (M. Kohut, M. Janak, 1994, 1996):

1. High Tatra type (biotitic tonalite to muscovite-biotite granodiorite).
2. Common Tatra type (granodiorite to granite).
3. Goryczkowa type (porphyric granite to granodiorite).
4. Biotite-amphibole quartz diorite.

The heterogeneity of the granite body seems to be produced by the melting of the heterogeneous source crustal rocks and differentiation of the magma chamber but of minor importance (M. Kohut, M. Janak, 1996).

Preliminary geochemical data suggesting the plate tectonic setting of the Tatra granitoid pluton are unequivocal: some features pointed to the volcanic arc environment (VAG according to M. Kohut, M. Janak, 1994; J. Degenhart *et al.*, 1996; A. Wilamowski, 1998) on the other hand the continental arc and collisional granite origin is suggested (M. Kohut, M. Janak, 1996).

It is assumed that tectonic correction is necessary to restore the crystalline core to its original orientation (M. Książkiewicz, 1972; J. Piotrowski, 1978; M. Bac-Moszaszwili *et al.*, 1984). The rotation by 20–30°to the N around horizontal, sub-latitudinal axis was supposed to take place in the Neogene during the uplift of the Tatra Mts. (J. Piotrowski, 1978; B. Sperner, 1996). It is supported by a very uniform moderate northern dips of the autochthonous Lower Triassic strata on the northern slopes of the Tatra crystalline core. The analyses of seismic profiles (Č. Tomek, 1993) and geoelectric data (J. Lefeld, J. Jankowski, 1985) confirm the assumptions of geologists (e.g. D. Plašienka, 1991; M. Putiš, 1992) that the Tatra crystalline massif is allochthonous and was uplifted and transported to the North during the Alpine orogeny.

SAMPLING AND EXPERIMENTAL METHODS

Sampling was carried out in the Polish part of the High Tatra Mts., along the touristic path from Morskie Oko to Czarny Staw (MO1), Czarny Staw to Kazalnica Mięguszowiecka (MO2), in the Za Mnichem Valley (M), at the footwall

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of Świstówka Roztocka (RZ), at Wodogrzmoty Mickiewicza (WM), and in the Gąsienicowa Valley (HG and K) (Fig. 1b). 31 samples of massive rocks from 7 localities, without the visible foliation, net of joints, faults and without the hydrothermal alterations were taken.

Standard palaeomagnetic specimens with 2.5 cm in diameter and 2.2 cm in height were drilled from hand samples. Usually 3–6 specimens were obtained from each hand sample. Natural remanent magnetisation (NRM) was measured by means of the JR-5 spinner magnetometer while magnetic susceptibility was monitored with KLY-2 bridge. Anisotropy of magnetic susceptibility (AMS) was computed using the Aniso program (V. Jelinek, 1977). The following parameters which characterise the AMS ellipsoid were examined (D. H. Tarling, F. Hrouda, 1993):

1. Mean susceptibility $K_m = (K1 + K2 + K3)/3$, where $K1 > K2 > K3$ are the principal susceptibilities in SI units.

2. Corrected anisotropy degree $P' = \exp \mathrm{sqrt}\{2[(\eta_1 - \eta_m)^2 + (\eta_2 - \eta_m)^2 + (\eta_3 - \eta_m)^2]\}$, where: $\eta_1 = \ln K1$, $\eta_2 = \ln K2$, $\eta_3 = \ln K3$ and $\eta_m = (\eta_1 + \eta_2 + \eta_3)/3$.

3. Shape parameter $T = [2(\eta_2 - \eta_3)/(\eta_1 - \eta_3)] - 1$, defining the shape of the AMS ellipsoid. The ellipsoid is oblate if $0 < T < 1$ and prolate if $-1 < T < 0$.

The rock specimens were thermally demagnetised with MMTD non-magnetic oven. Demagnetisation experiments and the NRM measurements were performed inside a Helmholz coils that reduced the geomagnetic field by 95%. Characteristic directions were calculated using the principal component analysis (J. L. Kirschvink, 1980). Fortran plot package (M. Lewandowski *et al.*, 1997) was used for orthogonal and stereographic projection of demagnetisation path. Palaeomagnetic poles were plotted using the GMAP for Windows package (T. H. Torsvik, M. A. Smethurst, 1994).

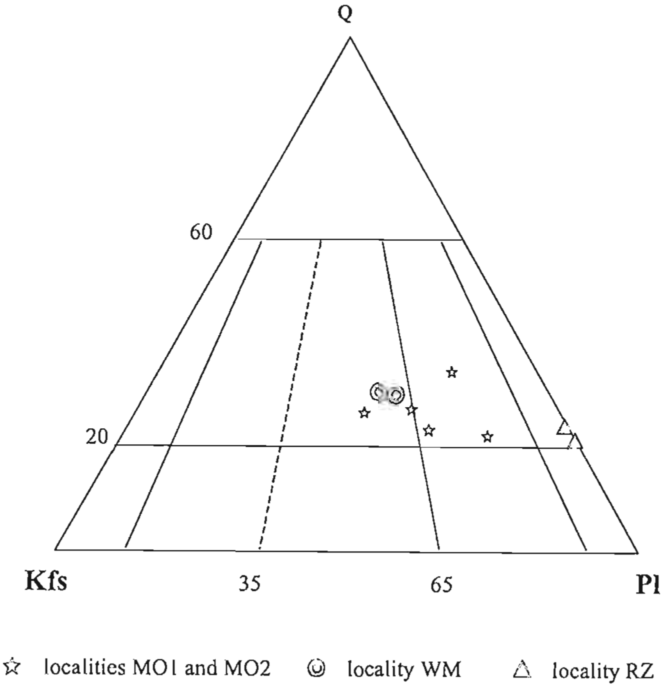

Fig. 2. Plot of modal composition of the granitoid samples in the quartz (Q), plagioclase (Pl) and orthoclase (Kfs) triangle

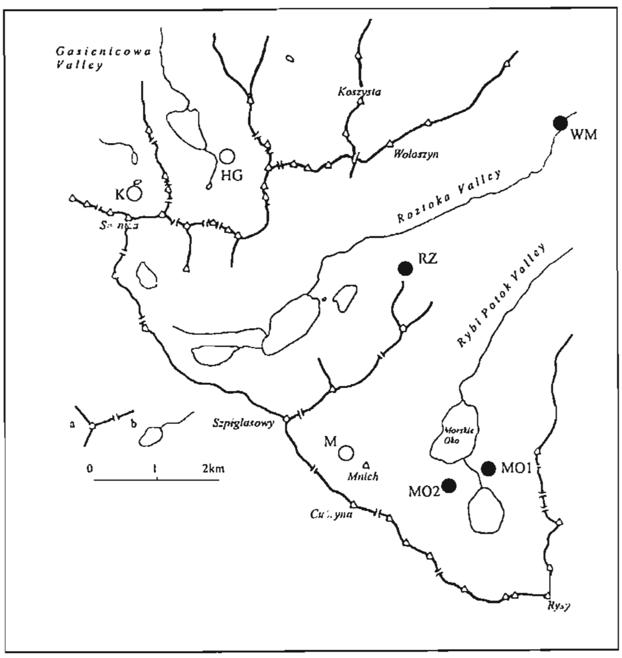

Fig. 1b. Topographic sketch of the Polish part of the High Tatra (simplified after K. Grochocka-Piotrowska, 1970) with sampling localities

Full dots — localities with good clustering of characteristic palaeomagnetic directions; open dots — localities with dispersed characteristic directions; a — mountain ranges with more important peaks and passes; b — lakes and streams; other explanations in the text

Several methods were used to identify the magnetic minerals: reflected light microscopy, stepwise acquisition of the isothermal remanence magnetisation (IRM) up to 2.7 T and thermal demagnetisation of the 3 axes IRM acquired in the fields of 0.1, 0.4 and 2.7 T (W. Lowrie, 1990). Some of the IRM experiments were carried out in the GeoForschungsZentrum in Potsdam. Standard thin sections were examined to evaluate the petrography of the studied rocks.

PETROGRAPHY

Petrographical analysis refers to localities which yielded good palaeomagnetic properties (see next paragraph and Fig. 1b).

Analysed granite samples differ in petrographical characteristics.

A. Samples from Morskie Oko (MO) show the widest range of compositions: they plot in the field of granite B and granodiorite, differing both in proportion of feldspars and quartz (Fig. 2). Mafic index is in the wide range of 5.6 to 10.7. Three generations of plagioclases could be distinguished: andesine (An_{35}–An_{50}) is enclosed in the idiomorphic to subidiomorphic oligoclase (An_{14}–An_{30}); interstitial albite (An_5–An_8) is the youngest generation. K-feldspar (microcline) is mainly xenomorphic, sometimes the larger (up to 10 mm in

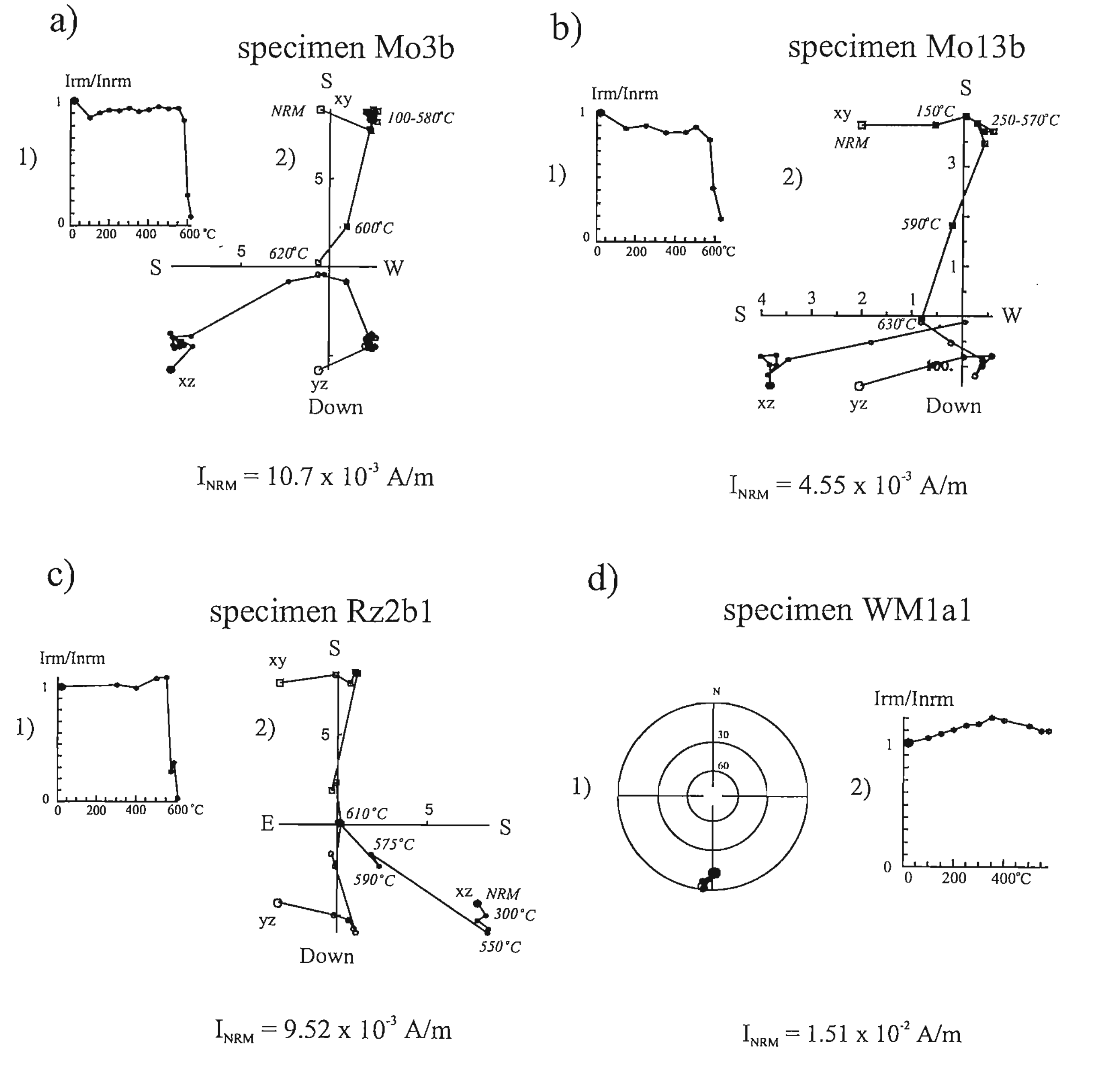

Fig. 3. Results of thermal demagnetisation of typical specimens; **a** — locality MO1, **b** — locality MO2, **c** — locality RZ; normalised intensity decay curve (1) and orthogonal plots (2); **d** — locality WM (after heating to 600°C specimens disintegrated) — stereographic projection of the demagnetisation path (1) and normalised intensity decay curve (2)

I_{NRM} — intensity of the natural remanent magnetisation; Irm/Inrm — intensity of the remanent magnetization after thermal treatment/intensity of the natural remanent magnetization

diameter) subidiomorphic grains of perthitic microcline could be found. Microcline grains usually contain the inclusions of andesine and quartz. At the contacts of plagioclase and microcline the myrmekitic intergrowths are common. Xenomorphic quartz grains show the deformation-induced wavy extinction. Biotite grains show the brown-beige-dark green pleochroism and are weakly chloritised. At the contacts of plagioclase and biotite epidote crystals can be found as the secondary phase. Secondary muscovite grows at the expense of K-feldspars and biotite. Zircon and apatite are the common accessory minerals. Among the ore minerals one can find hematite-ilmenite intergrowths with predominance of hematite (hematite A) and the abundant hydrothermal hematite (hematite B) as the individual mineral phase. MO samples characteristics are adequate to common Tatra type distinguished in the Slovak Tatras.

B. Samples from Wodogrzmoty Mickiewicza (WM) plot in the field of granite (Fig. 2) and show the narrow variation of mineral composition, either. The mafic index *M* is 12.5–13%. Large plagioclase crystals (0.9–3.4 mm), andesine-oligoclase in composition (An_{24}–An_{35}) are corroded both by albite (An_3–An_5) and microcline. Biotite is fully chloritised. Primary muscovite is weakly deformed, sometimes cut by the

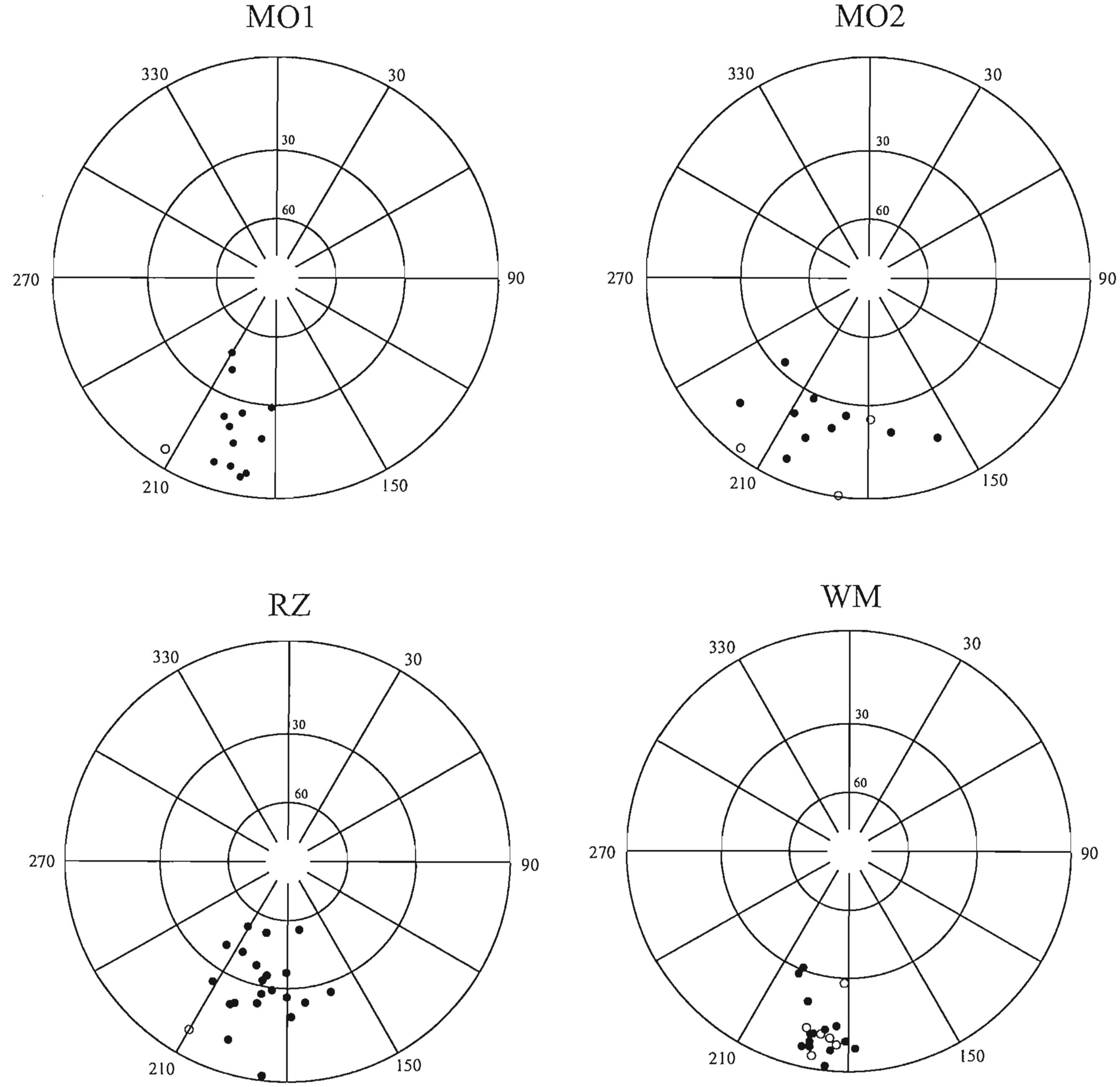

Fig. 4. Stereographic projection of the characteristic palaeomagnetic directions

Full dots — lower hemisphere projections; open dots — upper hemisphere projection; other explanations in the text

secondary muscovite. Among the ore minerals the most common is hydrothermal hematite B (sometimes coexisting with chlorite) and pyrite. The hematite-ilmenite intergrowths show very different proportions of both mineral phases. The overgrowths of hematite on the ilmenite-hematite intergrowths are sometimes associated with titanite.

C. Samples from Świstówka Roztocka (RZ) show tonalitic compositions with low quartz content and with very narrow compositional variation (Fig. 2). The mafic index *M* plots in the range of 5–6%. Subidiomorphic plagioclases (An_{18}–An_{30}) are the main constituents. Sheared, elongated quartz crystals show the mosaic internal structure and undulose extinction. Biotite is chloritised in 80–90%. Fine-grained primary muscovite shows ductile deformations, absent in secondary muscovite. Ore minerals are represented by ilmenite-hematite intergrowths with the rare remnants of ulvospinel.

The opaque minerals present in the investigated samples were of special interest. The primary mineral is Ti-magnetite (ulvöspinel), forming idiomorphic crystals or their aggre-

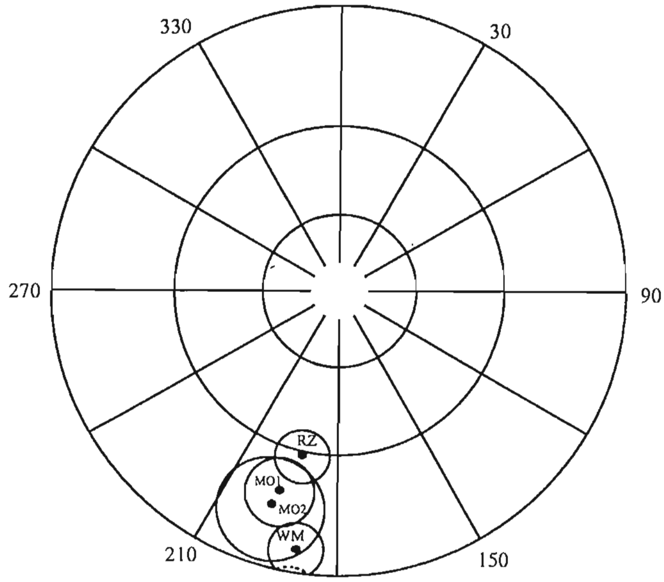

Fig. 5. Stereographic projections of the locality mean directions with their ovals of 95% confidence

Explanations in the text

gates, usually in association with biotite and/or andesine. The subsolidus exsolution textures are common in the magnetic minerals under consideration. The lenslike or podlike intergrowths of ilmenite-hematite show crystalographically controlled sharp contacts, along the (111) planes of ilmenite. The intergrowths showing different proportions of hematite/ilmenite were found in all investigated samples (Pl. I, Figs. 1–4). The Ti-rich phase forms 20–31% vol. of the whole crystal, but in some cases the ilmenite content changes from 71 to 5–8% vol. in different parts of the same grain. The enrichment in hematite can be interpreted as the result of influx of a late Fe-rich oxidised fluid penetrating the primary exsolution structures — similarly to catastrophic coarsening in perthites, described in feldspars (I. Parsons, W. L. Brown, 1984; W. L. Brown, I. Parsons, 1984). The calculated temperature of exsolution for sample RZ2 ranges from 670 to 720°C. That range of temperatures indicates rather low oxygen fugacity: from $10^{-15,5}$ to 10^{-18} (methods of calculation according to K. J. Spencer, D. H. Lindsley, 1981).

Separate grains of secondary hematite B are dissipated in all the studied granitoid samples. They are usually xenomorphic, rarely trigonal or pseudohexagonal plates can be seen. Its presence correlates with the secondary alterations of granitoids (chloritisation, epidotisation, sericitisation) and allows to suppose, that hematite crystallised at the late-hydrothermal stage of granite evolution.

DEMAGNETISATION RESULTS

Characteristic remanent magnetisation could be isolated in four localities only: MO1, MO2, RZ and WM. Samples from other three localities were not suitable for palaeomagnetic studies: they were either not stable during demagnetisation or there was no consistency of directions within a single hand sample. Therefore only palaeomagnetic properties of samples from successful localities are described here. On average NRM intensities were about 1 x 10^{-2} A/m (Tab. 1). Thermal demagnetisation was applied to all samples. Alternating field method was not effective due to presence of high coercivity magnetic minerals. Little part of the NRM was removed in the temperature range of 100–400°C but removed vectors displayed chaotic orientations and it was not possible to calculate any characteristic direction. Generally the NRM vector was very stable during thermal demagnetisation. A major decrease of the NRM intensity occurred in the temperature range of 550–630°C (Fig. 3). The characteristic directions are clustered in the southern part of the stereonet with shallow and moderately shallow inclinations (Fig. 4). It should be noted here that the distribution of characteristic directions is not Fisherian but they are streaked in the N–S direction. The mean directions of sampled localities also slightly differ from each other (Fig. 5, Tab. 2).

Table 1

Mean values of some rock magnetic parameters of the High Tatra granitoids

Locality	NRM intensity [x 10^{-3} A/m]		K_m [x 10^{-6} SI]		P'		T		n
	mean	range	mean	range	mean	range	mean	range	
WM	17.34	7.55–30.15	505	224–776	1.15	1.064–1.252	–1.139	–0.835–0.483	11
RZ	13.37	5.35–19.97	625	213–1133	1.11	1.079–1.165	0.186	–0.177–0.733	9
MO1	11.21	0.15–20.86	576	173–1092	1.10	1.037–1.153	0.004	–0.699–0.382	8
MO2	7.13	1.88–18.02	482	309–674	1.09	1.043–1.141	–0.223	–0.698–0.780	20

K_m — mean susceptibility, P' — mean degree of anisotropy, T — shape parameter, n — number of samples

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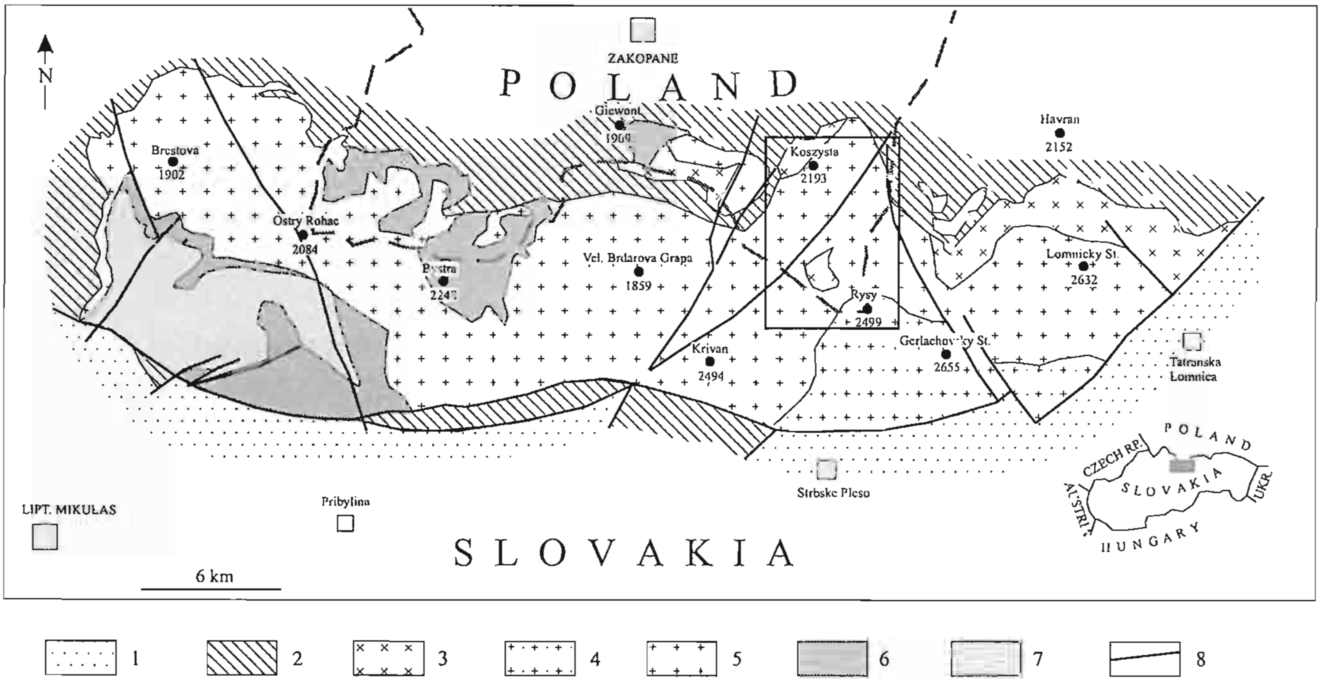

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2. Corrected anisotropy degree $P' = \exp \mathrm{sqrt}\{2[(\eta_1 - \eta_m)^2 + (\eta_2 - \eta_m)^2 + (\eta_3 - \eta_m)^2]\}$, where: $\eta_1 = \ln K1$, $\eta_2 = \ln K2$, $\eta_3 = \ln K3$ and $\eta_m = (\eta_1 + \eta_2 + \eta_3)/3$.
3. Shape parameter $T = [2(\eta_2 - \eta_3)/(\eta_1 - \eta_3)] - 1$, defining the shape of the AMS ellipsoid. The ellipsoid is oblate if $0 < T < 1$ and prolate if $-1 < T < 0$.

The rock specimens were thermally demagnetised with MMTD non-magnetic oven. Demagnetisation experiments and the NRM measurements were performed inside a Helmholz coils that reduced the geomagnetic field by 95%. Characteristic directions were calculated using the principal component analysis (J. L. Kirschvink, 1980). Fortran plot package (M. Lewandowski *et al.*, 1997) was used for orthogonal and stereographic projection of demagnetisation path. Palaeomagnetic poles were plotted using the GMAP for Windows package (T. H. Torsvik, M. A. Smethurst, 1994).

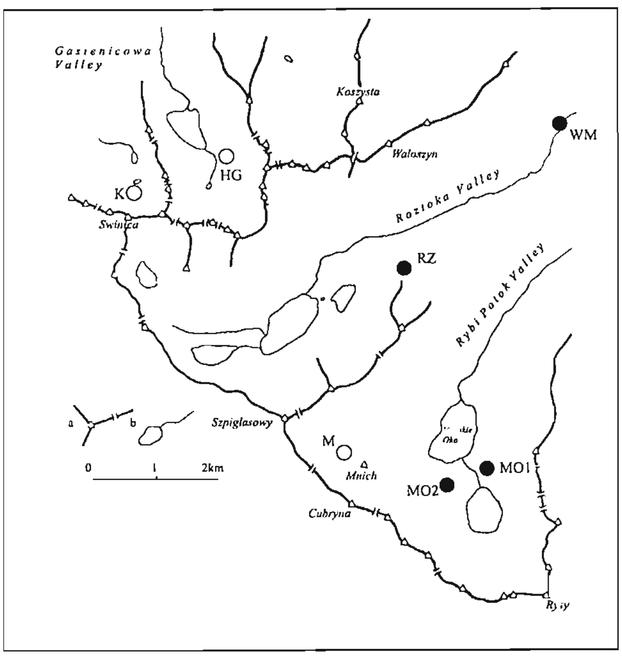

Fig. 1b. Topographic sketch of the Polish part of the High Tatra (simplified after K. Grochocka-Piotrowska, 1970) with sampling localities

Full dots — localities with good clustering of characteristic palaeomagnetic directions; open dots — localities with dispersed characteristic directions; a — mountain ranges with more important peaks and passes; b — lakes and streams; other explanations in the text

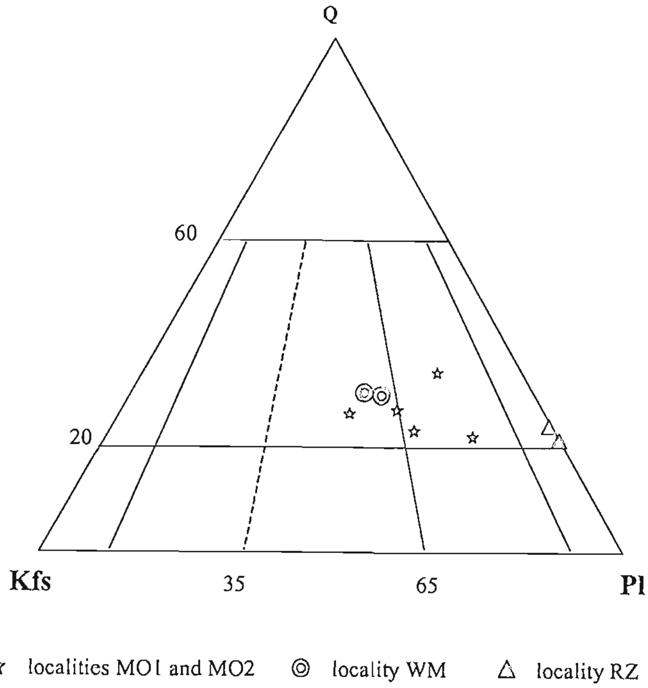

Fig. 2. Plot of modal composition of the granitoid samples in the quartz (Q), plagioclase (Pl) and orthoclase (Kfs) triangle

Several methods were used to identify the magnetic minerals: reflected light microscopy, stepwise acquisition of the isothermal remanence magnetisation (IRM) up to 2.7 T and thermal demagnetisation of the 3 axes IRM acquired in the fields of 0.1, 0.4 and 2.7 T (W. Lowrie, 1990). Some of the IRM experiments were carried out in the GeoForschungsZentrum in Potsdam. Standard thin sections were examined to evaluate the petrography of the studied rocks.

PETROGRAPHY

Petrographical analysis refers to localities which yielded good palaeomagnetic properties (see next paragraph and Fig. 1b).

Analysed granite samples differ in petrographical characteristics.

A. Samples from Morskie Oko (MO) show the widest range of compositions: they plot in the field of granite B and granodiorite, differing both in proportion of feldspars and quartz (Fig. 2). Mafic index is in the wide range of 5.6 to 10.7. Three generations of plagioclases could be distinguished: andesine (An_{35}–An_{50}) is enclosed in the idiomorphic to subidiomorphic oligoclase (An_{14}–An_{30}); interstitial albite (An_5–An_8) is the youngest generation. K-feldspar (microcline) is mainly xenomorphic, sometimes the larger (up to 10 mm in

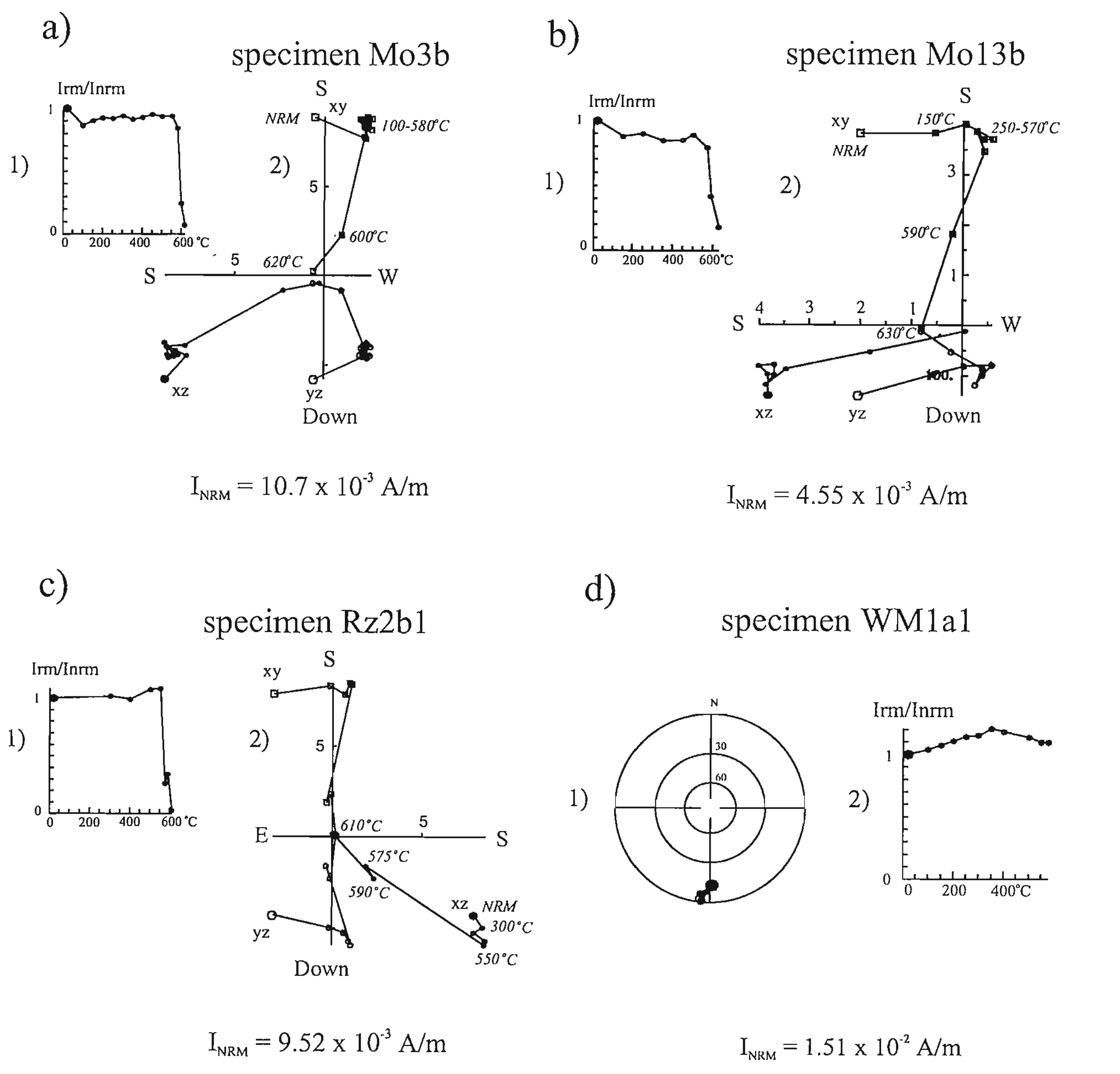

Fig. 3. Results of thermal demagnetisation of typical specimens; **a** — locality MO1, **b** — locality MO2, **c** — locality RZ; normalised intensity decay curve (1) and orthogonal plots (2); **d** — locality WM (after heating to 600°C specimens disintegrated) — stereographic projection of the demagnetisation path (1) and normalised intensity decay curve (2)

I_{NRM} — intensity of the natural remanent magnetisation; Irm/Inrm — intensity of the remanent magnetization after thermal treatment/intensity of the natural remanent magnetization

diameter) subidiomorphic grains of perthitic microcline could be found. Microcline grains usually contain the inclusions of andesine and quartz. At the contacts of plagioclase and microcline the myrmekitic intergrowths are common. Xenomorphic quartz grains show the deformation-induced wavy extinction. Biotite grains show the brown-beige-dark green pleochroism and are weakly chloritised. At the contacts of plagioclase and biotite epidote crystals can be found as the secondary phase. Secondary muscovite grows at the expense of K-feldspars and biotite. Zircon and apatite are the common accessory minerals. Among the ore minerals one can find hematite-ilmenite intergrowths with predominance of hematite (hematite A) and the abundant hydrothermal hematite (hematite B) as the individual mineral phase. MO samples characteristics are adequate to common Tatra type distinguished in the Slovak Tatras.

B. Samples from Wodogrzmoty Mickiewicza (WM) plot in the field of granite (Fig. 2) and show the narrow variation of mineral composition, either. The mafic index *M* is 12.5–13%. Large plagioclase crystals (0.9–3.4 mm), andesine-oligoclase in composition (An_{24}–An_{35}) are corroded both by albite (An_3–An_5) and microcline. Biotite is fully chloritised. Primary muscovite is weakly deformed, sometimes cut by the

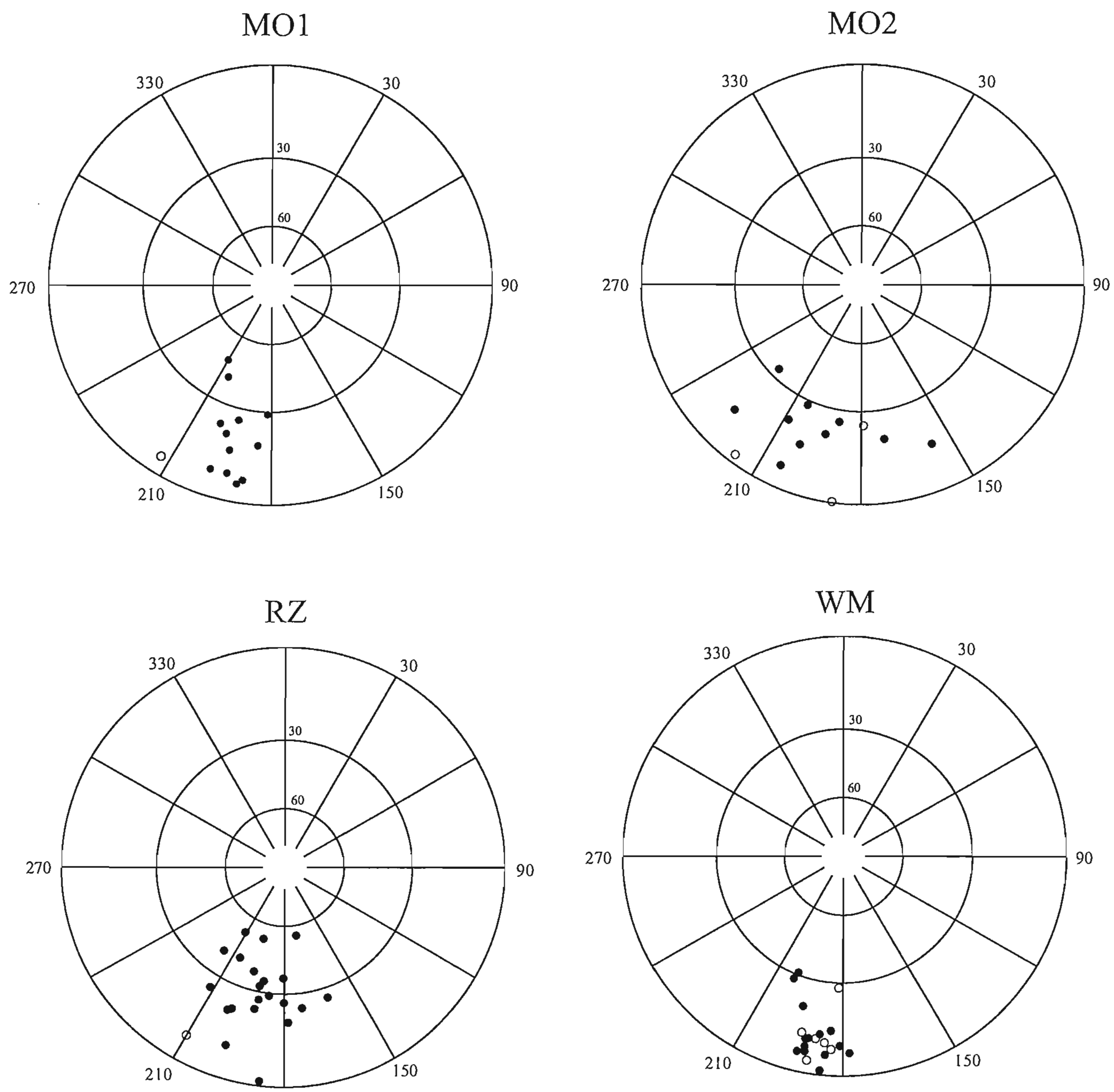

Fig. 4. Stereographic projection of the characteristic palaeomagnetic directions

Full dots — lower hemisphere projections; open dots — upper hemisphere projection; other explanations in the text

secondary muscovite. Among the ore minerals the most common is hydrothermal hematite B (sometimes coexisting with chlorite) and pyrite. The hematite-ilmenite intergrowths show very different proportions of both mineral phases. The overgrowths of hematite on the ilmenite-hematite intergrowths are sometimes associated with titanite.

C. Samples from Świstówka Roztocka (RZ) show tonalitic compositions with low quartz content and with very narrow compositional variation (Fig. 2). The mafic index *M* plots in the range of 5–6%. Subidiomorphic plagioclases (An_{18}–An_{30}) are the main constituents. Sheared, elongated quartz crystals show the mosaic internal structure and undulose extinction. Biotite is chloritised in 80–90%. Fine-grained primary muscovite shows ductile deformations, absent in secondary muscovite. Ore minerals are represented by ilmenite-hematite intergrowths with the rare remnants of ulvospinel.

The opaque minerals present in the investigated samples were of special interest. The primary mineral is Ti-magnetite (ulvöspinel), forming idiomorphic crystals or their aggre-

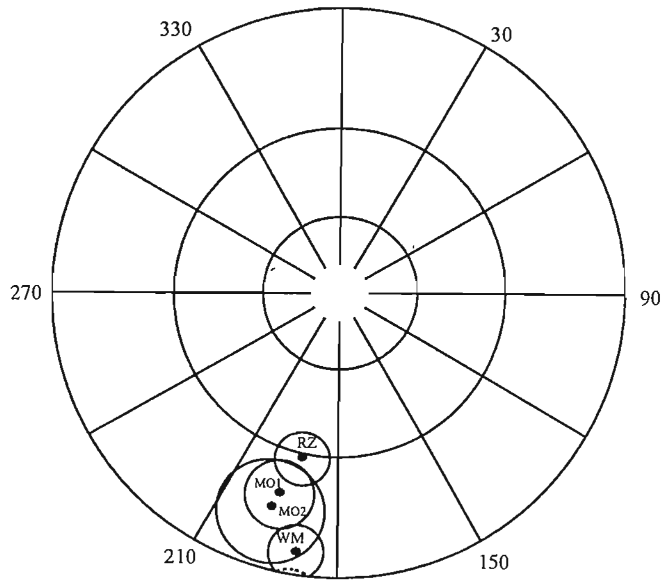

Fig. 5. Stereographic projections of the locality mean directions with their ovals of 95% confidence

Explanations in the text

gates, usually in association with biotite and/or andesine. The subsolidus exsolution textures are common in the magnetic minerals under consideration. The lenslike or podlike intergrowths of ilmenite-hematite show crystalographically controlled sharp contacts, along the (111) planes of ilmenite. The intergrowths showing different proportions of hematite/ilmenite were found in all investigated samples (Pl. I, Figs. 1–4). The Ti-rich phase forms 20–31% vol. of the whole crystal, but in some cases the ilmenite content changes from 71 to 5–8% vol. in different parts of the same grain. The enrichment in hematite can be interpreted as the result of influx of a late Fe-rich oxidised fluid penetrating the primary exsolution structures — similarly to catastrophic coarsening in perthites, described in feldspars (I. Parsons, W. L. Brown, 1984; W. L. Brown, I. Parsons, 1984). The calculated temperature of exsolution for sample RZ2 ranges from 670 to 720°C. That range of temperatures indicates rather low oxygen fugacity: from $10^{-15,5}$ to 10^{-18} (methods of calculation according to K. J. Spencer, D. H. Lindsley, 1981).

Separate grains of secondary hematite B are dissipated in all the studied granitoid samples. They are usually xenomorphic, rarely trigonal or pseudohexagonal plates can be seen. Its presence correlates with the secondary alterations of granitoids (chloritisation, epidotisation, sericitisation) and allows to suppose, that hematite crystallised at the late-hydrothermal stage of granite evolution.

DEMAGNETISATION RESULTS

Characteristic remanent magnetisation could be isolated in four localities only: MO1, MO2, RZ and WM. Samples from other three localities were not suitable for palaeomagnetic studies: they were either not stable during demagnetisation or there was no consistency of directions within a single hand sample. Therefore only palaeomagnetic properties of samples from successful localities are described here. On average NRM intensities were about 1 x 10^{-2} A/m (Tab. 1). Thermal demagnetisation was applied to all samples. Alternating field method was not effective due to presence of high coercivity magnetic minerals. Little part of the NRM was removed in the temperature range of 100–400°C but removed vectors displayed chaotic orientations and it was not possible to calculate any characteristic direction. Generally the NRM vector was very stable during thermal demagnetisation. A major decrease of the NRM intensity occurred in the temperature range of 550–630°C (Fig. 3). The characteristic directions are clustered in the southern part of the stereonet with shallow and moderately shallow inclinations (Fig. 4). It should be noted here that the distribution of characteristic directions is not Fisherian but they are streaked in the N–S direction. The mean directions of sampled localities also slightly differ from each other (Fig. 5, Tab. 2).

Table 1

Mean values of some rock magnetic parameters of the High Tatra granitoids

Locality	NRM intensity [x 10^{-3} A/m]		K_m [x 10^{-6} SI]		P'		T		n
	mean	range	mean	range	mean	range	mean	range	
WM	17.34	7.55–30.15	505	224–776	1.15	1.064–1.252	–1.139	–0.835–0.483	11
RZ	13.37	5.35–19.97	625	213–1133	1.11	1.079–1.165	0.186	–0.177–0.733	9
MO1	11.21	0.15–20.86	576	173–1092	1.10	1.037–1.153	0.004	–0.699–0.382	8
MO2	7.13	1.88–18.02	482	309–674	1.09	1.043–1.141	–0.223	–0.698–0.780	20

K_m — mean susceptibility, P' — mean degree of anisotropy, T — shape parameter, n — number of samples

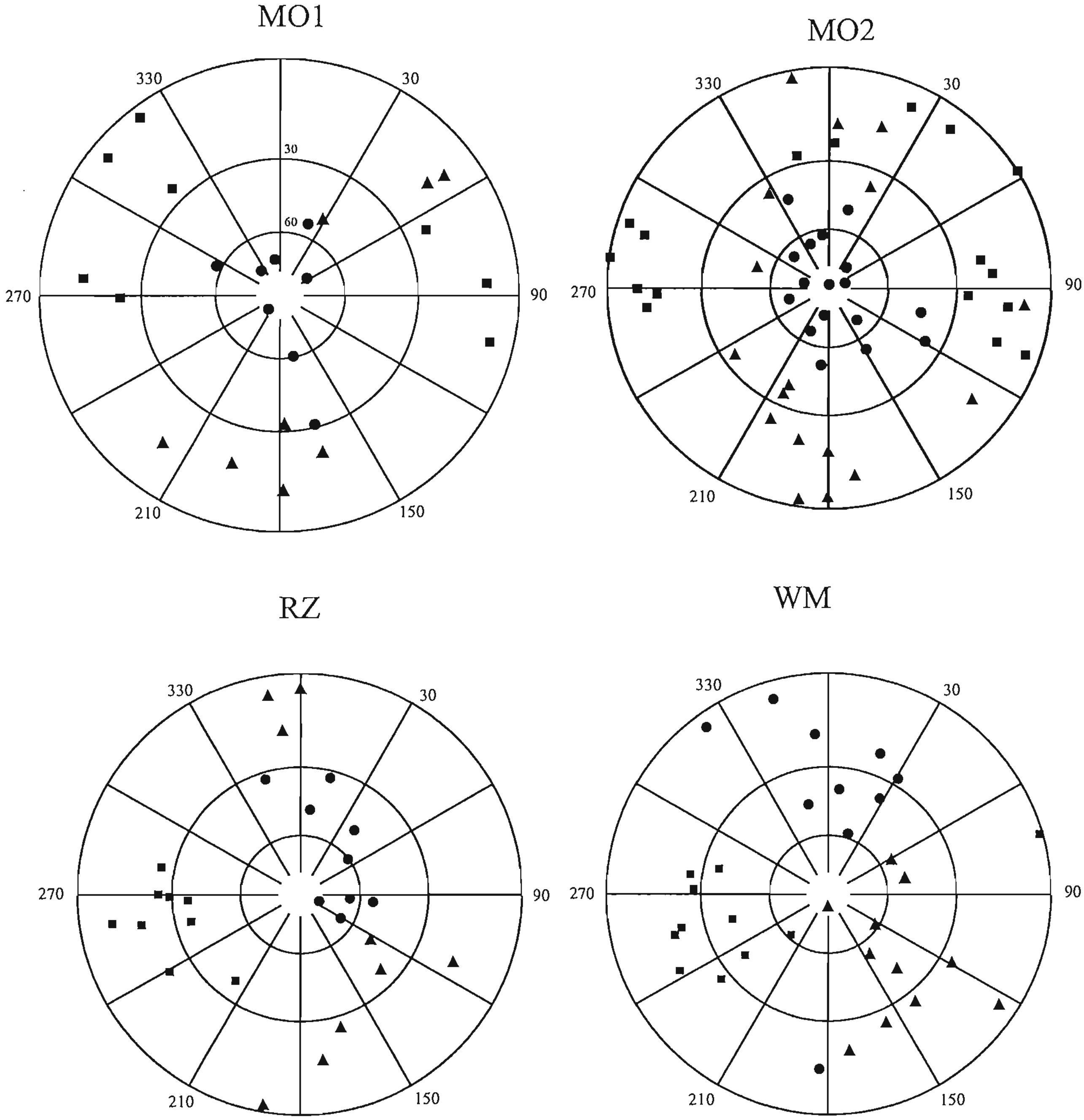

Fig. 6. Stereographic projection of the main axes of the AMS

Squares — maximum susceptibility axes; triangles — intermediate susceptibility axes; dots — minimum susceptibility axes; lower hemisphere projection; other explanations in the text

MAGNETIC SUSCEPTIBILITY AND ITS ANISOTROPY

According to the previous investigations (M. Kohut, M. Janak, 1994) the "High Tatra type" of granitoid have the features of "magnetic" granites (D. H. Tarling, F. Hrouda, 1993) with high susceptibilities between 10^{-3} and 10^{-2} SI units. The "common Tatra type" and "Goryczkowa type" reveal the intermediate properties between "magnetic" and "non-magnetic granites" with mean susceptibilities close to 10^{-3} SI units (M. Kohut, M. Janak, *op. cit.*). Mean susceptibilities of the granitoids from the MO1, MO2, RZ and WM localities ranging from 460×10^{-6} to 630×10^{-6} SI (see Tab. 1) are even lower. They are closer to the values of susceptibility of granitoids investigated by F. Hrouda and S. Kahan (1991, Table 2). Such a low value susceptibility is carried mostly by ilmenites and/or ilmeno-hematites (D. H. Tarling, F. Hrouda, 1993) possibly with contribution of paramagnetic matrix.

Anisotropy of magnetic susceptibility (AMS) in most samples is well defined. Mean degree of anisotropy *P'* for

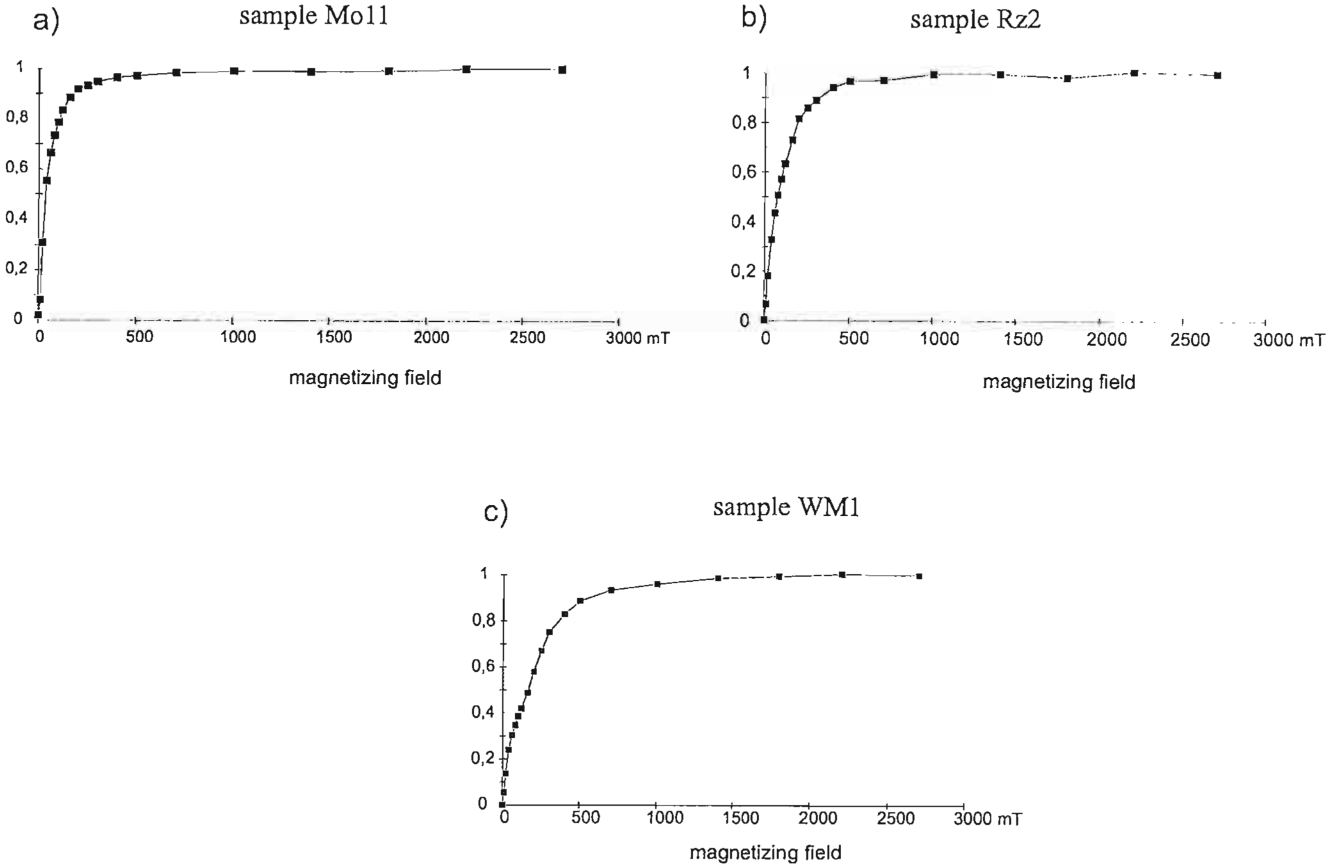

Fig. 7. IRM acquisition curves for typical specimens: a — locality MO2, b — locality RZ, c — locality WM

each locality varies between 1.09 (MO) and 1.15 (WM) (Tab. 1). High variation of the shape parameter *T* indicates that magnetic fabric is complex, consisting of planar and linear elements. The maximum susceptibility axes tend to group at the azimuths between 260 and 300°, while minimum and intermediate axes create a N–S or NNW–SSE directed girdle (Fig. 6). The magnetic fabric is just very similar to that described by F. Hrouda and S. Kahan (1991) from the Slovak part of the Tatra granite. The authors interpret it as deformational magnetic fabric of Alpine age that overprinted and locally obliterated the Variscan intrusive magnetic fabric.

Table 2

Palaeomagnetic directions from the High Tatra granitoids (modified after J. Grabowski, 1996)

Locality	*D*	*I*	α_{95}	*k*	Pole lg E/lt N	*n/N*
MO1	196	18	8.8	22.9	2/–30	13/4
MO2	197	14	13.1	10.9	0/–32	13/4
RZ	192	29	7.7	17.3	7/–25	22/3
WM	189	5	6.1	29.9	9/–38	20/3
Mean *Tbtc*	193	17	12	58.9	4/–31	–
Mean *Tatc*	194	47	12	58.9	7/–12	-

D — declination of palaeomagnetic direction, *I* — inclination of palaeomagnetic direction, α_{95}, *k* — Fisher statistics parameters, lg — longitude of palaeomagnetic pole, lt — latitude of palaeomagnetic pole, *n* — number of specimens, *N* — number of hand samples, *Tbtc* — component *T* before tectonic correction, *Tatc* — component *T* after tectonic correction

ROCK MAGNETIC STUDIES

Stepwise acquisition of the IRM reveals complex magnetic mineralogy of the studied rocks (Fig. 7). Low and high coercivity minerals are present in different ratios. In the WM1 sample the amount of high coercivity minerals is the highest and in the MO11 their contribution to the IRM is very small.

More detailed picture of magnetic minerals could be inferred from demagnetisation of the 3 axes IRM (Fig. 8). Minerals with low coercivity reveal maximum unblocking temperature close to the Curie point of magnetite (soft component in the Fig. 8). These features can be attributed to

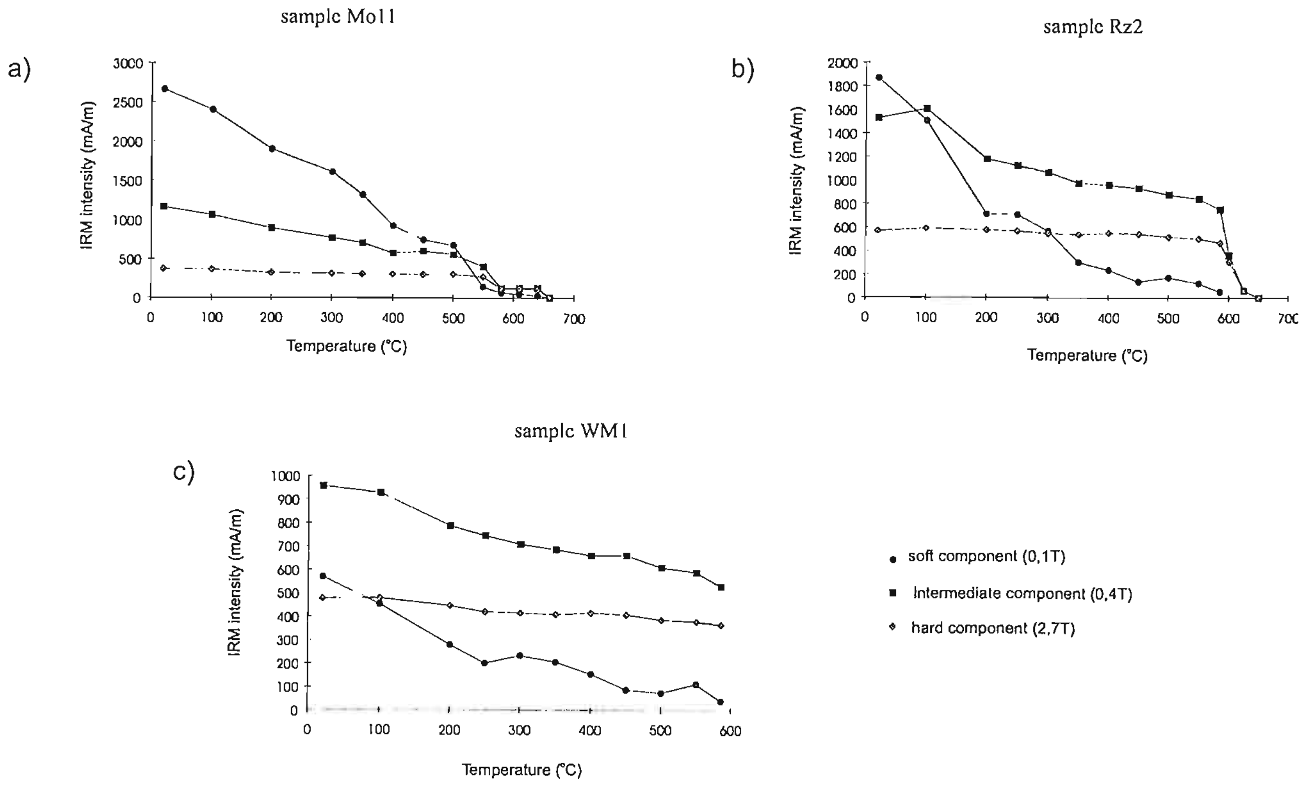

Fig. 8. Thermal demagnetisation of the 3 axis IRM (method described in: W. Lowrie, 1990): a — locality MO2, b — locality RZ, c — locality WM

ilmenite-magnetite series (titanomagnetites). High and intermediate coercivity fraction reveals unblocking temperatures between 550 and 630°C (intermediate and hard components in the Fig. 8). This indicates the presence of minerals of hematite-ilmenite series. The unblocking temperature spectrum is lower than maximum unblocking temperature of pure hematite (675°C). Obviously these minerals carry the characteristic component of magnetisation, because demagnetisation curves of the NRM (Fig. 3) and medium/high coercivity fraction in the Figure 8 are alike.

DISCUSSION

Palaeomagnetic pole *T*, calculated as a mean pole from the MO1, MO2, RZ and WM localities (Tab. 2), is situated between the Palaeozoic segments of the European and African Apparent Polar Wander Paths (APWP) (Fig. 9a, b). Hence, it certainly represents the Late Palaeozoic magnetisation. Several problems should be pointed out here.

1. Mean components from four studied localities are streaked along the N–S direction (Fig. 5). Do they represent a time-sequence of magnetisation? What magmatic/postmagmatic process should they be attributed to?

2. Which APWP, European or African (Western Gondwanian), should serve as a reference for the Tatra Mts.?

3. Should any correction for tectonic tilt be applied?

Any interpretation must start with the identification of magnetic carriers and origin of the ChRM. There should be no doubt that ChRM resides in hematite grains as indicated by unblocking temperature spectra. Petrographical analysis revealed that hematite was present in hematite/ilmenite intergrowths (hematite A) and in separate grains of possibly hydrothermal origin (hematite B).

The hematite-ilmenite exolutions, observed in all studied samples, are thought to be an effect of oxidation during magma cooling (A. F. Buddington, D. H. Lindsley, 1964). The Ti-magnetite or ulvöspinel behaves as binary solid solution only above 800°C. When the temperature drops down below 800°C and the magmatic fluids, rich in water and oxygen, penetrate the cooling granitoid body, the hematite-ilmenite miscibility gap occurs (K. J. Spencer, D. H. Lindsley, 1981). In the gap-field Ti-magnetite (and most of the "magnetic" phases) becomes unstable and exsolves into two components Fe_2O_3 (hematite) and $FeTiO_3$ (ilmenite) according to the reaction:

$$4Fe_2TiO_4 + O_2 = 4FeTiO_3 + 2Fe_2O_3$$

The calculated temperature of exsolution (670–720°C) seems to be the highest possible temperature in which the observed mineral structure started to form. Thus magnetic remanence residing in hematite A should be a typical ther-

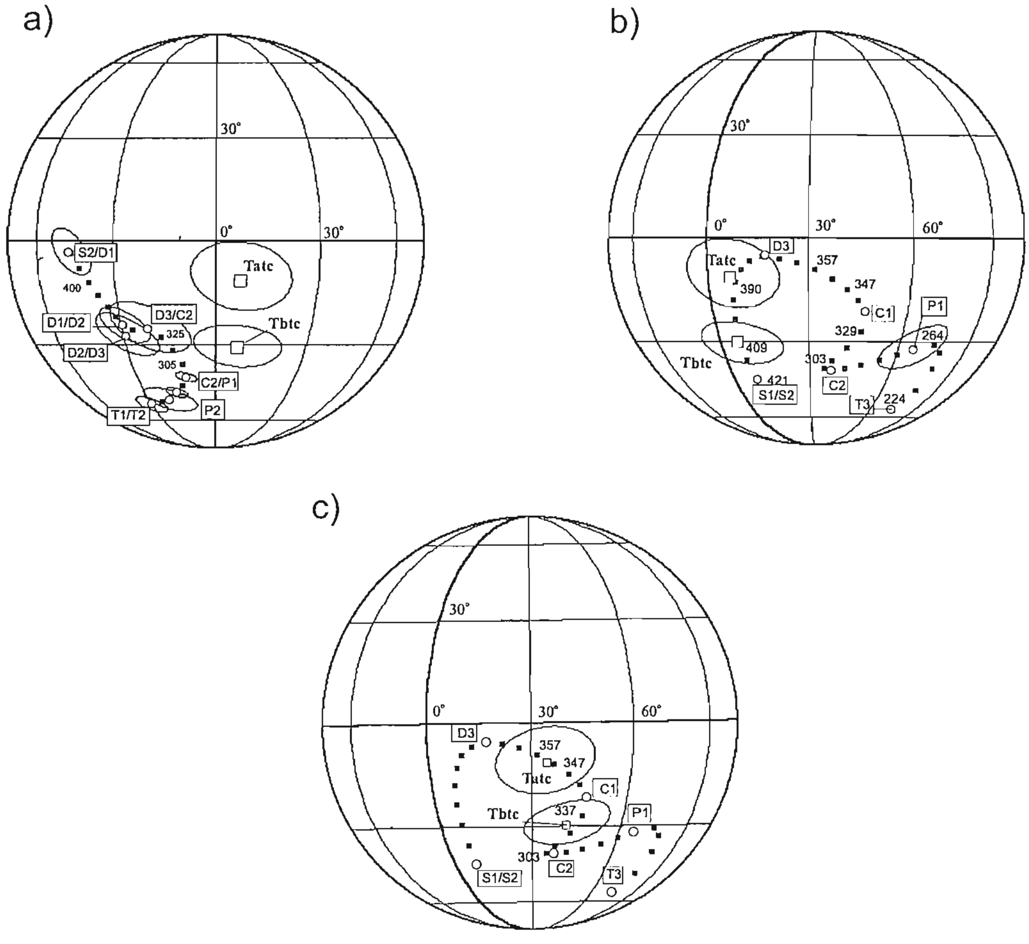

Fig. 9. Palaeopole *T* before (*Tbtc*) and after (*Tatc*) tectonic correction at the background of the APWP for: **a** — Europe, **b** — Africa; **c** — palaeopole *T* matched with the African APWP after 30° counter-clockwise rotation

European and African APWP after data of R. Van der Voo (1993); age calibration in Ma; S1, S2 — Early, Late Silurian, D1, D2, D3 — Early, Middle, Late Devonian, C1, C2 — Early, Late Carboniferous, P1, P2 — Early, Late Permian, T1, T2, T3 — Early, Middle, Late Triassic

moremanent magnetisation acquired in the high temperature possibly close to the Neel temperature of hematite.

On the other hand hematite B was formed during late hydrothermal processes in relatively low temperatures (most probably below 400°C). It should carry the remanence which is younger than that based on hematite A. The distinguishing between magnetisations residing in hematite A and B is difficult because their unblocking temperature spectra might be overlapping. It should be noted that proportions of the both hematite generations differ between localities. This refers especially to the WM locality which is situated in the zone affected by postmagmatic fluids causing the pegmatitisation/autometasomathosis of the granite (A. Michalik, 1951). That process was, by all means, younger than the granite itself. The amount of secondary hematite B seems the highest in that locality. Thus it might be possible that observed palaeomagnetic vectors are resultant between two components with overlapping unblocking temperature spectra: one of them lies closer to the mean WM direction while the other is situated in the vicinity of the RZ direction. It is quite remarkable that the sequence of mean palaeomagnetic directions seems to correlate with mineral composition of the granitoid rocks. The inclination value is the lowest at the WM (granite) and highest at the RZ locality (tonalite) with intermediate values at both MO localities (granodiorites). Thus the locality means might reflect different time of remanence acquisition for each petrographical type of the studied granitoid rocks.

Another possibility is that tectonic deformations of the rock structure disturbed the isochronous component in four localities in variable degree. It should be noted that the suggested direction of tectonic deformation (N–S simple shear and lateral shortening) inferred from the AMS ellipsoid (F. Hrouda, S. Kahan, 1991) is concordant with the N–S dispersion of characteristic directions at the sample and locality level (A. M. Hirt *et al.*, 1986). Apparently, there is a lack of correlation between the degree of anisotropy and amount of deviation of palaeomagnetic vector from the mean direction. It should not be surprising, because AMS parameters are influenced also by paramagnetic minerals, especially biotite and chlorite, whose response to tectonic stress can be different than that of magnetic minerals.

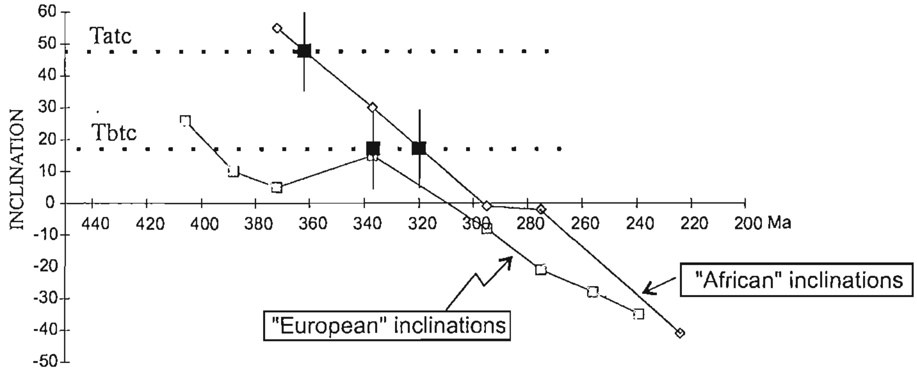

Fig. 10. Reference palaeoinclinations for the Tatra Mts. calculated from the European (squares) and African (diamonds) APWP and inclinations of the *T* component before (*Tbtc*) and after (*Tatc*) tectonic correction

Note the agreement of the *Tbtc* inclination with Middle/Late Carboniferous inclinations of Africa and Europe and *Tatc* inclination with Late Devonian/Early Carboniferous inclinations of Africa; error bars for inclinations of the *T* component are indicated

It is difficult to evaluate which factor is responsible for N–S dispersion of characteristic directions. For convenience a mean direction (labelled *T*) is calculated for all studied localities (Tab. 2) and used for further tectonic interpretations.

Tectonic tilt of the Tatra granite must be considered before final interpretation of palaeomagnetic directions. Tectonic correction applied in this study (azimuth 10°/dip 30°) corresponds to approximate bedding dip and azimuth of autochthonous Lower Triassic sandstones, that overlay directly the crystalline rocks in the Eastern and Western Tatras. Before tectonic correction the palaeomagnetic pole *T* is situated in the intermediate position between the Carboniferous poles of Eurasian and African APWP (Fig. 9a, b). Clockwise and counter-clockwise rotations respectively (around the local vertical axis) are required to match it with one of the reference curves (Fig. 9c). Mean inclination of the component *T* (before tectonic correction) fits the expected Middle/Late Carboniferous palaeoinclinations in both European and African coordinates (Fig. 10). This is in agreement with the Late Carboniferous/Permian palaeogeographical reconstructions (i.e. P. O. Yilmaz *et al.*, 1996) — in those times continental plates tend to converge forming a Pangean supercontinent. Thus a good agreement between the value of palaeoinclination and the radiometric cooling age (330–300 Ma) of the High Tatra granitoid is observed.

If the tectonic correction is applied palaeopole *T* is no more comparable with the Late Palaeozoic segment of European APWP. On the other hand it could be matched after counter-clockwise rotation with the Late Devonian/Early Carboniferous dates of African reference curve (Fig. 9c). It means that inclinations of tilt corrected direction *T* are concordant with expected Late Devonian/Early Carboniferous palaeoinclinations for Western Gondwana (Fig. 10). This interpretation, however, is apparently at variance with the radiometric data. More than 30 Ma gap would appear between the acquisition time of the *T* component and the isotopic cooling age of the High Tatra granite.

The magnetisation of the High Tatra granite is apparently older than the one described from the Sudetic granites (E. Halvorsen *et al.*, 1989). Component *T* before tectonic correction resembles the Late Carboniferous overprint (component A1) that affected some Variscan plutons in Bohemian Massif (J. Reisinger *et al.*, 1994) as well as in Spessart and Odenwald (J.-B. Edel, F. Wickert, 1991). Agreement of the *in situ* inclinations of the component *T* with expected European and African Late Carboniferous inclinations would be a premise that the High Tatra batholith was not tilted in the Alpine orogeny. In this case the models of Neogene rotational uplift of the crystalline core (J. Piotrowski, 1978; M. Bac-Moszaszwili *et al.*, 1984; B. Sperner, 1996) should be revised. On the other hand the relationship between the mean isotopic cooling age and time of acquisition of the *T* component is not clear. The exsolution (670–720°C) and acquisition of the remanence carried by hematite A in hemoilmenites (600–650°C) took place earlier than the unblocking of the Ar-Ar system (*ca.* 300–350°C). According to M. Kohut and M. Janak (1996) the recalculated whole rock Rb/Sr isochron of J. Burchart (1968) indicates age *ca.* 370 Ma. Therefore, it might be postulated that the characteristic component *T* was acquired *ca.* 360 Ma which is 30 Ma earlier than the Late Variscan uplift recorded by Ar-Ar mineral cooling ages (M. Janak, 1994). Radiometric dating and very detailed petrographical investigations performed exactly in the palaeomagnetically sampled localities are desired during further studies. Last but not least uncertainties in the African APWP (R. Van der Voo, 1993) and their bearing on dating of palaeomagnetic components should also be taken into account.

CONCLUSIONS

1. Variscan granites from the High Tatra Mts. reveal very complex palaeomagnetic properties. Characteristic magnetisation resides in the two phase hemoilmenite grains (hematite A) and hematite B of hydrothermal origin.

2. Characteristic magnetisation was acquired in the Late Palaeozoic between 360 and 320 Ma. Its exact age based on palaeoinclination estimations apparently fits the isotopic

cooling age of the Tatra granite (300–330 Ma) when no tilt correction is applied. However, a magnetisation age 360 Ma might be postulated, accepting the model of Neogene tilting and assuming that magnetisation related to hematite A preceded the cooling ages recorded by Ar-Ar method.

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WSTĘPNE WYNIKI BADAŃ PALEOMAGNETYCZNYCH GRANITOIDÓW POLSKIEJ CZĘŚCI TATR WYSOKICH

Streszczenie

Przeprowadzono badania paleomagnetyczne waryscyjskich granitoidów polskiej części Tatr Wysokich. Pobrano 31 próbek ręcznych z następujących 7 stanowisk: próg między kotliną Morskiego Oka i kotliną Czarnego Stawu (stanowiska MO1 i MO2), ściana Świstówki Roztockiej w Dolinie Roztoki (RZ), Wodogrzmoty Mickiewicza (WM), Dolinka za Mnichem (M), rejon Kościelca (K) i próg powyżej Czarnego Stawu w Dolinie Gąsienicowej (HG).

Badane skały mają skład granodiorytów i tonalitów, ze śladami przeobrażeń hydrotermalnych (epidotyzacja, chlorytyzacja). We wszystkich stanowiskach stwierdzono obecność lineacji magnetycznej o przebiegu W–E. Granitoidy tatrzańskie wykazują duże zróżnicowanie własności magnetycznych. Stabilne kierunki paleomagnetyczne wieku późnopaleozoicznego wyróżniono w stanowiskach MO1, MO2, RZ i WM (kierunek średni: $D = 193°$, $I = 17°$, $\alpha_{95} = 12$, $k = 59$, paleobiegun: 4°E, 31°S). Kierunki te są oparte na hematycie, który występuje jako przerosty w dwufazowych ziarnach hematytowo-ilmenitowych, odmieszanych w wysokich temperaturach (670–720°C) oraz jako minerał hydrotermalny. Średnie kierunki z poszczególnych stanowisk różnią się nieco i nie można wykluczyć, że namagnesowanie nie jest jednoczasowe. Wydaje się, że istnieje pewna korelacja między wiekiem namagnesowania a składem mineralnym granitoidu (kierunki starsze występują w granitoidach o składzie tonalitu), problem ten wymaga jednak dalszych badań.

Interpretacja wieku namagnesowania zależy od zastosowania korekcji tektonicznej. Wiek charakterystycznego kierunku namagnesowania, określony na podstawie parametru paleoinklinacji, wykazuje lepszą zgodność z wiekiem izotopowym stygnięcia intruzji tatrzańskiej (330–300 mln lat) bez stosowania korekcji tektonicznej. Paleobiegun leży między środkowo/późnokarbońskimi odcinkami krzywych referencyjnych dla płyty europejskiej i afrykańskiej i, po uwzględnieniu niewielkich rotacji wokół osi pionowej, może być dopasowany do obu krzywych na odcinku 340–320 mln lat. Po zastosowaniu korekcji tektonicznej (określonej na podstawie zapadania warstw dolnego triasu autochtonicznej pokrywy trzonu krystalicznego na jego północnej krawędzi) paleobiegun można dopasować jedynie do krzywej afrykańskiej w punkcie 360 mln lat. Interpretacji takiej nie można jednoznacznie odrzucić, gdyż namagnesowanie związane z przerostami hematytowo-ilmenitowymi mogło utrwalić się znacząco wcześniej niż zamknięcie systemu ^{40}Ar-^{39}Ar, wykazujące wiek stygnięcia intruzji (300–350°C). Jednoznaczna interpretacja wyników badań paleomagnetycznych granitoidu tatrzańskiego nie jest możliwa bez określenia wieku radiometrycznego badanych próbek.

EXPLANATIONS OF PLATE

PLATE I

Hematite/ilmenite exsolution structures in the High Tatric granites, locality RZ

Fig. 1. Exsolution lamellae of hematite (light grey) and ilmenite (grey) in hematite-rich grain; reflected light, 1 nicol; scale bar — 0.05 mm

Fig. 2. Irregular patches of catastrophically coarsened hematite (light grey) intergrowths in ilmenite (grey); reflected light, 1 nicol; scale bar — 0.25 mm

Fig. 3. Exsolved and coarsened hematite (light grey) intergrowths in ilmenite (grey); reflected light, 1 nicol; scale bar — 0.25 mm

Fig. 4. Exsolved and coarsened irregular patches of hematite (light grey) in ilmenite (grey). Near the margin and near the fissure small grains of titanite (dark grey) could be seen; reflected light, 1 nicol; scale bar — 0.25 mm

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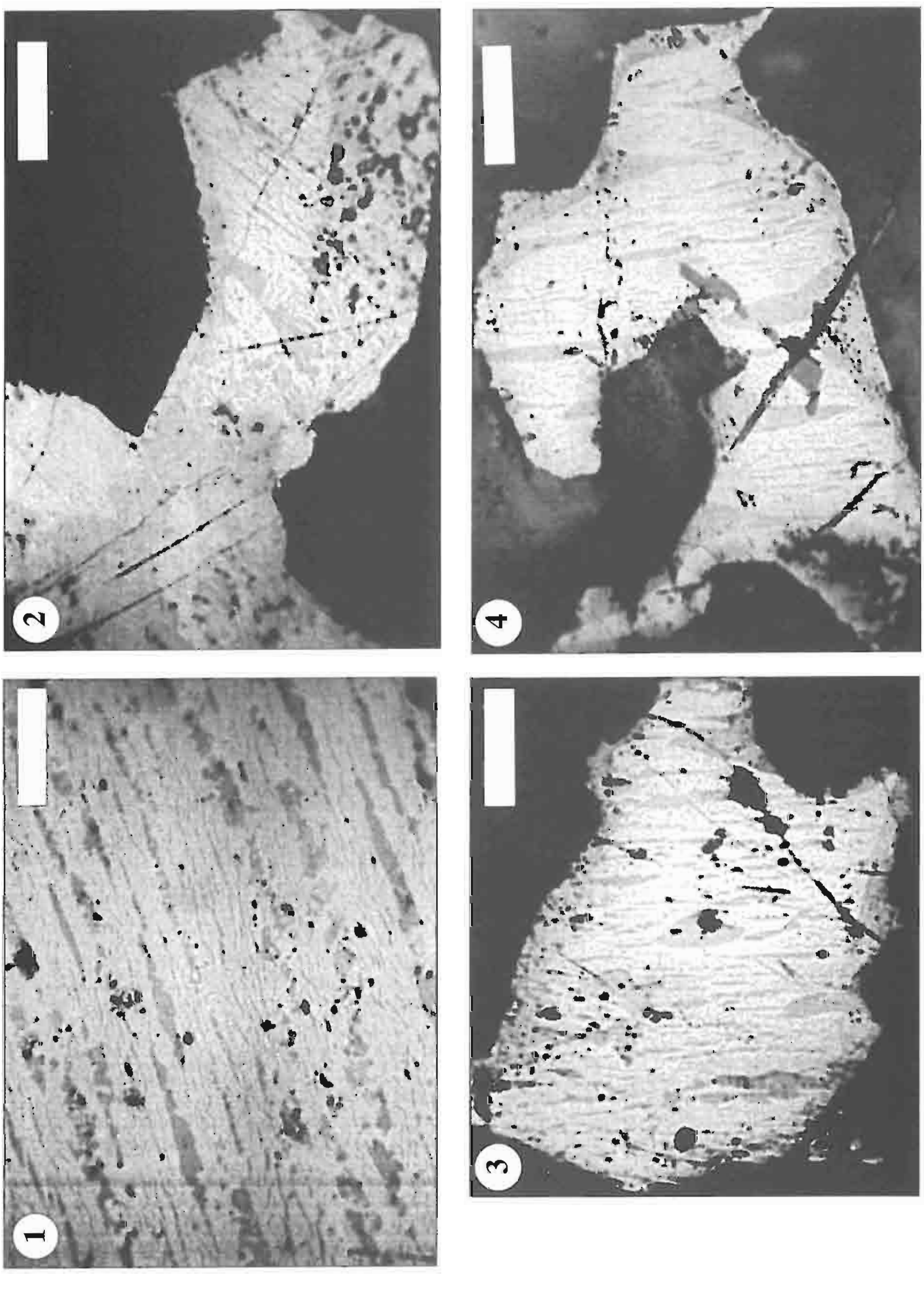

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