

Original paper

Geochemical characteristics of the Late Proterozoic Spitz granodiorite gneiss in the Drosendorf Unit (southern Bohemian Massif, Austria) and implications for regional tectonic interpretations

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The Spitz Gneiss, located near the Danube in the southern sector of the Variscan Bohemian Massif, represents a ~13 km² large Late Proterozoic Bt ± Hbl bearing orthogneiss body in the Lower Austrian Drosendorf Unit (Moldanubian Zone). Its formation age (U–Pb zircon) has been determined previously as 614 ± 10 Ma. Based on 21 new geochemical analyses, the Spitz Gneiss can be described as a granodioritic I-type rock (64–71 wt. % SiO₂) with medium-K composition (1.1–3.2 wt. % K₂O) and elevated Na₂O (4.1–5.6 wt. %). Compared to average granodiorite, the Spitz Gneiss is slightly depleted in Large-Ion Lithophile (LIL) elements (Rb 46–97 ppm, Cs 0.95–1.5 ppm), Sr (248–492 ppm), Nb (6–10 ppm), Th (3–10 ppm), the LREE (e.g. La 10–30 ppm), Y (6–19 ppm) and first row transitional metals (e.g. Cr 10–37 ppm). The Zr content (102–175 ppm) is close to average granodiorite. The major- and trace-element signature of the Spitz Gneiss is similar to some Late Proterozoic granodiorite suites in the Moravo–Silesian Unit (e.g. the Passendorf-Neudegg suite in the Thaya Batholith). However, granodiorites of such type and age do not occur elsewhere in the Moldanubian Zone of the Bohemian Massif. This observation fits existing tectonic models in which the Austrian Drosendorf Unit is considered allochthonous and part of the Moravo–Silesian Unit and the Avalonian Superterrane. Mineral chemistry data for amphibole, plagioclase and biotite allow an estimation of the Variscan peak regional metamorphic conditions for the Spitz Gneiss at ~700 °C and 7 kbar. Amphibole and plagioclase show hardly any signs of retrograde reequilibration, implying a fast late-Variscan exhumation. Partial chloritization of biotite indicates late fluid activity at T ~ 250 °C.

Keywords: Spitz Gneiss, Moldanubian, Bohemian Massif, Moravo–Silesian, Cadomian, Peri-Gondwana terranes

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1. Introduction

The central European Variscan fold belt hosts many remnants of Late Proterozoic to Cambrian (Cadomian) granitoids that once belonged to an Andean-type orogen along the northern Gondwana margin (Dörr et al. 2002; Linnemann et al. 2008). In the Early Paleozoic, the northern fringe of Gondwana underwent a phase of extension and fragmented into a number of terranes which drifted towards Laurentia and Baltica. These so-called peri-Gondwana terranes (Nance et al. 2008) became later incorporated into the Devonian–Carboniferous Variscan fold belt (Winchester and PACE TMR Network Team 2002; Kroner and Romer 2013) and partly underwent strong deformation and metamorphism.

The recognition and delineation of individual Peri-Gondwana terranes within the complex tectonic framework of the Variscan Orogen presents a great challenge (Tait et al. 1997; Finger et al. 2000; Franke and Żelaźniewicz 2000; Friedl et al. 2000; Kalvoda et al.

2008). A potential approach for tackling this problem is a systematic collection and comparison of geochemical and geochronological data from the remnant Cadomian granitoid bodies, in the hope that they are distinctive for individual terranes.

Here we present a geochemical study for the Late Proterozoic Spitz Gneiss in the south-eastern Bohemian Massif. This orthogneiss body has been repeatedly mentioned in the literature in connection with regional tectonic issues. In particular, it has been controversially discussed whether the Spitz Gneiss and surrounding rocks (summarized as the Drosendorf Unit) are a true part of the Moldanubian core region of the Variscan Orogen or rather a tectonically emplaced slice from the Moravo–Silesian foreland plate (Frasl 1970, 1991; Matura 1976, 2003; Finger and Steyrer 1995; Finger et al. 2007b) that collided with the Moldanubian during the Variscan Orogeny (Suess 1926; Fuchs 1976; Dudek 1980).

Based on U–Pb zircon dating, Friedl et al. (2004) have constrained the magmatic formation age of the Spitz

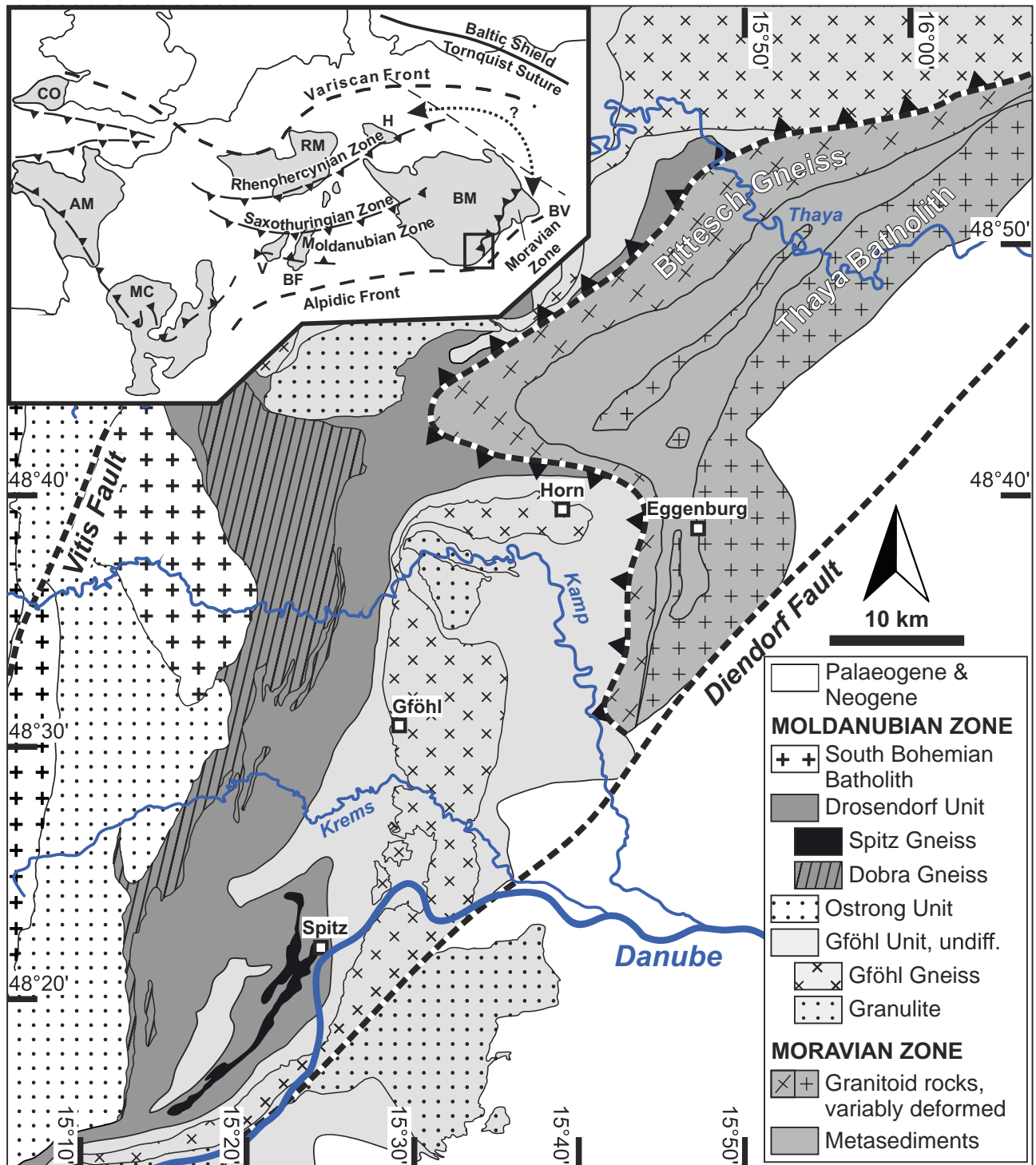


Fig. 1 Simplified geological sketch map of the study area in Lower Austria and its position in the Variscides (inset). Abbreviations: AM – Armorican Massif; BF – Black Forrest; BM – Bohemian Massif; BV – Brunovistulian (mostly covered); CO – Cornwall; H – Harz; MC – Massif Central; RM – Rhenish Massif; V – Vosges. Compiled after Fuchs and Matura (1976), Thiele (1984) and Matura (2003).

Gneiss to 614 ± 10 Ma. Conversely, the Cadomian plutonism in the Moravo–Silesian Unit was long believed to be comparably younger, i.e. ~ 580 Ma old (Finger et al. 2000). This age difference could be used as an argument against a Moravian affiliation of the Spitz Gneiss and the

Drosendorf Unit, respectively. However, recent age dating work in the Brno Massif has shown that the time span of Cadomian granitic activity in the Moravo–Silesian Unit was broader, covering a period from ~ 630 Ma to 570 Ma (Soejono et al. 2017). Comparable geochronological

dates were recently also obtained from the Thaya Batholith (Svojtka et al. 2017; W. Dörr, pers. comm.). The age of the Spitz Gneiss (614 ± 10 Ma) is, therefore, not at all in conflict with a Moravo–Silesian affiliation.

Only scarce petrographic and geochemical data have been available from the Spitz Gneiss so far (Fuchs and Matura 1976; Finger 1990), and any comparative studies with Moravo–Silesian granitoids were lacking. By introducing 21 new geochemical analyses, we are now able to comprehensively characterize the Spitz Gneiss and to compare it with Cadomian granitoids from the Moravo–Silesian Unit. Furthermore, we present a set of mineral analyses for the Spitz Gneiss, which can be used for estimating the Variscan metamorphic conditions.

2. Regional geological background

The Bohemian Massif is one of the major outcrops of the Variscan Orogen (Franke 1989, 2000; Kroner and Romer 2013). The southern half of the massif exposes the Moldanubian Zone (Kossmat 1927), the high-grade metamorphic core of the Variscan Orogen (Fig. 1). This zone represents a complex tectonic *mélange* (Matte et al. 1990; Franke and Żelaźniewicz 2000; Finger et al. 2007b; Lardeaux et al. 2014; Žák et al. 2014). In Austria, the Moldanubian has been subdivided into three tectonostratigraphic units that were juxtaposed during the Variscan Orogeny: the Gföhl, Ostrong and Drosendorf units.

The Gföhl Unit (Fig. 1) is mainly represented by felsic, variably migmatitic orthogneiss (Gföhl Gneiss), associated with paragneiss, amphibolite, felsic granulite, and locally eclogite and peridotite (Fuchs and Matura 1976; Thiele 1984; Fritz 1995; Schulmann et al. 2005; Racek et al. 2006; Hasalová et al. 2008). After a stage of subduction-related UHP and HP–HT metamorphism (O’Brien and Carswell 1993; Cooke et al. 2000; Faryad et al. 2011), the Gföhl Unit became steeply exhumed during the Variscan Orogeny (Schulmann et al. 2005).

The Ostrong Unit (Fig. 1) consists mainly of a LP–HT metamorphic monotonous series of cordierite-bearing paragneiss (Fuchs 1976; Linner 1992, 1996), likely derived from Early Paleozoic greywacke (Finger et al. 2007a; Košler et al. 2014). There are some intercalations of orthogneiss (e.g. Blaník Gneiss – Breiter et al. 2005; René and Finger 2016) and small exotic relics of eclogite which experienced severe recrystallization during the LP–HT event (O’Brien and Vrána 1995; Scott et al. 2013). It is likely that the whole Ostrong Unit is polymetamorphic (Linner 1992, 1996), but the strong Variscan LP–HT overprint has erased most of the preexisting mineral assemblages.

The Drosendorf Unit (Fig. 1) comprises a variegated series of metamorphic rocks including paragneiss, mar-

ble, quartzite, graphite schist, amphibolite, calc-silicate and orthogneiss (Fuchs and Matura 1976). All these rocks experienced Variscan MP–MT metamorphism followed by a decompression-related LP overprint (Högelsberger 1989; Petrakakis 1997). The two major orthogneiss bodies of the Drosendorf Unit are the Spitz Gneiss (614 ± 10 Ma) and the Dobra Gneiss. For the latter, a Mesoproterozoic formation age (1377 ± 10 Ma) has been postulated based on zircon dating (Gebauer and Friedl 1994). Note that we use the term Drosendorf Unit here strictly in its original sense as defined by Austrian workers, i.e., only for a distinct rock assembly in the Lower Austrian part of the Bohemian Massif.

At the end of the Variscan Orogeny, the Moldanubian Zone was intruded by the 330 to 300 Ma old granitoids of the South Bohemian Batholith (Finger et al. 1997; Gerdes et al. 2003; Breiter and Scharbert 2006; Breiter 2010; Žák et al. 2014 and references therein).

The easterly Moravian Zone (Suess 1926) was overthrust by the Moldanubian nappes in the Carboniferous (Fritz and Neubauer 1993). It is generally interpreted as the Variscan-reactivated margin of the large Moravo–Silesian or Brunovistulian foreland plate (Dudek 1980) that forms the eastern end of the Variscides (Finger et al. 2000; Friedl et al. 2000). The Moravo–Silesian Unit exposes many Late Proterozoic granitoid rocks in the Thaya Batholith and the Brno Batholith, including granitic, granodioritic and tonalitic varieties (Finger et al. 1989, 1995; Hanžl and Melichar 1997; Leichmann and Höck 2008). Orthogneisses in the higher grade western part of the Moravian Zone (e.g., Bittesch Gneiss) have Late Proterozoic formation ages as well (Friedl et al. 2004; Soejono et al. 2017). The Moravo–Silesian granitoids are generally characterized as I-type and volcanic-arc type granites (Finger et al. 1989; Jelínek and Dudek 1993; Hanžl and Melichar 1997).

3. Methodology

3.1. Mineral chemistry

Two polished thin sections, one from a more mafic, amphibole-bearing type of Spitz Gneiss (ML 14-10), and a second from a more felsic amphibole-free type (ML 14-1B), were studied by scanning electron microscopy (SEM) and energy dispersive X-ray spectrometry (EDX) at the Department of Chemistry and Physics of Materials, University of Salzburg. The used SEM (ZEISS ULTRAPLUS) is equipped with an Oxford X-MAX 50 EDX detector plus INCA software. Mineral analyses were carried out at 15 kV and 3 nA and calibrated against natural mineral standards (quartz, garnet) and synthetic oxides.

Tab. 1 XRF analyses for the GSJ granodiorite standard JG-1a (Imai et al. 1995), including information on typical analytical errors and detection limits for individual elements

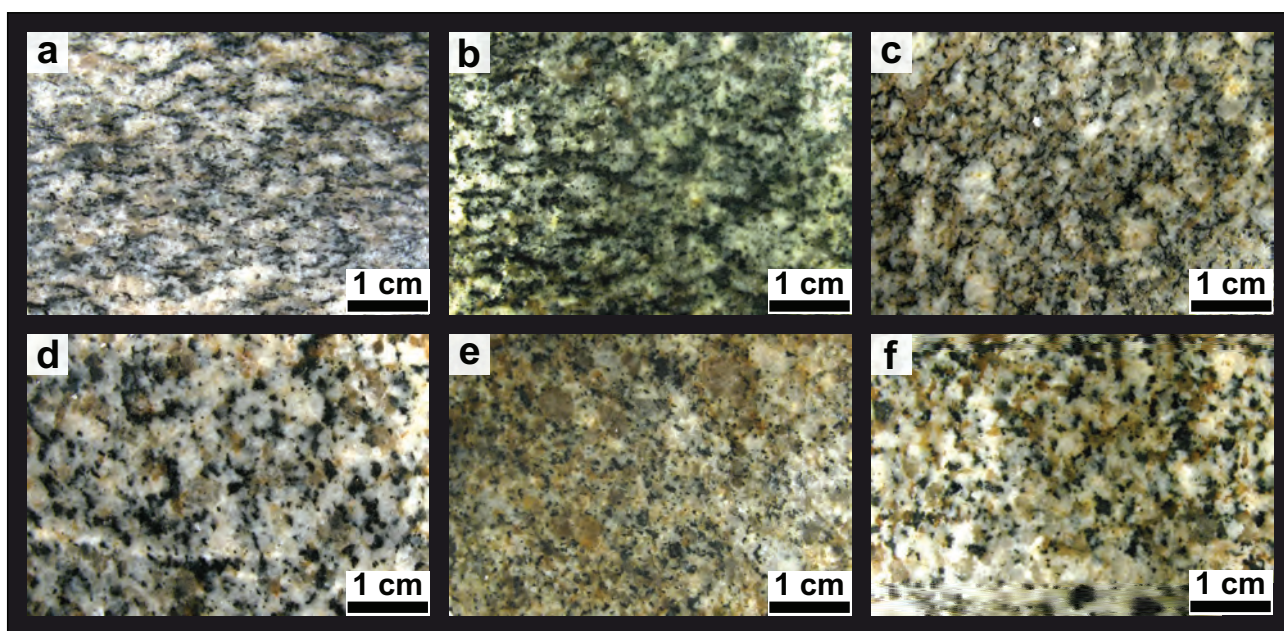
	Recommended	Measured	LLD [ppm]
SiO ₂	72.30 ± 0.51	72.40 ± 0.50	94.3
TiO ₂	0.25 ± 0.03	0.26 ± 0.01	10.0
Al ₂ O ₃	14.30 ± 0.41	14.11 ± 0.20	75.2
Fe ₂ O _{3t}	2.00 ± 0.10	1.97 ± 0.05	71.9
MgO	0.69 ± 0.07	0.72 ± 0.05	66.6
MnO	0.06 ± 0.01	0.06 ± 0.01	3.9
CaO	2.13 ± 0.08	2.15 ± 0.05	37.7
Na ₂ O	3.39 ± 0.13	3.54 ± 0.10	43.2
K ₂ O	3.96 ± 0.16	3.98 ± 0.05	28.2
P ₂ O ₅	0.08 ± 0.01	0.10 ± 0.01	24.9
H ₂ O ⁺	0.59 ± 0.13		
Sum	99.75	99.28	
Ba	470 ± 38	456 ± 10	16.5
Co	5.90 ± 1.54	5 ± 1	2.2
Cr	17.6 ± 4.4	17 ± 1	3.7
Ga	16.5 ± 0.7	15 ± 1	1.1
Gd	4.08 ± 0.63	6 ± 2	1.2
Nb	11.40 ± 0.98	13 ± 1	1.7
Ni	6.91 ± 1.90	12 ± 1	1.6
Pb	26.4 ± 2.8	22 ± 3	6.0
Rb	178 ± 10	176 ± 1	2.0
Sc	6.21 ± 0.49	6 ± 2	5.5
Sn	4.47 ± 0.56	7 ± 2	4.0
Sr	187 ± 12	178 ± 1	1.8
Th	12.8 ± 1.5	12 ± 1	2.9
V	22.7 ± 3.8	23 ± 2	6.5
Y	32.1 ± 3.8	32 ± 1	3.0
Zn	36.5 ± 2.2	39 ± 1	1.2
Zr	118 ± 13	119 ± 2	4.3

LLD = lower limit of detection

3.2. Whole-rock geochemistry

Large (2–3 kg), fresh and representative samples from 12 outcrops were collected. Weathered surfaces were removed and samples were ground to a fine powder in an agate ball mill. Twenty-one analyses were conducted by XRF methods on lithium tetraborate glass beads and pressed powder pellets using a Bruker Pioneer S4 crystal spectrometer at the Department for Chemistry and Physics of Materials, University of Salzburg. Obtained net count rates on single X-ray lines were recast into concentration data (wt. % and ppm), based on an in-house calibration routine that involves measurements of ~ 30 international geostandards (USGS and GSJ). The calibration is based on the Bruker AXS software SPECTRA^{plus} FQUANT (v1.7) and corrects absorption, fluorescence and line overlap effects. In addition, a monitor standard (GSJ Granodiorite JG-1a) was measured together with the samples. Results are given in Tab. 1 and include information on detection limits and typical analytical uncertainties for single elements. Reported errors are conservative and refer not only to the XRF counting statistics, but consider also the uncertainty of the linear fit of the calibrations. Loss on ignition (LOI) was determined gravimetrically after heating the dried powders to 1100 °C for three hours.

For four representative samples, rare earth elements, Cs, Hf, Ta, Th, and U contents were determined by ICP-MS techniques at ALS Laboratories in Laughrea, Ireland. Analyses were carried out on acid digested lithium borate fusion beads. Lower limits are between 0.01 ppm and 0.1 ppm.

**Fig. 2** Sand down surfaces of a representative selection of samples showing the macroscopic variability of the Spitz Gneiss. **a** – Sample ML 16-6, Eitental. **b** – ML 16-20, Gießhübel. **c** – ML 14-10, Eitental. **d** – ML 16-17, Gießhübel. **e** – ML 14-13, Eitental. **f** – ML 16-16, Gießhübel.

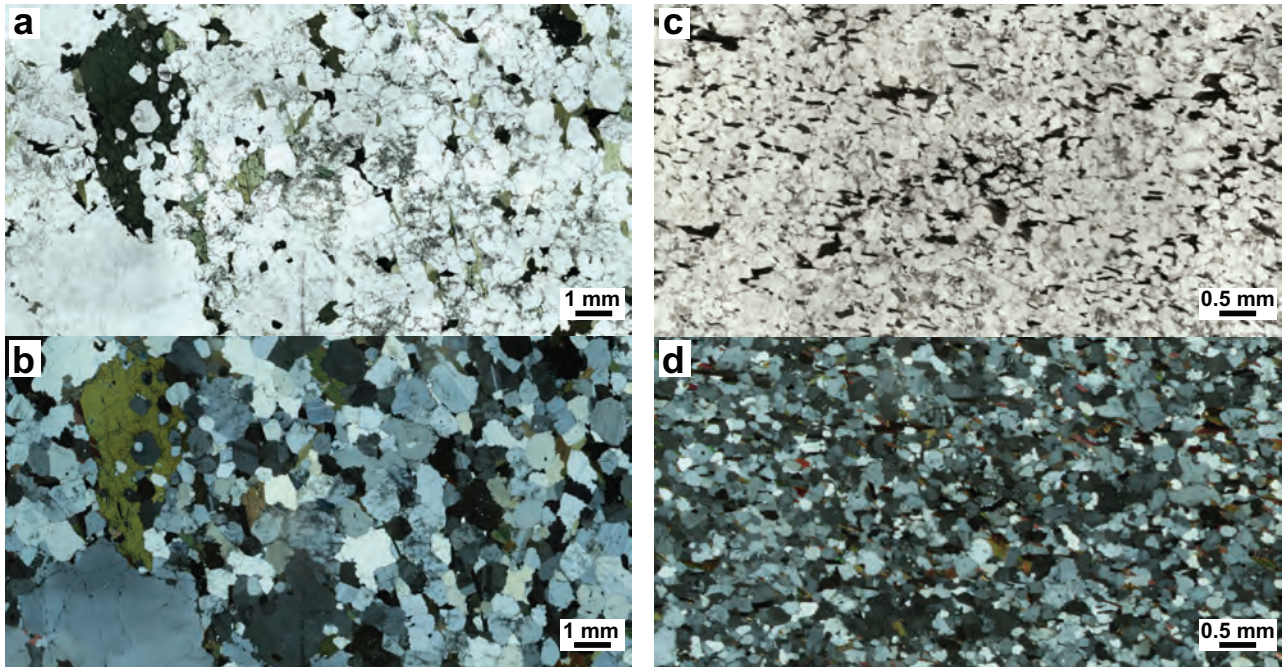


Fig. 3 Thin-section photographs of a medium-grained, amphibole-bearing variety (a, b) and a finer grained, biotite-bearing variety (c, d) of Spitz Gneiss. Plane- (a, c) and cross-polarized light (b, d).

4. Petrography of the Spitz Gneiss

Figure 2 gives an overview of the macroscopic variability of the Spitz Gneiss. The rock has a fine to medium grain size. The fabric and the minerals are metamorphic with the exception of some diffuse textural relics of magmatic feldspar phenocrysts. The foliation is weak to moderate and mainly defined by oriented biotite, more rarely by elongated feldspar or quartz (Fig. 3). The quartz–feldspar fabric is typically granular. Some variants of Spitz Gneiss carry amphibole (Fig. 3a).

Biotite is partly chloritized with fine exsolution lamellae of ilmenite having formed in the chlorite domains. Amphibole, plagioclase and K-feldspar appear homogenous under the optical microscope and in backscattered-electron (BSE) imagery. Small retrograde epidote and muscovite rarely occur along grain boundaries. Accessory minerals are zircon, apatite and allanite.

5. Mineral chemistry and geothermobarometry

5.1. Mineral compositions

Representative analyses are given in Tab. 2. The anorthite content of *plagioclase* ranges between An_{23} and An_{31} in the amphibole-bearing thin section and from An_{19} to An_{25}

in the more felsic sample. Single plagioclase crystals are compositionally widely homogenous. *K-feldspar* is Na-poor (Ab_{5-10}) and shows BaO contents of ~0.6 to 0.8 wt. %; CaO remained below the detection limit (0.1 wt. %).

Amphibole compositions are uniformly calcic with CaO contents around 11 wt. % and moderate alkali-element concentrations (Na_2O 1.1–1.6 wt. %; K_2O 1.3–1.9 wt. %). Having $Mg/(Mg+Fe^{2+})$ below 0.5 and Si (*apfu*) around 6.5, the amphiboles can be classified as ferroedenite to ferropargasite (Fig. 4b, 5a).

Biotite (Fig. 4a–d, 5b) from the more felsic sample shows $Mg/(Mg + Fe)$ ratios of ~0.5, those in the amphibole-bearing sample ~0.45. The Al^{IV} content (*apfu*) is relatively low (2.0–2.6) in both cases.

The secondary *chlorite* (Fig. 4d) classifies as chamosite (Fig. 5c), with $Mg/(Mg + Fe)$ of 0.49 in the amphibole-free sample and 0.43 in the amphibole-bearing sample. Si (*apfu*) is around 5.65 in the amphibole-free sample and between 5.9 and 6.8 in the amphibole-bearing one. Table 2 includes also an analysis of a rare secondary epidote (Fig. 4b) and a secondary muscovite.

5.2. Mineral thermobarometry

The mineral paragenesis amphibole plus plagioclase (+ Kfs, Qz) is suitable for various geothermobarometric calculations. Using the hornblende–plagioclase geothermometer introduced by Holland and Blundy (1994), in connection with the Al-in-amphibole geobarometer of

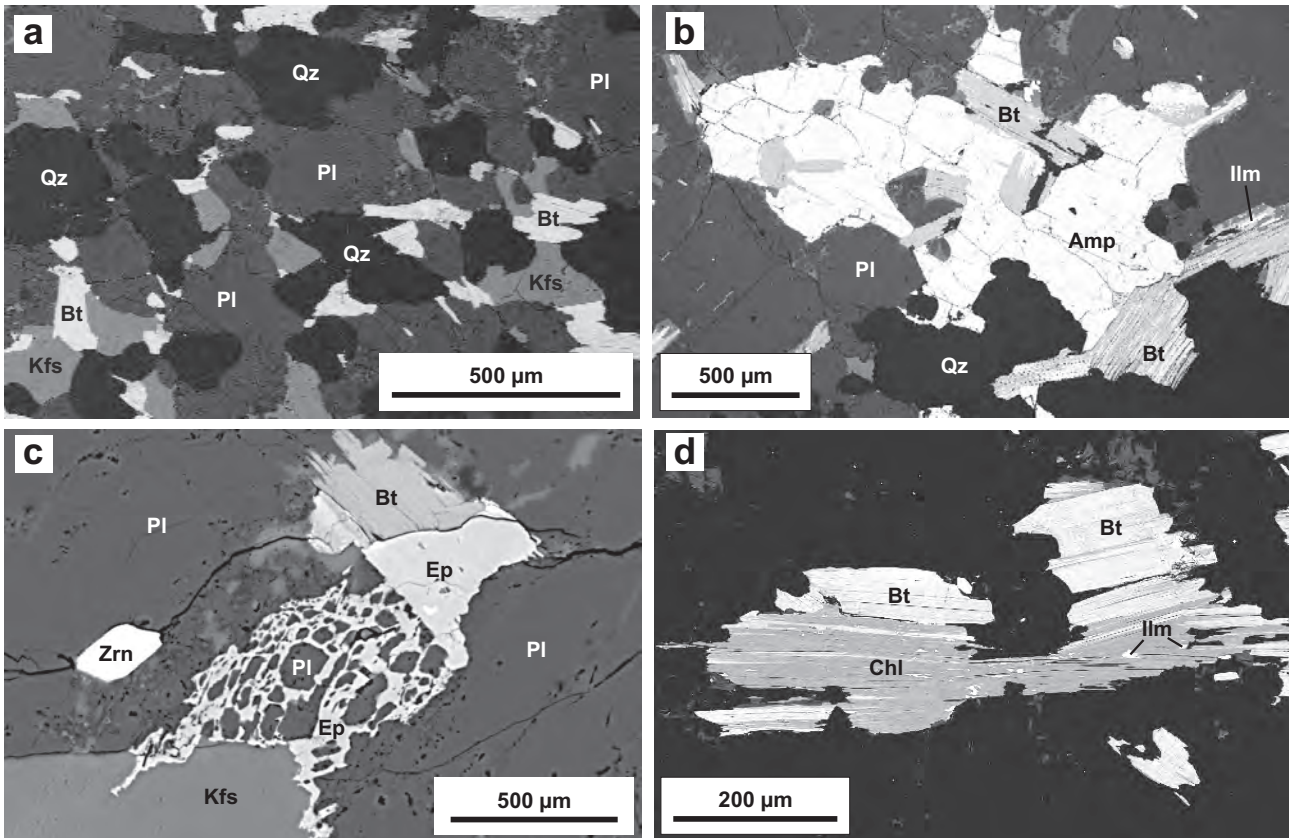


Fig. 4 Scanning electron microscope images (BSE, from thin sections ML 14-1B and ML 14-10. **a** – Typical metamorphic fabric of the Spitz Gneiss with plagioclase, quartz, biotite and K-feldspar (ML 14-1B). **b** – Large amphibole grain with biotite inclusions intergrown with plagioclase; ilmenite exsolution lamellae in chloritized biotite. **c** – Rare example of secondary epidote. **d** – Chloritized biotite with ilmenite exsolution in chlorite. Mineral abbreviations according to Whitney and Evans (2010).

Anderson and Smith (1995), an average temperature of $698 \pm 10^\circ\text{C}$ (2σ) at an average pressure of 6.51 ± 0.25 kbar can be deduced. The application of the empirical Ti-in-

amphibole thermometer of Otten (1984) yields almost the same temperature estimate of 697°C . The Ti-in-biotite thermometer of Henry et al. (2005) gives an average

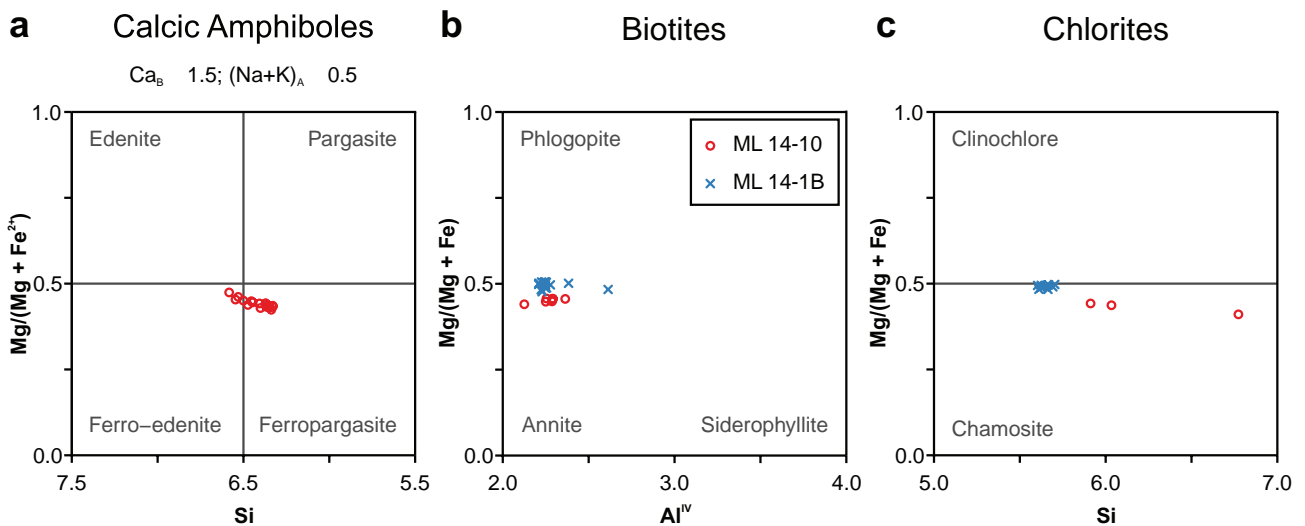


Fig. 5 Mineral chemistry classification diagrams (*apfu*). **a** – Amphibole classification diagram for calcic amphiboles ($\text{Ca}_B \geq 1.5$; $(\text{Na} + \text{K})_A \geq 0.5$) after Leake et al. (1997). Fe^{2+} calculated from FeO_{tot} after Holland and Blundy (1994). **b** – Biotite classification diagram (Rieder et al. 1999). Total iron calculated as Fe^{2+} . **c** – Chlorite classification diagram (Bayliss 1975). Total iron calculated as Fe^{2+} .

Tab. 2 Representative mineral analyses for Spitz Gneiss samples ML 14-1B and ML 14-10

Sample	ML 14-10		ML 14-1B		ML 14-1B		ML 14-10		Sample	ML 14-10		ML 14-1B		Kfs		
	Mineral	Amp	Amp	Bt	Bt	Ms	Chl	Chl		Mineral	Ep	Pl	Pl		Pl	Kfs
SiO ₂	42.17	42.74	41.76	38.37	36.84	38.01	47.96	27.88	27.01	37.44	60.71	61.48	64.85	61.47	65.19	65.02
TiO ₂	1.02	1.09	1.13	3.36	2.97	2.85	0.59	0.28	0.13	0.64	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.
Al ₂ O ₃	12.55	11.36	12.27	14.77	14.38	15.57	30.00	21.86	20.90	22.45	24.27	23.60	22.84	22.98	17.94	18.08
FeO _{tot}	20.15	20.54	20.38	21.78	18.67	19.32	4.60	24.88	25.89	12.90	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.
MgO	7.34	7.46	7.26	9.94	10.32	10.70	3.23	13.82	13.58	0.22	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.
MnO	0.66	0.68	0.70	0.46	0.34	0.27	b.d.l.	0.30	0.47	0.31	n.a.	n.a.	n.a.	n.a.	n.a.	n.a.
Na ₂ O	1.55	1.31	1.61	0.23	0.39	0.33	0.19	b.d.l.	b.d.l.	b.d.l.	7.95	8.22	9.13	8.78	0.91	0.61
K ₂ O	1.70	1.57	1.87	9.75	8.89	8.57	10.56	0.41	0.09	0.11	0.39	0.67	0.37	0.31	14.76	15.01
CaO	10.84	11.16	10.97	b.d.l.	b.d.l.	0.15	b.d.l.	b.d.l.	b.d.l.	22.67	6.59	5.22	4.16	4.98	0.00	0.00
BaO	b.d.l.	b.d.l.	b.d.l.	b.d.l.	b.d.l.	b.d.l.	b.d.l.	b.d.l.	b.d.l.	b.d.l.	0.13	b.d.l.	b.d.l.	b.d.l.	0.73	0.71
Total	98.00	97.91	97.96	98.64	92.80	95.76	97.13	89.44	88.08	96.74	100.04	99.20	101.36	98.52	99.53	99.44
Number of ions on the basis of:																
Si	23 O	23 O	23 O	22 O	22 O	22 O	22 O	28 O	28 O	12.5 O	8 O	8 O	8 O	8 O	8 O	8 O
Al ^{IV}	6.38	6.48	6.35	5.71	5.75	5.72	6.38	2.85	2.83	3.02	2.71	2.75	2.83	2.77	3.02	3.02
Al ^{VI}	1.62	1.52	1.65	2.29	2.25	2.28	1.62	1.15	1.17	3.02	1.28	1.25	1.17	1.22	0.98	0.99
Sum	8.00	8.00	8.00	8.00	8.00	8.00	8.00	4.00	4.00	6.04	3.98	4.00	4.00	3.99	4.00	4.00
Al ^{VI}	0.62	0.51	0.55	0.30	0.40	0.49	3.08	1.49	1.41	2.13	0.32	0.25	0.19	0.24	0.00	0.00
Ti	0.11	0.12	0.13	0.38	0.35	0.32	0.06			0.78	0.69	0.71	0.77	0.77	0.08	0.05
Fe ^{3+*}	0.43	0.41	0.39							2.91	0.02	0.04	0.02	0.02	0.87	0.89
Mg	1.66	1.69	1.65	2.21	2.40	2.40	0.64	2.11	2.12	1.93	0.00	0.00	0.00	0.00	0.01	0.01
Mn	0.08	0.09	0.09	0.06	0.04	0.03	0.00	0.03	0.04	0.04	0.00	0.00	0.00	0.00	0.01	0.01
Fe ^{2+*}	2.10	2.18	2.19	2.71	2.44	2.43	0.51	2.13	2.27	0.01	1.03	1.00	0.99	1.02	0.97	0.96
M1,2,3	5.00	5.00	5.00	5.66	5.63	5.68	4.29	4.29	4.29	0.01	30.74	24.99	19.68	23.44	0.00	0.00
Ca	1.76	1.81	1.79	0.00	0.00	0.02	0.00			0.03	67.10	71.18	78.23	74.84	8.60	5.81
Na	0.22	0.17	0.20	0.07	0.12	0.10	0.05			0.02	2.16	3.83	2.09	1.72	91.40	94.19
Fe	0.02	0.02	0.01	1.85	1.77	1.65	1.79	0.05	0.01	2.03						
M4	2.00	2.00	2.00	1.92	1.89	1.77	1.84	10.09	10.15							
Na	0.23	0.21	0.28													
K	0.33	0.31	0.36													
A	0.56	0.52	0.64													
1-A	0.44	0.48	0.36													

* Fe²⁺ and Fe³⁺ calculated from FeO_{tot} after Holland and Blundy (1994)
 Mineral abbreviations according to Whitney and Evans (2010)

temperature of 676 and 678 °C for the two investigated samples. Al-in-amphibole barometry results in 7.11 ± 0.52 kbar (Schmidt 1992) or 6.78 ± 0.47 kbar (Anderson 1996).

Measured chlorite compositions point to a low chlorite formation temperature, for instance 225–240 °C using the chlorite thermometer of Xie et al. (1997) that is based on the temperature dependency of Al^{IV} (Cathelineau 1988) and the linear relationship between Al^{IV} and $Fe/(Fe + Mg)$. Average temperatures of 211 °C at 1 kbar, 221 °C at 2 kbar, or 236 °C at 3 kbar resulted from the model of Lanari et al. (2014) that is based on the Fe–Mg,

Tschermak and di-trioctahedral substitution, assuming all measured Fe is Fe^{2+} .

6. Geochemistry of the Spitz Gneiss

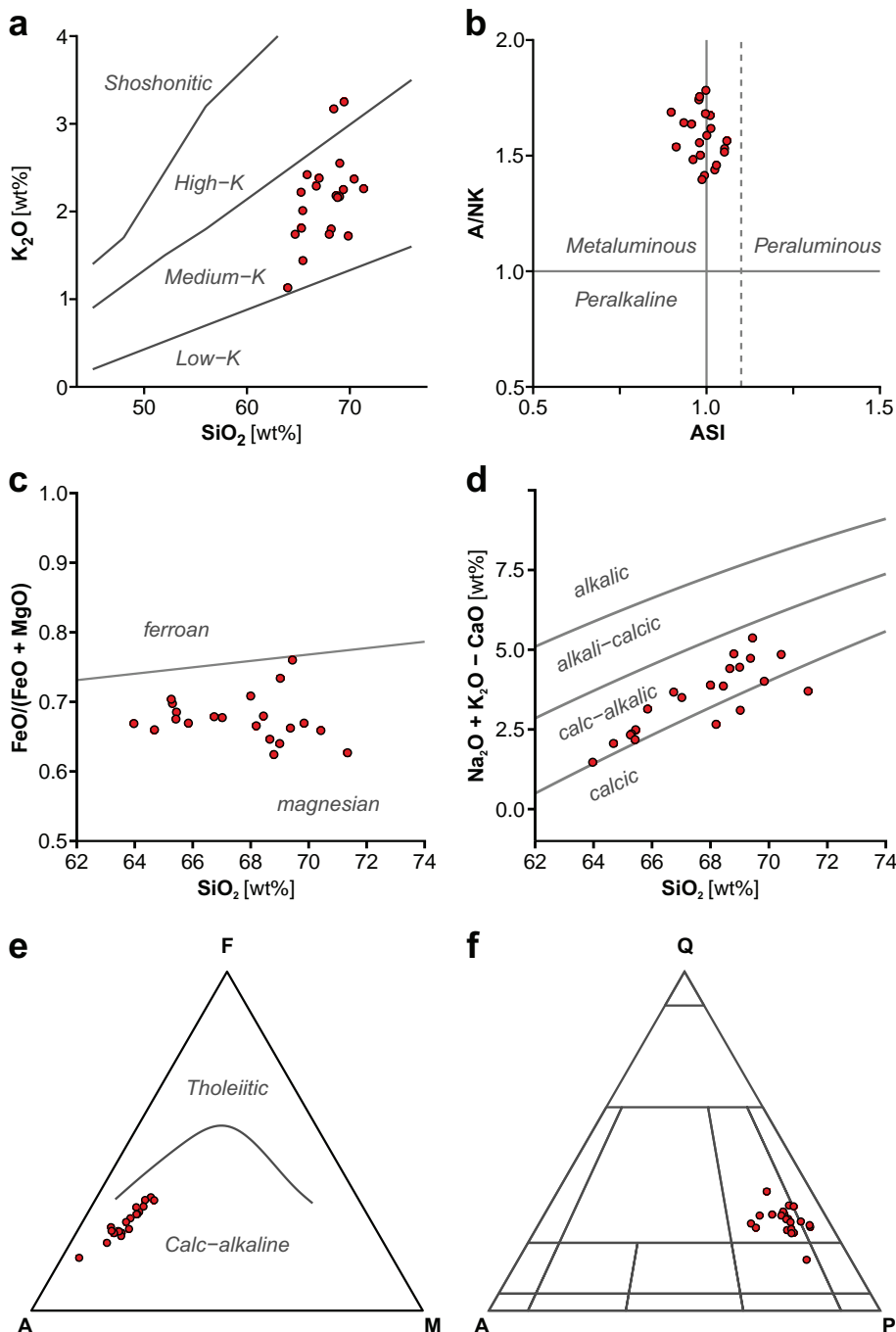
6.1. Major-element data

The Spitz Gneiss covers a SiO_2 range from ~64 to ~71 wt. % (Fig. 6, Tab. 3) and it thus represents an intermediate to moderately acidic rock. Amphibole-bearing samples

have commonly less than 68 wt. % SiO_2 , amphibole-free samples more than 68 wt. % SiO_2 . Many major elements (e.g. FeO , MgO , TiO_2 , CaO , Al_2O_3 , P_2O_5) show a negative covariation with SiO_2 ; exceptions are K_2O (rising with increasing SiO_2) and Na_2O (no clear trend). The K_2O/Na_2O ratios are typically 0.3–0.8. Based on the moderate K_2O content (Fig. 6a), the Spitz Gneiss can be classified as “medium-K” (Peccerillo and Taylor 1976); only a few samples are “high-K”.

Applying the classification schemes of Frost et al. (2001), the Spitz Gneiss is magnesian, and subaluminous, i.e. with an aluminium saturation index (ASI) around 1 (Fig. 6b–c). Amphibole-bearing samples are in general slightly metaluminous, amphibole-free samples

Fig. 6 Various major-element-based classification diagrams for igneous rocks with data from the Spitz Gneiss. **a** – SiO_2 vs. K_2O plot of Peccerillo and Taylor (1976). **b** – ASI vs. A/NK plot of Frost et al. (2001) showing the metaluminous to slightly peraluminous nature. **c** – SiO_2 vs. $FeO/(FeO_i + MgO)$ plot of Frost et al. (2001) illustrating the magnesian character. **d** – SiO_2 vs. $Na_2O + K_2O - CaO$ diagram with fields for alkali, alkali-calcic, calc-alkalic and calcic associations after Frost et al. (2001). **e** – AFM diagram (A = $Na_2O + K_2O$, F = FeO_i , M = MgO) after Irvine and Baragar (1971) displaying the calc-alkaline trend. **f** – QAP-plot after Streckeisen (1974) from the calculated mesonormative compositions (Mielke and Winkler 1979). Most samples fall within the granodiorite field.



Tab. 3 Representative whole-rock major- (wt. %) and trace-element (ppm) compositions for Spitz Gneiss samples

Sample		a	b	c	d	e	f	g	h	
		Amphibole-bearing				Amphibole-free				
		ML 16-28	ML 16-33	ML 14-8	ML 16-6	ML 14-10	ML 16-27	ML 14-1A	ML 16-16	
Coordinates	N	48.379555	48.401523	48.359276	48.258377	48.262578	48.358024	48.359179	48.316165	
	E	15.350658	15.389478	15.365541	15.275041	15.275880	15.384578	15.365783	15.360836	
X-ray fluorescence analysis (University of Salzburg, Austria)	SiO ₂	63.97	64.68	65.30	66.74	67.02	68.00	68.66	68.44	
	Al ₂ O ₃	17.78	16.99	16.51	16.73	16.62	16.55	16.40	16.23	
	MnO	0.05	0.06	0.07	0.07	0.07	0.05	0.04	0.04	
	MgO	1.46	1.87	1.57	1.30	1.35	0.95	1.25	1.05	
	CaO	5.32	4.42	4.37	3.65	3.56	3.49	2.85	3.44	
	Na ₂ O	5.66	4.74	4.92	5.03	4.68	5.64	5.08	4.13	
	K ₂ O	1.13	1.74	1.81	2.29	2.38	1.74	2.18	3.17	
	TiO ₂	0.54	0.56	0.54	0.42	0.41	0.33	0.34	0.36	
	P ₂ O ₅	0.17	0.16	0.14	0.13	0.14	0.12	0.13	0.11	
	Fe ₂ O ₃	3.64	4.48	4.47	3.39	3.50	2.85	2.82	2.75	
	SO ₃	0.02	0.01	0.03	0.02	0.02	0.01	0.01	0.01	
	F	0.12	0.14	0.13	0.11	0.12	0.11	0.11	0.09	
	LOI	0.87	0.62	0.64	0.52	1.01	0.73	1.12	1.00	
	Sum	100.73	100.47	100.50	100.40	100.88	100.57	100.99	100.82	
	X-ray fluorescence analysis (University of Salzburg, Austria)	Rb	22	66	62	73	79	46	87	97
		Sr	492	417	358	269	248	422	331	358
		Ba	396	354	455	418	422	541	426	912
		Ga	22	20	18	19	19	19	20	17
		Nb	9	10	9	7	7	8	7	7
		Zr	172	166	136	125	138	133	119	131
Y		11	12	16	13	19	11	11	7	
Sc		11	8	11	7	8	7	bdl	bdl	
Pb		8	8	7	6	11	10	7	bdl	
Zn		34	67	58	61	67	69	29	46	
V		43	45	68	47	43	28	49	22	
Co		4	8	7	3	3	4	4	4	
Cr		23	37	18	22	16	17	30	20	
Ni		9	13	8	6	5	6	9	9	
ICP-MS (ALS Laboratories, Galway, Ireland)		La		10.3	19.1		22.1		10.6	
		Ce		44.4	45.6		45.9		25.1	
	Pr		2.79	4.85		5.33		2.56		
	Nd		10.0	17.1		18.7		8.2		
	Sm		2.43	3.57		3.54		1.57		
	Eu		1.03	0.99		1.03		0.74		
	Gd		2.57	3.09		3.38		1.54		
	Tb		0.42	0.49		0.51		0.28		
	Dy		2.41	2.86		3.29		1.69		
	Ho		0.47	0.68		0.69		0.38		
	Er		1.20	1.66		2.01		1.01		
	Tm		0.15	0.23		0.31		0.16		
	Yb		1.02	1.61		1.97		1.14		
	Lu		0.18	0.24		0.34		0.20		
	Hf		4.5	4.2		4.0		3.3		
	U		0.93	2.61		1.38		1.81		
Th		7.89	7.03		6.76		5.85			
Cs		1.47	0.95		1.45		1.14			
Ta		0.4	0.6		0.5		0.3			

a–d – amphibole-bearing samples; e–h – amphibole-free samples
 bdl = below detection limit, LOI = loss on ignition

slightly peraluminous. A/NK ratios range between 1.36 and 1.63 (Fig. 6b). The $\text{Na}_2\text{O} + \text{K}_2\text{O} - \text{CaO}$ values are mainly in the calc-alkalic range *sensu* Frost et al. (2001) (Fig. 6d). In the AFM diagram after Irvine and Baragar (1971), the Spitz Gneiss follows the calc-alkaline differentiation trend (Fig. 6f). All these geochemical parameters are characteristic of I-type granitoids (Chappell and White 1974).

Primary modal compositions, calculated using the Mesonorm model (Mielke and Winkler 1979; implemented in the *GCDkit* by Janoušek et al. 2006), indicate strong

plagioclase dominance. The Spitz Gneiss protolith can thus be described as quartz-poor granodiorite, transitional to tonalite (Fig. 6e). The primary biotite contents are estimated between 6 and 15 % and a small amount of normative amphibole is present (up to 6 %) in the low SiO_2 samples, consistent with microscopic observations.

6.2. Trace-element data

Some trace elements like Zr or Nb show a negative correlation with SiO_2 (Fig. 7a–b). The Zr content falls from ~170 ppm to ~100 ppm, documenting zircon saturation of the magma and early zircon precipitation during magmatic crystallization (Chappell et al. 1998; Miller et al. 2003).

Strontium, although generally a compatible element in granitoids, lacks a clear negative correlation trend with SiO_2 . A possibly bimodal data distribution in a SiO_2 vs Sr diagram (Fig. 7c) may indicate that the more mafic and the more felsic samples of Spitz Gneiss had different parental magmas or even sources. On the other hand, Sr is potentially mobile during metamorphism and the original magmatic covariations could be disturbed.

Barium shows concentrations consistently around 400 ppm in the less felsic samples, but scatters widely between 300 and 800 ppm in the more felsic samples (Fig. 7d). Rubidium increases with the SiO_2 content, and shows a good correlation with K_2O (Fig. 7e), as is often observed in igneous systems.

In the trace-element discrimination diagram of Pearce et al. (1984), the Spitz Gneiss falls within the field of volcanic-arc

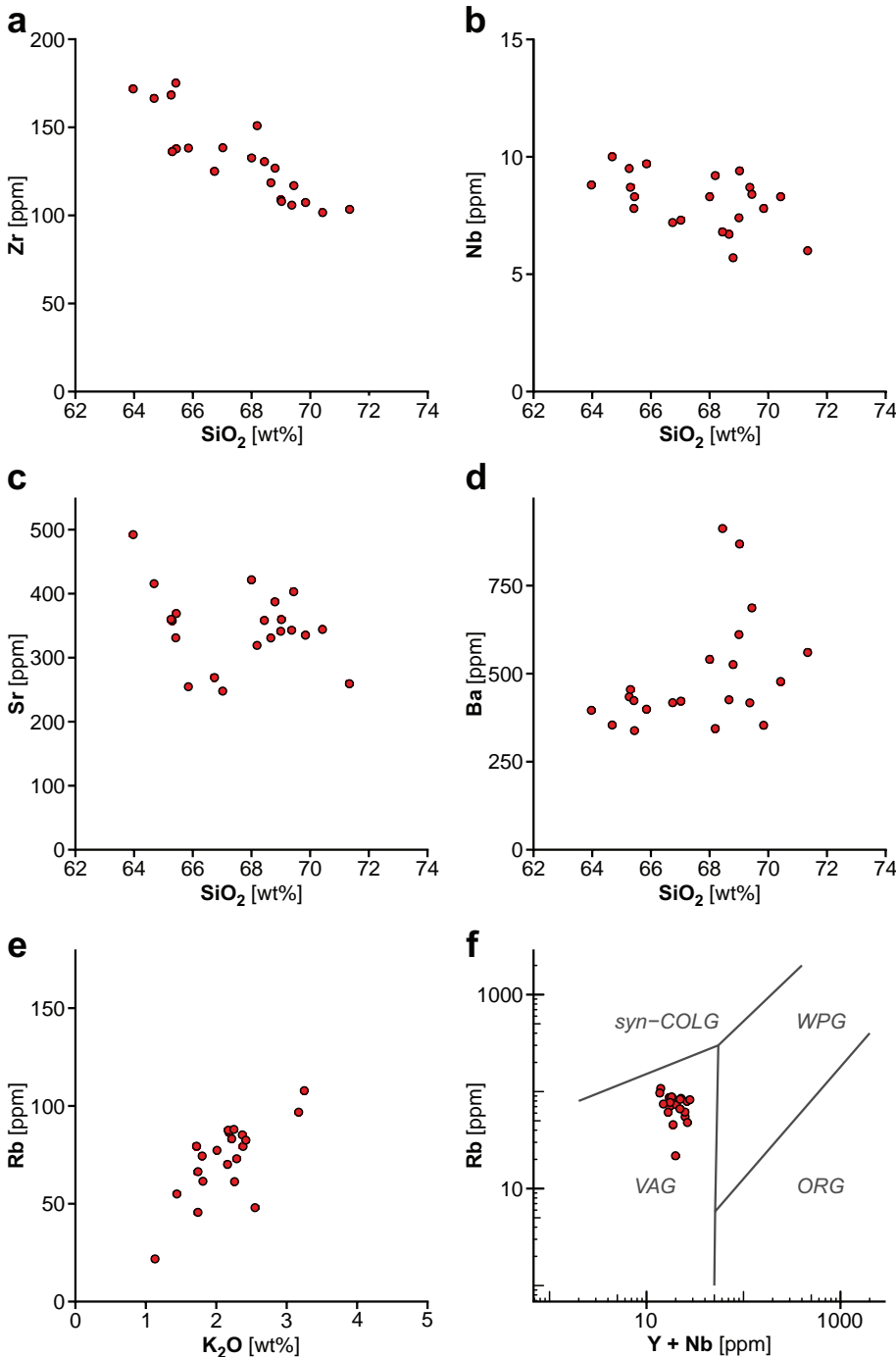


Fig. 7a–d – Binary plots of SiO_2 vs. Zr, Nb, Sr and Ba. **e** – K_2O vs. Rb diagram showing the positive correlation of both elements. **f** – $\text{Y} + \text{Nb}$ vs. Rb diagram after Pearce et al. (1984). Spitz Gneiss samples fall within the volcanic-arc field.

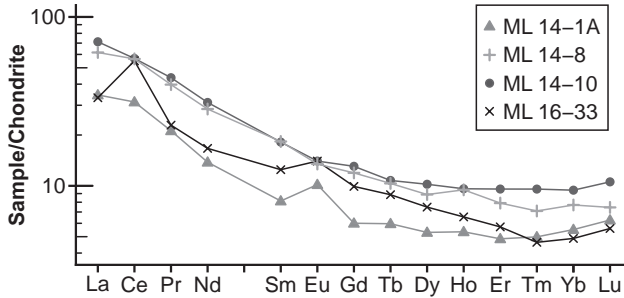


Fig. 8 Chondrite-normalized REE multi-element plot for Spitz Gneiss samples (normalizing values from Boynton 1984).

granites (Fig. 7f), which is in accordance with its I-type characteristics.

Chondrite-normalized Rare Earth Element (REE) patterns (Boynton 1984) for four representative samples of Spitz Gneiss are shown in Fig. 8. All samples have moderate LREE enrichment, decreasing from La (33–70 La_N) to Sm (8–18 Sm_N). The HREE tails of the patterns are flat (5.9–10.7 Tb_N ; 5.6–10.6 Lu_N). Samples ML 14-1A and ML 16-33 have a slight positive Eu anomaly and thus likely have experienced magmatic feldspar accumulation. The rare feature of a positive Ce anomaly is observed in sample ML 16-33. It may result from the addition of Ce^{4+} from a fluid (Starijaš Mayer et al. 2014). Finds of cerianite (CeO_2) in a pegmatite in the vicinity of sample ML 16-33 (Walter et al. 2017) are an indication for a local fractionation and mobility of Ce^{4+} . The problem regarding the Ce anomaly appears interesting and would deserve a more detailed future study including LA-ICP-MS measurements on single rock-forming minerals.

Figure 9 shows an N-MORB normalized multi-elemental plot for the Spitz Gneiss. There is a moderately strong enrichment of the Large Ion Lithophile Elements (LILE; e.g. Cs, Rb, Ba, K) at an only slight to moderate enrichment of the HFSE and the LREE. We observe troughs for Nb, Ta, P and Ti, as well as spikes for K, Pb and Sr. The HREE (Gd–Lu) and Y are slightly depleted relative to N-MORB. The enrichment of LILE over Nb and Ta, in combination with a relative depletion in HFSE, is characteristic of most granodioritic rocks world-wide and the Spitz Gneiss is no exception in this respect. We can state though, that the enrichment of Cs in the Spitz Gneiss is less pronounced than in many other granodiorites (e.g. Variscan granodiorites from the Bohemian Massif – Janoušek et al. 2000; Laurent et al. 2014). Conversely, the LREE/HREE enrichment is rather modest in the Spitz Gneiss, while the depletion in the HREE is pronounced relative to the average granodiorite (Taylor 1968).

7. Discussion and conclusions

7.1. Geochemical comparison of the Spitz Gneiss and the Moravo–Silesian granitoids

The Late Proterozoic Moravo–Silesian granitoids are well investigated with some 200 geochemical data being available in the literature (van Breemen et al. 1982; Finger et al. 1989, 1995, 2000; Souček et al. 1992; Jelínek and Dudek 1993; Hanžl and Melichar 1997; Leichmann

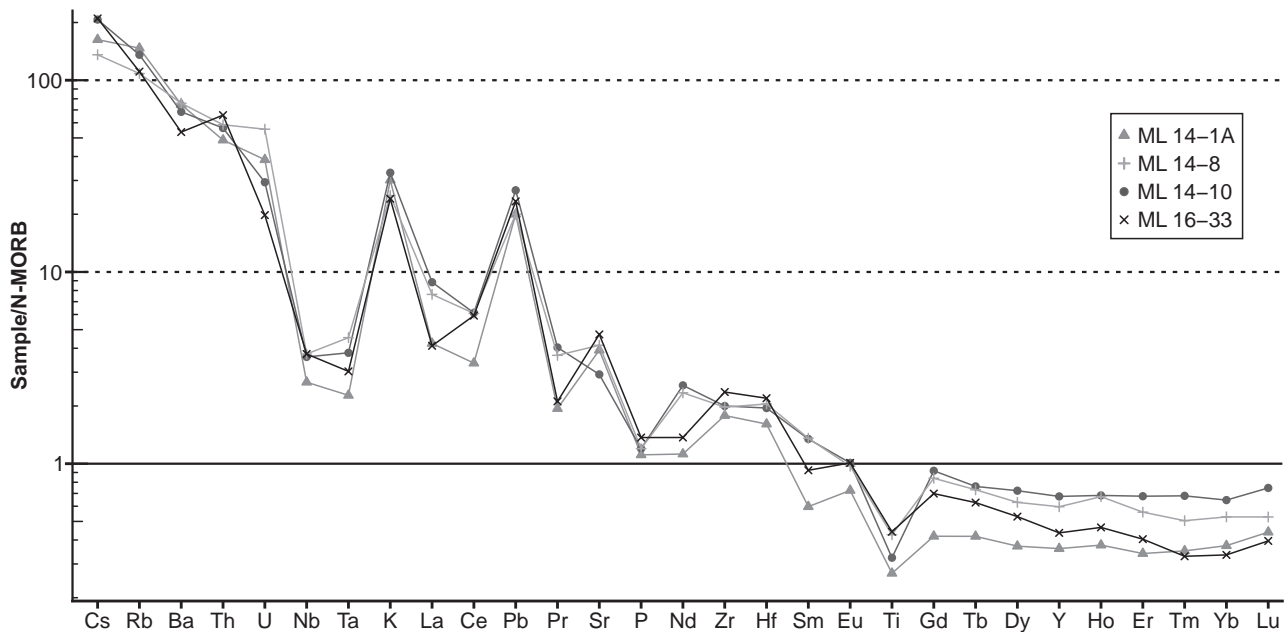


Fig. 9 Multi-element plot normalized by N-MORB (normalizing values from Sun and McDonough 1989).

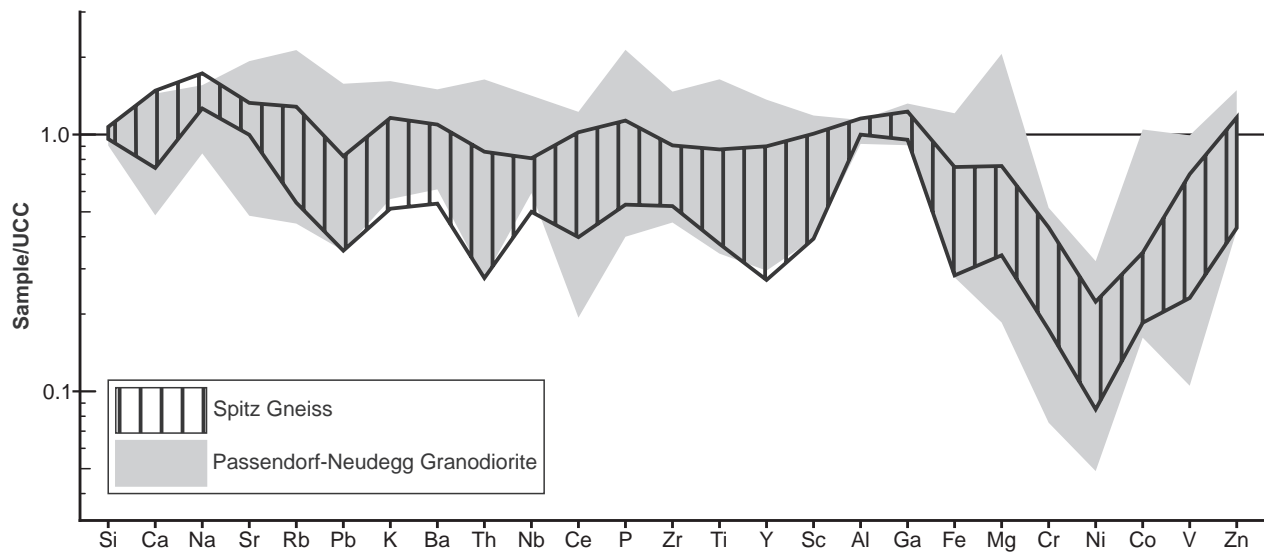


Fig. 10 Multi-element plot illustrating the similarities of Spitz Gneiss (hatched field; $n = 21$) and Passendorf-Neudegg Granodiorite (grey field; $n = 42$). Normalization to average upper continental crust (UCC; Rudnick and Gao 2003).

and Höck 2008; Finger and Riegler 2012a, b; Matzinger and Finger 2017; Soejono et al. 2017). Discrete bodies of I-type biotite ± hornblende granodiorite were described from several places (Štelcl and Weiss 1986; Finger et al. 1989; Leichmann and Höck 2008; Soejono et al. 2017). Some of these Moravo–Silesian granitoid occurrences are geochemically distinct from the Spitz granodiorite gneiss, as for instance the high-K, Ba- and Cr-rich Buttendorf granodiorite gneiss (Finger and Riegler 2012b, c), the Sr-rich Blansko granodiorite (Hanžl and Melichar 1997;

Leichmann and Höck 2008), the Ba-rich granodiorite from Anenský mlýn (Soejono et al. 2017), or the Zr-rich, high-K Gumping granodiorite (Finger and Riegler 2009, 2012c). However, a good geochemical match is found between the Spitz Gneiss and the Passendorf-Neudegg granodiorite suite in the Thaya Batholith (Finger and Riegler 2006, 2012a), as is shown in Fig. 10.

7.2. Geochemical comparison of Spitz Gneiss and Dobra Gneiss

The term Spitz Gneiss was first used by Waldmann (1938), but a “hornblende-bearing granodioritic gneiss in the surroundings of Spitz” has already been mentioned in earlier work (Becke 1917). The Spitz Gneiss and the Dobra Gneiss were long considered to be the same rock type (e.g. Waldmann 1951; Desphande and Özpeker 1965). However, Fuchs and Matura (1976) defined the Spitz Gneiss as a separate unit due to the fact that it has a significantly lower K-feldspar content than the Dobra Gneiss.

Notably, intercalations of amphibolite (0.5–10 m) are sometimes observed in the Spitz Gneiss (Lindner 2016). They have been interpreted as pre-Variscan mafic dykes (Finger and Steyrer 1995), but their precise age is unknown.

As has been shown in a recent MSc thesis (Lindner 2016), the Dobra Gneiss is also an I-type granitoid, but differs from the Spitz Gneiss in a number of geochemical parameters. Most significant is its higher SiO_2 content (Fig. 11) and its more ferroan nature (Frost et al. 2001). A striking geochemical affinity exists between parts of the Dobra Gneiss and the Moravian Bittesch Gneiss (Lindner and Finger 2016).

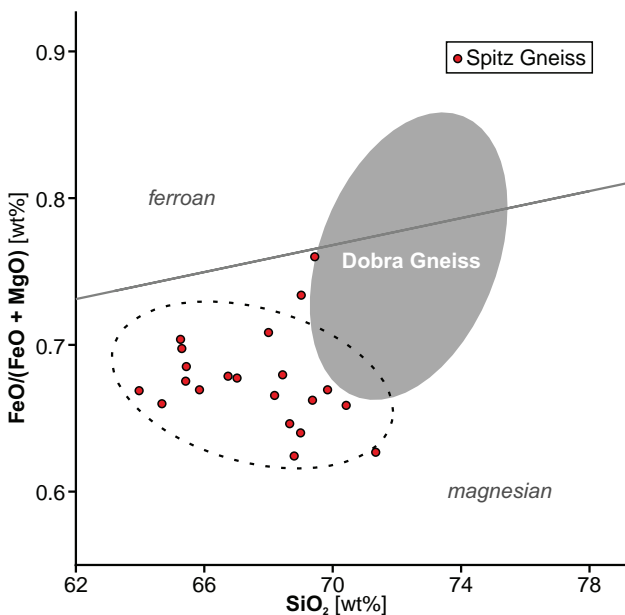


Fig. 11 The SiO_2 – $\text{FeO}_t/(\text{FeO}_t + \text{MgO})$ diagram after Frost et al. (2001) showing the distinct chemical characters of the Dobra Gneiss (grey field) and the Spitz Gneiss (dashed field). The two ellipses outline 90 % of the Spitz and Dobra Gneiss data, respectively.

7.3. Comparing the Cadomian to Early Paleozoic magmatism in the Moravo–Silesian and the Moldanubian domains

Based on available literature and unpublished University of Salzburg data, we made an attempt to discriminate the Cadomian Moravo–Silesian granitoid province from the Cadomian to Early Palaeozoic granitoids that are preserved as orthogneiss in the Moldanubian Zone (i.e. Gföhl Gneiss, Blaník Gneiss, Stráž Gneiss, Vír/Svratka Gneiss, Orlice–Sněžník Gneiss; van Breemen et al. 1982; Souček et al. 1992; Vellmer 1992; Breiter et al. 2005; Hasalová et al. 2008; Buriánek et al. 2009; René and Finger 2016). The Moldanubian orthogneisses have, on average, higher A/CNK values, higher Y and HREE and lower Ba values. The best discrimination is provided in the Rb vs. Sr diagram (Fig. 12). Most Moravo–Silesian granitoids have Sr > 200 ppm and Rb < 200 ppm, while the Moldanubian orthogneisses have invariably Sr < 200 ppm and Rb often exceeds 200 ppm. Both, the Spitz and the Dobra Gneiss, fall clearly within the field of the Moravo–Silesian granitoids.

7.4. Tectonic implications

The fact that the Spitz Gneiss fits reasonably well into the compositional and age spectrum of the Cadomian granitoids of the Moravo–Silesian Unit corroborates geological models claiming that the Lower Austrian Drosendorf Unit belonged to the Moravian Zone (and to Avalonia) before it was tectonically incorporated into the Moldanubian nappe system (Fritz and Neubauer 1993; Finger et al. 2007b). Notably, temporal and compositional equivalents to the Spitz Gneiss (as well as to the Dobra Gneiss) have not yet been found elsewhere in the Moldanubian Zone, whereas the most characteristic Moldanubian S-type orthogneisses (Gföhl Gneiss and Blaník Gneiss) are totally absent in the Drosendorf Unit. All this highlights the exotic nature of the Lower Austrian Drosendorf Unit.

During the Variscan Orogeny, the Drosendorf Unit inevitably experienced significantly higher regional metamorphic conditions than the Moravian Unit in the Austrian part of the Thaya Dome. Our peak P–T estimates for the Spitz Gneiss (~700 °C, 7 kbar) are in accord with previous geothermobarometric data from other parts of the Lower Austrian Drosendorf Unit (Högelsberger 1989), whereas maximum P–T conditions of around 600 °C, 6 kbar have been reported from the hanging-wall parts of the Moravian Unit in Austria (Höck et al. 1990). Based on the Variscan metamorphic conditions it would make sense to interpret the Drosendorf Unit as a westerly and thus hotter part of the subducted Moravian Plate that was ripped off and steeply exhumed into the overlying

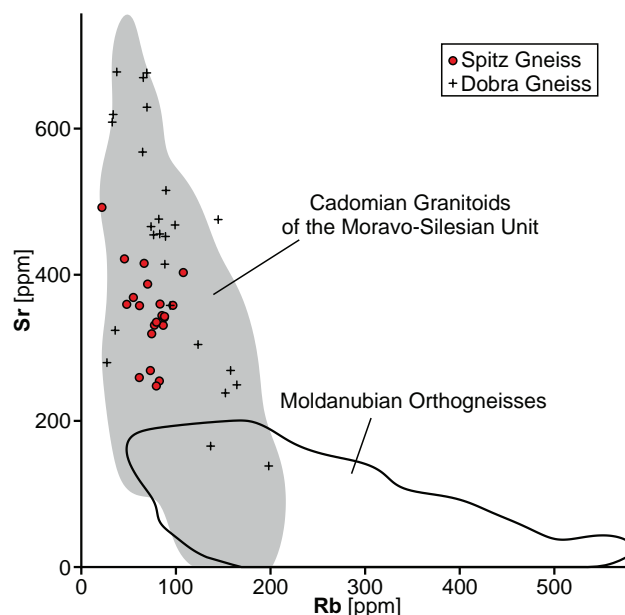


Fig. 12 Rb vs. Sr scatterplot for Moldanubian orthogneisses and the Cadomian granitoids of the Moravo–Silesian Unit. Dobra and Spitz Gneiss are independently plotted and fall within the Moravo–Silesian field. Fields represent 90 % of the input data, based on a 2D kernel density estimation (Venables and Ripley 2002). (Data sources: van Breemen et al. 1982; Souček et al. 1992; Vellmer 1992; Jelínek and Dudek 1993; Hanžl and Melichar 1997; Finger et al. 2000; Breiter et al. 2005; Leichmann and Höck 2008; Hasalová et al. 2008; Buriánek et al. 2009; Finger and Riegler 2012a, b, c, 2013, 2014; René and Finger 2016; Matzinger and Finger 2017; Soejono et al. 2017 and additional unpublished data of the University of Salzburg).

Moldanubian Plate during the Variscan Orogeny, possibly through processes of backthrusting (Fritz and Neubauer 1993). In terms of metamorphic conditions, the Drosendorf Unit may correspond to the Vranov-Olešnice Unit in the Czech part of the Moravian Unit.

The Variscan exhumation of the Spitz Gneiss (and probably of the entire Lower Austrian Drosendorf Unit) to upper crustal levels was likely a fast process, because minerals like hornblende and plagioclase lack retrograde reequilibration at lower amphibolite- and upper greenschist-facies conditions. A fast cooling of the unit at ~340 Ma has also been inferred by Dallmeyer et al. (1992) and Racek et al. (2017) based on Ar–Ar mica and hornblende dating. The observable chloritization of biotite in the Spitz Gneiss represents a post-exhumation event at ~250 °C. It was probably triggered by a regional fluid introduction into the post-collisional Variscan crust (Jaweck 1996), during the Carboniferous or Permian.

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