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Martin Labuta

Vztah zemětřesení a vulkanismu v západních Čechách a na Islandu
Relation of earthquakes and volcanism in West Bohemia and Iceland

Typ závěrečné práce
Bakalářská práce

Vedoucí bakalářské práce: Ing. Josef Horálek CSc.
Konzultant bakalářské práce: doc. RNDr. Tomáš Fischer, Ph.D.

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Prohlášení

Prohlašuji, že jsem závěrečnou práci zpracoval samostatně a že jsem uvedl všechny použité informační zdroje a literaturu. Tato práce ani její podstatná část nebyla předložena k získání jiného nebo stejného akademického titulu.

V Praze, 26.05.2015

Podpis

English abstract

Most earthquakes occur as accompanying activity of large tectonic earthquakes at the edges of lithospheric plates. Another type of seismic activity are earthquake swarms, which are characterized by series of earthquakes with several dominant shocks of similar strength. The origin of earthquake swarms is mostly combination of tectonic stress and movement of hydrothermal fluids along the fault plane. Mainly, swarms are observed within volcanic areas but even in areas without recent volcanic activity, like in West Bohemia.

This region lies above an intra-continental rift zone – the Eger Rift and it has been seismologically active in past years, including swarms in 2008, 2011, 2013 and 2014 monitored by the local seismic network WEBNET with dominant shocks of magnitudes below ML 4.5. All these events are located in depths between 7 and 11 km and they create 8 km long focal zone with N-S orientation. However, the latter swarm changed it's character from a continuous occurrence with a dominant shock to a main shock and aftershock activity.

Contrary, the region of Reykjanes peninsula in Iceland is located above the Mid-Atlantic Ridge, which results in shallower depths between 2 to 9 km. Additionally, the seismic energy in this area is released as a typical swarm-like activity and migrates along the rift in certain clusters.

Main purpose of this study is to clarify the processes of earthquake swarms and to study possible relations in different tectonic settings. This could support further research concerning the better understanding of spatial distribution of releasing accumulated stress.

Acknowledgments

Firstly, I would like to thank my supervisor Ing. Josef Horálek CSc. for his assistance during writing this thesis. Secondly, I am grateful to Doc. RNDr. Tomáš Fischer PhD. for his help with MATLAB. Lastly, I would like to thank all people at the Institute of Geophysics of the Academy of Sciences, Czech Republic, who gave me helpful advices.

Contents

1 Introduction

The bachelor thesis focuses on earthquake swarms and volcanism in different tectonic areas represented by intra-continental rift zone in West Bohemia/Vogtland (the Czech republic/Germany) and by the active continental margin on the Reykjanes peninsula, which is the most south-western part of Iceland. Both areas are volcanically active since the beginning of Quaternary, moreover, they have frequent occurrence of seismic activity. This activity releases stress in form of earthquake swarms, which can last from days to months. The one of the possible causes for this time distribution is mobility of crustal fluids.

The main goal is to locate occurrence of these swarms and make a conception about possible shared features of this seismicity in both areas and its origin. These features include speed of releasing stress, depths of foci and spatial distribution of earthquakes.

This text is divided into following parts - a review, which include description of geology, tectonics, hydrothermal activity, volcanism and earthquake swarms and their history.

2 West Bohemia and Vogtland

Seismicity in this region is part of the Regensburg-Leipzig-Rostock seismoactive zone, which has the north-south orientation (Bankwitz et al., 2003). The first observations were conducted during medieval times and since the 19th century they were documented based on macroseismic data - 100 earthquakes were analysed in 1824 by Knett. In the end of the 19th century to beginning of the 20th, there were swarms with maximum intensity I_o of 5.5 to 7.0. First sufficiently recorded swarm was in 1908 with estimated maximum magnitude of 4.4 (Fischer et al., 2014 and reference therein).

In 1962, seismic monitoring was commenced using a network of analogue stations in the Vogtland area(Neunhöfer and Meier, 2004). During the swarm in 1985/86, two seismic stations were installed - VAC and TIS, followed by NKC station set up in 1989. Between 1991 and 2008, there was also a seismic network named KRASNET. The WEBNET seismic network, which is currently operated by the Institute of Geophysics of the Academy of Sciences, was established in 1994 using these three previous stations installed in 85 and 89. Presently, it includes 13 permanent and 10 temporary seismic stations (Fischer et al., 2014).

2.1 Geologic and tectonic settings

The area of West Bohemia and Vogtland lies in the western part of the Bohemian Massif (BM) on the transition section of its three structural units: the Saxo-Thuringian zone (S-T), the Teplá-Barrandian block (T-B) and the Moldanubian zone (MZ) (Fischer et al., 2014); which were consolidated during the Variscian orogeny. The Bohemian Massif includes the fourth unit - the Moravo-Silesian (M-S) unit (Babuška and Plomerová, 2013), which occupies the eastern part of BM (Fig.1 on page 3).

The Saxo-Thuringian zone creates the north-western section of the BM and it is separated on the south-east from the Teplá-Barrandian by the Teplá suture with NE-SW orientation (Babuška and Plomerová, 2013), which is a relic of subduction of the Saxothuringian ocean under the T-B block in the end of the Middle Devonian. On the south, the border between MZ and S-T is also tectonic. The S-T zone is divided into three parts: the Ore Mountains, the Elbe lineament and the Sudetes zone based on the intensity of metamorphosis HP-UHP (high pressure - ultra high pressure) due to the different thermal gradient. This intensity decreases away from the suture to the north-west. Additionally, it influenced Cadomian magmatic rocks and pre-Variscian sedimentary sequences (Kachlík, unpublished; Babuška and Plomerová, 2013).

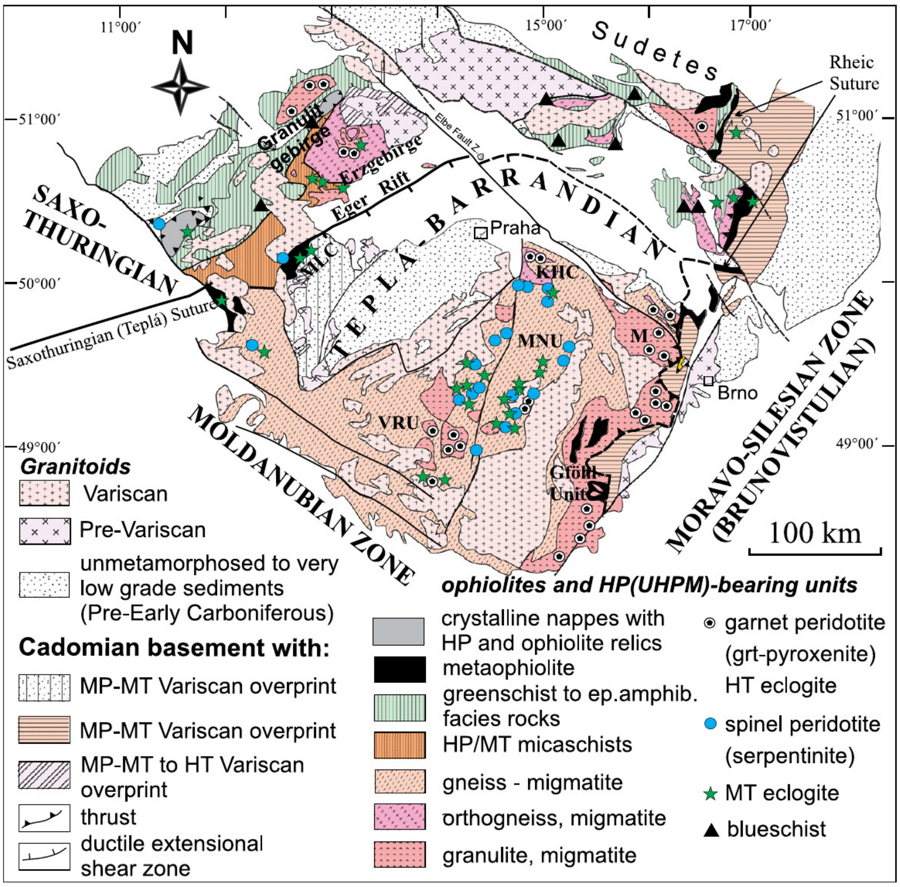

Figure 1: The Bohemian massif (Babuška and Plomerová, 2013; modified after Faryand and Kachlík, 2013). M - Mohelno peridotite; MLC - Mariánské Lázně Complex; KHC—Kutná Hora Complex; MNU—Monotonous Unit; VRU—Varied Unit.

The Moldanubian zone is located in the southern part of the BM. Its boundaries are determined by tectonic lines, which were marked during the Variscian orogeny. On the west, the MZ subsides under the S-T unit in the Western Bohemia shear zone, and under the Kutna Hora-Svratecko area on the north. However, it overrides the margin area of the M-S on the east. Lastly, on the south, the MZ is covered by platform sedimentary rocks of Eastern Alps (Kachlík, unpublished). The MZ is divided into three parts based on the lithotectonic structure and metamorphosis. Firstly, the medium- to high-grade Monotonous (Ostrong) and Varied (Drosendorf) Units, which contains gneiss, migmatites with sparsely distributed block of eclogites and peridotites. Secondly, the high-grade Gföhl Unit, which also includes garnet peridotites (Faryand and Kachlík, 2013).

The Teplá-Barrandien unit occupies the central part of the BM. It comprises two structural floors: anchimetamorphic Cadomian sedimentary basement with volcaniclastic rocks and early Paleozoic sedimentary sequences, which were later folded by Variscian orogeny. According to geophysical measurements, its crust has higher density than surrounding structural units and magnetic anomalies caused by volcaniclastic rocks. The TB is marked by Teplá and Ghöhl suture (Kachlík, unpublished).

Lastly, the Moravo-Silesian unit as the eastern part of the BM consists of Neoproterozoic Brunovistulian fundament and late Paleozoic pre- and synorogenic sedimentary succession (Babuška and Plomerová, 2013).

2.1.1 West Bohemia

The studied area is situated above the cross section of the Cenozoic Eger Rift visible on surface as the Eger Graben, which is a 300 km long and 50 km wide zone, ENE-WSW oriented; and the Regensburg-Leipzig-Rostock (RLR) Zone (Fischer et al., 2014). This zone is 700 km long and 40 km wide with frequent N-S faults, resulting in the same orientation of the seismoactive sinistral fault zone (the Počátky-Plesná Zone) at Nový Kostel in the Cheb Basin. The Basin was formed by reactivation of faults in the fundament (Bankwitz et al., 2003).

In the eastern part of the RLR Zone is located the Cheb-Domažlice Graben, which is delimited on its east side by the Mariánské Lázně Fault (MLF) heading NNW-SSE. The MFL is a 100 km long zone with a 50-400 m high steep slope, the western part of the graben has more gradual tomography (Fischer et al., 2014).

In a terms of geology, this area consists of eight basic units: the Slavkov forest crystalline basement, the Ore Mountains crystalline basement, the Cheb phyllites, the Kladská unit, the Carlsbad pluton, the Fichtel Mountains massif, the Cheb Basin and the Sokolov Basin. These units includes rocks of the Neoproterozoic, Paleozoic and Cenozoic Eras (Müller et al., 1998). In this study, only the Cheb Basin will be described as the area of interest.

Filling of the Cheb basin comprises the Cenozoic sediments such as lignite, clay and sand followed by gravel, sand and clay sedimentary formations. The overall thickness does not exceed 300 m. Sedimentary process was initiated in Late Eocene, lasted to Lower Oligocene with a hiat until Late Oligocene. Sedimentation continued to Miocene, followed by a 12 Ma long hiat to Upper Pliocene, when it restarted. (Bankwitz et al., 2003, Fischer et al., 2014).

2.2 Volcanic activity and fluids

2.2.1 Volcanism

Volcanism is represented by Komorní hůrka (KH) and Železná hůrka (ZH), which are two scoria cones of Quaternary age. Additionally, the Mýtina maar with depth of more than 140 m is situated near the ZH. These volcanoes are located at the Tachov fault, which creates the western boundary of the Cheb-Domažlice Graben. Their activity is divided into two stages: the phreatomagmatic initial stage and the eruptive final stage with lava determined as an olivine nephelinite. Age studies indicated, that this activity occurred in the middle Pleistocene 0.78-0.12 Ma ago (Fischer et al., 2014 and reference therein).

2.2.2 Activity of fluids

Fluids and their changes in chemical composition are connected to seismic activity. Their migration along faults is supposed as a possible initial process of triggering the earthquakes. In West Bohemia, there is a large CO_2 degassing in form of mineral waters saturated with CO_2 and wet or dry moffetes. It is estimated, that the gas flow in the Cheb Basin is about thousands of cubic meters per hour (Faber et. al., 2009), followed by another degassing areas: Mariánské Lázně, Carlsbad and Bad Brambach (Fischer et al., 2014, Koch et al., 2003).

Therefore, the monitoring stations were installed in the Cheb Basin: Oldřišská (established in 2004); Nový Kostel (operated since 2006), both of them extract gas from depth of 2.5 m to mitigate external effects; the station Soos (installed 2007, 8 km SSW from Nový Kostel), which measures deep seated source of CO_2 emission at a mofette (Faber et. al., 2009); and the station Bublák near Františkovy Lázně in West Bohemia. Moreover, the Wettinquelle spring as the one of the other six mineral springs at Bad Brambach is monitored. It is also known for its strong radioactivity, which is used for therapeutic applications (Koch et al., 2003). Additional monitored sites are mofettes at Hartoušov and Dolní Častkov; and mineral springs at Kopanina, Plesná, U Mostku and the Císařský spring (Fischer et al., 2014) (Fig.2 on page 6).

Flows of CO_2 help to carry isotopes of helium. The ratio of the $^3He/^4He$ is used to determine origin either in the crust or in the mantle. The highest ratio of 6 R_a (1 R_a corresponds to atmospheric ratio) was monitored in the Cheb Basin, which indicates a mantle source of helium. Releasing of helium from those two parts of Earth varies in time, but the gradual increase of the $^3He/^4He$ support the idea of a hidden magmatic process underneath the crust. In Mariánské Lázně and Carlsbad area, the isotope ratio is lower - 4.9 R_a and 2.5 R_a respectively, but it is more constant in time. This is caused by lower seismic activity (Fischer et al., 2014 and reference therein).

Monitoring of the mineral springs indicates a possible connection between changes of water temperature, chemistry, groundwater level and the occurrence of earthquakes. For example, an increase of water temperature of 3° C was observed few months before the swarm in 1985. Furthermore, a decrease of groundwater level was noticed in all wells during the swarm in 2008 (Fischer et al., 2014).

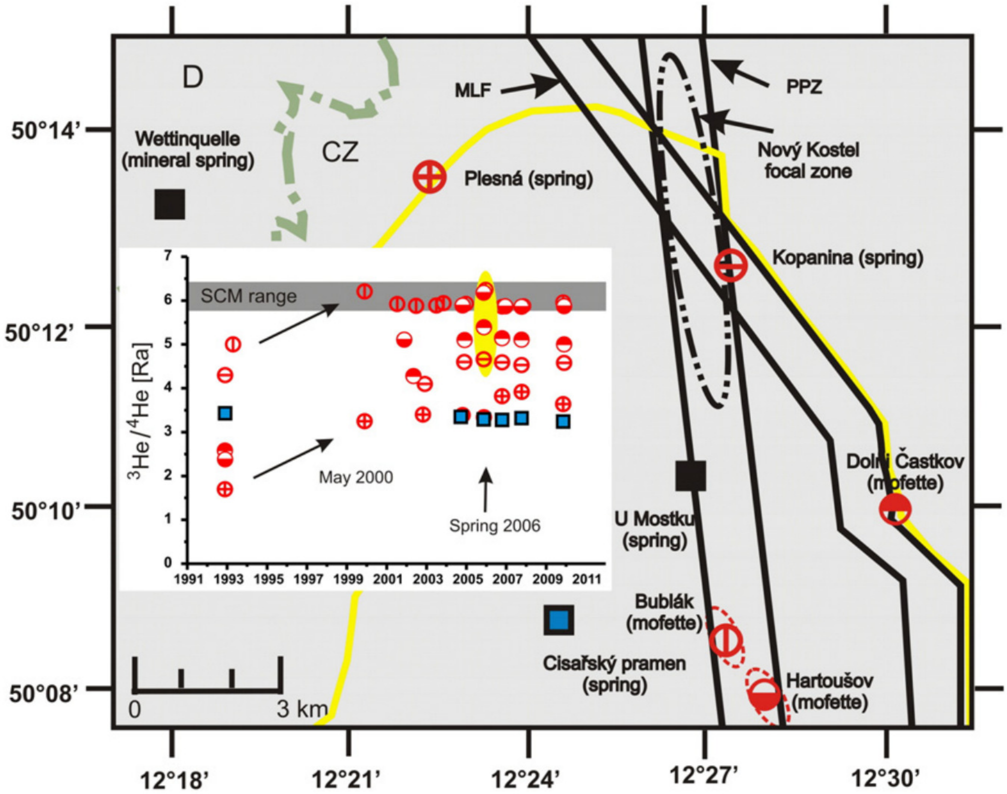

Figure 2: Monitored locations of hydrothermal activity (after Fischer et al., 2014) Yellow line - border of the Cheb Basin, MLF - Mariánské Lázně fault, PPZ - Počátky–Plesná zone, the dotted ellipses at Bublák and Hartoušov are the mofettes. The yellow ellipse displays the increase of $^3He/^4He$ ratios at locations along the PPZ and MLF in spring 2006.

2.3 Earthquakes and earthquake swarms

The seismicity in West Bohemia is marked by the area between 49.9° and 51°N and 12.0° and 12.8°E. The Nový Kostel (NK) focal zone is created by a NNW heading belt, which is 2 km wide and 12 km long. Over 18 thousands events have been detected in the NK focal zone during the period of 1991-2011 (Fischer et al., 2014) (Fig.3 on page 8).

In the timespan from 1991 to 2000, two earthquake swarms, which released most of the seismic energy (January 1997 and August–December 2000 with 10 weeks long main activity), and 27 micro-swarms were monitored in the Nový Kostel (NK) focal zone. The hypocenters formed two clusters along the planar fault zone in depths between 6 and 11 km. This main fault zone (MFP) is striking 169° N and dipping 80° W and accommodates the foci of 2000 swarms and microseismicity, however, the January 1997 swarm has occurred across the fault plane. The focal belt's position and direction is in agreement with the Nový Kostel-Počátky-Zwota tectonic line (Fischer and Horálek, 2003).

The space distribution groups larger events $M_L > 1.5$ into planar clusters, while the micro-swarms form two parallel seismogenic lines dipping to the south at an angle of 32°. Reactivation of the MFP occurred several times, mostly at the clusters of 1985/1986; January 1997, which was characteristic by frequent micro-swarms; and 2000 swarms, however, a large area of MFP was activated only once. Migration of seismic activity suggests, that the whole seismicity from 1991 to August 2000 have been located in the northern cluster of the MFP, whereas the swarm in 2000 occurred in the southern cluster (Fischer and Horálek, 2003).

In 2004 another episode of seismic unrest was initiated in a form of micro-swarms with magnitudes below 2.0. The hypocenter clusters were located at the flanks of the active fault area, and migrated to the south and to the greater depths of 10 to 13 km. All the focal mechanisms show the compatible pattern with NNW-SSE geometry, which results in the left-lateral strike slip with varying dip-slip component. Small amount of the seismic energy was released compare to swarm in 2000 (Fischer and Michálek, 2008).

Following earthquake swarm began on 6 October 2008 and ended in December. The largest events include two earthquakes of magnitude M_L 3.8 and nine events of magnitude $M_L > 3.0$. About 25 thousands events of magnitude $-1.5 < M_L < 3.8$ were detected. Using Gutenberg–Richter law, the b-value close to 1 was determined for the whole 2008 swarm. During this swarm, there were an excess of the largest events and lack of events of magnitude 3.0 to 3.5. This suggests a set of multiple mainshock and aftershocks series. The main seismic activity lasted for only 4 weeks, which led to the fastest

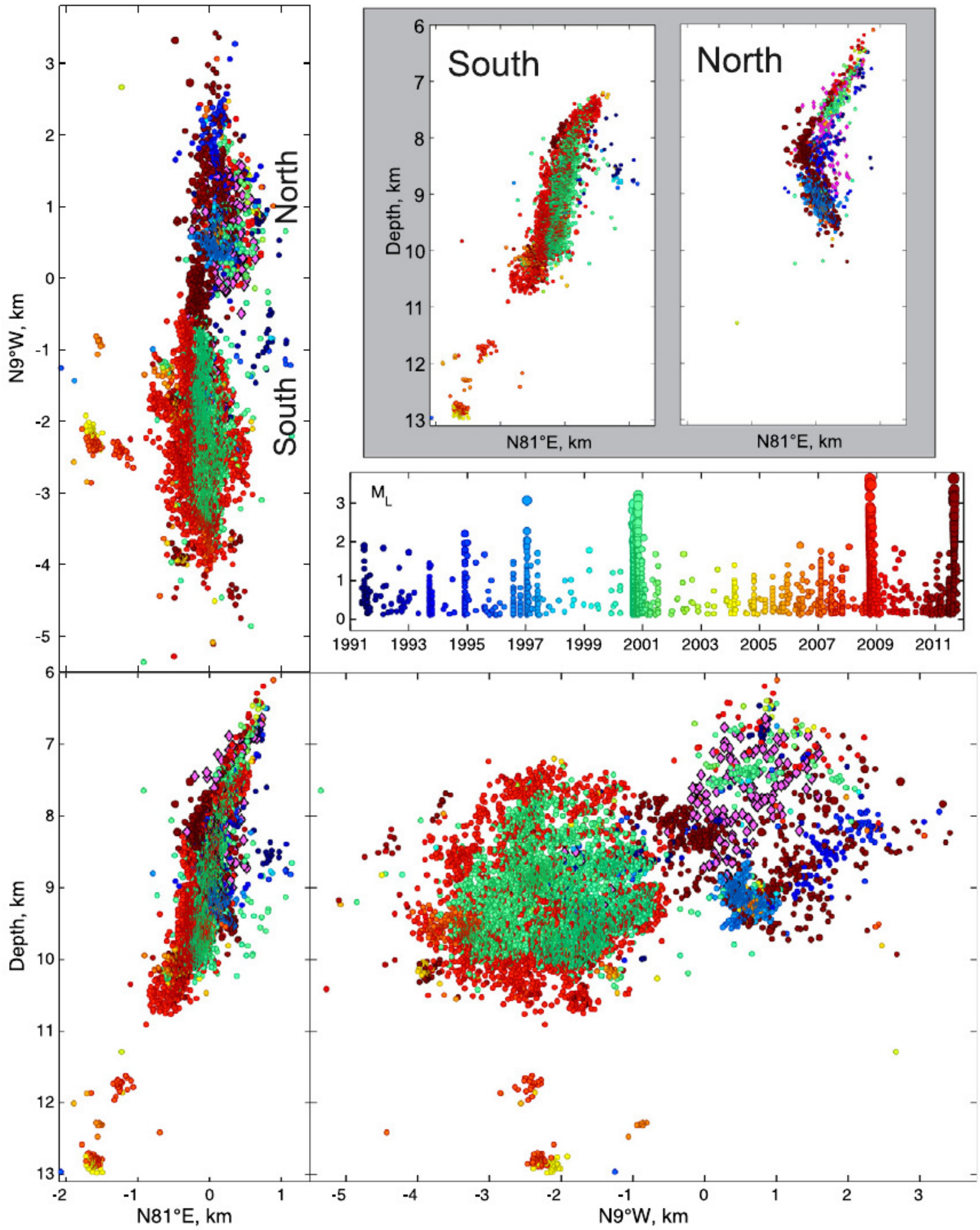

Figure 3: Hypocenters in the Nový Kostel focal zone during the period 1991-2011 and the 1985/86 swarm displayed as violet diamonds. Map view (top-left), two perpendicular sections (bottom) and sections of the southern and northern cluster (top-right) (after Fischer et al., 2014).

seismic moment release since the swarm in 1985/86. The hypocentres were located at the same part of the fault at the same depth (6 to 11 km) with identical focal mechanism (strike 170° and dip 80°) as the hypocentres of the 2000 swarm (Fischer et al., 2010).

The 2011 swarm (M_L 3.5) is characteristic by even shorter timespan of 2 weeks in August/September, that causes faster releasing of the seismic energy. Hypocentres are lined along the northern cluster in depths about 6 to 11 km. It started only 34 months after the 2008 swarm, which was the shortest inter-swarm time (Fischer et al., 2014). The most recent swarms occurred in 2013, May/June 2014 (M_L 4.5) and August 2014 (M_L 3.5) (Horálek et al., 2015).

3 Iceland

Iceland lies on the Mid-Atlantic Ridge, where the asthenosperic flow interacts with a mantle seated plume. The plume was seismically imaged to depths about 400 km. In the last 60 Ma, the north-east Atlantic plate and Eurasia plate boundary have shifted relatively to the surface expression of the plume at rate of 1-3 cm/year to the north-west. This results in positioning the plume under the Vatnajökull glacier. However, the Icelandic rift zones migrated in the east direction to stay in the vicinity of the surface expression of the plume, which led to a different pattern of the particular rift zones and transform fault zones (Trønnes, unpublished).

Most of earthquakes are bounded to these zones, the largest earthquakes with magnitude up to 7.1 have been observed in two fracture zones: the Tjörnes Fracture Zone (TZF) and the South Iceland Seismic Zone (SISZ). The TZF is located between the Northern Volcanic Zone (NVZ) and Kolbeinsey Ridge and the SISZ connects the Eastern Volcanic Zone (EVZ) to the Western Volcanic Zone (WVZ) and Reykjanes Peninsula (Jakobsdóttir et al., 2002) (Fig.5 on page 13).

Seismicity in Iceland is monitored by the South Iceland Lowland (SIL) system, which started as a network of 8 seismic stations located in the SISZ and operated since 1990 (Stefánsson et al., 1993). In 1993, additional stations were installed in the TZF and became a regional network for Iceland. Further expansion was realised in the following years, resulting in 51 stations at the end of 2007. The SIL system comprises automated three component data acquisition, monitoring system and automatic processing software (Jakobsdóttir, 2008).

The SIL uses two formulas for calculations of magnitude. The first one - the local magnitude M_l is based on the maximum peak-to-peak amplitude in a 10 seconds interval around the arrival of the S wave:

$$M_l = log_{10}A + 2.1 * log_{10}D - 4.8$$

where A is the maximum amplitude of velocity, which is filtered by high-pass filter with a frequency of 2 Hz. D is the distance in kilometres between the hypocenter and the station. The second formula is used for determining the local moment magnitude M_{lw} of earthquakes M_{lw} <2:

$$M_{lw} = log_{10}(M_0) - 10$$

where M_0 is the seismic moment in Nm (Jakobsdóttir, 2008).

The network utilize a 1D gradient velocity model named SIL and proposed by Stefánsson et al., (1993) with ratio of P-velocities to S-velocities equal to 1.78 (Table 1 on page 10). However, Vogfjörd et al., (2002) suggested for the SISZ and the Reykjanes Peninsula another gradient model P1 based on the earthquake source arrays (Table 2 on page 11).

depth [km]	P [km/s]	P gradient	S [km/s]	S gradient
0.00	3.53	0.940	1.98	0.530
1.00	4.47	0.690	2.51	0.390
2.00	5.16	0.440	2.90	0.250
3.00	5.60	0.360	3.15	0.200
4.00	5.96	0.260	3.35	0.140
5.00	6.22	0.280	3.49	0.160
6.00	6.50	0.100	3.65	0.060
7.00	6.60	0.060	3.71	0.030
8.00	6.66	0.070	3.74	0.040
9.00	6.73	0.045	3.78	0.025
15.0	7.00	0.040	3.93	0.022
20.0	7.20	0.017	4.04	0.010
32.0	7.40	0.000	4.16	0.000

Table 1: The SIL 1D velocity model modified after Stefánsson et al., (1993).

The south-west Iceland is additionally monitored by the local seismic network named REYKJANET, which is operated by the Institute of Geophysics of the Academy of Sciences and consists of 15 stations installed in August/September 2013. The stations were previously ran by University of Uppsala during 2009 - 2013 (http://www.ig.cas.cz/reykjanet). There are currently installed 3-component digital seismometers: Guralp CMG-40T or Lennartz LE-3D 1Hz with logging unit GAIA on these stations (Horálek,

depth [km]	P [km/s]	P gradient	S [km/s]	S gradient
0.00	3.30	0.871	1.85	0.490
3.10	6.00	0.160	3.37	0.089
7.80	6.75	0.750	3.79	0.450
8.00	6.90	0.033	3.88	0.019
17.0	7.20	1.500	4.05	0.800
17.2	7.50	0.056	4.21	0.033
19.0	7.60	0.008	4.27	0.050
25.0	7.65	0.010	4.30	0.050
35.0	7.75	0.000	4.35	0.000

Table 2: The P1 1D velocity model modified after Vogfjörd et al., (2002).

personal statement). The question is, which velocity model will provide better results with lower time residuum (Fig.4 on page 11) . It is the object of further research.

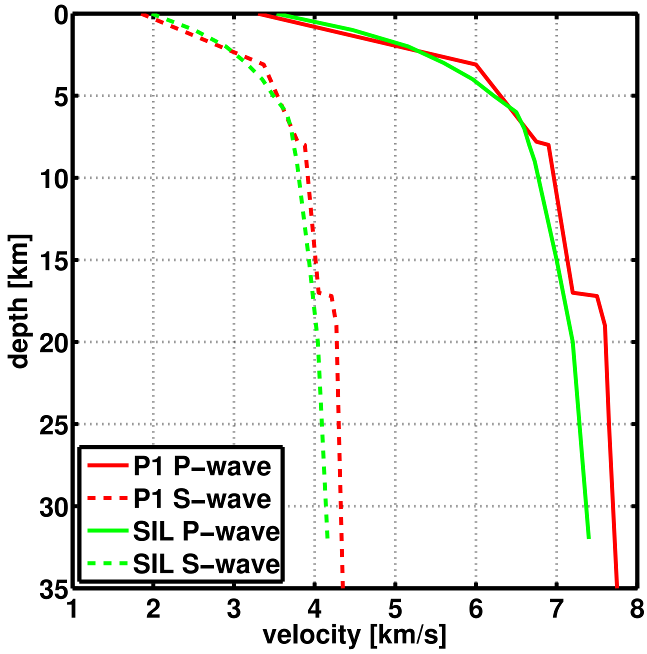

Figure 4: Comparison of velocity models: the SIL (green lines) (Stefánsson et al., 1993) and the P1 (red lines) (Vogfjörd et al., 2002) for the south-west Iceland. Full lines represent velocity of the P-wave, while dashed lines indicate velocity of the S-wave.

3.1 Geologic and tectonic settings

3.1.1 The South Icelandic Seismic Zone and the Reykjanes Peninsula

The Reykjanes Peninsula is volcanically active zone and connects the Reykjanes Ridge with the WVZ and SISZ. The boundary of plates is defined as a 2 km wide seismic zone, which runs across the peninsula (Fig.6 on page 14). Hypocenters in this zone are mostly located at a depth of 1-5 km. The minimum compressive stress is horizontal in a north-west direction. The maximum stress varies between the vertical orientation and the horizontal north-east direction, which causes strike-slip faulting on north and east striking faults or normal faulting on north-east striking faults in case of vertical stress. Normal faulting is the most frequent mechanism. Consequently, this stress regime results in the north-west oriented minimum stress. The other principal stress is spatially and time dependent (Einarsson, 1991).

The transition zone between the peninsula and the WVZ with SISZ creates the Hegill triple junction. This volcanic system is placed in the centre of 70 km long fissure zone with graben structure. Quaternary activity occurred in two parallel system - the Hengill massif and the Mount Hrómundartindur to the east. The latest rifting process in this area has developed in 1789 without any volcanism (Tryggvason et al., 2002).

The South Icelandic Seismic Zone forms a connection between two subparallel rift zones: the Western Volcanic Zone, and the Eastern Volcanic Zone, which has width about 60 km. This junction with EVZ is determined by the Hekla, Vatnafjöll and Torfajökull volcanic systems. The SISZ is oriented east-west and marked by surface ruptures and historical earthquakes that created areas of destruction. These areas strike in the north-south direction and reveal right-lateral faulting that accompanies the overall left-lateral transform motion along the main zone (Einarsson, 1991) (Fig.5 on page 13).

Structural imaging of the brittle crust beneath this area propose a large low-velocity anomaly at depths between 4 to 10 km under the Hengill volcanic system. Both velocities are low, but the S-wave velocity has greater reduction than P-wave velocity, which suggest a fractured volcanic fissure system filled with fluids of magmatic origin and meteoric water. In the centre of the peninsula, lower P- and S-wave velocities are observed, on the contrary, higher velocities surrounds the lower velocities in the volcanic zones. This is explained by presence of rocks with higher metamorphic grade or intrusive rocks in the shallower depth in the volcanic zone. At the Krisuvík volcanic system below 6 km depth, an increased V_p/V_s ratio is present, which is interpreted as a zone with elevated temperatures. Thickness of the brittle crust varies from 5 km in the Reykjanes Peninsula to 12 km at the eastern end of the SISZ (Tryggvason et al., 2002).

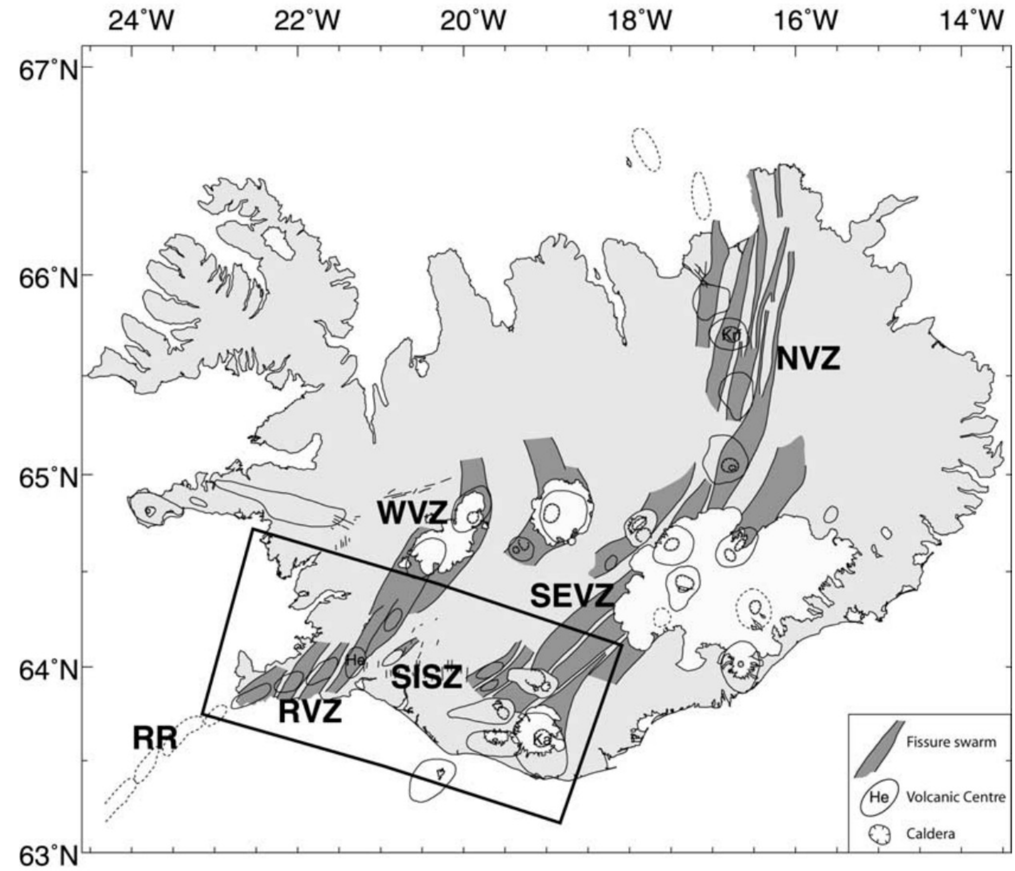

Figure 5: Map of the plate boundary represented by particular seismic zones (after Tryggvason et al., 2002). RR - Reykjanes Ridge, RVZ - Reykjanes Peninsula, SISZ - South Icelandic Sesmic Zone, WVZ - Western Volcanic Zone, SEVZ - South-Eastern Volcanic Zone, NVZ - Northern Volcanic Zone

3.2 Volcanism

Volcanism in Iceland is presented in form of fissure swarms, central volcanoes, dyke swarms and sheet swarms. Moreover, it could be divided into subglacial or subaerial volcanism. The form of the subglacial eruption depends on the hydrostatic pressure and the internal volatile pressure in the magma. Decreased pressure changes pillow lava via pillow breccia to hyaloclastite tuff. Most Icelandic subglacial volcanoes have bottom layer consisted of pillow lava, which overlays breccia and covers hyaloclastite tuff on the top.

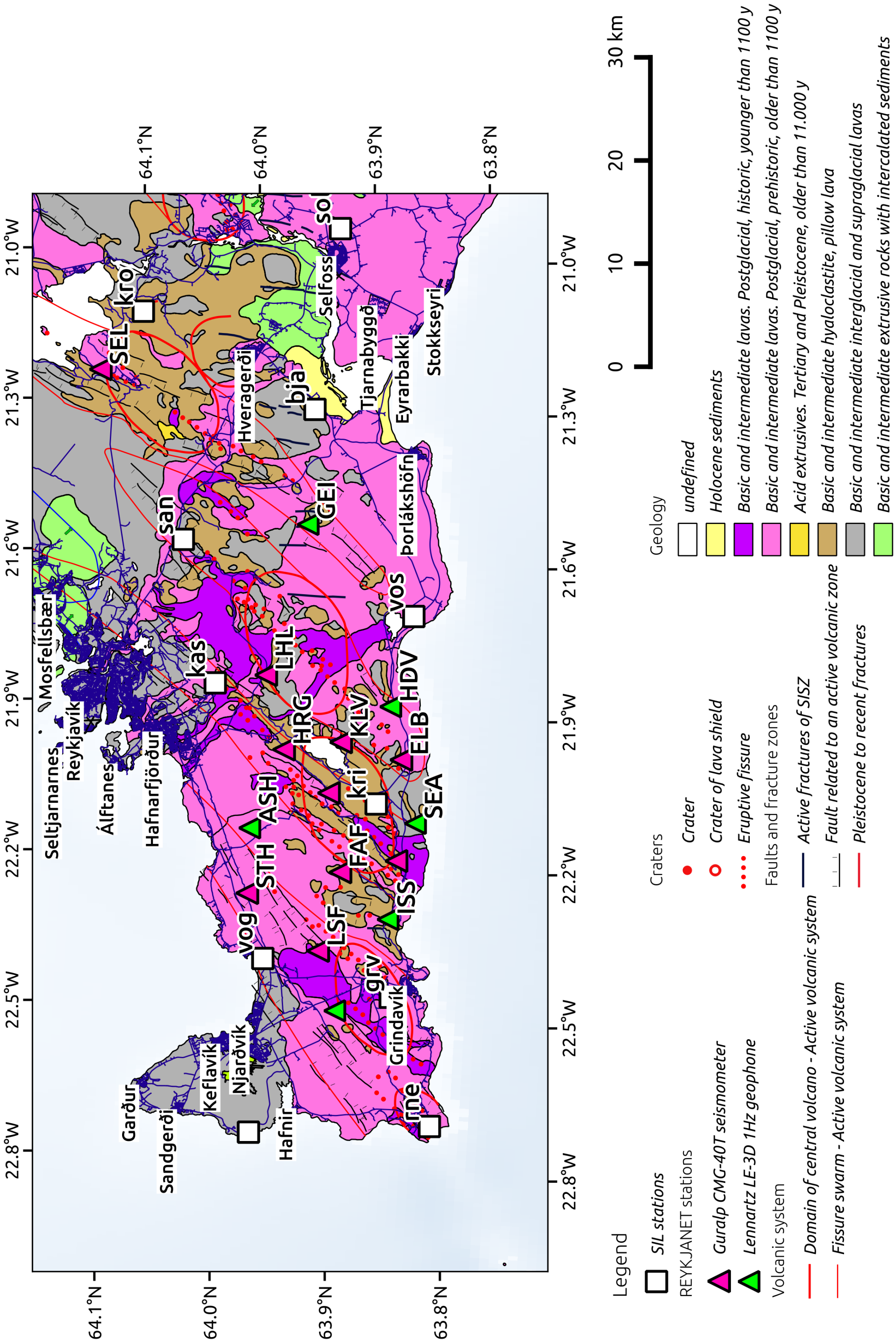

Figure 6: Geology and tectonics of the Reykjanes Peninsula. Based on data from the Icelandic Institute of Natural History and the National Land Survey of Iceland. The blue lines are roads.

These layers are determined by decreasing hydrostatic pressure during the growth of the mountain. Subglacial volcanism has presumably built high topography (Trønnes, unpublished).

In subaerial conditions, the volcanism has possibility to change to lava eruptions, that create different topography - steep sides and flat top. These mountains are referred as table mountains. The products of eruption are mostly basaltic lava flows from fissure swarms or shield volcanoes (Trønnes, unpublished).

Fissure swarms are 5-10 km wide and 40-80 km long and contain tectonic fractures. These swarms comprises tension fractures (100 m long), normal faults (1 km long) and volcanic fissures (1 km long). One of the recent rifting activities occurred in the Krafla fissure swarm in timespan of 1975 to 1984, when 9 eruptions filled the fissures and formed pseudodykes. The total dilatation was 9 m horizontal and 1-3 m vertical across the swarm (Gudmundsson, 1995).

Central volcanoes are developed in many volcanic systems alongside the fissure swarms. Their sources are mostly shallow-seated crustal magmatic chambers, which can propagate into dykes or channels of magma heading towards the surface. Chambers enlarge by melting of the host rocks around them and additionally by melting the roof of the reservoirs.

Icelandic volcanoes are characteristic by frequent eruptions and total timespan varies from 300 thousand to 2 million years. The frequency is determined to several hundreds of years. However, some particular volcanoes could erupt more frequently, for example the Hekla volcano has period of 55 years, or the Grimsvötn volcano has erupted every 11 years in the last 400 years. Additionally, the most volcanoes produce basaltic rocks. Intermediate and acid rocks are mostly bounded to central volcanoes due to the connection to magmatic chambers (Gudmundsson, 1995).

The recent historic eruptions in the south part of Iceland occurred for example in the Katla volcanic system in 1918 and 1955, which were preceded by seismicity. Even though Katla lacked eruptions in the last few decades, it is the most seismically active volcano of the area. Another central volcano - Torfajökull is also seismically active. Its extrusive volcanism included acid rhyolitic volcanism, which is accompanied by geothermal activity. Despite the missing eruptions in the last few hundred years, geothermal activity is located in the western part of the caldera, which is presently seismically active. The Vestmannaeyjar volcanic system was volcanically active from 1964 to 1967 and during 1973, when the eruption was preceded by an earthquake swarm. Hypocenters were located in depths between 15 to 25 km and created a spherical volume that is believed to be a storage for magma (Einarsson, 1991).

In 2010, there was an explosive eruption of the Eyjafjallajökull volcano, which lasted from April to May and emitted ash and tephra into the atmosphere (Cioni et al., 2014). The latest eruption and rifting event in Iceland started in the Bardarbunga caldera and expanded as a 45 km long dyke to Holohraun lava field, where it started as an effusive eruption of basaltic lava. This phase started on 31 August 2014 and by January 2015 had formed a lava field of volume of one cubic kilometer. Seismic activity has started on 16 August 2014 beneath the Bardarbunga caldera (Sigmundsson et al., 2015).

3.3 Earthquakes and earthquake swarms

The first event recorded by the SIL system was eruption of Hekla volcano in 1991. Following earthquake swarms were monitored in south-west Iceland during the first half of 1991 with 2 earthquakes of magnitude greater than 2. Other greater earthquakes occurred in the south of Langjökull of magnitude M_l 4.6 and near Ölfus with magnitude M_l 4.0, which was the largest event of the related swarm in June. During the first half of 1991, between 100 and up to 500 events were detected every day, however, after this period until 1994, about 5-20, but no more than 100 events during swarms per day were registered. The next swarm activity occurred at Kleifarvatn on the Reykjanes Peninsula in Novermber 1992 with M_l 4.0 earthquake. This activity migrated towards the Hengill triple junction, where was monitored second earthquake of magnitude M_l 4.3 in a small swarm in December 1992. In 1993 there were 3 earthquakes of $M_l > 3$, related to a small swarm near Krýsuvík on the Reykjanes Peninsula, and one event of magnitude 4 in Eyjafjördur in the end of this year (Jakobsdóttir, 2002).

During 1994-2007, about 250 thousands earthquakes were located in the Iceland and the surrounding continental shelf. About 10 thousands of events had greater magnitude than M_l 2. The Tjornes Fracture Zone on the north of Iceland contains the Grímsey lineament and the Húsavík-Flatey fault, which are the most active areas in this region. Two earthquakes of magnitude M 5-5.5 were observed in the TZF. First earthquake occurred on 8 February 1994 in the western part and the second on 16 September 2002 in the northern part of the zone, 53 km north-west from Grímsey. About 450 aftershocks were observed after this second event in the next 5 days (Jakobsdóttir, 2008).

The June 2000 earthquake swarm started on 17 June in the middle of the South Iceland Lowland. The largest event was magnitude 6.6 and it was forecasted to occur in the seismic gap, which preceded the epicenter area. The result of this event was triggering of additional earthquakes in the SISZ and on the Reykjanes Peninsula in following 30 seconds. These consequent earthquakes were magnitude 3.5 to 5.5. Moreover, M 5 were

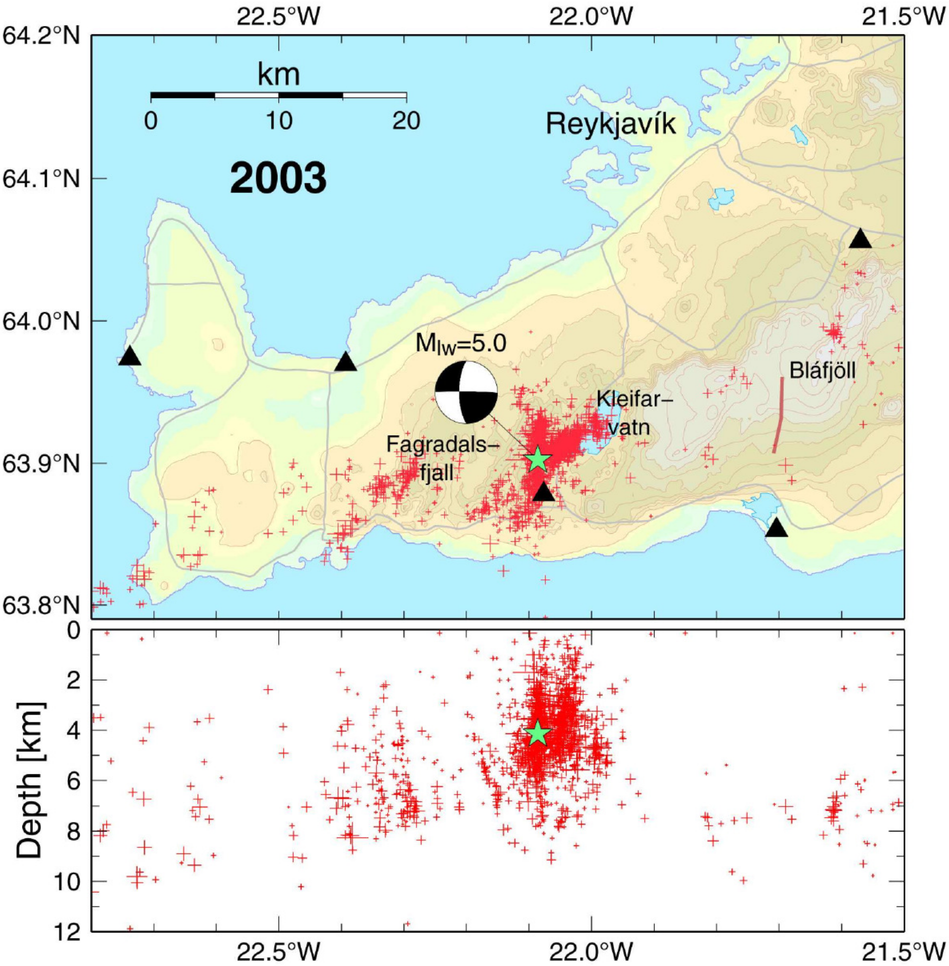

Figure 7: Epicenters and an E–W cross section of the focal depth distribution on the Reykjanes Peninsula in 2003 (after Gudmunsson et al., 2004).

triggered along the zone in the next minutes. These large events were followed by thousands of aftershocks along the 100 km long belt. On 21 June the M 6.6 earthquake struck the area of Hestvatn, where one of the previous aftershock has occurred. Structural analysis of this region revealed, that depth of focal planes reaches to 10 km and the faults are vertical and have overall strike of 7° and 179° (Jakobsdóttir, 2008).

The another episode of larger events occured at Hengill – Ölfus and lasted from August 1994 to November 1998. The first swarm of more than 5000 events was observed north of Hveragerdi. It was initiated on 13 August with an earthquake of M_l 3.5 and followed in the next 8 days by 6 more larger earthquakes of magnitude M_l >3.5. This activity lasted for 10 days and culminated in event of M_l 4 on 20 August. This swarm determined the beginning of sesmically active 5 years long period at the Hengill triple junction. This period was ended by two M_l 5 earthquakes in June and November 1998. The mapped faults tends to have vertical dip and north-east or eastern orientation. The NE faults are mostly right lateral, while the eastern are mostly left lateral faults (Jakobsdóttir, 2008).

On 23 August 2003 an M_{lw} 5 earthquake was observed near Kleifarvatn on the Reykjanes Peninsula at depth about 4 km. Additionally, about 1200 aftershocks occurred in the same day with 3 larger events of magnitude greater than 3. This aftershock activity continued for the next week. The focal mechanism of the largest event suggest right lateral strike slip (Gudmunsson et al., 2004 and reference therein) (Fig.7 on page 17).

3.3.1 Monitored seismicity related to volcanism

Most of monitored eruptions reacted according to similar seismic pattern. Firstly, seismicity related to prospective eruption site increased its frequency and culminated into main shock during of before the beginning of eruption, followed by series of aftershock swarm until the end of volcanic activity. After that, seismicity has been reduced or even disappeared for several months. That includes volcanic episode at Gjálp in 1996 with the largest earthquake of magnitude 5.6 on 29 September; Hekla in 2000 (M_l 2.2 on 26 February), where after the first hours of the eruption a continuous tremor were monitored; and Grímsvötn in 1998 and 2004 (M_l 3 on 31 October 2004) (Jakobsdóttir, 2008).

4 Discussion and Conclusion

Relation between West Bohemia and Iceland varies from similar geochemistry of volcanism to releasing accumulated stress in form of swarms and mainshock - aftershock activity. Both regions have mainly basaltic volcanism with presence of olivine rocks. Seismicity related to volcanism in Iceland occurs in form of short seismic episodes with large release of the seismic energy. Unfortunately, this connection wasn't observed in West Bohemia due to recently missing volcanic eruptions. However, the volcanic related processes in WB as degassing and hydrothermal activity may indicate movement of magmatic masses, which triggers seismicity in the crust.

In terms of seismic structure, the WB has generally greater depths of hypocenters compare to rift seismicity, however, the seismicity related to volcanism mentioned above could reach into depths of 15 to 25 km in the magma chambers under particular volcanoes. The 5 years long Hengill seismic episode could be correlated to seismic unrest in 2004 in WB, which resulted in earthquake swarm in 2008. However, the initial 10 days long Hengill swarm, together with 1 week long swarm near Kleifarvatn in 2003, are more similar to the most recent swarms in WB (in terms of speed of releasing stress) than to swarms, which occurred in the last century.

In conclusion, West Bohemia lies in the transition zone of three structural units of the Bohemian Massif above the Eger intraplate rift zone. It comprises Neoproterozoic, Paleozoic and Cenozoic rocks. Volcanic eruptions are presented in form of two scoria cones and one maar. Their latest activity dates to middle Pleistocene. Fluids are represented by degassing sites of CO_2 and ration of $^3He/^4He$ from mantle. Additionally, monitoring of the groundwater level indicates possible precursor parameters (water temperature) to following seismic activity. Earthquakes in this area tend to have higher speeds and shorter inter-swarm time than in the past. The depths of foci are about 6 to 11 km.

Secondly, Iceland lies on the the Mid-Atlantic Ridge. It is volcanically active with several recent eruptions. These eruptions are mostly basaltic. Most of earthquakes are related to seismic and volcanic zones. Seismic activity bounded to volcanoes has higher rate of releasing stress, followed by quiescence. Focal depths in Reykjanes Peninsula are located mostly 1 to 5 km under the surface. Two larger episodes at Hengill 1994 and Kleifarvatn 2003 occurred during one or two weeks. Regional stress in this area is characterised by the north-west oriented minimal stress.

References

Babuška, V., Plomerová, J. (2013). Boundaries of mantle-lithosphere domains in the Bohemian Massif as extinct exhumation channels for high-pressure rocks. Gondwana Research, 23(3), 973–987. http://doi.org/10.1016/j.gr.2012.07.005

Bankwitz, P., Schneider, G., Kämpf, H., Bankwitz, E. (2003). Structural characteristics of epicentral areas in Central Europe: Study case Cheb Basin (Czech Republic). Journal of Geodynamics, 35(1-2), 5–32. http://doi.org/10.1016/S0264-3707(02)00051-0

Cioni, R., Pistolesi, M., Bertagnini, a., Bonadonna, C., Hoskuldsson, a., Scateni, B. (2014). Insights into the dynamics and evolution of the 2010 Eyjafjallajökull summit eruption (Iceland) provided by volcanic ash textures. Earth and Planetary Science Letters, 394(May 2010), 111–123. http:

//doi.org/10.1016/j.epsl.2014.02.051

Einarsson, P. (1991). Earthquakes and present-day tectonism in Iceland. Tectonophysics. http://doi.org/10.1016/0040-1951(91)90501-I

Faber, E., Horálek, J., Boušková, A., Teschner, M., Koch, U., Poggenburg, J. (2009). Continuous gas monitoring in the West Bohemian earthquake area, Czech Republic: First results. Studia Geophysica et Geodaetica, 53(3), 315–328. http://doi.org/10.1007/s11200-009-0020-z

Faryad, S. W., Kachlík, V. (2013). New evidence of blueschist facies rocks and their geotectonic implication for Variscan suture(s) in the Bohemian Massif. Journal of Metamorphic Geology, 31(1), 63–82. http://doi.org/10.1111/jmg.12009

Fischer, T., Horálek, J. (2003). Space-time distribution of earthquake swarms in the principal focal zone of the NW Bohemia/Vogtland seismoactive region: Period 1985-2001. Journal of Geodynamics, 35(1-2), 125–144. http://doi.org/10.1016/S0264-3707(02)00058-3

Fischer, T., Michálek, J. (2008). Post 2000-swarm microearthquake activity in the principal focal zone of West Bohemia/Vogtland: Space-time distribution and waveform similarity analysis. Studia Geophysica et Geodaetica, 52(4), 493–511. http://doi.org/10.1007/s11200-008-0034-y

Fischer, T., Horálek, J., Michálek, J., Boušková, A. (2010). The 2008 West Bohemia earthquake swarm in the light of the WEBNET network. Journal of Seismology, 14(4), 665–682. http://doi.org/10.1007/s10950-010-9189-4

Fischer, T., Horálek, J., Hrubcová, P., Vavryčuk, V., Bräuer, K., Kämpf, H. (2014). Intra-continental earthquake swarms in West-Bohemia and Vogtland: A review. Tectonophysics, 611, 1–27. http://doi.org/10.1016/j.tecto.2013.11.001

Gudmundsson, A. (1995). Infrastructure and mechanics of volcanic systems in Iceland. Journal of Volcanology and Geothermal Research, 64(1-2), 1–22. http://doi.org/10.1016/0377-0273(95)92782-Q

Gudmundsson, G. B., Jakobsdóttir, S. S., Thorbjarnardóttir, B. S. (2004). Seismicity in Iceland 2003. JOKULL, (54), 67–74.

Horálek, J., Čermáková H., Fischer, T. (2015). Earthquake statistics, spatiotemporal distribution of foci and source mechanisms as a key to understanding of causes leading to the West Bohemia/Vogtland earthquake swarm. EGU General Assembly 2015, Vienna, Austria, April 12 - 17. Abstract EGU2015-10037 Jakobsdóttir, S. S., Gudmundsson, G. B., Stefánsson, R. (2000). Seismicity in Iceland 1991 – 2000 monitored by the SIL seismic system. JOKULL, (51), 87–94. http://hraun.vedur.is/ja/skyrslur/Jokull51skj2000.pdf

Jakobsdóttir, S. S. (2008). Seismicity in Iceland : 1994 – 2007. JOKULL, (58), 75–100. http://earthice.hi.is/sites/jardvis.hi.is/files/Pdf_skjol/Jokull58_pdf/jokull58-jakobsdottir.pdf

Kachlík V., unpublished, Geologický vývoj České republiky. SURAO Praha http://web.natur.cuni.cz/ugp/kachlik/RegionalniGeologie/

Koch, U., Heinicke, J., Voßberg, M. (2003). Hydrogeological effects of the latest Vogtland-NW Bohemian swarmquake period (August to December 2000). Journal of Geodynamics, 35(1-2), 107–123. http://doi.org/10.1016/S0264-3707(02)00057-1

Müller, V., Burda, J., Dubec, O., Hrazdíra, P., Hrkal, Z., Jinochová, J., Lochmann, Z., Majer, V., Manová, M., Mlčoch, B., Rejchrt, M., Sáňka, V., Schovánek, P., Skácelová, D., Šalanský, K., Šantrůček, P. (1998). Vysvětlivky k souboru geologických a ekologických účelových map přírodních zdrojů v měřítku 1 : 50 000. Listy 11 - 13 Hazlov, 11 - 14 Cheb. Český geeologický ústav. ISBN 80-7075-263-7

Neunhöfer, H., Meier, T. (2004). Seismicity in the Vogtland/Western Bohemia earthquake region between 1962 and 1998. Studia Geophysica et Geodaetica, 48(3), 539–562. http://doi.org/10.1023/B:SGEG.0000037471.18297.07

Sigmundsson, F., Hooper, A., Hreinsdottir, S., Vogfjörd, K. S., Ofeigsson, B. G., Heimisson, E. R., Eibl, E. P. S. (2015). Segmented lateral dyke growth in a rifting event at Bardarbunga volcanic system, Iceland. Nature, 517(7533), 191–195. Retrieved from http://dx.doi.org/10.1038/nature14111

Stefánsson, R., Böðvarsson, R., Slunga, R., Einarsson, P., Jakobsdóttir, S., Bungum, H., Korhonen, H. (1993). Earthquake prediction research in the South Iceland Seismic Zone and the SIL project. Seismol. Soc. Am.

Bull., 83(3), 696–716.

Trønnes, R. G., unpublished. Geology and geodynamics of Iceland. Unpub. Ms., (2000). Retrieved from https://notendur.hi.is/oi/Pdf_reprintlibrary/GeologyandgeodynamicsofIceland.pdf

Tryggvason, A., Rögnvaldsson, S. T., Flóvenz, Ó. G. (2002). Three-dimensional imaging of the P- and S-wave velocity structure and earthquake locations beneath Southwest Iceland. Geophysical Journal International, 151(3), 848–866. http://doi.org/10.1046/j.1365-246X.2002.01812.x

Vogfjörd K. S., G. Nolet, W.J. Morgan, R. M. Allen, R. Slunga, B.H. Bergsson, P. Erlendsson, G. Foulger, S. Jakobsdottir, B. Julian, M. Pritchard and S. Ragnarsson (2002). Crustal profiling in Iceland using earthquake source arrays. AGU Fall Meeting, San Francisco, California, December 6-10, 2002. Abstract S61C-1161.